7. Mountain Glaciers and Ice Caps

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Contents

Key Findings

- Mountain glaciers and ice caps in the Arctic cover an area of around 402 000 km^2 and contain about 0.41 m sea-level equivalent of water.

- Over the past century, nearly all have retreated from maximum extents reached during the Little Ice Age, which ended in the late 19th century. This period of glacier retreat has been associated with an overall reduction in glacier mass during the period of record, which extends to more than 60 years in some cases. Surface mass balance measurements showed negative or nearly balanced conditions and generally showed no trend until the mid-1990s, but since then reveal significantly higher rates of mass loss in Alaska, the Canadian Arctic, and Iceland.

- The fraction of ablation that occurs through iceberg calving can be as much as 40% in regions where it has been measured, but it has not been measured over large areas of the Arctic. Estimation of calving fluxes is therefore a major source of uncertainty in estimates of current and future rates of mass loss from mountain glaciers and ice caps in the Arctic.

- Mass loss (surface mass balance plus calving) from Arctic glaciers probably exceeded 150 Gt/y in the past decade, when it was similar to mass loss from the Greenland Ice Sheet. This suggests that glacier and ice sheet change in the Arctic is probably now the dominant contributor to the eustatic (water mass) component of global sea level rise.

- Under the IPCC A1B emissions scenario, the total volume of Arctic glaciers is projected to decline by between 13% and 36% by 2100 (corresponding to an increase of 51 to 136 mm sea-level equivalent), depending on the choice of general circulation model. These projections are a lower bound since they do not include mass losses by iceberg calving. Regardless, mountain glaciers and ice caps will continue to influence global sea-level changes beyond the 21st century.

- In many parts of the Arctic, climate warming should cause glacier runoff to increase for a few decades or longer, but glacier area reduction will ultimately cause glacier runoff to decline. These changes in glacier runoff will have impacts on water supplies; water quality; hydroelectric power generation; flood hazards; freshwater, estuarine and coastal habitats; and ocean circulation patterns.

- Iceberg hazards to shipping and offshore activities related to exploration for and exploitation of offshore hydrocarbon and mineral resources may increase if changes in tidewater glacier dynamics result in more iceberg production and/or larger bergs, and reductions in sea-ice cover allow icebergs to become more mobile.

Summary

In addition to the Greenland Ice Sheet, the Arctic contains a diverse array of smaller glaciers ranging from small cirque glaciers to large ice caps with areas up to 20 000 km^2. Together, these glaciers cover an area of more than 400 000 km^2, over half the global area of mountain glaciers and ice caps. Their total volume is sufficient to raise global sea level by an average of about 0.41 m if they were to melt completely.

These glaciers exist in a range of different climatic regimes, from the maritime environments of southern Alaska, Iceland, western Scandinavia, and Svalbard, to the polar desert of the Canadian Arctic. Glaciers in all regions of the Arctic have decreased in area and mass as a result of the warming that has occurred since the 1920s (in two pulses – from the 1920s to the 1940s and since the mid-1980s). A new phase of accelerated mass loss began in the mid-1990s, and has been most marked in Alaska, the Canadian Arctic, and probably Greenland. Current rates of mass loss are estimated to be in the range 150 to 300 Gt/y; comparable to current mass loss rates from the Greenland Ice Sheet. This implies that the Arctic is now the largest regional source of glacier contributions to global sea-level rise.

Most of the current mass loss is probably attributable to a change in surface mass balance (the balance between annual mass addition, primarily by snowfall, and annual mass loss by surface melting and meltwater runoff). Iceberg calving is also a significant source of mass loss in areas such as coastal Alaska, Arctic Canada, Svalbard, and the Russian Arctic. However, neither the current rate of calving loss nor its temporal variability have been well quantified in many regions, so this is a significant source of uncertainty in estimates of the total rate of mass loss. It is, however, clear that the larger Arctic ice caps have similar variability in ice dynamics to that of the Greenland Ice Sheet. That is to say, areas of relatively slow glacier flow (which terminate mainly on land) are separated by faster-flowing outlet glaciers (which terminate mainly in the ocean). Several of these outlet glaciers exhibit surge-type behavior, while others have exhibited substantial velocity changes on seasonal and longer timescales. It is very likely that these changes in ice dynamics affect the rate of mass loss by calving both from individual glaciers and the total ice cover.

Projections of future rates of mass loss from mountain glaciers and ice caps in the Arctic focus primarily on projections of changes in the surface mass balance. Current models are not yet capable of making realistic forecasts of changes in losses by calving. Surface mass balance models are forced with downscaled output from climate models driven by forcing scenarios that make assumptions about the future rate of growth of atmospheric greenhouse gas concentrations. Thus, mass loss projections vary considerably, depending on the forcing scenario used and the climate model from which climate projections are derived. A new study in which a surface mass balance model is driven by output from ten general circulation models (GCMs) forced by the IPCC (Intergovernmental Panel on Climate Change) A1B emissions scenario yields estimates of total mass loss of between 51 and 136 mm sea-level equivalent (SLE) (or 13% to 36% of current glacier volume) by 2100. This implies that there will still be substantial glacier mass in the Arctic in 2100 and that Arctic mountain glaciers and ice caps will continue to influence global sea-level change well into the 22nd century.

As glaciers and ice caps shrink in a warming climate, runoff initially increases in response to higher rates of surface melting. Ultimately, however, runoff will decline as reductions in glacier area outweigh the effect of more rapid melting. This phase of declining runoff does not yet seem to have begun in most regions of the Arctic, but it may begin soon in the Russian Arctic mountains. In the Yukon River basin, Greenland, Iceland, and Norway, glacier runoff is an important resource for hydroelectric power generation, and the viability of hydroelectric projects may ultimately be compromised by runoff decreases associated with glacier shrinkage.

Changes in glacier runoff also result in changes in stream temperature, sediment load, and nutrient export (both magnitude and type) that can be expected to initiate changes in the ecology and productivity of downstream river, lake, and fjord environments. Increased rates of glacier melt may accelerate the release of a range of 'legacy' pollutants stored in firn (partially compacted snow that is the intermediate stage between snow and glacier ice) and glacier ice back into the environment. Increasing freshwater fluxes to fjords and other nearshore marine environments will alter the characteristics of surface water masses and drive changes in circulation. Circulation changes may also follow the retreat of tidewater glaciers onto land, particularly in regions of upwelling close to the termini of these glaciers. These changes can decrease the availability of feeding and resting habitats that are important for marine mammals and seabirds.

The number and size of icebergs produced is likely to change as tidewater glaciers retreat, ultimately reaching zero as their termini emerge onto land. Break-up of floating glacier tongues and ice shelves, a process that has accelerated in the past decade along the northern coast of Ellesmere Island, results in large tabular bergs, while accelerated flow of tidewater glaciers tends to result in accelerated production of small bergs unless flotation of the glacier terminus occurs, when large tabular bergs may be produced. Cessation of small berg production when tidewater termini retreat onto land can reduce the number of such bergs that become grounded in fjords, decreasing the availability of important resting habitat for seals. Circulating icebergs are a potential hazard for shipping, drilling platforms, and seafloor pipelines in the Arctic. Circulation patterns and longevity of bergs may change as the Arctic sea-ice cover declines. Thinner and less extensive sea ice is likely to result in greater berg mobility, while warmer surface waters may result in more rapid melting and disintegration of bergs. A knowledge of the size distributions of bergs produced and circulating in different regions is critical to evaluating the risk of bergs contacting the sea floor and damaging seafloor pipelines.

Glacier retreat will be associated with changes in the magnitude and frequency of a range of geomorphological hazards, most notably outburst floods from ice-marginal and moraine-dammed proglacial lakes, and mass movements from newly deglaciated valley walls. The degree of risk from such phenomena will, however, be highly variable, depending upon the nature of the deglaciated terrain, the size of local populations, and the amount of infrastructure present in individual regions.

Finally, it should be emphasized that the ability to monitor and predict changes in the Arctic's mountain glaciers and ice caps is still quite limited. Basic inventory data for Arctic glaciers are lacking. The number of glaciers on which mass balance is measured is small and declining, and the distribution of measurement sites is highly non-uniform. There are no measurement sites at all in some areas with large glacier areas (such as the Russian Arctic and the Yukon). There is no routine monitoring of mass losses by iceberg calving, and understanding of what controls calving rates is rudimentary. This severely constrains the ability to model calving losses into the future. Studies of the socio-economic impacts of Arctic glacier change are currently few and limited in scope, so most statements about such impacts are based solely on general principles. As such, they do not provide a strong basis for either the formulation or enactment of a policy response to Arctic glacier change. Some of the largest impacts of the ongoing changes in Arctic glaciers (such as global sea-level rise) will be felt in regions of the world that are very far from the Arctic. They may, however, still have social, political, and economic repercussions for Arctic nations – repercussions that need to be explored more thoroughly.

7.1. Introduction

- Mountain glaciers and ice caps cover an area of nearly 402 000 km^2 in the Arctic. Over half of this area is in western North America and the Canadian Arctic.

- The combined volume of these glaciers is sufficient to raise sea level by around 0.41 m if they all melted.

- Glacier types range from small cirque glaciers to large ice caps, with areas of up to 20 000 km^2. These ice caps are dynamically complex, and are drained in part by fast-flowing outlet glaciers that often reach the ocean and lose mass by calving icebergs.

- In many regions, a proportion of the glaciers exhibit 'surge-type' behavior, in which long periods of flow at relatively low speeds are punctuated by short-lived episodes of very rapid flow.

- Long-term changes in the extent, volume, and mass of these glaciers are driven by changes in climate and oceanographic conditions that alter their 'mass balance' – the annual balance between mass gains (due mainly to snowfall) and mass losses (due mainly to surface melting and runoff, and iceberg calving).

- Changes in the thermal structure and flow of glaciers can play an important role in how, and how rapidly, they respond to changing climate and oceanographic conditions.

7.1.1. Background

Glacier ice (including mountain glaciers, ice caps and the Greenland Ice Sheet) occupies about 2.16 million km^2 of the Arctic land surface. This chapter deals with all mountain glaciers and ice caps (hereafter referred to as glaciers), including those in Greenland that are not connected to the ice sheet. The Greenland Ice Sheet itself is addressed in Chapter 8. Mountain glaciers are ice bodies whose geometry and boundaries are controlled by bedrock topography, while ice caps are dome-shaped ice bodies that submerge the underlying bedrock topography. Table 7.1 presents the regional distribution of the

Table 7.1. The areas (rounded to three significant figures) and estimated volumes, in mm sea-level equivalent (SLE), of ice caps and glaciers in the Arctic compared to global values and the ice sheets in Greenland and Antarctica.

Region	Ice-covered area, km²	Volume[a], mm SLE	Source, area
Canadian Arctic	1 51 000	199 ± 30	Ommanney, 1970
Northwestern North America	91 800	71 ± 8	Berthier et al., 2010; ESRI, 2003
Russian Islands[b]	56 700	44 ± 8	Radić and Hock, 2010
Greenland[c]	48 600	44 ± 7	Weng, 1995
Svalbard	36 500	26 ± 2	Hagen et al., 1993
Iceland	11 000	12 ± 6	Björnsson and Pálsson, 2008
Scandinavia	3 100	6 ± 0	Østrem et al., 1973, 1988
Russian Arctic	2 900	4 ± 0	Radić and Hock, 2010
Total (Arctic)	401 600	410 ± 30	
Mountain glaciers and ice caps (global)	741 400	600 ± 70	Radić and Hock, 2010
Greenland Ice Sheet	1 755 600	7500	Bamber et al., 2001
Antarctic ice sheet Antarctic ice shelves	12 348 000 1 555 000	57 000	Fox and Cooper, 1994 (areas) Lythe et al., 2001 (volume)

[a] Volumes of all mountain glaciers and ice caps are from Radić and Hock (2010); [b] Franz Josef Land (13 700 km²), Novaya Zemlya (23 600 km²), Severnaya Zemlya (19 400 km²); [c] Greenland excluding the ice sheet: only the glaciers physically disconnected from the ice sheet are considered here. Areas over 70 000 km² have been reported (Holtzscherer and Bauer, 1954; Weidick and Morris, 1998); however, these estimates have included some glaciers that are connected to the ice sheet but considered independent of the ice sheet in a dynamic sense or that were, by their morphology, discernible as units independent of the ice sheet. Nevertheless, the estimate by Weng (1995) is a minimum estimate because it is based on a 1:2 500 000 map, a scale too coarse to allow identification of many small glaciers.

area and volume of glacier ice in the Arctic. Here, the ice volume is given in units of mm sea-level equivalent (SLE), which is the volume of water (in m³) stored as glacier ice, divided by the surface area of the global ocean (in m²) multiplied by 1000.

7.1.2. Context: What did the Arctic Climate Impact Assessment report say about mountain glaciers and ice caps?

The SWIPA report is an update of knowledge on the state of the Arctic cryosphere relative to the Arctic Climate Impact Assessment (ACIA, 2005). This dealt with mountain glaciers, ice caps, and the Greenland Ice Sheet in a single section of chapter 6, '*Cryosphere and Hydrology*'. It gave a regional summary of glacier changes since around 1950 with an emphasis on measurements of surface mass balance and, for regions where mass balance measurements were lacking, changes in glacier area. Mass losses by iceberg calving were mentioned but not discussed in detail. The major ACIA contribution was a series of projections of changes in the surface mass balance of Arctic glaciers for the period until 2100. These mass balance projections were computed using regional seasonal sensitivity characteristics (a measure of the expected change in the annual surface mass balance due to prescribed changes in monthly mean air temperature and precipitation amounts) and projected changes in temperature and precipitation derived from a suite of five GCMs forced by the IPCC B2 emissions scenario. The mass balance projections were used to estimate the potential contributions of different regions of the Arctic to sea level rise over the 21st century. The ACIA report also commented on the potential impacts of changes in glaciers for other parts of the physical system,

ecosystems, and people, and gave an assessment of critical research needs that remains valid today.

7.1.3. Geographic setting

The glacierized areas (i.e. those presently overlain by glaciers) addressed in this chapter are shown in Figure 7.1. Most of the glaciers are located north of 60° N, but some more southerly glaciers, in Kamchatka, Alaska, and northwestern Canada are also included (Table 7.1). The total glacier-covered area is approximately 402 000 km², which is about 54% of all glaciers in the world excluding the large ice sheets (~741 400 km², Radić and Hock, 2010).

Arctic glaciers are irregularly distributed in space (Figure 7.1), and are located in a range of very different climatic regimes (Braithwaite, 2005). In southern Alaska, Iceland, western Scandinavia, and western Kamchatka, the climate is maritime with a relatively small annual temperature range and precipitation rates of a few metres per year, while in the Canadian High Arctic it is very dry, cold and continental, with short summers, a very large annual temperature range (greater than 50 °C), and annual precipitation that ranges from 0.1 to 0.7 m/y. Conditions on Svalbard and in the Russian Arctic islands fall between these two climatic extremes. In northwestern North America, most of the ice cover is in the mountain ranges adjacent to the Gulf of Alaska, while in Arctic Canada the most heavily glacierized regions are in the mountains on Devon and Ellesmere Islands, which are nourished in part by moisture from a large persistent polynya in northern Baffin Bay (the North Open Water).
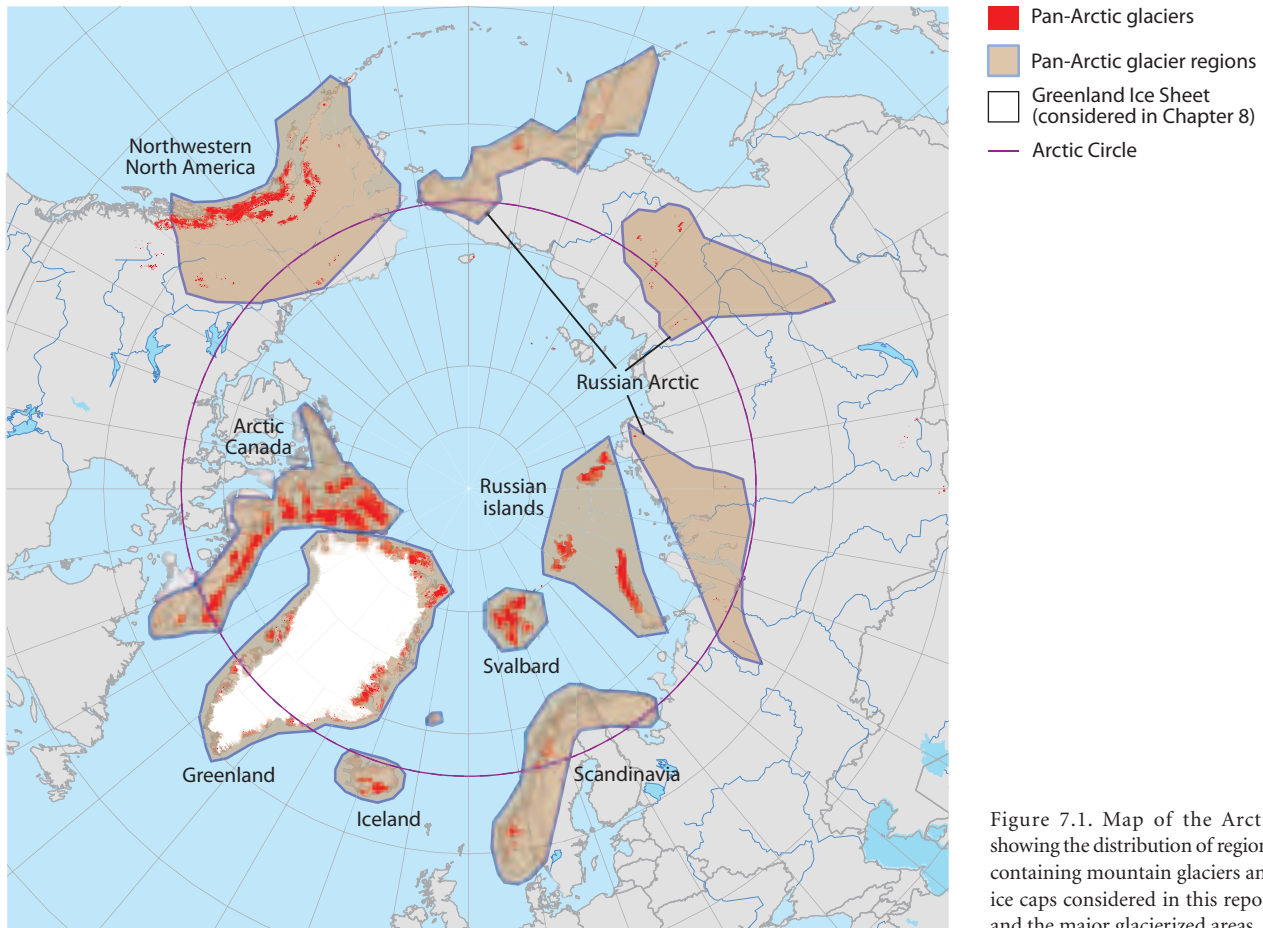
Figure 7.1. Map of the Arctic showing the distribution of regions containing mountain glaciers and ice caps considered in this report and the major glacierized areas.

7.1.4. Characteristics of Arctic mountain glaciers and ice caps

7.1.4.1. Morphology

Arctic glaciers have a range of forms. Dome-shaped ice caps have lobes and outlet glaciers that drain ice away from the accumulation area, where annual snowfall exceeds annual surface melt, to lower-lying regions (ablation areas) where the reverse is true, and in some cases to tidewater margins where the ice reaches the ocean and icebergs are calved. Large ice caps are found in the Canadian Arctic, Iceland, Svalbard, the Russian Arctic islands, and in Greenland beyond the margins of the ice sheet (Figure 7.2). In other regions, large glaciers originate from icefields that fill basins within mountain ranges (e.g., in southern Alaska). Many regions, for example Svalbard, also have a large number of individual valley glaciers that occupy valleys and basins in the landscape.

Much of the Arctic ice mass is contained in relatively large ice caps with areas of up to 20 000 km², although there are large numbers of independent glaciers with areas ranging from 0.1 to several thousand km² (Dowdeswell and Hagen, 2004). The largest ice masses in the Arctic are found in Arctic Canada, and include the Agassiz Ice Cap (17 300 km²) and Prince of Wales Icefield (19 400 km²) on Ellesmere Island, the Devon Island Ice Cap (about 14 400 km²), and the Barnes and Penny Ice Caps (each almost 6000 km²) on Baffin Island. On Greenland, the largest independent ice cap is the ~9000 km² Flade Isblink in the northeast, while the more northerly Hans Tausen Ice Cap,

at 3975 km², is probably the best studied (Reeh et al., 2001). Austfonna on Nordaustlandet in eastern Svalbard (8120 km²) is the largest ice cap in the Eurasian Arctic (Dowdeswell, 1986; Hagen et al., 1993). The largest Russian ice cap is the Academy of Sciences Ice Cap (5575 km²) on Severnaya Zemlya (Dowdeswell et al., 2002), although the northern island of Novaya Zemlya has a larger ice-covered area with many outlet glaciers. In Iceland, Vatnajökull has an area of 8100 km² (Björnsson and Pálsson,

Figure 7.2. Small plateau ice caps on Axel Heiberg Island, Arctic Canada, with a larger ice cap in the distance. Note the clear contrast between the snow-covered accumulation area and the greyer ablation area, and the outlet glaciers draining into valleys leading away from the ice cap. Source: Martin Sharp, University of Alberta.

2008). Among the mountain glaciers (Figure 7.3), the largest is Bering Glacier in Alaska, with an area of 3630 km².

The larger Arctic ice caps and icefields (Figure 7.4) have complex dynamics involving a mix of fast- and slow-flowing elements that can vary in how they respond to climate changes and variability. Vestfonna on Nordaustlandet in eastern Svalbard is a good example, where four fast-flowing outlet glaciers are embedded within a largely slower-flowing ice cap of 2500 km² (Dowdeswell and Collin, 1990). Typical surface velocities may be less than 10 m/y close to the equilibrium-line altitude of valley and cirque glaciers that terminate on land. They can reach many hundreds of metres per year on major calving outlet glaciers, on which seasonal velocity variations can be large, with summer velocities up to an order of magnitude greater than winter velocities (Williamson et al., 2008). Interannual velocity variations can also be significant on such glaciers.

7.1.4.2. Thermal regime

Most glaciers in Iceland, Scandinavia, and central and southern Alaska are predominantly temperate (composed of ice at the pressure melting point). Elsewhere in the Arctic, glaciers tend to have polythermal temperature regimes (composed of a mixture of ice at and below the pressure melting point), and their dynamics may be strongly affected by climate-driven changes in the thermal regime. In some parts of polythermal glaciers, the ice temperature at the glacier bed is below the melting point, implying that the glacier is frozen to its bed. In some smaller (cold) glaciers, all the ice is at temperatures below the melting point, except at the surface where the ice temperature may reach the melting point in summer. The thermal regime of glaciers is determined by the prevailing climatic conditions (snow accumulation and surface energy balance). Climatic change can alter the thermal regime of a glacier, and potentially also its dynamics, because ice deforms more rapidly at higher temperatures and glacier flow can be enhanced by the lubricating effect of meltwater at the glacier bed. Most cold-based and polythermal glaciers are found in dry regions with low accumulation rates. It takes a long time for a climate change signal to penetrate into such glaciers, and changes in the temperature regime are probably not very large on a 100-year timescale. However, in areas where penetration of surface meltwater into cold firn on the glacier surface increases, the release of latent heat when this meltwater refreezes can cause a more rapid change in the thermal regime.

Owing to the polythermal nature of many Arctic glaciers, the formation of superimposed ice by meltwater refreezing on the glacier surface and internal accumulation (where percolating meltwater freezes in cold snow and firn) can be important processes of mass accumulation. When these processes occur, the refrozen water has to be melted again to become meltwater runoff, complicating the measurement and modeling of the surface mass balance. Neglecting or inadequately accounting for these processes in mass balance measurements overestimates mass loss. Superimposed ice formation is important on many Arctic glaciers and is the dominant form of accumulation on some (Koerner, 1970).

Surge-type glacier behavior is common in many parts of the Arctic and can be a source of significant hazards. In a glacier surge, surface velocities increase by an order of magnitude or more and glacier fronts can advance many kilometres (sometimes more than 10 km) in a matter of years. When tidewater glaciers surge, iceberg production increases and can be an important process of mass loss from the glacier (Liestøl, 1973; Dowdeswell, 1989). While the occurrence of individual surges is not directly related to climate change, climate change may alter the frequency of surging (Dowdeswell et al., 1991; Eisen et al., 2001), or even cause glaciers to cease being of the surge type (Dowdeswell et al., 1995). Changes in the number of actively surging glaciers in a region that extend into tidewater can have a large short-term impact on the regional glacier mass balance through changes in the overall calving flux to the ocean.

7.1.4.3. Tidewater glaciers

Iceberg calving plays a major role in the overall mass balance of tidewater glaciers and Arctic ice caps that have significant tidewater margins (Dowdeswell et al., 1997, 2008; Dowdeswell and Hagen, 2004; Błaszczyk et al., 2009) (Figure 7.5). However, accurate calculation of the calving flux requires information that is frequently not available (e.g., ice thickness at the glacier terminus), and modeling of calving flux is challenging because
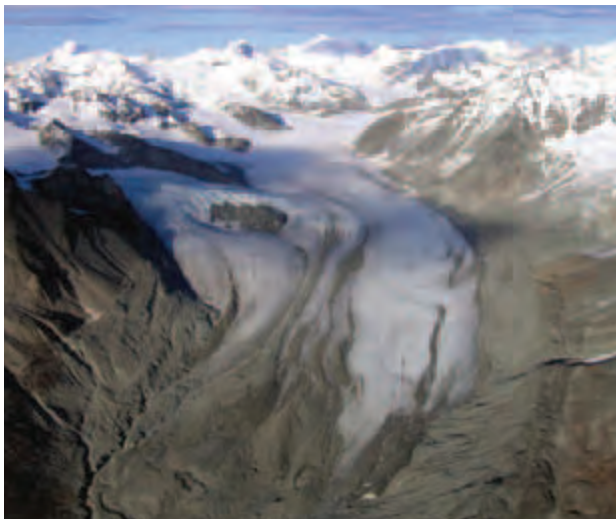

Figure 7.3. Gulkana Glacier in Alaska is a typical mountain glacier and a long-term mass balance monitoring site. Note the end moraines surrounding the glacier terminus that indicate that the glacier was more extensive and thicker in the past. Source: U.S. Geological Survey.

Figure 7.4. The Glacier Bay Icefield, Alaska. Outlet glaciers drain outward from a central accumulation area within the mountains. Source: Landsat imagery courtesy of NASA Goddard Space Flight Center and U.S. Geological Survey.

Figure 7.5. Time-lapse camera overlooking the calving terminus of the tidewater glacier, Kronebreen, in Svalbard. Kronebreen is a focus of study within the IPY GLACIODYN project. Source: Monica Sund, University Centre, Svalbard.

of the lack of a widely accepted calving law. Although individual ice caps may have multiple tidewater glacier outlets, the mass loss by calving is often dominated by one or two of these outlets (Krenke, 1982; Dowdeswell et al., 2002, 2008; Burgess et al., 2005; Williamson et al., 2008; Mair et al., 2009), but calving fluxes from individual glaciers can show large temporal variability.

Recent increases in calving flux from the Greenland Ice Sheet have been associated with acceleration, retreat, and thinning of major outlet glaciers (Howat et al., 2005; Rignot and Kanagaratnam, 2006). Similar changes are observed in Alaska and on some large Arctic ice caps, and may be caused by changes in glacier dynamics related to increased surface melting, penetration of meltwater to glacier beds, and subsequent lubrication of the ice-bed interface allowing increased flow by basal sliding. Other possible causes of calving flux increases include increased melt of the underwater part of terminal ice cliffs due to increasing ocean temperature (Holland et al., 2008), thermal transitions (from cold-based to warm-based) in polythermal glaciers, and break-up and removal of floating ice tongues that formerly restrained the flow of the glacier. Such dynamic changes may be larger and more rapid than those induced by changes in surface mass balance alone.

7.1.4.4. Response to climate change

Glaciers respond to climate change over very different timescales depending on their size, shape, and thermal regime. Among glaciers that terminate on land, smaller glaciers tend to respond more quickly, changing their shape, flow, and terminus position over years or decades. The hypsometry (area-altitude distribution) of a glacier plays an important controlling role in determining its response to changes in climate. In a given region, low-lying glaciers may shrink and retreat quickly at the same time as glaciers at higher elevations grow or maintain their size. Changes in the temperature and salinity of the ocean water adjacent to calving ice fronts can trigger rapid changes in the terminus position, flow velocity, and calving flux of tidewater glaciers, so that large tidewater glaciers may alter their form more rapidly than small glaciers that terminate on

land (Holland et al., 2008; Rignot et al., 2010; Straneo et al., 2010). Although Arctic glaciers show a variety of responses to changing climate, it is something to which they are all sensitive. It is worth noting that the surface mass balance of glaciers in the Arctic may be affected indirectly by changes in other components of the Arctic cryosphere, such as the regional snow cover and extent of sea ice, that affect the surface energy balance, atmospheric circulation, and availability of open water as a source of water vapor (e.g., Rennermalm et al., 2009).

7.1.5. Significance and impacts of glacier changes

According to recent estimates, glaciers contributed about 0.8 to 1 mm annually to global sea-level rise during the period 2001 to 2004. The upper limit of the estimates includes the contribution from glaciers surrounding the Greenland and Antarctic ice sheets (Kaser et al., 2006; Solomon et al., 2007). Meier et al. (2007) gave a slightly higher estimate of 1.1 mm for 2006. This is about one third of the total current sea-level rise, or about 60% of that component of the rise attributable to the addition of water mass to the oceans, as opposed to the thermal expansion of ocean waters. The remaining 40% of that component comes from the combined contribution of the Greenland and Antarctic ice sheets. Meier et al. (2007) predicted that a large contribution to sea-level rise over the period to 2100 will still come from glaciers and ice caps. As more than half of the world's glacier and ice cap area is found in the Arctic, it is very important to reduce the uncertainty in estimates of the current and future mass balance of the Arctic glaciers and ice caps.

It is also important to evaluate the impacts of ongoing and future changes in glacier extent and volume on regional water resources, water quality, and the incidence of glacier-related natural hazards. Glacier runoff is exploited as a source of hydroelectric power in Scandinavia, Iceland, Greenland, western Canada, and Alaska. Increasingly negative surface mass balance of the glaciers, and reductions in glacier area may have a direct impact on the water balance of basins used for hydroelectric power generation. Increased freshwater flux from glaciers to nearby fjords and oceans may have an impact on the marine ecosystem via freshening of ocean water and increased transport of nutrients and contaminants from land to sea. Freshening of ocean water may have an impact on fjord circulation and on global ocean thermohaline circulation. Changes in the frequency and magnitude of iceberg calving events may impact infrastructure development, marine transport, fisheries, and oil and gas exploration and production on Arctic continental shelves, where human activity is expected to increase significantly over the coming decades.

Changes in the extent and thickness of Arctic glaciers can destabilize the surrounding terrain and generate geomorphological hazards (rock slides, debris flows, ice avalanches, glacial mudflows, outburst floods), especially when combined with changes in the extent and thickness of permafrost on surrounding slopes. They also create a threat of floods (*jökulhlaups*) from ice-dammed and subglacial lakes, especially in volcanic areas such as Iceland, Alaska, and Kamchatka.

Glacier related tourism is important to some local economies, and shrinking or even disappearing glaciers may have a direct economic impact.

7.1.6. Challenges

The ACIA report (ACIA, 2005) stated that the most difficult task in a regional-scale assessment of glacier behavior is to generalize results from a few glaciers and ice caps to all ice masses in the Arctic. This statement is still valid today. Neither the remote sensing tools nor the glacier and ice sheet models currently available are well suited to studies of regional ice covers that comprise a multitude of glaciers of varying sizes in regions with complex surface topography.

Direct measurements of the mass balance of Arctic glaciers are limited to a small number of glaciers across the Arctic (50 in the 2000 to 2004 pentad); 31 of these were in Scandinavia, which contains only 3.5% of the total mountain glacier and ice cap area in the Arctic. In the past 35 years, the number of *in situ* glacier monitoring sites in the Canadian Arctic and Arctic Russia (regions which contain over 50% of the area of mountain glaciers and ice caps in the Arctic) has declined from 16 in the 1970 to 1975 period, to just six in the period 2000 to 2005, with no measurements in Arctic Russia since 1990 and only one set since 1980 (Dyurgerov and Meier, 2005). There is a pressing need for an updated regional-scale assessment of glacier and ice cap mass balance, the last assessment having been made in 2005. The uncertainties associated with such regional-scale assessments are large owing to the small number of *in situ* measurements and their uneven spatial distribution relative to the distribution of glacier ice.

In situ measurements are critical for quantifying regional mass balance, enhancing process understanding, and validating remote sensing techniques and model predictions. Given the limited number and distribution of such measurements, it is essential to improve and sustain remote sensing capabilities for monitoring ongoing changes in glacier extent, surface elevation and thickness, surface mass balance, ice dynamics, and iceberg production. Although there has been some progress on these issues, there is also a need for repeated, regional-scale mapping of parameters that provide simple indices of glacier mass balance (such as equilibrium-line altitude, glacier facies zones, and summer melt duration; see Section 7.4.2) to be used to evaluate year-to-year variability in climate effects on glacier health.

7.2. Climate evolution in glacierized regions of the Arctic

- While some Arctic glaciers have existed throughout the period since the end of the last glaciation, most disappeared during a warm period between 10 000 and 6000 years ago. Many of today's glaciers, therefore, formed in a cooler period after 5000 years ago.

- Although there were several warmer intervals within this cool period, the 20th century appears to have been the warmest century in the past 2000 years, with the warmest conditions occurring between the 1930s and early 1960s and since the mid-1980s.

- Across much of the Arctic, low winter precipitation means that most of the year-to-year variability in glacier mass balance arises from changes in summer temperature. In more maritime regions such as southern Alaska, Iceland, western Scandinavia and Svalbard, variability in winter precipitation can be an important influence on mass balance variability.

7.2.1. Holocene climate

Ice formed during the last glaciation is found in some ice caps in the Canadian Arctic (Koerner and Fisher, 2002; Zdanowicz et al., 2002) and Severnaya Zemlya (Kotlyakov et al., 1991). The absence of such ice in ice caps in Svalbard and other parts of the Russian Arctic is evidence for substantial retreat (or even disappearance) of glaciers and ice caps across much of the Arctic outside Greenland in the early Holocene (Koerner and Fisher, 2002). Therefore, the genesis of much of the present-day Arctic glacier cover may lie in the climate of the middle and late Holocene.

Many climate proxy records suggest that the warmest period of the Holocene in the Arctic was between 10 000 and 6000 years BP, after which the climate cooled, possibly because of a decrease in incident solar radiation at high latitudes due to changes in the Earth's orbital parameters (Koerner and Fisher, 1990; Vinther et al., 2009). Air temperature reconstructions based on ice-core oxygen isotope records from the Agassiz Ice Cap (Canadian Arctic) and Renland (eastern Greenland) suggest peak temperatures slightly in excess of 2 °C warmer than at present during this period (Vinther et al., 2009), which is often referred to as the Holocene Thermal Maximum. However, the exact timing of peak air temperatures during this period seems to have varied across the Arctic. In North America, it was later in the eastern Arctic than in the western Arctic, probably because the residual Laurentide Ice Sheet had a cooling effect on the climate in the east. During the Thermal Maximum, around 7500 years BP, summer temperatures were about 1.6 ± 0.8 °C warmer than in the 20th century (Kaufman et al., 2004). Records of summer melt and oxygen isotope ratios from ice cores in the Canadian and Russian Arctic indicate that cooling following the Thermal Maximum was underway by no later than 6000 years BP (Koerner and Fisher, 2002) (Figure 7.6).

A 2000-year, decadally-resolved, multi-proxy record of Arctic summer surface temperatures (which includes the temperature reconstructions from Agassiz Ice Cap and Renland) shows a cooling trend from 2000 to 100 years BP (Kaufman et al., 2009) (see Fisher, 2002, for an analysis of the reliability of multi-proxy temperature reconstructions). This trend (-0.22 °C/millennium) was probably driven by the orbitally-forced trend in summer insolation at high latitudes over the same period, but may have been amplified in some regions by feedbacks involving the progressive expansion of Arctic sea ice (England et al., 2008; Vare et al., 2009) and seasonal snow cover. The record also shows centennial-scale anomalies about the long-term trend, with the period 450 to 700 AD being cooler than the long-term trend and the period 900 to 1050 being warmer. The coldest period in the record occurred between 1600 and 1850, during the Little Ice Age (Kaufman et al., 2009). There is evidence that this cold period was associated with a persistently negative phase of the North Atlantic Oscillation (Trouet et al., 2009). Although the decline in orbitally-driven summer insolation continued throughout the 20th century, reconstructed Arctic summer temperatures rose sharply, reaching the highest values in the 2000-year record in the period after 1950. On the basis of this reconstruction and recent instrumental data, Kaufman et al. (2009) concluded that the decade 1999 to 2008 had the warmest summers in the 2000-year record (Kaufman et al., 2009).

Thousand years
Before Present (BP)
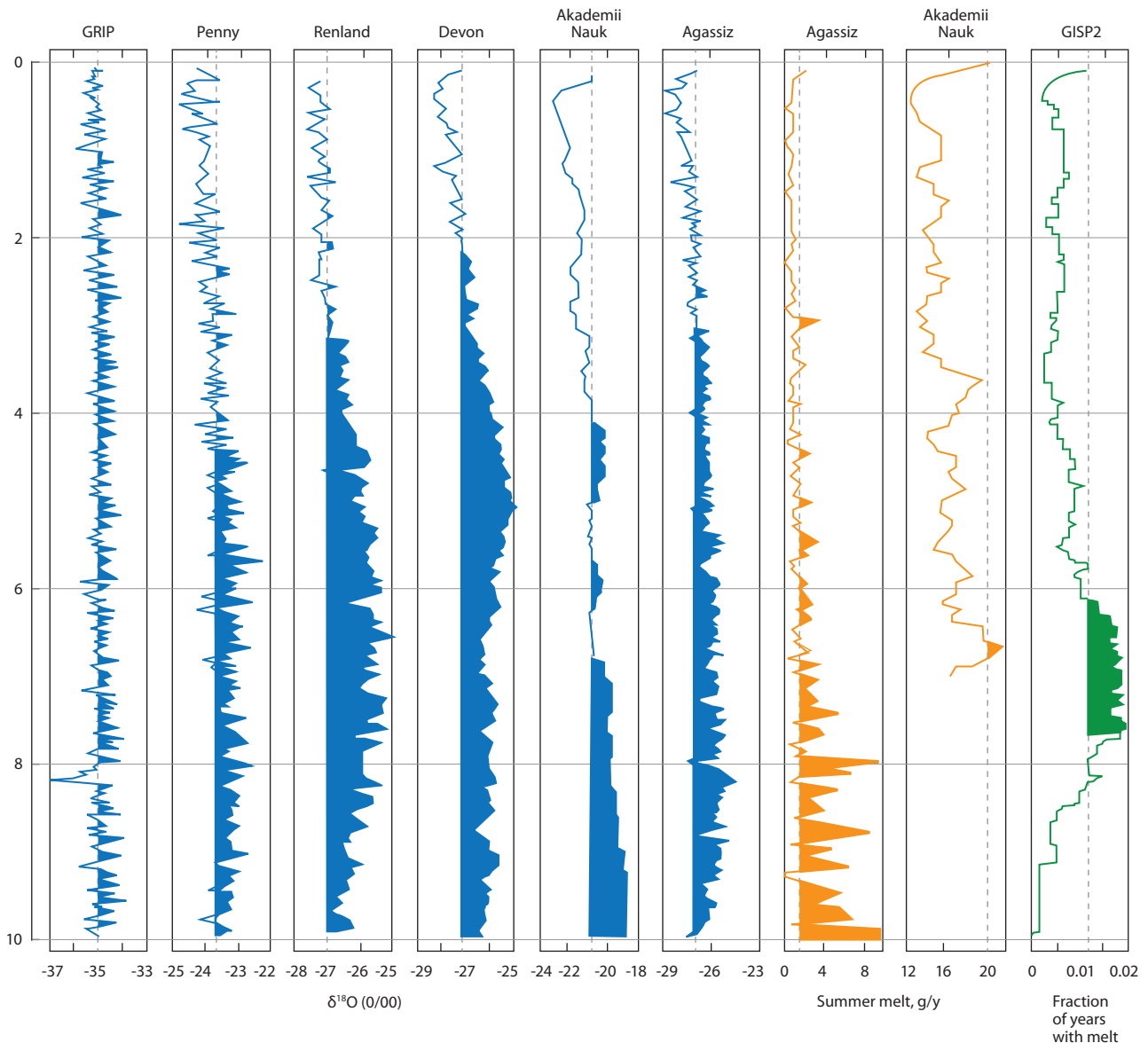
Figure 7.6. Holocene sections of ice cores from Greenland (GRIP, GISP2, Renland), Arctic Canada (Agassiz, Devon, Penny), and Severnaya Zemlya (Akademii Nauk) showing records of $\delta^{18}O$ (negative values) and summer melt (positive values), which are proxies for air temperature over the ice caps / ice sheet. Shaded sections suggest conditions warmer than those of today. Source: Koerner and Fisher (2002).

7.2.2. Arctic ice cores: Holocene climate records from glacierized regions

Ice cores from the larger Arctic ice caps yield records of past environmental conditions in glacierized regions that span the full Holocene. The longest records are from the Canadian Arctic, Mount Logan (Yukon), and Severnaya Zemlya (Koerner and Fisher, 1990, 2002; Kotlyakov et al., 1991; Fisher et al., 1995, 2008; Koerner, 1997; Zdanowicz et al., 2002; Kinnard et al., 2008). The melt layer record from the Agassiz Ice Cap and the oxygen isotope record from Severnaya Zemlya show a temperature maximum in the early Holocene, before 8000 BP, while oxygen isotope records from Arctic Canada typically show a later thermal maximum – as late as 5000 BP on Devon Island (Paterson et al., 1977). Melt layer records show a persistent cooling trend from about 8000 BP until the mid-19th century, after which there is renewed warming. Oxygen isotope records

also show cooling, but the timing of this is delayed until after about 5000 BP in some records (Koerner and Fisher, 2002). The melt records suggest maximum summer cooling from the Holocene Thermal Maximum of 2.0 to 2.5 °C, while the oxygen isotope records suggest cooling of 1.3 to 3.5 °C (Koerner and Fisher, 2002). Part of the early Holocene warming recorded in the ice core records must be due to the effects of surface lowering due to ice cap thinning, although this would have been partly offset by cooling induced by isostatic uplift of the land surface as the ice cover thinned and shrank (Koerner and Fisher, 2002; Vinther et al., 2009).

Oxygen isotope records from ice cores from smaller ice caps in Arctic Canada, northern Greenland, Svalbard, and the Russian Arctic islands show temporal variability but not the long-term cooling trend apparent in the records from the larger ice caps. This suggests that the small ice caps began to re-grow in the latter part of the Holocene after melting away during the

Thermal Maximum (Hammer et al., 2001; Landvik et al., 2001; Koerner and Fisher, 2002). The record from the upper 125 m of the ice core from Hans Tausen Iskappe in northern Greenland suggests a warm period between 900 and 1100 AD, and that the period 1700 to 1900 was the coldest in the past 2000 years. There was strong warming from the 1920s to the early 1960s (making the 20th century the warmest part of the record), but no clear trend in temperatures thereafter (Hammer et al., 2001). New oxygen isotope records from Austfonna and Lomonosovfonna in Svalbard span the past 600 to 800 years and show gradual cooling to a minimum in the 19th century, followed by rapid warming around 1900 that made the 20th century the warmest century in the past 600 years (Isaksson et al., 2003, 2005). A 275-year oxygen isotope record from Akademii Nauk Ice Cap on Severnaya Zemlya shows the coldest conditions around 1790, and then continuous warming until 1935. It also shows that the 20th century was the warmest period in this region (Fritzsche et al., 2005). A reconstruction of snow accumulation rates on Lomonosovfonna shows an increase of 25% in the late 1940s (Pohjola et al., 2002). A similar increase is recorded on Severnaya Zemlya after 1935 (Opel et al., 2009).

The oxygen isotope record from Mount Logan is unusual in that it appears to be a proxy for changes in the moisture source region for precipitation, rather than air temperature (Fisher et al., 2008). Major changes in moisture source region seem to be associated with switches between strong, frequent El Niño conditions (associated with meridional patterns of atmospheric flow and water vapor transport towards the Yukon) and strong, frequent La Niña conditions (associated with more zonal flow and moisture transport towards the Pacific Northwest). La Niña periods are associated with reduced precipitation (and snow accumulation) in the southern Yukon, while the reverse is true for El Niño periods. Periods of enhanced meridional flow seem to have occurred around 4200 BP and between 8000 and 7000 BP, while the modern El Niño Southern Oscillation regime seems to have been initiated after 4200 BP (Fisher et al., 2008).

Chemically-based melt indices from the Lomonosovfonna ice core show very high melt in the 12th century (Grinsted et al., 2006), as do melt layer records from the Canadian Arctic. This suggests a warm episode in medieval times in these parts of the Arctic. This warm episode was followed by a long period of reduced melt lasting to the mid- to late 19th century, after which melt increased sharply. For example, the melt layer record from Akademii Nauk Ice Cap shows maxima in the 1840s, 1880s, and from 1900 to 1970, after which melt layer content dropped sharply (Opel et al., 2009). However, the increase in melt was delayed until about 1925 in the new record from Prince of Wales Icefield, Ellesmere Island (Kinnard et al., 2008). The first principal component of seven Arctic melt layer records spanning the period 1551 to 1956 explains 34% of the variance in these records. It shows a strong increase in melt layers starting around 1830 and peaking in the mid-20th century (Kinnard et al., 2008) and has a similar form to a multi-proxy summer temperature reconstruction for the Arctic (Overpeck et al., 1997). Melt layer records from multiple sites on the Devon Island Ice Cap show an increase in ice fraction of over 50% after 1989 relative to the period 1963 to 1988 (Colgan and Sharp, 2008), consistent with a shift to increasingly negative surface mass balance after 1987 (Gardner and Sharp, 2007).

7.2.3. Climate records for glacierized regions of the Arctic derived from instrumental observations and climate re-analysis

Observational records of past surface air temperature from around 70 stations in the Arctic (north of 62° N) since 1875 show large multi-decadal fluctuations. Maxima in mean annual air temperatures occurred in the 1930s and 1940s and in the past two decades, with minima before 1920 and from about 1955 to 1985 (Polyakov et al., 2003). The same trends are apparent for the region 68° to 90° N in the 250-km smoothed air temperature dataset from the Goddard Institute for Space Studies. Mean summer 2-m air temperature anomalies in this dataset show a similar history to mean annual anomalies, but the range of summer anomalies is smaller than for the mean annual air temperature. In addition, there is a period of high mean summer air temperature in the 1950s that is not apparent in the record of mean annual air temperature (Figure 7.7).

Correlations between surface air temperature measured at coastal weather stations and the monthly North Atlantic Oscillation index are positive for a region that includes northern Scandinavia, Svalbard, and the Eurasian Arctic islands, and negative for a region that covers most of Greenland and the Canadian Arctic (Polyakov et al., 2003). Since the late 1990s, surface air temperature anomalies over the Arctic Ocean have become increasingly positive in autumn, especially over those regions where there has been a strong decrease in September sea-ice extent (Serreze et al., 2009). This trend may have resulted in longer melt seasons over glacierized regions of the Canadian and Eurasian Arctic.

In most regions of the Arctic, low winter precipitation means that interannual variability in annual glacier mass balance arises from variability in the summer balance, which is strongly controlled by summer air temperature. Gardner et al. (2009) found positive correlations between mean lower-troposphere summer air temperature from climate re-analyses and surface air temperature measured at the summit elevations
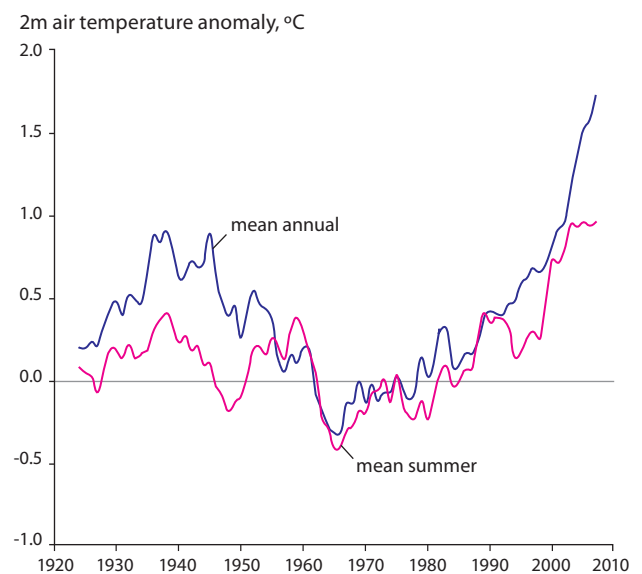
2m air temperature anomaly, °C

Figure 7.7. Five-year running means for mean annual and mean summer (June – August) 2-m air temperature anomalies (relative to the mean for 1951 to 1980) for the region 68° to 90° N from the Goddard Institute for Space Studies 250-km smoothed dataset.

of Arctic ice caps. Sharp and Wang (2009) reported similar correlations between lower tropospheric summer mean air temperature from climate re-analyses and melt season duration on Eurasian Arctic ice caps. For the period 1948 to 2008, NCEP/NCAR (U.S. National Weather Service National Centers for Environmental Prediction and the National Center for Atmospheric Research) Reanalysis (Kalnay et al., 1996) decadal mean 700 hPa air temperatures over glacierized regions of the Arctic in June to August were warmest in either the 1950s (Russian Arctic, northern Ellesmere and Axel Heiberg Islands) or the 2000s (Alaska, Iceland, Svalbard, and the rest of the Canadian Arctic). They were coldest in the 1990s in the Eurasian Arctic and the 1960s and 1970s over Iceland, Alaska, and the Canadian Arctic.

In the more maritime regions of the Arctic (such as Alaska, Iceland, Svalbard, and western Scandinavia), high variability in winter accumulation can also induce interannual variability in the annual mass balance. According to the NCEP/NCAR Reanalysis, decadal mean winter (September to May) precipitation anomalies were most negative in the 1950s over Alaska, the 1960s over Iceland, and the 1980s over Svalbard. They were most positive in the 2000s over Alaska and the 1950s over Iceland and Svalbard.

High summer precipitation keeps glacier surface albedo relatively high and reduces summer melt, while low summer precipitation has the opposite effect. In the Eurasian Arctic islands, decadal mean summer precipitation anomalies were typically positive from the 1950s through the 1970s and negative from the 1980s to 2000s. The opposite trend was observed in Iceland.

7.3. Changes in glacier extent, volume and total mass

- Following the early Holocene warm period, the cooler climate between 6000 years BP and the end of the Little Ice Age resulted in glacier growth and advances across the Arctic.

- In some cases, maximum glacier extents were reached as early as the mid-17th century, but in most regions glaciers were at or close to their maximum extents in the late 19th and/or early 20th centuries.

- Across the Arctic, glaciers began retreating and losing mass in the early 20th century. This trend continued through the mid-20th century, although glaciers in some regions experienced brief episodes of slower retreat, reduced mass loss, or even mass gain. Overall, there were substantial reductions in glacier area and mass across the entire Arctic over the 20th century.

- Mass loss has accelerated across most regions of the Arctic since the mid-1990s.

- In some regions, the timing and duration of episodes of relatively positive or negative mass balance were tied to major changes in atmospheric circulation related to phenomena such as the North Atlantic Oscillation (Iceland, western Scandinavia), the Pacific Decadal Oscillation (western Canada, southern Alaska), and the location and intensity of the summer circumpolar vortex (Arctic Canada).

7.3.1. Arctic glacier changes during the Holocene

Reconstructions of dated lateral and terminal moraine positions (and associated stratigraphy) are the primary source of information on glacier changes during the Holocene. In western North America, the Cordilleran ice sheet reached its maximum extent and thickness by 16 000 BP. Between 15 000 and 11 000 BP, climate conditions allowed some lobes of the ice sheet to advance (Menounos et al., 2009), and alpine glaciers also advanced in the southern and north-central ranges of the Cordillera, but by 11 000 BP glacier extent was comparable to that at present (Clague et al., 1982). Glacier extent reached a minimum between about 11 500 to 9000 BP, with temperatures 2 to 3 °C above present and generally drier conditions (Kaufman et al., 2004). Subsequently, periods of glacier advance occurred from 8600 to 8200 BP, 7400 to 6500 BP, 4400 to 4000 BP, 3500 to 2800 BP, and 1700 to 1300 BP (Menounos et al., 2009), and maximum Holocene extents were reached during the early 18th or mid-19th centuries AD.

A number of ice caps in Arctic Canada (Devon, Agassiz, Barnes, Penny) contain ice dating from the last glaciation (Paterson et al., 1977; Hooke and Clausen, 1982; Fisher et al., 1983, 1995, 1998; Zdanowicz et al., 2002) and must therefore have survived throughout the Holocene. During deglaciation, the margins of the last Innuitian ice sheet covering the Queen Elizabeth Islands reached the current margins of ice caps on Devon, Ellesmere, and Axel Heiberg Islands between 10 000 and 7500 ^{14}C years BP (England et al., 2006). Many of the smaller ice caps and glaciers in this region disappeared during this period (Koerner and Fisher, 2002). Glaciers began to grow again after 1000 ^{14}C years BP (Blake, 1981), reaching maximum extents in the late 19th or early 20th centuries. On Baffin Island, the Laurentide Ice Sheet retreated progressively throughout the Holocene to its current margin, that of the Barnes Ice Cap. Deglaciation was especially rapid around 7000 BP, when the ice cover in Foxe Basin collapsed (Briner et al., 2009). The early Holocene ice extent in this region is poorly constrained but glaciers began to advance again as early as 6000 BP, reaching extents similar to those of the Little Ice Age (the Holocene maximum for alpine glaciers on Baffin Island) by 3500 to 2500 BP (Briner et al., 2009). Small ice caps have been present in the northern interior of Baffin Island for 1600 to 2800 years, but they were absent for much of the middle Holocene. These ice caps started to expand after 1280 AD, and especially after 1450 (Anderson et al., 2008).

Most glaciers in Greenland were smaller during the early Holocene than at present and some areas that are currently occupied by mountain glaciers and ice caps may have become completely ice free in that period (Hammer et al., 2001). Neoglacial moraines have been reported (from western and southeastern Greenland), but in most regions historical advances were the most extensive after the early Holocene. In all areas apart from northern Greenland (where many ice margins are stationary), local glaciers are currently receding from their historical maximum extents (Kelly and Lowell, 2009).

Iceland was largely ice free during the early Holocene (8000 to 6000 BP), after which Neoglacial cooling began. Glacier extent increased between 4500 and 4000 BP, and again between 3000 and 2500 BP. The most extensive Holocene ice extent was

reached during the period 1250 to 1900 AD – probably in the latter part of that period between 1700 and 1850 (Geirsdóttir et al., 2009). However, some steep alpine glaciers reached their maximum extents earlier, probably around 2500 BP, when the present ice caps are also likely to have formed. A general glacier recession that began in the 1890s accelerated after 1930. Retreat slowed by the 1960s as a result of cooler summers beginning in the 1940s, and many steep glaciers were advancing in the 1970s. Renewed climate warming since 1985 has led to a new episode of retreat, and all major non-surging outlet glaciers have been retreating since 1995, at rates of up to 100 m/y. The southern outlets of Vatnajökull eroded down into soft sediments during the pre-1890 advance and have been especially susceptible to rapid retreat since then. The volume of Vatnajökull has decreased by about 300 km³ (10%) since 1890 (Björnsson and Pálsson, 2008).

In northern Scandinavia, retreat of the Scandinavian Ice Sheet during the Late Glacial and early Holocene was punctuated by numerous readvances in the period 11 200 to 8000 BP. Most glaciers disappeared completely at some point in the early to mid-Holocene, and ice cover is likely to have reached a minimum between 6600 and 6300 BP (Nesje, 2009). Subsequently, there was a period of renewed glacier growth during which there were numerous century- to millennium-long 'Neoglacial' events. Glaciers in northern Sweden were probably at their Little Ice Age maximum extent in the 17th and early 18th centuries (Karlén, 1988), while those in Norway reached their maxima in the mid-18th century (Winkler, 2003; Nesje, 2009), or even later in some parts of northern Norway (Bakke et al., 2005). Most Swedish glaciers reached an extent close to their Holocene maximum at the beginning of the 20th century (Holmlund, 1993). Since then, Scandinavian glaciers have generally retreated, particularly since the 1930s (Nesje, 2009). Increased snow accumulation caused some glaciers to advance in the 1990s (Dowdeswell et al., 1997).

In northeast Svalbard, many glaciers started to retreat after the Last Glacial Maximum by about 9500 BP (Blake, 1989). During this period of deglaciation, variations of the glaciers and ice caps in Svalbard were similar to those of glaciers in Scandinavia. However, the Younger Dryas advance was apparently small due to a very dry climate, and the Little Ice Age maximum extent was larger than the Younger Dryas advance (Mangerud and Landvik, 2007). In the early Holocene, glaciers were much smaller than at present and perhaps even completely absent. A period of glacier growth from 3000 to 2500 BP was followed by a warmer period with smaller glaciers (Svendsen and Mangerud, 1997). Dating of relict vegetation found under the cold-based Longyearbreen indicates that late Holocene glacier growth started about 1100 BP. This period of glacier growth culminated in the Little Ice Age maximum with the greatest Holocene glacier extent. Many ice-cored moraines were formed at this time, which may have been as recently as around 1900 (Humlum et al., 2005). Surging of Svalbard glaciers is common and the maximum extent of many glaciers may be a result of surge dynamics (Hagen et al., 1993). Since the Little Ice Age maximum, the glaciers have generally retreated in response to early 20th century warming (Lefauconnier and Hagen, 1991; Hagen et al., 2003a,b).

In the Russian Arctic, outlet glaciers in northwestern Novaya Zemlya retreated from the outer coast by the earliest Holocene. Prior to about 9500 BP the terminus of the tidewater Shokal'sky Glacier was over 1 km behind its present margin, allowing incursion of the sea (Zeeberg, 2001). A sediment core from Russkaya Gavan spans a period of 800 years and allows reconstruction of the recent history of Shokal'sky Glacier. A noticeable advance around 1400 AD was followed by a major retreat by about 1600. Increased melting in the 1900s is suggested by an increase in sedimentation rates (Polyak et al., 2004). On Franz Josef Land, glaciers were less extensive than at present throughout the period 10 300 to 4400 ¹⁴C years BP, but re-expanded to their present margins by 2000 ¹⁴C years BP (Lubinski et al., 1999). A major advance occurred during the past 1000 years, with maxima around 1200, 1400, and in the mid-17th century. Glaciers were more extensive than now at the start of the 20th century; since then there has been widespread glacier retreat.

7.3.2. Arctic glacier changes in the 20th and 21st centuries

7.3.2.1. Methods for determining glacier mass balance

Mass balance is the change in mass of a glacier, quantified as the difference between mass gains (accumulation) and mass losses (ablation), over a specified time period. A summary of the processes that determine the mass balance of a glacier is provided in Appendix 7.1 (see also Box 7.1). Over the 20th and beginning of the 21st centuries, many new methods have been developed for measuring glacier mass balance. Each method provides information at unique spatial and temporal resolutions, and samples specific parameters to characterize glacier evolution over time. Important challenges for global mass balance inventories include the integration of these different datasets and quantification of the errors in each approach.

Long-term records of surface mass balance that form the basis of existing global datasets have been acquired from direct sampling of individual glaciers using glaciological methods (Østrem and Brugman, 1991; Cogley, 2005). In this approach, measurements of snow accumulation are acquired from snow probing or digging snow pits to measure snow depth and density (Figure 7.8). Glacier ablation is measured by determining the

Box 7.1. Units for mass balance measurements

Mass balance measurements have units of mass per unit time (see Appendix 7.1). It is also common to report the rate of change of mass per unit area, for example, kg/m² per year, which is numerically equivalent to the millimetres of water equivalent (w.e.) per year. In this report, mass balances are usually presented in units of gigatonnes per year (Gt/y) or metres of water equivalent per year (m w.e./y). Exceptions are those studies reporting cumulative mass balances, in which case they are not converted to a rate because it is not always clear what exact time period was used for the measurements, and those studies reporting volume changes, in which case they are not converted to a mass equivalent unless the density of the changing snow and ice volume is clearly stated.
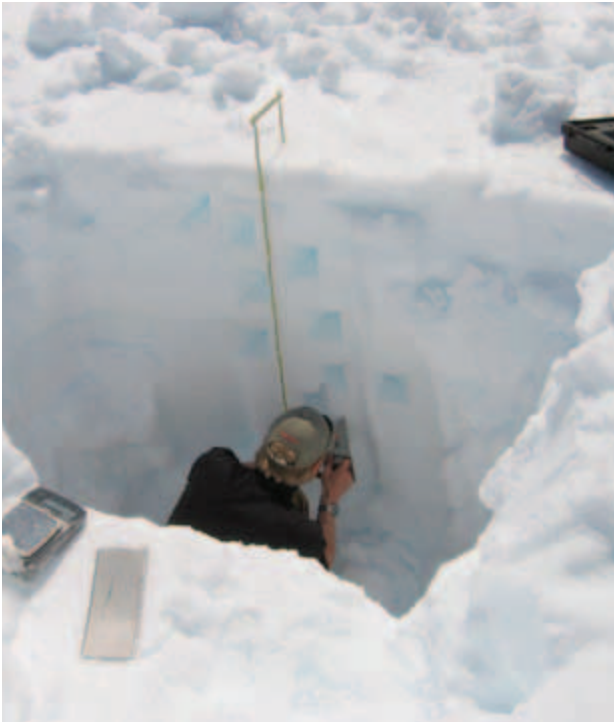
Figure 7.8. Measuring snow accumulation in a snow pit.

vertical change in distance from the glacier surface to a marker, fixed within the reference frame of the moving ice mass, and by assuming an appropriate density for the snow, firn, or glacier ice that melted. Annual mass balance is defined as the difference between the annual accumulation and the annual ablation. Point measurements of this quantity are extrapolated to the entire glacier. Two processes, internal accumulation and the formation of superimposed ice, involve the refreezing of some fraction of the annual melt and alter surface mass balance by reducing the amount of meltwater runoff. However, not all surface mass balance programs measure these processes owing to the difficulty of doing so, resulting in overestimates of the rate of mass loss. Regardless, glaciological mass balance measurements generally provide the most accurate sampling of mass balance at a point, but are limited in spatial extent.

Developments in ground surveying and remote sensing technologies have resulted in many geodetic mass balance methods. These calculate a volume balance determined from repeat measurements of surface elevation and area, from which the mass balance is obtained by estimating the density of the changing ice volume. Elevations are determined using overlapping aircraft or satellite imagery to determine the 3-dimensional position of the surface through stereoscopy, ground mapping of the glacier surface elevation using theodolite or Global Positioning System (GPS) surveying, or airborne / satellite altimetry using radar or laser ranging instruments. Geodetic methods provide a broad regional sampling of glacier changes, but can have large errors due to uncertainties in the density of the volume lost, especially over short periods, non-representative sampling of fast- and slow-flowing sectors of ice caps, or biases in the distribution of surface-elevation change measurements relative to the distribution of rates of elevation change (e.g., in aircraft altimetry where measurements are typically made along a glacier centerline where rates of thickness change may be

larger than the glacier-wide mean; Berthier et al., 2010).

Recent advances in spaceborne geodesy provide gravimetric data to determine glacier mass variations via measurement of changes in Earth's gravity field. At present, this is achieved using the Gravity Recovery and Climate Experiment (GRACE) satellites that measure variations in Earth's gravitational potential. Corrections must be made to the satellite measurements to account for all non-glacier sources of mass variation, such as glacial isostatic adjustment and changes in terrestrial water, atmospheric, and ocean mass. Gravimetric methods circumvent the need for density corrections of measured volume changes, but are limited in spatial resolution (about 50 000 km^2) and can only effectively sample broad regions undergoing significant change.

Measurements of glacier thickness are essential for estimating glacier volume and for computing mass losses by iceberg calving. They are also needed as input to models of the dynamic response of glaciers to climate forcing. They are usually obtained using the technique of radio echo sounding from either ground-based or airborne platforms (Dowdeswell and Evans, 2004). Airborne campaigns are typically required for measurements of the thickness of larger glaciers and ice caps, as well as for regional-scale studies (Dowdeswell et al., 2002, 2004).

Area/volume scaling (Bahr et al., 1997) is used to estimate glacier volume when glacier areas are known but direct measurements of volume are lacking. This method is based on the observed power-law relationship between a glacier's volume and area. Glacier surface areas measured at different times are used to estimate corresponding glacier volumes, from which changes in glacier volume can be determined. This method can have large errors when data are not available to constrain the scaling relationship and determine suitable scaling coefficients (Radić et al., 2008). Therefore, it is most appropriately applied to entire populations of glaciers rather than to individual ice masses.

Changes in proxy indicators are also used to approximate mass balance trends, and these approaches are summarized in Section 7.4. The distribution and length of glaciological and geodetic mass balance records in the Arctic region are highly variable, but the longest continuous records (from Scandinavia) extend back to the 1940s (Figure 7.9). Records from individual regions are discussed in the following sections.

7.3.2.2. Northwestern North America

Northwestern North America includes all glaciers of Alaska (United States) and Yukon Territory (Canada) and those glaciers in British Columbia and Alberta (Canada) that drain into the Alaska Panhandle or into the Arctic Ocean via the northern tributaries of the Mackenzie River.

During the 20th century most land-terminating glaciers in northwestern North America retreated extensively from their Little Ice Age maximum extent (Kaufman et al., 2004). Most of the recession occurred between 1900 and 1950 and from 1980 to present, with a period of slower recession, and in some cases advance, between 1950 and 1980. Since 1980, nearly all glaciers have been in a state of retreat. For example, in British Columbia, the extent of glaciers in the northern Coast and St. Elias mountains changed by -7.7 ± 2.9% and -7.9 ± 2.6%,

Figure 7.9. Glacierized regions in the Arctic and the locations where mass balance measurements have been made. The length of time of the geodetic and glaciological mass balance records are also indicated. Source: AMAP.
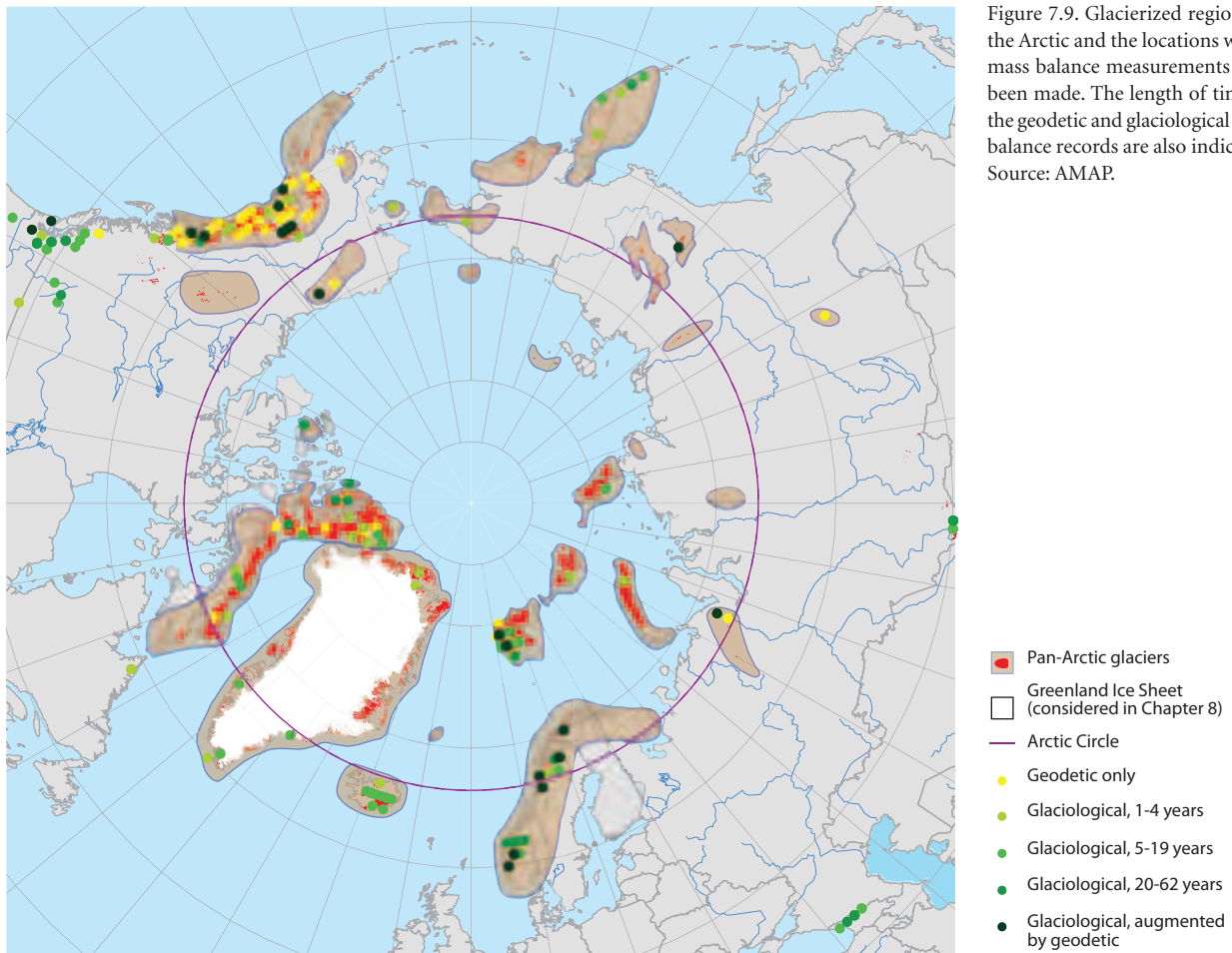
respectively, over the period 1985 to 2005 (Bolch et al., 2010). Between 1958/1960 and 2006/2008, the total glacier area in the Yukon Territory decreased by 2541 ± 189 km^2 (or 22% of the initial area of 11 622 km^2) (Barrand and Sharp, 2010).

Regular glaciological mass balance measurements in Alaska began during the 1960s and trends generally agree with observed variations in glacier extent. Wolverine Glacier lost mass from 1966 to 1976 and gained mass from 1976 to 1988, while Gulkana Glacier lost mass at a constant rate from 1966 to 1973 and at an increasing rate from 1974 to 1985 (Josberger et al., 2007). After about 1988, both glaciers entered a period of accelerating mass loss (Figure 7.10).

Aircraft laser altimetry measurements showed that about 85% of 67 glaciers in northwestern North America had a mass balance of -52 ± 15 Gt/y between 1950/1970 and the early 1990s (Arendt et al., 2002; Echelmeyer et al., 1996). A repeat sampling of 28 of these glaciers between the early 1990s and late 1990s / early 2000s indicated a doubling of the rate of mass loss (-96 ± 35 Gt/y). A recent analysis compared digital elevation models (DEMs) from the 1950s/1970s against those from the 2000s, and argued that estimates based on centerline aircraft altimetry overestimated the mass loss of Alaskan glaciers by 34% (Berthier et al., 2010). These new estimates suggest a mass loss rate of -41.9 ± 8.6 Gt/y for the period 1962 to 2006.

Other geodetic measurements have been acquired through differencing of 1950s U.S. Geological Survey (USGS) and NASA's Shuttle Radar Topography Mission (SRTM) DEMs. Glacier mass balances in regions south of 60.5° N in southeastern Alaska and the Kenai Mountains were -15 ± 4.0 Gt/y (Larsen et al., 2007) and -1.2 ± 0.25 Gt/y (VanLooy et al., 2006), respectively, during a 30- to 50-year period ending in 2000. Another study using the SRTM data found the glaciers of the northern Coast and St. Elias mountains had a mass balance of -11.1 ± 2.0 Gt/y during the period 1985 to 2000 (Schiefer et al., 2007). Recent altimetry measurements for 2004 to 2008 suggest an acceleration of mass loss relative to the 1990s, especially for glaciers at low elevations (Lingle et al., 2008).

There are three estimates for the mass balance of glaciers surrounding the Gulf of Alaska based on data from GRACE: -110 ± 30 Gt/y for the period 2002 to 2004 (Tamisiea et al., 2005); -101 ± 22 Gt/y for the period 2002 to 2005 (Chen et al., 2006); and -84 ± 5 Gt/y for the period 2003 to 2008 (Luthcke et al., 2008).

7.3.2.3. Canadian Arctic

The Canadian Arctic region includes the Queen Elizabeth Islands (including Ellesmere, Devon, and Axel Heiberg Islands), and Bylot and Baffin Islands. Glaciers cover an area of about 150 000 km^2 in this region. Current and past glacier extents are known from aerial photography taken during the period 1956 to 1961 and from satellite imagery collected since the mid-1970s (and especially since 1999) (Sharp et al., 2003, in press; Dowdeswell et al., 2007; Wolken et al., 2008). Comparison of ice extents from around 1960 with trimline and end moraine positions visible on satellite imagery shows that there has been extensive ice retreat across the region from glacier maximum positions reached by the end of the Little Ice Age. The ice-

covered area on the Queen Elizabeth Islands decreased by 37% from a Little Ice Age maximum of 170 124 km^2 to about 107 735 km^2 in 1960, and all permanent ice was removed from most of the western islands (Dyke, 1978; Wolken et al., 2008). Further glacier retreat has occurred since 1960, with ice area reductions by 2000/2001 of 5.1% (253 km^2) on Bylot Island (Dowdeswell et al., 2007) and 2.7% (2844 km^2) on the Queen Elizabeth Islands (Sharp et al., 2003). The 14 400 km^2 Devon Island Ice Cap decreased in area by 338 ± 40 km^2 (2.4%) (Burgess and Sharp, 2004). The area of small plateau ice caps in interior northern Baffin Island decreased by 55% (from 150 km^2) between 1958 and 2005 (Anderson et al., 2008), while the area of 264 glaciers on the Cumberland Peninsula in southeast Baffin Island decreased by about 294 km^2 (12.5%) between around 1920 and 2000 (Paul and Svoboda, 2009). In the latter study, proportional area reductions were greatest for the smallest glaciers, and the rate of area loss roughly doubled after 1975.

Glaciological mass balance measurements have been made in the Queen Elizabeth Islands since the late 1950s (Cogley et al., 1996; Koerner, 2005) (Figure 7.10), but there are no long-term records from Bylot or Baffin Islands, and no regular measurements since the mid-1980s (Hooke et al., 1987). In this region, variability in the annual mass balance arises almost entirely from variability in the summer balance (Koerner, 2005). The cumulative surface mass balance of all measured glaciers in the region has been negative over the measurement period (e.g., Devon Island Ice Cap -4.4 m w.e. from 1961 to 2007; White Glacier, Axel Heiberg Island -6.8 m w.e. from 1960 to 2007). Compared with the 1963 to 1987 period, the rate of mass loss accelerated significantly after 1987 (Gardner and Sharp, 2007). There have been a number of very negative mass balance years since 1997 (notably 1998, 2001, 2005, 2007, 2008, 2009) that suggest a further acceleration of mass loss since 1997. The

cumulative mass loss from the Devon Island Ice Cap from 1997 to 2007 exceeds that for the period 1963 to 1997. Consistent with this trend, geodetic measurements on the Barnes Ice Cap suggest a progressive increase in rates of thinning from 0.12 m/y (1970 to 1984), to 0.76 ± 0.35 m/y (1984 to 2006), and 1.0 ± 0.14 m/y (2004 to 2006) (Sneed et al., 2008).

Iceberg calving also contributes to mass loss from many of the larger ice caps (Burgess et al., 2005). Separate estimates of surface mass balance and iceberg calving fluxes are available for two large ice caps in Arctic Canada: the Devon Island Ice Cap and Prince of Wales Icefield. For the Devon Island Ice Cap, the estimates equate to an overall mean annual balance of either -2.3 Gt/y (1963 to 2000) or -5.2 ± 1.7 Gt/y (1980 to 2006), depending on whether the net surface balance estimates of Mair et al. (2005) (-1.8 ± 0.8 Gt/y) or Gardner and Sharp (2009) (-4.7 ± 1.7 Gt/y) are added to the Burgess et al. (2005) estimate of mass loss due to iceberg calving (-0.5 Gt/y). The surface mass balance of the Prince of Wales Icefield (1963 to 2003) appears to be within an error of zero, so the overall negative balance (-2.0 ± 0.5 Gt/y) is similar to the annual calving flux (1.9 ± 0.2 Gt/y) (Mair et al., 2009).

Repeat airborne laser altimetry flown over some of the larger ice caps in the region in 1995 and 2000 showed that most ice caps were thickening or maintaining a constant thickness at elevations above 1600 m, and thinning at lower elevations (Abdalati et al., 2004). Rates of thinning were greater on Baffin Island ice caps (>1 m/y) than in the Queen Elizabeth Islands (< 0.5 m/y). Based on these data, Abdalati et al. (2004) suggested that the total mass balance of Arctic Canada in the period 1995 to 2000 was -22.9 Gt/y.

Thus, measurements of glacier extent, thickness, and surface mass balance are all consistent in showing progressive and possibly accelerating shrinkage of the Canadian Arctic land

Cumulative area-averaged mass balance, 10^3 kg/m^2
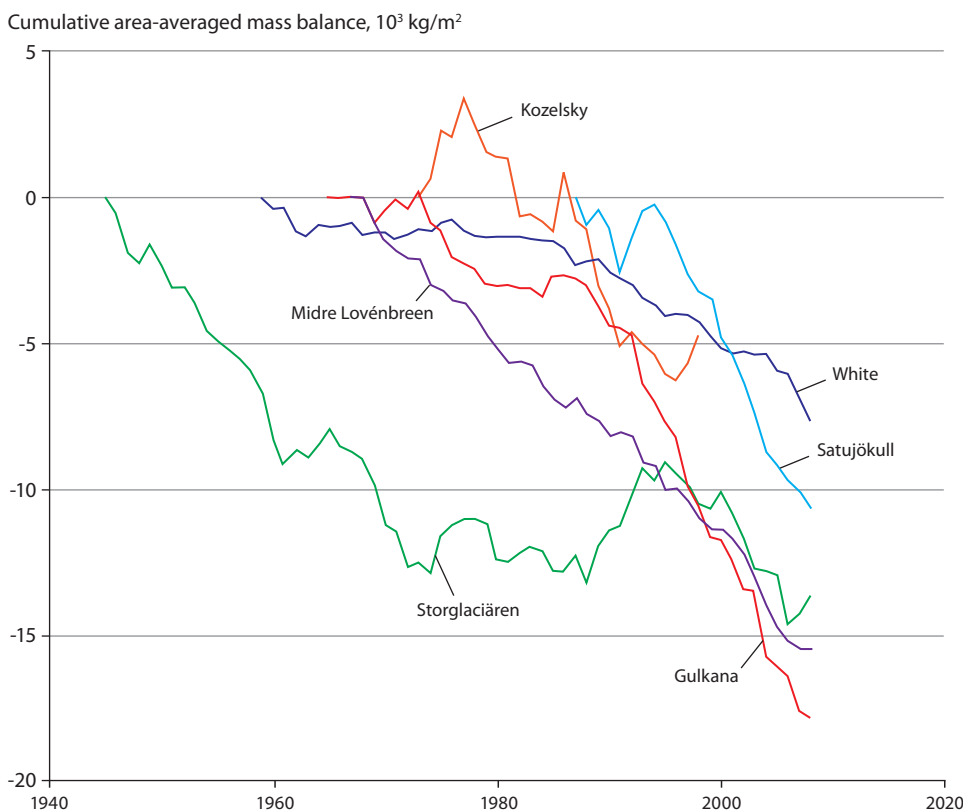
Figure 7.10. Cumulative area-averaged mass balance for six Arctic glaciers: White Glacier (Axel Heiberg Island), Gulkana Glacier (Alaska), Storglaciären (Sweden), Midre Lovénbreen (Svalbard), Satujökull (Hofsjökull, Iceland), and Kozelsky (Kamchatka) over the period of record. All records show a negative cumulative balance (net thinning) over the period of record and more rapid thinning after about 1990 (although the precise timing of this varies between regions). Source: AMAP.
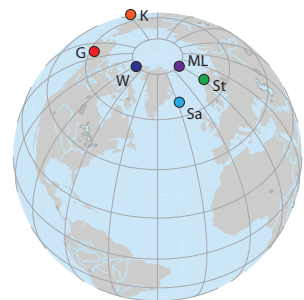
ice cover. Nonetheless, a precise and regionally comprehensive estimate of the magnitude of these changes and their implications for global sea level is not yet available.

7.3.2.4. Greenland (excluding Greenland Ice Sheet)

Glaciers occupy about one quarter of the peripheral area around the Greenland Ice Sheet (Thomsen and Weidick, 1992) (Figure 7.11). The glaciers of Disko Island (Qeqertarsuaq) and the Nuussuaq and Svartenhuk peninsulas, West Greenland, that covered an area of 3832 km^2 in 2001, lost between 16% and 21% of their area from the end of the Little Ice Age to 2001 (Citterio et al., 2009). There were no noticeable changes in the position of most calving glacier fronts between the late 19th century and the mid-1980s (Weidick, 1995). Between 1978 and 1991, major tidewater glacier termini along the northern coast of Kangerlussuaq and the southern part of Blosseville Kyst showed little change or a slight loss (0.1 to 0.5 km^2) in the glacier tongue area (Dwyer, 1995).

Extreme fluctuations of the terminus of the surge-type tidewater glacier Sortebrae involved a retreat of about 1.5 km between 1933 and 1943, a surge advance of about 10 km by 1950, a retreat of 8 km between 1950 and 1981, followed by stagnation between 1981 and 1992, and a second surge advance of over 5 km between 1992 and 1994 (Jiskoot et al., 2001). Another example of surge-related terminus fluctuations is shown by a small land-based Scoresby Sund glacier, which retreated 0.5 km between 1987 and 2000, and advanced by 2.8 km between 2000 and 2007 (Jiskoot and Juhlin, 2009). Neighboring glaciers did not change their terminus positions significantly during the same period.

7.3.2.5. Iceland

Glacier terminus variations in Iceland have been monitored regularly since 1930. Most non-surge-type glaciers retreated rapidly from 1930 to 1960, but more slowly in the 1960s. Glaciers began to advance in the 1970s and rates of advance peaked in the mid-1980s. In the 1990s, glaciers began to retreat again and all monitored glaciers were retreating by 2000. Terminus fluctuations follow changes in mean summer temperature with a delay of only a few years (Sigurðsson et al., 2007).

Mass balance measurements were initiated on Hofsjökull in 1987/1988 (Figure 7.10; Satujökull), Vatnajökull in 1991, and Langjökull in 1996. Vatnajökull's balance was positive from 1991 to 1994 and near zero in 1994 to 1995, but it has been negative ever since, with a cumulative specific balance of about -9.2 m w.e. (84 Gt, or 2.7% of the ice cap mass) from 1994 to 2006 (Björnsson and Pálsson, 2008). In the 1990s, mass loss from Vatnajökull due to basal melting resulting from geothermal activity and volcanic eruptions averaged 0.55 Gt/y, about 4% of the total surface ablation in a year of zero surface balance. However, the Gjalp volcanic eruption in 1996 melted about 4.0 Gt of ice. The net balance of Langjökull has been negative since measurements began and the cumulative specific net balance from 1996 to 2006 was about -12.8 m w.e. (13.1 Gt, 7% of the ice cap mass) (Björnsson and Pálsson, 2008). The current period of negative mass balance reflects a combination of high summer temperatures and low winter snowfall.

Figure 7.11. Mountain glaciers and ice caps occupy much of the peripheral areas of Greenland that lie beyond the margins of the Greenland Ice Sheet. Source: Anthony Arendt, University of Alaska.

7.3.2.6. Scandinavia

After their Little Ice Age maximum, Scandinavian glaciers retreated slightly until the 1930s/1940s, when there was a period of more rapid retreat that lasted into the 1950s for maritime outlet glaciers and the 1970s for more slowly responding outlets in the interior. Increased winter precipitation between 1988/1989 and 1994/1995 triggered a readvance of some glaciers in western Norway and Sweden, but higher summer air temperatures after 2001 induced an episode of rapid glacier retreat (Andreassen et al., 2005; Nesje, 2005, 2009). The longest mass balance records from Scandinavia date to the late 1940s (e.g., Holmlund et al., 2005), but the number of measurements increased in the early 1960s. Since that time, glaciers in inland Scandinavia (for which net balance variability is due largely to variability in the summer balance) have had a cumulative negative mass balance (Figure 7.10), while maritime glaciers in western Norway and some others such as Storglaciären in Sweden (for which net balance variability is due largely to variability in the winter balance) have had a cumulative positive net balance (especially in the 1990s when winter precipitation was unusually high) (Nesje, 2009). Since 2001, however, the mass balance of Norwegian glaciers has become predominantly negative. The North Atlantic Oscillation has a strong influence on the variability of the mass balance of maritime glaciers in western Norway (Reichert et al., 2001).

7.3.2.7. Svalbard

Over the past century, there has been a general trend toward retreat and thinning of Svalbard glacier fronts (Dowdeswell, 1986), excluding those that have surged during the period of observation, and thickening of higher elevation regions (Bamber et al., 2004; Nuth et al., 2007, 2010; Kohler et al., 2007; Moholdt et al., 2010a,b). Mass balance has been measured on two Svalbard glaciers (Austre Bröggerbreen and Midre Lovénbreen) since 1967 and on two more (Kongsvegen and Hansbreen) since the late 1980s (Hagen et al., 2003b). Almost all annual net balances are negative, the most notable exceptions being in 1987 and 1991 when all the measured balances were positive. There is no significant long-term trend in any of the annual balance records. Large annual variations are driven mainly by variations in the summer ablation. The cumulative balances of Austre Bröggerbreen (1967 to 2007) and Midre

Lovénbreen (1968 to 2007) were -20.3 and -15.3 m w.e., respectively, while that of Hansbreen (1989 to 2007) was -6.2 m w.e. (WGMS, 2009). Kongsvegen had a much smaller negative annual balance and a cumulative balance of -1.6 m w.e. (1987 to 2007) (WGMS, 2009), indicating that larger glaciers with high-elevation accumulation areas are experiencing more favorable mass balance conditions than smaller, low-lying glaciers. This is also the case for Austfonna in northeastern Svalbard, which had a surface net balance of -0.05 m/y w.e. for the five-year period 2004 to 2008 (Moholdt et al., 2010a). Shallow ice cores from Austfonna also indicated a net surface mass balance close to zero for the period between the Chernobyl fallout in 1986 and 1999 (Pinglot et al., 1999).

Hagen et al. (2003b) used the hypsometry (area-altitude distribution) of the glaciers and curves of net balance versus altitude for 13 regions to estimate the surface balance of Svalbard glaciers over the period 1968 to 1998. They found it to be slightly negative (-0.5 ± 0.1 Gt/y), or a mean specific surface net balance of -0.014 ± 0.003 m w.e./y. Mass loss by iceberg calving was estimated to be -4.0 ± 1.0 Gt/y (Dowdeswell and Hagen, 2004), making the total net balance of the archipelago significantly more negative, in total -4.5 Gt/y, or -0.12 ± 0.03 m w.e./y. A more recent estimate of the Svalbard calving flux (2000 to 2006) is -6.8 ± 1.7 Gt/y (Błaszczyk et al., 2009).

Geodetic measurements suggest that the rate of volume loss from a number of glaciers, especially low-lying glaciers close to the west coast, has increased by up to a factor of four since the early 1960s (Kohler et al., 2007). This is consistent with summer temperatures recorded at Longyearbyen, which increased by 0.04 °C/y from 1967 to 2006 and 0.17 °C/y from 1997 to 2006 (Førland and Hanssen-Bauer, 2003). Nuth et al. (2010) compared satellite altimetry from ICESat (2003 to 2007) to older topographic maps and digital elevation models (1965 to 1990) and estimated the geodetic mass change over the past 40 years for Svalbard, excluding Austfonna and Kvitøya, to be -9.7 ± 0.6 Gt/y (surface mass balance of -0.36 ± 0.02 m w.e./y). When the Austfonna estimate is added, this gives an overall mass balance for Svalbard of nearly -13 Gt/y. Moholdt et al. (2010b) used repeat track ICESat laser altimetry to assess mass and volume changes of Svalbard glaciers from 2003 to 2008. During this period, most regions experienced thinning at low elevations and thickening at high elevations. The geodetic mass balance for all of Svalbard's glaciers (not taking into account the effects of advance and retreat of calving glacier termini) was estimated to be -4.3 ± 1.4 Gt/y. The largest mass losses were in western and southern Svalbard, while Austfonna and glaciers in northeastern Spitsbergen gained mass over the period. Thus, available estimates of Svalbard net mass balance range from -4.3 to -14 Gt/y, depending upon the period of measurement and which calving flux estimate is used.

7.3.2.8. **Russian Arctic islands**

The major glacierized regions in the Russian Arctic are in the archipelagos of Franz Josef Land, Severnaya Zemlya, and Novaya Zemlya. Between 1952 and 2001, the glacierized areas decreased by 2.7% on Franz Josef Land, 0.4% on Severnaya Zemlya, and 1.4% on Novaya Zemlya. The total area reduction was about 725 km² (-375 km² on Franz Josef Land, -284 km² on Novaya Zemlya, and -65 km² on Severnaya Zemlya) (Glazovsky and

Macheret, 2006). In Novaya Zemlya, net glacier recession has been greater on the west coast than on the east coast (Glazovsky, 2003). Using local area-volume scaling relationships, volume losses from the three archipelagos between 1952 and 2001 were estimated at -71 km³ (Franz Josef Land), -24 km³ (Severnaya Zemlya), and -150 km³ (Novaya Zemlya). Assuming the volume was lost as ice, this gives a mean annual mass loss of -4.5 Gt/y from the three archipelagos (Glazovsky and Macheret, 2006).

Limited measurements of the surface mass balance of Shokal'sky Glacier on Novaya Zemlya, supplemented by reconstructions based on regression analyses between winter temperature / accumulation and summer temperature / ablation, suggest predominantly negative balances between 1930 and the mid-1990s (Zeeberg and Forman, 2001). Direct measurements on Vavilov Ice Cap (Severnaya Zemlya) between 1974 and 1988 also showed a slightly negative surface balance (-0.03 m w.e./y) (Barkov et al., 1992).

7.3.2.9. **Russian mountains**

The USSR glacier inventory mapped the glaciers of northeastern Siberia from aerial photography. Glacier changes have been determined by comparison with 2002/2003 Landsat7 ETM+ imagery. The area of glaciers of the Suntar Khayata mountain range (Figure 7.12) decreased from 200 km² in 1945 to 162 km² in 2002/2003, while that of glaciers in the Chersky mountain range decreased from 156 km² in 1970 to 113 km² in 2002/2003 (Ananicheva et al., 2006). The area of the Byrranga glaciers on the Taymir Peninsula decreased from 30.5 km² in 1967 to 24.4 km² in 2003, a reduction of 20%. South- and east-facing glaciers had the largest areal reductions, consistent with greater warming trends and smaller influence of westerly-derived cyclones on the eastern side of the mountain range (Ananichava and Kapustin, 2010).

On the Koryak Upland, a peninsula between Kamchatka and Chukotka, 248 glaciers were mapped from maps and aerial photographs dating from the 1950s and using Landsat 7 ETM+ and ASTER data from 2003. Their combined area decreased by 66.5%, from 176.6 km² to 54.4 km², over the intervening period. Glaciers with a northeasterly aspect showed the largest proportional area reductions. These glacier area changes are the largest recorded in the Russian sub-Arctic and reflect a

Figure 7.12. Mountain glaciers in the Suntar Khayata Range of northeastern Siberia. Note the well-developed meltwater drainage systems on the glacier surfaces. Source: Maria Ananicheva, Russian Academy of Science.

response to a combination of climate warming and decreased snowfall from Pacific sources (Ananichava and Kapustin, 2010).

At the end of the 1950s, there were 143 glaciers in the Ural Mountains with a total area of 28.7 km^2 (Osipova, 2006). From 1958 to 2001, mass balance observations were carried out on two Polar Ural glaciers (IGAN Glacier, area 0.73 km^2, and Obruchev Glacier, area 0.35 km^2). These records have been extended to cover the entire 20th century by correlation with local measurements of summer air temperature and winter precipitation. The resulting series show periods of negative balance between 1900 and 1920, in the late 1950s and 1960s, and since 1980. Positive balances occurred between 1920 and the mid-1950s (Kononov et al., 2005). From 1953 to 2000 (1981 for IGAN Glacier), the average area decrease rate of the four glaciers in the region (Obruchev, MGU, Chernov, IGAN) was 0.9% per year (Glazovsky et al., 2007).

7.3.2.10. Kamchatka

There are currently 448 glaciers in Kamchatka, with an area of about 874 km^2. The climate of Kamchatka is strongly influenced by intense storm activity from the Pacific Ocean, the thermal influence of the Sea of Okhotsk on the western side of the Peninsula, and the topography of the region. The mass balance of Kozelsky Glacier was positive from 1973 to 1978 and generally negative thereafter, with a cumulative mass balance of about -5 m w.e. over the period of record (1973 to 1997; Dyurgerov and Meier, 2005) (Figure 7.10).

7.3.2.11. Summary of regional changes

In general, glaciers across the Arctic began retreating and losing mass in the early 20th century. Area changes and mass loss continued through the mid-20th century, with some regions experiencing brief episodes of slower retreat, reduced mass loss, or even mass gain. The rate of mass loss increased during the past decade across most regions of the Arctic. This marks a change from the situation reported by Dowdeswell et al. (1997), who found that although most Arctic glaciers had experienced predominantly negative net surface mass balance over the previous few decades, there was no uniform trend in mass balance.

7.3.3. Links between mass balance and climate

Annual glacier mass balances often correlate over spatial scales as large as 1200 km (Cogley and Adams, 1998; Rasmussen, 2004) as a result of trends in climate that are consistent over broad regions. These trends often occur due to synoptic-scale teleconnections between the positioning of the polar jet stream, the phase of the El Niño Southern Oscillation (ENSO), and regional patterns in temperature and precipitation. The mass balance of glaciers in Alaska, western Canada, and the U.S. Pacific Northwest is strongly affected by a decadal ENSO-like phenomenon, otherwise known as the Pacific Decadal Oscillation (PDO), which is measured by variations in the sea-surface temperature of the Gulf of Alaska (Bitz and Battisti, 1999; Josberger et al., 2007). Mass balance measurements from Gulkana Glacier in interior Alaska, Wolverine Glacier in maritime Alaska, and South Cascade Glacier in coastal Washington State show the complex reaction of glacier mass balance to variations in the PDO that result in major changes

in the positioning of the North Pacific storm track and the distribution of winter precipitation. From 1966 to 1977, the PDO was in a strong cold phase and the net balance of the three glaciers was slightly negative. In 1977, the PDO switched into a strong warm phase, and the mass balance of South Cascade became very negative, a trend that continues to present. The mass balance of Gulkana and Wolverine Glaciers began to increase as the storm track was diverted toward Alaska. From 1987 to present, the PDO warm phase has weakened, and the net balance of all three glaciers has become uniformly negative.

Gardner and Sharp (2007) showed that accelerated glacier surface mass loss in the Canadian Arctic after 1987 was associated with increased July mean air temperature and a shift in the mean position of the center of the July circumpolar vortex from the western hemisphere to the eastern hemisphere. This allowed increased intrusion of warm air from continental North America into the Queen Elizabeth Islands in summer. July vortex types associated with warm temperatures and negative mass balance anomalies were also more common from 1948 to 1962, so predominantly negative surface mass balances may also have characterized the 1950s.

Occasional periods of positive balance on Novaya Zemlya (especially in the late 1940s and early 1950s) occurred when the North Atlantic Oscillation (NAO) was in a positive phase, sea-surface temperatures in the southern Barents Sea were anomalously warm, and winter precipitation was increased (Zeeberg and Forman, 2001). Negative surface balances occurred whenever summer temperatures rose more than 1 °C above the 1933 to 1991 mean, most notably from the early 1950s to the late 1960s. Positive NAO conditions also coincided with the period of positive mass balances in western Scandinavia in the early 1990s (Pohjola and Rogers, 1997).

7.4. Proxy indicators of surface mass balance

- The number of *in situ* measurements of the surface mass balance of glaciers in the Arctic is small, and the distribution of monitored glaciers is uneven. Therefore, it is difficult to estimate accurately the surface mass balance of the regional ice cover (the quantity that needs to be known to determine the contribution of Arctic glaciers to global sea level change) on the basis of these measurements alone.

- There is, therefore, considerable interest in using remote sensing methods to repeatedly measure regional-scale patterns in parameters that vary in ways that are well-correlated with surface mass balance in order to generate records that may reflect the temporal trends and variability in surface mass balance.

- Parameters of interest include the length of the summer melt season, the distribution of snow and ice facies across mountain glaciers and ice caps, the equilibrium-line altitude, the accumulation area ratio, the surface albedo, and the surface temperature. These can be measured using active microwave methods or passive monitoring in the visible and infrared parts of the electromagnetic spectrum. To date, such measurements have been made more widely on the large ice sheets than on mountain glaciers and ice caps, but this is changing with improvements in sensor resolution.

- Pan-Arctic mapping of summer melt season duration and the end-of-summer distribution of snow and ice facies has been possible for the 1999 to 2009 decade due to the availability of enhanced resolution data products from the Seawinds scatterometer on QuikSCAT. That satellite is no longer functional and there is as yet no new data series to replace the one it generated.

7.4.1. Introduction

Glaciological (*in situ*) measurements of the surface mass balance of Arctic glaciers and ice caps are limited to a small number of glaciers that are unevenly distributed across the Arctic and are biased toward small, land-terminating glaciers. For many of these glaciers, the time series of surface mass balance measurements are discontinuous, of short duration, and not recent. In some regions of the Arctic, *in situ* surface mass balance data are simply nonexistent. Such limitations are mainly due to the remoteness of the Arctic and the intense logistical demands associated with establishing and maintaining *in situ* surface mass balance observations. However, common practice has been to produce regional-scale estimates of surface mass balance based on interpolation of the small number of *in situ* observations. This results in large uncertainties where data are sparse, so that it is not clear whether these estimates truly represent regional surface mass balance trends over the Arctic. Because of these limitations, there is a pressing need to develop alternative methods for estimating trends in surface mass balance and to be able to apply them repeatedly over large areas in order to monitor changes in surface mass balance at regional scales.

Proxy indicators of surface mass balance are parameters or indices that are directly related to the surface mass balance process and that can be used to infer changes in surface mass balance. For example, variations in the abundance of melt layers in ice cores have been used for this purpose (e.g., Koerner and Fisher, 1990; Kinnard et al., 2008). Currently, regional-scale proxy indicators of annual surface mass balance are typically developed from satellite-derived remote sensing data. Some of these proxies provide an indication of the sign of the surface mass balance change between consecutive mass balance years (e.g., Wolken et al., 2009), while others may be useful in the development of methodologies for up-scaling *in situ* surface mass balance measurements to the regional-scale glacier cover (e.g., Wang et al., 2005) or for validating predictions from surface mass balance models.

7.4.2. Proxy indicators, data sources, and methods of measurement

7.4.2.1. Melt duration

Spatial and temporal variability in the surface mass balance of many Arctic glaciers is controlled by the variability in surface melt (Koerner, 2005; Gardner and Sharp, 2007). Surface melt characteristics (extent and duration) can provide valuable information about the melt process and can be detected at regional scales by satellite remote sensing. Because melt usually occurs in all regions of Arctic glaciers and ice caps every summer, melt extent provides little information about variability in the summer melt process on these ice masses.

Melt duration is a much better descriptor and is well correlated with the annual positive degree-day total (a parameter used to simulate snow and ice melt rates in temperature index melt and mass balance models) (Wang et al., 2005), and with summer mean temperatures and geopotential heights in the lower troposphere as derived from climate re-analyses (Wang et al., 2005, 2007; Sharp and Wang, 2009).

Microwave remote sensing is well suited for mapping melt duration on Arctic ice masses. Passive microwave remote sensing is commonly used for melt mapping on Greenland, but its coarse spatial resolution (≥25 km) precludes its use on the smaller Arctic glaciers and ice caps. Data from Ku- and C-band microwave scatterometers have been used successfully to map annual melt duration on ice caps and glaciers in the Arctic (Smith et al., 2003; Wang et al., 2005; Sharp and Wang, 2009; Sharp and Wolken, 2009, 2010; Wolken et al., 2009) (Figure 7.13). Annual melt duration is determined by calculating the number of days between the melt onset and freeze-up dates minus the duration of any intervening periods when melt was not detected. Wang et al. (2005) used time series of enhanced resolution data from the SeaWinds scatterometer aboard QuikSCAT (Early and Long, 2001; Long and Hicks, 2005) to determine melt onset and freeze-up dates and to determine annual melt season durations on ice caps in Arctic Canada (2000 to 2004). On these ice caps, annual mean melt duration varies with ice surface elevation (r = -0.80) and distance from Baffin Bay (r = -0.44) and is strongly linked to variations in geopotential height at the 500 hPa level. For the period 2000 to 2004, average melt duration for individual ice caps ranged from 26 days on the Agassiz Ice Cap to 62 days on the Manson Icefield. For the Queen Elizabeth Islands as a whole, the shortest melt season was 2002 and the longest was 2001.

In the Eurasian Arctic (Svalbard, Severnaya Zemlya, Novaya Zemlya), average melt duration (2000 to 2004) is well correlated with latitude, longitude, and elevation and is consistent with the steep climatic gradient across the region (i.e., warm and moist in the west to cool and dry with persistent sea ice in the east; Dowdeswell et al., 2002; Sharp and Wang, 2009). Average melt duration on Severnaya Zemlya (51 days) is significantly shorter than on Novaya Zemlya (75 days) and Svalbard (77 days). Data show that 2000 was the shortest melt season and 2001 the longest during the 2000 to 2004 period (Sharp and Wang, 2009). Variations in annual mean melt duration were strongly correlated with the mean (June and August) 850 hPa air temperature.

7.4.2.2. Snow and ice facies

Glacier surfaces are organized into zones on the basis of differences in the physical properties of surface and near-surface snow and ice (Benson, 1962). The differences between snow and ice facies are related to variations in the magnitude of summer melt, which are strongly correlated with elevation. Snow and ice facies on glaciers and ice caps in the Arctic include the following (from high to low elevation): dry snow facies – formed where no melt occurs and snow is converted to ice solely by compaction; percolation facies – formed where meltwater percolates into the snowpack and refreezes as lenses, layers and pipes, but the snowpack never reaches the melting point throughout the melt season; saturation facies – formed where
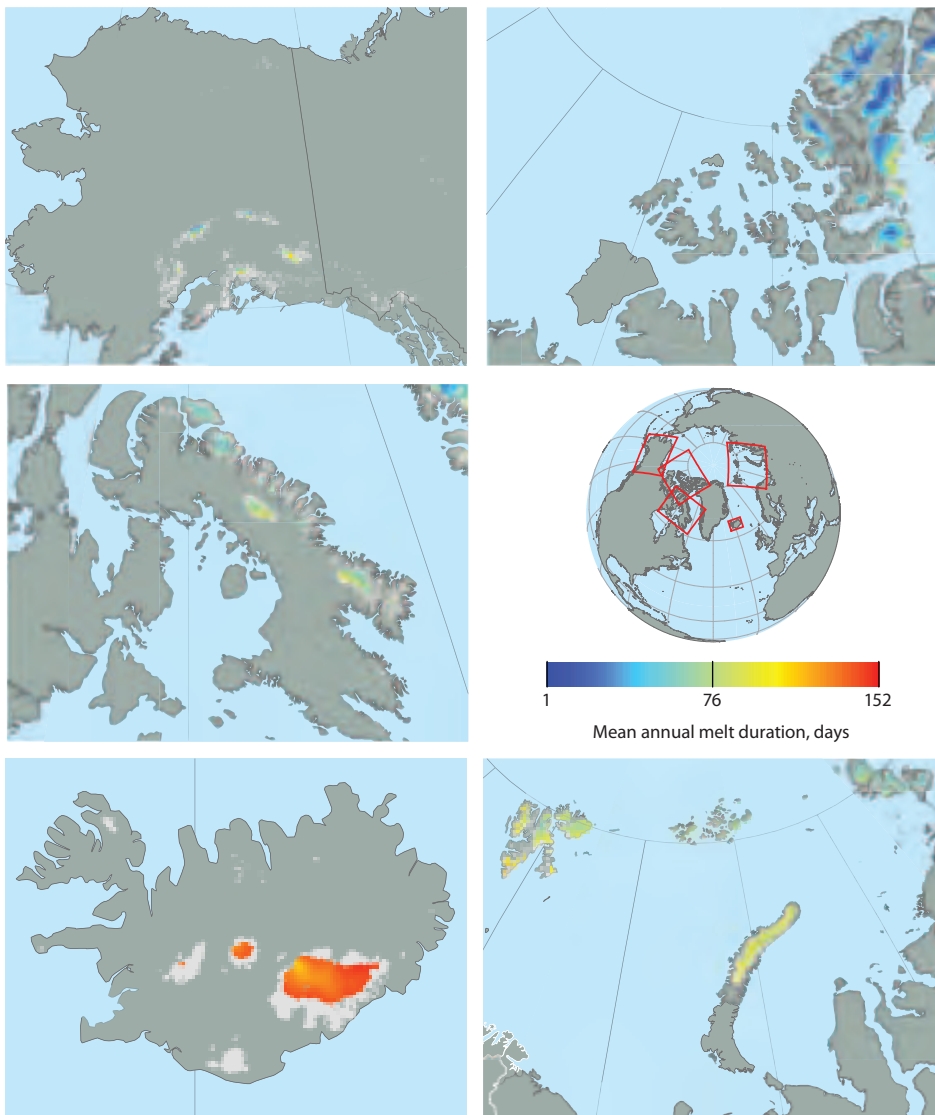
Figure 7.13. Mean annual melt duration on larger Arctic ice masses for the period 1999 to 2008 based on QuikSCAT scatterometer data. Source: Gabriel Wolken, Alaska Division of Geological and Geophysical Surveys.

the temperature of the entire snowpack reaches the melting point sometime during the melt season and wet snow refreezes at the end of the melt season; superimposed ice facies – formed in a transitional zone in the lowest part of the accumulation area where meltwater freezes directly onto the glacier ice surface; and glacier ice facies – exposed in the ablation zone in summer when summer melt removes the annual snow accumulation and exposes glacier ice at the surface (Figure 7.14). There is often a transitional zone between snow and ice facies in which a mixture of these facies can be found. The distribution of snow and ice facies can change annually, and not all facies are found on all ice masses in all years. Because the distribution of snow and ice facies is directly linked to the balance between the processes of accumulation and surface melt, it can provide valuable proxy information about spatial and temporal variations in the surface mass balance.

Snow and ice facies have been mapped at regional scales using optical and microwave remote sensing. Mapping with optical sensors has had limited success, largely due to the often subtle contrast between snow and ice facies and to the occurrence of fresh snow (which conceals the underlying facies) and cloud cover (which conceals the glacier) at the end of the ablation season (Østrem, 1975a; Nolin and Payne, 2007; Dowdeswell et al., 2007). Microwave radiation is unaffected by cloud cover, but

is sensitive to the surface and volumetric physical properties of snow and ice. Snow and ice facies on Arctic glaciers and ice caps have been linked to specific signatures acquired by active microwave sensors, and mapped on the glaciers Kongsvegen and Lovénbreen (Svalbard) using synthetic aperture radar (SAR; ERS-1/2) data (Engeset et al., 2002; König et al., 2002). Firn, superimposed ice, and glacier ice facies were identified

Figure 7.14. Snow and ice facies zones on a glacier or ice cap. Source: after Benson (1962).

in these studies, but the complicated nature of backscatter in the superimposed ice zone precluded the identification of the equilibrium line. Further work on Kongsvegen using *in situ* data and SAR concluded that the superimposed ice zone represents 35% of the glacier (Obleitner and Lehning, 2004). Langley et al. (2008) and Brandt et al. (2008) also investigated snow and ice facies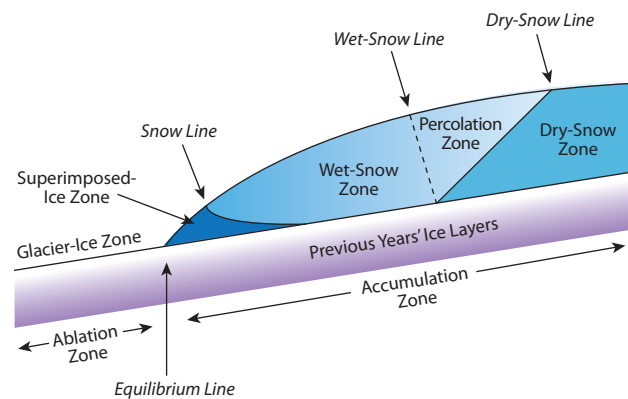 on Kongsvegen by combining data from SAR and ground-penetrating radar (GPR) and concluded that SAR zones were in very good agreement with GPR-derived glacier facies. On Austfonna (Svalbard), Dunse et al. (2009) used GPR to determine the multi-year (2004 to 2007) snow and ice facies distribution and linked variability in facies over the study period to changes in surface mass balance.

Regional-scale mapping and monitoring of snow and ice facies have been performed using microwave scatterometer data. Wolken et al. (2009) used a classification of end-of-summer, enhanced resolution (2.225 km²) data from QuikSCAT (tuned using available field observations) to map the distribution of snow and ice facies in the Canadian High Arctic and its interannual variability (1999 to 2005) (Figure 7.15). Increases in summer air temperature and melt duration were associated with a decrease in the areas of the percolation and saturation zones, an increase in the area of the glacier ice zone, and an increase in the elevation of inter-facies boundaries.

7.4.2.3. Equilibrium-line altitude

The equilibrium line is the line connecting points on a glacier surface where annual accumulation equals annual ablation. Above the equilibrium line (the accumulation area) there is a net gain in mass for a particular year, whereas below the equilibrium line (the ablation area) there is a net loss in mass. The equilibrium-line altitude (ELA) is the average altitude of the equilibrium line at the end of the ablation season in any balance year. The ELA is related to the surface mass balance because it rises and falls as the annual mass balance of the whole glacier becomes more or less negative (Østrem, 1975a; Hagen and Liestøl, 1990). It can be used to identify variability and trends in surface mass balance when monitored annually. The long-term ELA represents the average position of the ELA over many years. Change in the long-term ELA, generally linked to a change in glacier geometry, serves as an important indicator of regional climate change.

The ELA is most accurately determined using standard *in situ* glaciological techniques for measuring surface mass balance. Dyurgerov (2002), Dyurgerov and Meier (2005) and Dyurgerov et al. (2009) compiled all available time series (around 280) of *in situ* mass balance data and related variables, including ELA, for glaciers outside the two ice sheets. For the few glaciers in the Arctic where ELAs are determined through *in situ* measurements, ELA time series are characterized by a high degree of interannual variability. However, while these measurements are robust and invaluable to the glaciological community, they are also spatially restricted and limited in their ability to provide a meaningful measure of ELA variability on a regional scale.

Alternative methods have been developed for monitoring regional-scale changes in the annual ELA, the most effective of which involve the use of satellite microwave remote sensing. In Arctic Canada, regional-scale variability in the
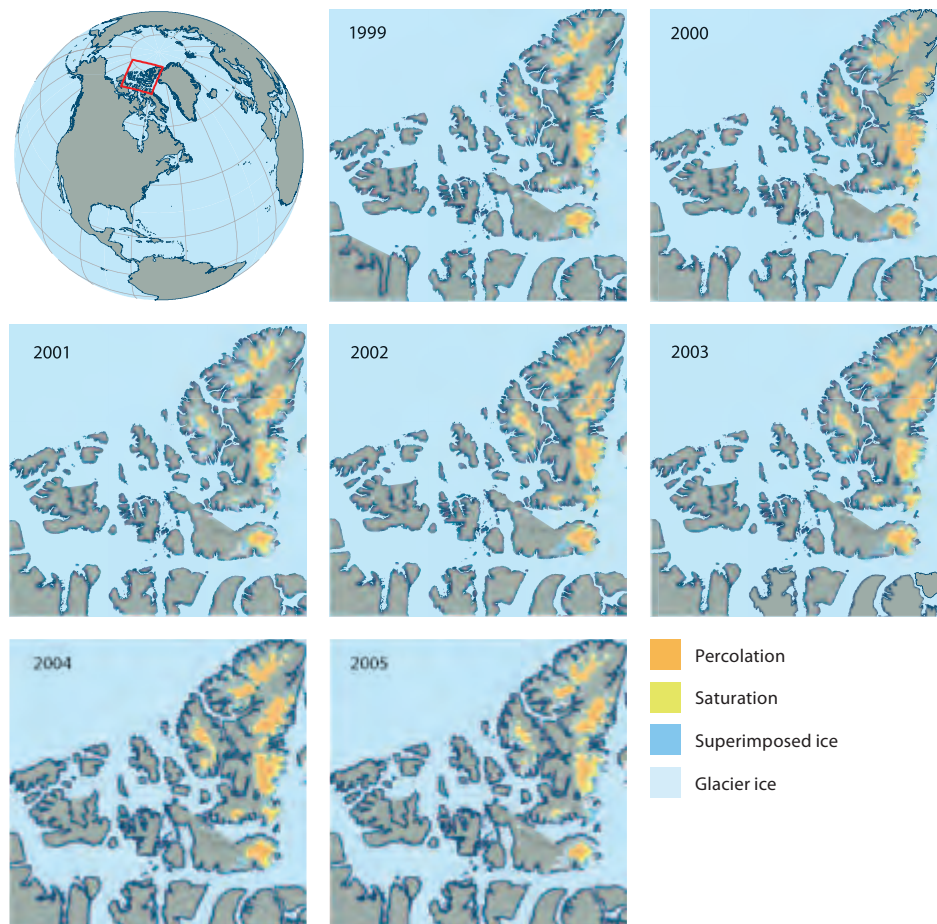

Figure 7.15. Annual distribution of snow and ice facies at the end of the melt season on ice caps in the Canadian High Arctic (1999-2005) as derived from QuikSCAT scatterometer data. Source: Wolken et al. (2009).

ELA can be inferred from changes in the lower boundary of the superimposed ice zone, which is highly correlated with changes in surface mass balance measurements (Wolken et al., 2009). Scatterometer-derived changes in the annual average ELA for each ice cap show a generally coherent pattern of change during the study period (1999 to 2005), with the highest ELAs occurring in 2001, the year with the longest melt season.

7.4.2.4. Accumulation area ratio

The accumulation area ratio (AAR) is the ratio of the area of the accumulation area to the area of the entire glacier. Because the lower limit of the accumulation area is determined by the position of the equilibrium line, the AAR and the ELA are directly related. If a record of the AAR is combined with *in situ* surface mass balance data, then it is possible to produce quantitative estimates of changes in the surface mass balance with respect to changes in the AAR (Dyurgerov et al., 1996, 2009; Hock et al., 2007b). Quantification of regional-scale changes in surface mass balance from AAR changes in the Arctic is limited by the sparseness of *in situ* surface mass balance data for use in calibration. However, regional-scale changes in surface mass balance can be qualitatively assessed based on the relative changes in the AAR between regions. Annual AARs for ice caps in the Canadian High Arctic were determined from maps of the distribution of snow and ice facies derived from QuikSCAT scatterometer data for the period 1999 to 2005 (Wolken et al., 2009). Results from this study indicate considerable spatial and temporal variability in the annual AAR during this period. For the Queen Elizabeth Islands as a whole, the mean AAR was 0.75 over the period 1999 to 2005. The smallest mean AAR (0.50) occurred in 2001and the largest (0.82) in 2004 (Wolken et al., 2009).

7.4.2.5. Albedo

The albedo (the fraction of incoming solar radiation reflected by a surface) of snow and ice is a function of snow/ice grain size, surface water content, solar incidence angle (Wiscombe and Warren, 1980), and snow impurities (Warren and Wiscombe, 1980), including black carbon from natural and industrial sources (McConnell et al., 2007; Forsstrom et al., 2009). On Arctic glaciers, solar radiation is usually the main source of energy for surface melt, and its magnitude and distribution are controlled in part by the glacier surface albedo. As a result, and because factors such as grain size and water content are coupled to surface melt rates, interannual variations in surface albedo and their spatial patterns over individual glaciers can serve as a proxy indicator of surface mass balance.

In situ observations of albedo can be scaled-up with satellite optical remote sensing measurements (Williams, 1987; Reijmer et al., 1999; Greuell and Knap, 2000; Nolin and Payne, 2007), producing data that can be used to identify reductions in albedo that provide an indication of melt onset (Hall et al., 1987; Bindschadler et al., 2001). De Ruyter de Wildt et al. (2002, 2003) used Advanced Very High Resolution Radiometer (AVHRR) data for Vatnajökull, Iceland to determine the evolution of the surface albedo over the melt season and estimate the potential absorbed radiation, which was shown to correlate well with the mean specific mass balance ($r = 0.87$ and 0.94 for different outlet glaciers). This methodology was further developed by

Greuell et al. (2007), who generated estimates of mass balance for the whole of Vatnajökull.

7.4.2.6. Land (glacier) surface temperature

Because winter precipitation is consistently low in the Arctic (excluding southern Alaska, Scandinavia, Iceland, and Kamchatka), variability in glacier mass balance is largely controlled by summer temperature. While precipitation events during summer can be important as summer snowfalls result in short-term increases in surface albedo, temperature tends to drive the ablation process and control variations in surface mass balance (Koerner, 2005; Gardner and Sharp, 2007). In the absence of systematic *in situ* temperature observations, which are available for only a few areas across the Arctic, satellite-borne thermal infrared sensors can be used to monitor clear-sky land surface temperature (LST) and surface melt at regional scales over the terrestrial cryosphere (Key and Haefliger, 1992; Key et al., 1997; Stroeve and Steffen, 1998; Hall et al., 2004, 2006, 2008a,b). In the Eurasian Arctic, Sharp and Wang (2009) used Moderate-Resolution Imaging Spectroradiometer (MODIS) LST data in place of *in situ* near-surface air temperature data to tune melt detection algorithms applied to QuikSCAT scatterometer data. However, there has been little other use of LST data on Arctic glaciers and ice caps. This is probably because reliable satellite-derived LST data can only be obtained under cloud-free conditions and because there are errors and calibration inconsistencies associated with many of these data. Owing to the great potential in using LST as a proxy for regional-scale surface mass balance, there is a need for more *in situ* validation work over ice-covered regions in order to assess the accuracy of satellite-derived LST over Arctic glaciers and ice caps.

7.5. Ice dynamics and iceberg calving

- Iceberg calving from ocean-terminating glaciers can be an important process of mass loss. Rates of calving loss can change rapidly for reasons that are influenced, but not directly controlled, by climate. These changes can generate anomalously high rates of mass loss, but not anomalously high rates of mass gain.

- Possible controls on calving rates include changes in lubrication from seasonal meltwater input, changes in resistance to flow from thinning and flotation, loss of contact with a stabilizing moraine, and break-up of floating ice tongues, as well as from changes in upstream dynamics, including surging.

- A tidewater glacier cycle is recognized. It involves prolonged periods of slow (~1000 years) glacier advance into the ocean, alternating with shorter periods (around decades) of accelerated glacier flow and retreat.

- Although measurements of iceberg calving fluxes are now available for many regions of the Arctic, the ability to model the processes that determine these fluxes is rudimentary. There is currently no ability to predict how calving fluxes will evolve in the future.

- The character of a glacier terminus and the icebergs produced are linked. Small, hard-to-detect bergs are produced mainly by grounded termini and glaciers in rapid retreat, while large tabular bergs come from floating ice tongues and ice shelves.

- The largest remaining ice shelves in the Arctic fringe the northern coastline of Ellesmere Island. These entered a new phase of break-up after 2000. As a result, several fjords in the region are now free of a permanent ice cover for the first time in at least 3000 years.

7.5.1. Overview

Glaciers that terminate in the ocean or in lakes lose mass by iceberg calving as well as by melting and sublimation, and iceberg calving can be the dominant mass-loss term for many glaciers and ice caps in the Arctic. Rates of iceberg calving from marine-ending glaciers (referred to as 'tidewater glaciers' if their termini are grounded below sea level) are modulated by stress and flow conditions near the glacier's terminus, which will vary as the glacier changes shape and volume in response to climate change. Rates of calving can also change in response to internally controlled processes that alter stresses and may be entirely independent of climate. In either case, glacier mass loss through calving may be sudden and rapid, and may be far greater in magnitude than losses due to climatically controlled mass balance. Rapid dynamic changes can cause anomalously rapid mass loss, but not anomalously rapid mass gain, because alterations in flow can force fast ablation (by moving ice either to lower and warmer elevations or into the ocean), while accumulation rates are limited by precipitation.

7.5.1.1. Hydrology and basal sliding

Glaciers that slide over their beds typically exhibit significantly larger velocities and fluxes than glaciers that flow by internal (ice) deformation alone. Sliding is facilitated by the presence of water and deformable sediments at the ice/bed interface, although the relationship between water input and sliding speed is highly non-linear. Warmer surface conditions that increase surface melting in summer will accelerate glacier flow, but there is no simple relationship between the amount of warming and the magnitude of the flow acceleration. Higher water fluxes through subglacial drainage systems may result in greater enlargement of individual drainage channels, lower subglacial water pressures, and less flow enhancement than is generated by smaller fluxes. Thus, it is by no means clear that climate warming will, in the long term, result in more rapid glacier flow. As on Greenland, however, supraglacial lakes often form on Arctic glaciers in summer (Figure 7.16), and may drain suddenly when water-pressure induced fracturing creates drainage connections between the glacier surface and the glacier bed (Boon and Sharp, 2003). These lake drainage events may result in short-lived accelerations of glacier flow. As yet, however, there is no general rule to link increased sliding speed with increased englacial or subglacial water input, and no validated, reliable numerical model for basal sliding exists (Fountain and Walder, 1998; Hooke, 1998; Marshall et al., 2002).

Figure 7.16. Supraglacial lakes forming on the surface of Columbia Glacier, Alaska, during the melt season. Source: W. Tad Pfeffer, Institute for Arctic and Alpine Research, University of Colorado

7.5.1.2. Glacier surges

Glacier surges are episodes of anomalously rapid motion (about 10 to 100 times 'normal' or non-surging speeds) that last for months to a few years and alternate with longer periods (decades to centuries) of normal, or quiescent, glacier behavior. Glacier velocities during the active phase of the surge cycle may reach 100 m per day and rates of terminus advance may approach 1 km per month. Glacier surges are not a direct response to climate forcing, but surging can alter a glacier's sensitivity to climate, especially in the immediate post-surge phase when its accumulation area is anomalously depleted (Dowdeswell et al., 1995).

7.5.1.3. Calving and marine-ending outlet glacier instability

'Marine-ending' or 'ocean-terminating' glaciers reach the ocean shore with sufficient flux to maintain an ice tongue that is either grounded in water shallower than the tongue's flotation depth or may be floating under polar conditions (subfreezing and, therefore, strong when in tension). Floating tongues can also form during periods of rapid terminus retreat. The term 'tidewater glacier' is often used interchangeably with these labels, although it was originally restricted to glaciers with grounded termini. Marine-ending glaciers can undergo a periodic growth/shrinkage instability with long (roughly centuries) periods of slower motion and slow advance

alternating with shorter periods (decades) of rapid (~5 to 10 km/y) motion, high calving flux, and rapid (~1 km/y) retreat (Meier and Post, 1987) (Figure 7.17). This is significant in that during rapid retreat a glacier can lose mass into the ocean far faster (by loss of its marine-grounded tongue and drawdown of its source basin by rapid flow) than is possible by direct surface mass balance forcing alone.

In its advancing phase, a tidewater terminus is stabilized in part by back stress against a moraine at its advancing margin. If it retreats from the moraine, a tidewater glacier in an advanced position can switch abruptly (within years) into an unstable phase of rapid flow and calving accompanied by thinning in its ocean-grounded reach that results in terminus retreat. Once initiated, rapid flow and retreat appear to be generally irreversible and continue until the glacier's terminus retreats inland to a point where the water depth provides no significant buoyancy force. The causes of unstable retreat are not completely understood, but involve high basal water pressures tied to the depth of the glacier bed below sea level, rapid sliding, vertical thinning, and positive feedback between thinning and rapid flow. The processes governing the size and rate of production of icebergs are extremely poorly known. Climate-induced thinning can play a critical role in initiating unstable retreat, but once initiated, unstable retreat appears to be modulated by channel geometry and englacial and basal hydrology and is essentially independent of climate and surface mass balance. Melting of the submarine faces of grounded tidewater glaciers can be rapid (Motyka et al., 2003) and can result in undercutting of the terminal ice cliff with consequences for iceberg production and terminus stability (Rignot et al., 2010; Straneo et al., 2010).

7.5.1.4. **Lake-terminating glaciers**

The total area of lake-terminating calving glaciers is much smaller than the area of marine-terminating glaciers, and the dynamics of their calving appear to be significantly different. Thinning rates on lake-calving glaciers are among the highest observed anywhere (e.g., -3.0 m/y basin-averaged thinning at Yakutat Glacier, Alaska) (Larsen et al., 2007).

7.5.1.5. **Ice shelf / ice tongue break-up**

Stable floating ice tongues (or ice shelves) are held in place by stabilizing stresses supplied by lateral attachments to valley walls or embayment margins (and, to a lesser degree, back stress from sea ice and confined iceberg fragments: *sikkusaq*), rather than by basal drag or back stress on the glacier terminus (Dowdeswell et al., 2000; Reeh et al., 1999). Submarine melt can be a significant source of mass loss from floating tongues (Mayer et al., 2000). Spreading and thinning of floating ice produces small tensile stresses, crevassing, and calving of icebergs. Under warming conditions, a floating ice tongue can be destabilized by (i) increased submarine melt, (ii) increased internal ice temperature, and (iii) percolation of surface meltwater into open crevasses and fractures. Loss of an ice shelf or floating tongue and the associated back stress on the feeding glacier(s) affect the force balance on the ice upstream from the grounding line and in general cause an acceleration of the grounded ice and increased ice discharge to the ocean. The loss of the floating tongue has no influence on sea level,

The Tidewater Glacier Cycle
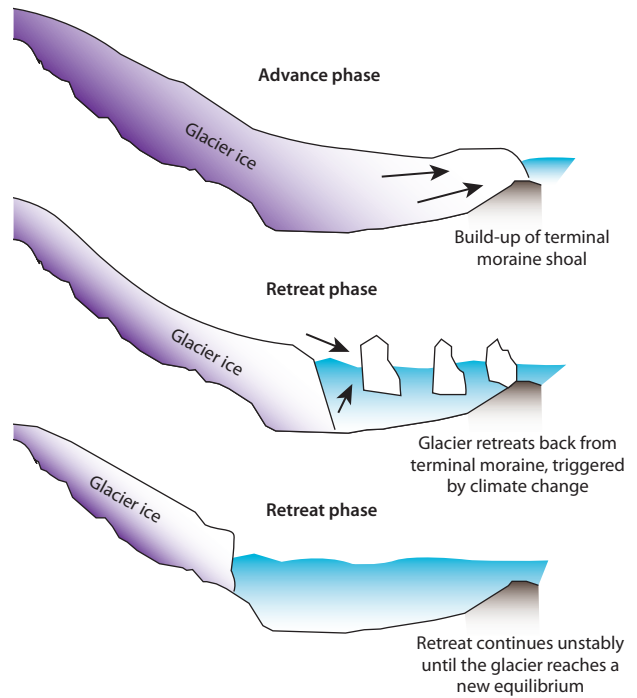
Figure 7.17. Schematic diagram showing the typical advance-retreat cycle of a tidewater glacier. Source: after Meier and Post (1987).

but increased discharge from the grounded part of the glacier contributes to eustatic sea-level rise.

7.5.2. **Measurement methods**

Glacier dynamic changes are manifested principally in changes in speed, surface elevation, ice discharge, and iceberg calving. Consequently, surface observations to track moving features or measure surface altitude are critical to understanding the processes involved. Glacier velocity is determined from repeat measurements of the position of markers on the glacier surface made by optical survey (Harper et al., 2007), GPS (Howat et al., 2008), and by feature tracking using aerial photography (Krimmel, 2001) and ground-based photography (O'Neel et al., 2005). Velocities are also measured remotely by manual and automated feature tracking applied to a variety of visible satellite imagery (Scambos et al., 1992; Joughin, 2002; Dowdeswell and Benham, 2003) and by radar interferometry and speckle tracking (Joughin et al., 1999). Elevation changes are measured by photogrammetry, by repeat altimetry from airborne or spaceborne platforms (Arendt et al., 2002, 2006; Larsen et al., 2007; Howat et al., 2008), and to a lesser degree by ground-based GPS and optical surveys.

The terminus advance/retreat rate ($\partial x/\partial t$) is the difference between the calving rate (u_c) and the ice flow speed at the terminus (u_T): ($\partial x/\partial t = u_c - u_T$). Iceberg calving events are episodic and very short in duration, and the volume of individual icebergs is hard to determine from imagery. The calving rate is therefore difficult to measure directly, and it is typically calculated as $u_c = \partial x/\partial t + u_T$ (O'Neel et al., 2005), but this is a weak approach in predictive models where $\partial x/\partial t$ is often the variable being sought. New methods are being developed to use seismic records to identify individual calving events and estimate iceberg discharge (O'Neel and Pfeffer, 2007).

7.5.3. Calving and surging glaciers in the Arctic

Surging and marine calving glaciers are found throughout the Arctic, although there are currently none in Norway or Sweden. Table 7.2 summarizes recent estimates of mass loss by calving from major regions and/or major ice masses within the Arctic (but note that no estimates are available for some regions and ice masses).

7.5.3.1. Alaska and western Canada

The greatest concentrations of glacier area in Alaska and northwestern Canada lie along the coast of the Gulf of Alaska, where land-terminating glaciers on the interior side drain predominantly into the Matanuska, Copper, and Yukon Rivers, while seaward-facing glaciers drain into, or terminate in, the Gulf of Alaska.

7.5.3.1.1. Calving

A significant but undetermined fraction of the area of Gulf of Alaska glaciers is drained through 54 calving tidewater outlets. No estimates have been made of calving flux for the entire Gulf, but a 1982 study (Brown et al., 1982) compared geometry and calving speed for 13 of the largest tidewater glaciers and found a total flux of 7.2 Gt/y draining from an aggregate basin area of 8138 km², estimated for various periods within the interval 1942 to 1979. No more recent or more complete assessment has been made.

In the past 200 years, the 54 tidewater glaciers terminating in the Gulf of Alaska all retreated from extended stable positions. Columbia Glacier was the last to start retreating and as of 2009 had retreated roughly two-thirds of the way up the 30-km long fjord in which it was grounded up to about 1982. The majority of Gulf of Alaska tidewater glaciers are now either stable in retreated positions or slowly advancing (e.g., Hubbard and Mears Glaciers). Columbia Glacier is undergoing the most rapid change, calving an average 3.3 Gt/y of ice (1982 to 2001), with a maximum annual calving flux of 6.6 Gt in 2001 (O'Neel et al., 2005), and thinning in the terminus region by 20 m/y. GRACE gravity measurements suggest that the Gulf of Alaska glaciers were losing mass at a rate of 84 ± 5 Gt/y in the period 2003 to 2007 (Luthcke et al., 2008). A study of the southeastern (panhandle) Gulf of Alaska sub-region (not including Columbia Glacier) found a loss of 16.7 ± 4.4 Gt/y from 14 580 km². Two-thirds of this was from calving glaciers, although the fraction of this volume loss that was attributable to calving (as opposed to surface melting) is unknown (Larsen et al., 2007).

Table 7.2. Estimated mass losses by iceberg calving of mountain glaciers and ice caps. Specific balances due to calving are derived by dividing the calving mass loss by the total area of the glacier.

	Period	Area, km²	Calving loss, Gt/y	Specific balance, m/y w.e.	Source
Alaska					
Columbia Glacier	1982 – 2001	1000	3.3	-3.3	O'Neel et al., 2005
Canadian High Arctic					
Devon Ice Cap	1960 – 1999	14 000	0.53±0.12	-0.04	Burgess et al., 2005
Prince of Wales Icefield	1963 – 2003	19 300	1.9 ±0.02	-0.10±0.00	Mair et al., 2009
Agassiz Ice Cap	1999 – 2002	19 500	0.60±0.14	-0.03±0.01	Williamson et al., 2008
W Grant Ice Cap[a]	1999 – 2003	2 000	~0.23	~ -0.12	Williamson et al., 2008
Russian Arctic					
Academy of Sciences Ice Cap	1995	5 575	0.59	-0.10	Dowdeswell et al., 2002
Franz-Josef Land	1929 – 1959	13 700	2.1	-0.15	Govorukha, 1989
	1952 – 2001		4.3	-0.31	Glazovsky and Macheret, 2006
Novaya Zemlya	1930 – 1960	23 600	1.8	-0.08	Govorukha, 1989
	1952 – 2001		1.4	-0.06	Glazovsky and Macheret, 2006
Severnaya Zemlya	1929 – 1972	19 400	0.45	-0.02	Govorukha, 1989
	1952 – 2001		0.7	-0.04	Glazovsky and Macheret, 2006
Svalbard					
All glaciers	1968 – 1998	36 600	4±1	-0.1±0.0	Hagen et al., 2003b
	2000 – 2006	36 600	6.75±1.7	-0.18±0.05	Błaszczyk et al., 2009
Austfonna	[b]	8 000	2.5±0.5	-0.31±0.06	Dowdeswell et al., 2008

[a] Calving estimates only from Otto Glacier which is likely to represent the majority of the mass loss by calving from the ice cap; [b] a combination of various data sets from the period 1973 to 2001.

Here, and elsewhere, episodic discharges from individual glaciers can exceed the aggregate discharge from the rest of the region, making averages over time difficult to interpret or extrapolate.

7.5.3.1.2. Surging

Surging glaciers are abundant in the central Alaska Range (Denali massif), eastern Alaska Range, and throughout the St. Elias mountains, where 'Alaskan-type' surging was defined (Figure 7.18). Bering Glacier, in this region, is the largest glacier (and largest surging glacier) in Alaska (Shuchman and Josberger, 2010). Its last major surge took place in 1995 to 1996, although there was a small surge in 2009 to 2010. Bering Glacier terminates in a large proglacial lake at sea level, the size of which changes dramatically as the glacier advances and retreats. Some surging glaciers in the St. Elias Range terminate in the ocean, allowing surge behavior to influence calving discharge. No surging glaciers are known in the Chugach or Brooks Ranges.

7.5.3.2. Arctic Canada

7.5.3.2.1. Calving

There is no comprehensive summation of calving flux or inventory of calving glaciers for the Canadian Arctic Archipelago, but estimates of calving fluxes for some of the ice caps exist for mixed time periods in the interval 1960 to 2003, totaling 3.0 Gt/y (Devon Ice Cap, 0.5 Gt/y; Agassiz Ice Cap, 0.6 Gt/y; Otto Glacier (western Grant ice complex), 0.2 Gt/y; Prince of Wales Icefield, 1.9 Gt/y). As in Alaska, individual glaciers can dominate the regional flux estimates.

7.5.3.2.2. Surging

At least 51 surge-type glaciers have been identified in the Queen Elizabeth Islands (Copland et al., 2003). Of these, 15 were surging in 1959/1960 and/or 1999/2000. The largest recorded surge-type advances (4 to 7 km) were observed on Axel Heiberg Island and the Manson Icefield (Ellesmere Island). The active phase of the surge cycle of these glaciers may last more than a decade (Mueller, 1969; Copland et al., 2003), typical of 'Svalbard-type' surging (Mueller, 1969; Murray et al., 2003). Glacier velocities during the active phase of the surge cycle may range from several hundred metres to over 1 km per year. Because many of the known surge-type glaciers terminate in the ocean, surging may cause large short-term changes in the loss of mass to the oceans by iceberg calving. Little is known about the duration of the quiescent phase of the surge cycle in this region owing to the short period of observations.

7.5.3.3. Greenland (excluding the Greenland Ice Sheet)

Little is known about the history of area and volume changes of the roughly 48 600 km² of glaciers and ice caps that surround the Greenland Ice Sheet, or about their mass balance, calving discharge, or the fraction of their area drained through marine-terminating outlets. Mass balance estimates for these glaciers have been made (Dyurgerov and Meier, 2005) using correlated mass balance records from Svalbard and the Canadian Arctic

Figure 7.18. Hubbard Glacier, which drains into Russell Fjord in the Yakutat region of southern Alaska, is a surging tidewater glacier. Source: Anthony Arendt, University of Alaska.

Archipelago. Net losses from 1992 to 2002 are estimated to be about 18 Gt/y, but the fraction due to calving is completely unknown.

7.5.3.4. Iceland

In Iceland, 26 surge-type glaciers and 80 surge advances have been identified (Björnsson et al., 2003), and surges account for a significant fraction of the mass transported by major outlet glaciers of all the main ice caps. Surges affected 38% of the area of Vatnajökull in the 1990s, and accounted for about 25% of the mass transfer from the accumulation zone of the ice cap to the ablation zone (Björnsson and Pálsson, 2008). Calving losses are unknown but probably not significant.

7.5.3.5. Svalbard

Glacier geometry, flow velocity, and front position changes of Svalbard glaciers were determined from ASTER images acquired in 2000 to 2006 (Błaszczyk et al., 2009). A total of 163 grounded tidewater glaciers drain an area of around 21 200 km² (more than 60% of the total glacier area of Svalbard) through a combined calving terminus length of 860 km.

7.5.3.5.1. Calving

Mass loss due to calving from the whole archipelago was initially estimated to be about 4 Gt/y for the period 1968 to 1998 (Hagen et al., 2003b), but a more recent estimate for the period 2000 to 2006 is 5.0 to 8.4 Gt/y, with a mean value of 6.75 ± 1.7 Gt/y (Błaszczyk et al., 2009). The mean retreat rate ($\partial x/\partial t$) of the calving glaciers is about 30 m/y, and terminus retreat ($\dot{M}_L$) accounts for about 30% (~2.1 Gt/y) of the total calving flux from the archipelago. Calving constitutes about 20% of the overall mass loss from Svalbard glaciers (Błaszczyk et al., 2009).

Dowdeswell et al. (2008) estimated the calving flux from Austfonna to be 2.5 ± 0.5 Gt/y from an aggregate terminus length of about 230 km. This represents 30% to 40% of the annual ablation from this ice cap. Terminus retreat reduced the ice cap area by about 10 km²/y during the period 1991 to 2001 and accounted for about 50% of the calving loss.

7.5.3.5.2. Surging

Surge-type glaciers are very common in Svalbard and comprise a range of glacier types from small land-terminating cirque glaciers to large calving outlets. Although the overall fraction of surge-type glaciers is unknown, estimates range from 13% to 90% (Liestøl, 1969; Dowdeswell et al., 1991; Lefauconnier and Hagen, 1991; Hamilton and Dowdeswell, 1996; Jiskoot et al., 1998). Surging of marine-grounded outlets can interact with calving dynamics. The 1250-km² Hinlopenbreen calved about 2 km³ of icebergs in a single year during a surge (Liestøl, 1973). Bråsvellbreen (1100 km²) advanced up to 20 km along its 30-km marine-grounded margin during a surge in 1936 to 1938 (Schytt, 1969). In the inner part of Storfjorden, east Spitsbergen, Negribreen advanced about 12 km in less than a year (about 35 m/d) along a 15-km long section of its front during a surge in 1935 to 1936 (Liestøl, 1969). Recently, the tidewater glacier Nathorstbreen advanced about 7.6 km or about 20 m/d from September 2008 to September 2009 (Sund and Eiken, 2010).

7.5.3.6. Russian Arctic islands

7.5.3.6.1. Calving

Govorukha (1989) estimated calving fluxes of 1.8 Gt/y from Novaya Zemlya (1930 to 1960), 0.45 Gt/y from Severnaya Zemlya (1929 to 1972), and 2.1 Gt/y from Franz Josef Land during the period 1929 to 1959 (a total of 4.35 Gt/y). Using measurements of terminus ice thickness derived by radio echo sounding, ice surface velocities at glacier termini derived by radar interferometry, and measurements of terminus length and terminus advance/retreat, Glazovsky and Macheret (2006) estimated calving fluxes from the Russian Arctic islands to be 6.4 Gt/y for the period 1952 to 2001 (Novaya Zemlya = 1.4 Gt/y; Severnaya Zemlya = 0.7 Gt/y; Franz Josef Land = 4.3 Gt/y). Losses due to terminus retreat were estimated to be about 1.35 Gt/y, or roughly 21% of the total calving flux. It is not possible to determine the significance of the difference between the two calving flux estimates, so it should not be taken to imply a temporal trend in calving fluxes.

7.5.3.6.2. Surging

Surge-type glaciers are rare in the Russian Arctic Archipelago (Dowdeswell and Williams, 1997). None are known to exist in Franz Josef Land, while until recently only three had been identified in Novaya Zemlya and two in Severnaya Zemlya. However, a new analysis based on high-resolution satellite imagery identified 32 potential surge-type glaciers on Novaya Zemlya, representing 18% of the total glacier area (Grant et al., 2009). These glaciers are generally relatively long (median length 18.5 km), large (median area 106.8 km²) outlet glaciers that terminate in water. They occupy large, complex catchment areas that may be receiving increased precipitation from the Barents Sea.

7.5.4. Recent ice shelf break-up events

Most of the largest floating ice shelves in the Arctic, fringing the northern coast of Ellesmere Island, are not predominantly glacier-fed, but form mainly by basal accretion of sea ice and surface accumulation of precipitation (Jeffries, 1992). They developed around 5500 to 3000 years BP (England et al., 2008) and have lost over 90% of their area during the 20th century, with only four major ice shelves now remaining. Much of the area loss occurred in the 1950s (Koenig et al., 1952) and 1960s (Hattersley-Smith, 1963, 1967). A new phase of disintegration began in 2000, with renewed fracturing of the Ward Hunt Ice Shelf (Mueller et al., 2003) and continued in 2005 with the abrupt loss of almost all of the Ayles Ice Shelf and major calving from the Petersen Ice Shelf (Copland et al., 2007). In summer 2008, the total ice shelf area in Arctic Canada decreased further (by 23% relative to 2007; Mueller et al., 2008) due to the loss of the entire Markham Ice Shelf and 60% of the area of the Serson Ice Shelf. As a result, several fjords on the northern coast of Ellesmere Island are now ice-free for the first time in 3000 to 5500 years (England et al., 2008). In 2010, large new fractures in the Ward Hunt Ice Shelf were first detected in Radarsat-2 images from 7 and 14 August. A MODIS image from 18 August shows that break-up of the eastern part of the ice shelf was underway (Figure 7.19), and some 65 to 70 km² of the shelf had been lost by the end of August. Meanwhile, fragments of the ice islands that calved from the Ayles, Serson, Peterson, Ward Hunt and Markham ice shelves in 2005 and 2008 have drifted into Canada Basin and the Sverdrup and Queen Elizabeth Islands and are beginning to enter the Northwest Passage via Barrow Strait.

Causes of ice shelf break-up appear to include rising summer air temperature and the loss of semi-permanent landfast sea ice along the northern coast of Ellesmere Island, although strong southerly winds seem to play an important role in individual break-up events. Ice shelf break-up events produce large 'ice islands' which may become a hazard to shipping and offshore resource exploration if they drift westward into the Beaufort Sea. They can also result in the drainage of epishelf lakes and the loss of their unique ecosystems (Veillette et al., 2008).

As with floating ice tongues, the loss of ice shelves does not directly influence sea-level rise. In the case of the Ellesmere ice shelves, because most of them do not interact with terrestrial glaciers, there is no indirect effect on sea level.

The Matusevich Ice Shelf in Severnaya Zemlya, the only major ice shelf in the Eurasian Arctic, has also undergone periodic calving events during roughly the past 70 years (Williams and Dowdeswell, 2001).
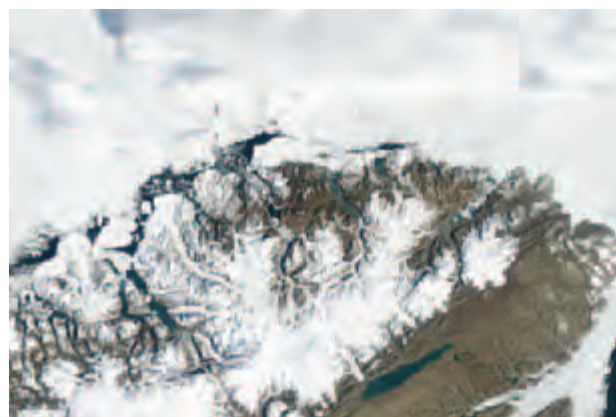

Figure 7.19. MODIS image from 18 August 2010 showing the Ward Hunt Ice Shelf, northern Ellesmere Island, and the extensive fracturing that had developed in the ice shelf to the east and south of Ward Hunt Island. Source: Canadian Ice Service.

7.5.5. Iceberg characteristics and relationship to ice dynamics

There are certain relationships between iceberg characteristics (size, number, morphological type) and parent glacier type and behavior. These relationships can be broadly described as follows (Dowdeswell, 1989; Dowdeswell et al., 1992):

- The largest tabular icebergs (large horizontal dimension relative to depth) are calved from floating glacier tongues and ice shelves. Horizontal sizes reach hundreds of metres to kilometres. The number of individual icebergs calved each year is usually small.

- Fast-flowing tidewater glaciers that are grounded have high rates of discharge, but often produce icebergs of relatively small size (tens to hundreds of metres) and irregular shape that tend to break up quickly into still smaller bergs and bergy bits (Dowdeswell and Forsberg, 1992) (Figure 7.20). Slower-flowing glaciers calve fewer icebergs, but their sizes may be larger than those of icebergs calved from more active glaciers.

- Surging glaciers and tidewater glaciers in rapid retreat tend to produce a very large number of small icebergs (10 to 50 m in length) with irregular shapes (Schytt, 1969). If such glaciers start to float, however, the mode of calving may change to one in which episodic flow-perpendicular rift formation and propagation result in the release of a few very large icebergs (Walter et al., 2010).

The systematic classification of iceberg types made by Dowdeswell for glaciers of Spitsbergen can be applied to all sources of icebergs in the Arctic. It is supported by descriptions of icebergs made at different times near Spitsbergen, Franz Josef Land, and Novaya Zemlya (Sandford, 1955; Kubyshkin et al., 2009), and by observations in the northern Barents Sea in 2003 to 2007. For example, tidewater outlet glaciers in Russkaya Gavan Bay and Inostrantsev Bay in northern Novaya Zemlya have strong fracturing of the front and produce a moderate quantity of small- and medium-sized icebergs of irregular shape (Buzin and Glazovsky, 2005). Floating margins on some of the larger ice caps of Franz Josef Land produce large tabular icebergs (Dowdeswell et al., 1994; Zubakin et al., 2007).

Figure 7.20. Small iceberg off the coast of Franz Josef Land. Note the debris-rich layers within the iceberg. Source: Photo courtesy of Arctic-shelf laboratory, AARI.

7.5.6. Controls on calving fluxes

Although the detailed physics of iceberg calving are poorly understood, certain controlling factors can be identified. The tensile strength of ice is strongly dependent on temperature, and temperate ice (ice at its pressure melting point, where water and ice exist in thermal equilibrium) has essentially no tensile strength (Schulson, 1999). Along-flow extension or compression of ice influences the opening or closing of transverse crevasses, but no clear relationship between calving rate and along-flow strain rate has been demonstrated observationally on meaningful timescales (Alley et al., 2008). A statistical relationship between water depth at grounded glacier termini and calving rate has been demonstrated (Brown et al., 1982), but this relationship is purely empirical and possibly applicable only to calving margins near steady-state. Sea ice, floating brash ice, and iceberg debris can exert substantial restraining forces (or back stresses) that are transmitted to the fjord walls and inhibit calving (Amundson et al., 2010). The presence or absence of a moraine shoal at the terminus providing resistive back stress against extension of ice in the direction of flow has a strong influence on calving rate (Meier and Post, 1987; Fischer and Powell, 1998; Nick et al., 2007). However, it is unclear whether the detachment of a glacier terminus from its terminal moraine is a cause or a consequence of thinning and retreat.

7.5.7. State of theory and modeling

A number of attempts have been made to formulate calving models, but there is still no universal theoretical framework that can explain the full suite of available observations. Using data from twelve Alaskan glaciers, Brown et al. (1982) proposed an empirical linear relationship between the calving rate and the water depth at the terminus. However, the correlation is not broadly applicable because there is no obvious physical control that can explain any depth dependency of the calving rate (Benn et al., 2007). A drawback to such empirical models is that they are often based on data from glaciers that are either in or near a steady-state (Van der Veen, 1996). Indeed, based on observations at Columbia Glacier in Alaska, Meier and Post (1987) suggested that a linear calving relationship might not be valid during periods of rapid advance or retreat. Various approaches to a mechanically based process model of calving have been pursued since the mid-1960s, based on several different lines of analysis. These include, most notably, an elastic bending beam model developed by Reeh (1968), analyses of crevasse penetration with and without water (Nye, 1957; Weertman, 1973), and analyses based on linear elastic fracture mechanics (Van der Veen, 1998a,b). Other observations and analyses include work by Holdsworth (1969) and Hughes (1992).

Based on observations from Columbia Glacier, Van der Veen (1996) proposed a flotation model, wherein the terminus of the glacier retreats to the position where the ice thickness exceeds the flotation thickness by some threshold value, H_0. On Columbia Glacier, H_0 was determined to be about 50 m. Vieli et al. (2001) pointed out that for smaller glaciers (water depths below 150 m and flow velocities less than 200 m/y), frontal cliff heights do not usually exceed 50 m, so the height above flotation will be much less than was suggested for Columbia

Glacier. As a result, they proposed a modified flotation criterion, with the minimum height above flotation, H_0, replaced by a fraction, q, of the flotation thickness at the terminus. Benn et al. (2007) noted that one significant weakness of flotation models is that they do not allow floating termini to form, thus limiting their use to certain types of calving margins. As a result, they proposed a new criterion based on a simple model of crevasse formation, in which the calving margin is located where crevasse depth is equal to the glacier surface height above sea level, and the determination of the calving rate is reduced to determining ice thickness, velocity distribution, and changes in thickness and velocity through time.

Hanson and Hooke (2000) treated calving as a multivariate problem, using water depth, longitudinal strain rate, and ice temperature as the primary factors controlling the calving rate. Nick and Oerlemans (2006) compared the water depth and flotation models using synthetic simplified glacier geometries and found the latter to be superior, but they failed to simulate a full cycle of glacier length variations when the glacier terminates in very deep water. Nick et al. (2007) elaborated on their model by including a simple sediment transport scheme in the ice flow model and concluded that sedimentation at the glacier front needs to be included for the glacier to advance into deepening water.

Most calving models presented to date are either wholly empirical or based on mechanics. Other englacial processes, including glacial hydraulics, have not been incorporated, although there is evidence that water flow along cracks may play an important role in the calving process (O'Neel and Pfeffer, 2007).

No validated, broadly applicable, and robust calving model exists at this point. The calving relationships and models proposed so far have been tailored and applied to specific glaciers and data sets, and the proposed models tend to require data that are not readily available on regional scales. Larger models, either focused glacier/ice sheet models or comprehensive coupled GCMs, tend to include calving only in an extremely simplistic fashion (e.g., specifying a fixed calving boundary location) or do not treat calving at all. Hence, there is a pressing need for robust calving formulations of intermediate complexity that are capable of being applied in a wide variety of settings and for regional-scale mass change predictions. There is also a need for the basic data required to constrain models based on such a formulation.

7.5.8. Requirements for improved predictions of calving fluxes

Because calving can deplete terrestrial ice reservoirs very rapidly, calving dynamics are a key element in understanding and predicting the mass balance and sea level rise contributions of all glaciers, ice caps, and ice sheets with calving margins. Knowledge of present-day calving rates is extremely limited owing to a lack of fundamental observations. Since changes in flow dynamics and calving are not controlled directly by externally observed variables (such as climate), robust predictions depend on an understanding of glacier physics that is still lacking and on observations that can be difficult to make. The most fundamental observations needed to better constrain potential calving losses are (i) more complete observations of calving rate and marine-ending terminus changes, and (ii) radio echo sounding of ice thicknesses and basal topography to identify marine-grounded outlets and the marine-grounded fractions of glacier and ice cap areas. Without these measurements, the potential for dynamically forced calving losses cannot be constrained let alone predicted.

7.6. Projections of Arctic glacier changes

- Most attempts to simulate the response of Arctic glaciers to future climate change involve evaluation of the response of the surface mass balance to prescribed changes in climate (usually air temperature and precipitation); simulations generally do not include changes in glacier dynamics or mass loss by iceberg calving. Climate change can be imposed as either step changes from an initial state or as transient changes over some period of time derived from climate models.

- Projections for individual glaciers throughout the Arctic show substantial mass loss or even disappearance of smaller glaciers by the end of the 21st century in response to imposed temperature and precipitation scenarios.

- Very few studies have attempted to model the response of Arctic glaciers on larger scales. A new simulation of the surface mass balance of the global population of mountain glaciers and ice caps to 2100 uses a temperature index mass balance model driven by downscaled output from ten GCMs, all forced by the IPCC A1B emissions scenario. For Arctic glaciers, projected volume loss due to surface mass balance ranges between 0.051 to 0.136 m sea-level equivalent, or 13% to 36% of their current volume by 2100 depending on the choice of GCM. Most models suggest that the largest mass losses are from glaciers in Alaska and Arctic Canada.

7.6.1. Downscaling climate model projections

General circulation models (Solomon et al., 2007) driven by standardized IPCC emissions scenarios (B1, A2, A1B; see Chapter 3) generally predict that warming and precipitation increases in the Arctic beyond 2030 will be larger than the global mean (Chapman and Walsh, 2007; Kattsov et al., 2007). Only part of the projected increases in precipitation will lead to increases in snow accumulation because the fraction of precipitation falling as rain will increase with the projected air temperature increases.

Direct use of data from GCMs for glacier mass balance projections is not currently feasible, owing to the generally large biases in model data on regional scales. Individual glaciers are typically much smaller than climate model grid boxes, and often occupy complex terrain that is only coarsely resolved in the model's underlying topography. Mass balance modeling is particularly sensitive to biases and offsets in air temperature, which controls the energy available for melt and the snow-rain ratio of precipitation (and thus controls both ablation and accumulation). Hence, some form of downscaling is required to transfer global-scale climate information to local glacier scales prior to making glacier mass balance projections.

Statistical downscaling is often used, whereby statistical relationships are established between meteorological

quantities determined at the GCM and the local scales. This can be done using either field observations or other suitable meteorological data such as data from climate re-analyses. Radić and Hock (2006) applied 'local scaling' to correct for the biases in climate model air temperature and precipitation for projecting the mass balance of Storglaciären until 2100. Downscaled temperature series were produced from GCMs and regional climate models (RCMs) by shifting the series by the averaged monthly differences between climate model and local-scale data (here ERA-40 re-analysis data) over a baseline period for which both GCM and local-scale data were available. Hence, the average seasonal cycle from ERA-40 was used as a reference by which seasonal cycles from the climate model could be 'corrected'. The results highlight the importance of including seasonally varying biases in the downscaling instead of assuming a constant bias throughout the year. Precipitation was scaled by the ratio of precipitation in the reference data set (here ERA-40) summed over the baseline period and the corresponding climate model precipitation sum. A drawback of statistical downscaling is the inherent assumption that the statistical relationships established over a baseline period will continue to hold in future climates.

An alternative approach to statistical downscaling is to use changes in GCM variables between a defined time slice in a GCM simulation of future climate and a baseline period corresponding to a period for which local climate data are available. These changes are then used to perturb the observed local climate data to drive a mass balance model and project future mass balance changes (e.g., Schneeberger et al., 2003). The changes in climate variables are often simply linearly interpolated between time slices to allow for transient simulations. Due to generally large interannual variability, however, projected climate variable changes can be sensitive to the choice of baseline period. This is especially true when large fluctuations occur around the baseline period or when the climate variable shows a trend during this period (Aðalgeirsdóttir et al., 2006).

Zhang et al. (2007a) used a dynamical downscaling approach, employing the high resolution regional model Polar MM5 driven by global atmospheric re-analysis to obtain temperature and precipitation data on a 10-km resolution grid to force a glacier mass balance model. The results of mass balance simulations using dynamically downscaled data and simulations based on observed temperature and precipitation data were in reasonably good agreement when calibration was used to minimize systematic biases in the MM5 downscaling.

7.6.2. Modeling mountain glacier and ice cap mass balance in the 21st century

While numerous studies have projected the response of individual glaciers to climate change, very few studies have attempted to project the response of Arctic glaciers on a regional scale. Zhang et al. (2007b) used dynamically downscaled daily maximum and minimum air temperatures and precipitation driven by the IPCC A1B emissions scenario of the CCSM3 climate model to force a temperature-index mass balance model for Hubbard Glacier (2460 km^2) and Bering Glacier (3630 km^2) in Alaska for the period 2010 to 2018. Both glaciers are projected to have increased accumulation, particularly

on the upper reaches of the glaciers, and increased ablation, particularly on the lower parts. The modeled mass balance of Bering Glacier is projected to become more negative on average (-2.0 m w.e./y compared to -1.3 m w.e./y during the 1994 to 2004 baseline period) and the modeled mass balance of Hubbard Glacier, which was positive during the baseline period, is projected to decrease slightly (to 0.3 m w.e./y from 0.4 m w.e./y).

Hock et al. (2007a) modeled cumulative mass balance changes of the 3 km^2 Storglaciären in Sweden until 2100 in response to model-projected regional temperature changes. They predicted total changes ranging from -81 to -121 m w.e., depending on the choice of the mass balance model. A fully distributed energy balance melt model produced a greater mass balance response than simpler temperature-index models. Projections based on the IPCC B2 emissions scenario showed a 50% to 90% decrease in ice volume by 2100 (Radić and Hock, 2006) (Figure 7.21). The volume change projections vary by 40% of the initial ice volume for six different GCM inputs to the mass balance model, and by 10% depending on the details of the mass balance model used. This is in agreement with the conclusion of Oerlemans et al. (2005) that the uncertainties due to a simple representation of glacier processes are less than those associated with the output from GCMs.

Huss et al. (2008) modeled the complete disappearance of the 4.4 km^2 Laika Glacier in Arctic Canada by 2100 in response to temperature and precipitation trends derived from a climate model forced by the IPCC A1B emissions scenario. Bassford et al. (2006) modeled complete wastage of Pioneer Ice Cap (~200 km^2) and Vavilov Ice Cap (~1770 km^2) on Severnaya Zemlya within about 370 and about 1080 years, respectively, by imposing a linear change in temperature and precipitation over the period 1990 to 2100 based on the IPCC IS92a 'business-as-usual' scenario, after which the climate was held constant. Although these ice caps have quite low static sensitivities of mass balance to changes in temperature and precipitation, they are nonetheless highly sensitive to long-term climate change owing to their hypsometry and location on relatively flat beds close to sea level.

Ananicheva and Krenke (2007) modeled mass changes of 17 glacier regions in northeastern Siberia and Kamchatka (1040 km^2) for the period 2040 to 2069 using climate projections from the ECHAM4 GCM driven by the IPCC A2 emissions scenario. The equilibrium-line altitude (ELA) on these glaciers is predicted to rise by 230 m in the northern parts and 500 m in the southern parts of northeastern Siberia. The upward shift is largest in Kamchatka. Based on known correlations between the ELA and the elevation of the glacier terminus, only 22% and 31% of the glacier areas are predicted to remain in northeastern Siberia and Kamchatka, respectively, in 2070 (Ananichava et al., 2010).

On larger regional or global scales, mass changes have been projected using either an 'indirect' approach based on mass balance sensitivity, defined as the change in mass balance caused by an instantaneous change in a climatic variable such as air temperature or precipitation, or a 'direct' approach based on transient mass balance modeling. The indirect approach involves computation of sensitivities, often for each month individually ('seasonal sensitivity characteristic'; Oerlemans and Reichert, 2000), which are used to calculate the change in
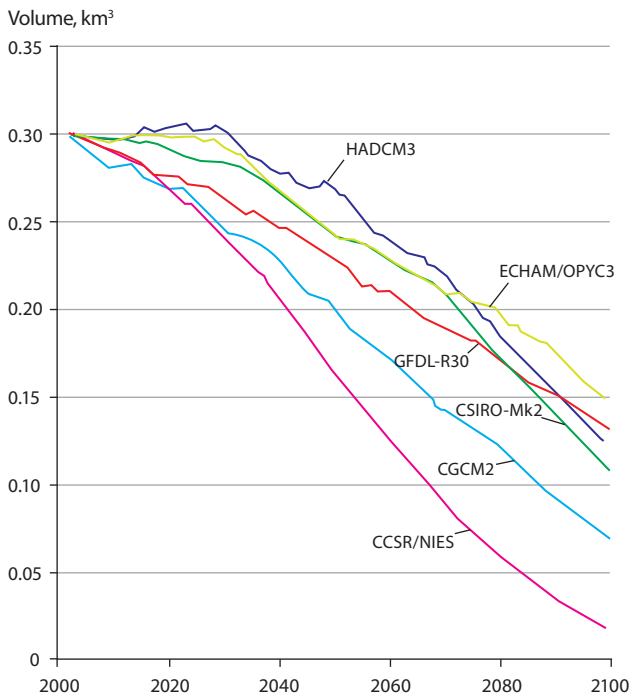
Figure 7.21. Volume projections for Storglaciären, Sweden, in the 21st century derived from six general circulation models (GCMs). Source: Radić and Hock (2006).

surface mass balance for given anomalies of temperature and precipitation. Mass balance sensitivities are determined for individual glaciers by energy balance or temperature-index modeling. Mass balance sensitivities to temperature computed on a 1° by 1° grid exhibit large spatial variations across the Arctic, with the largest sensitivities in Iceland, southern Greenland, and coastal Alaska, and the smallest sensitivities in Arctic Canada and northern Greenland (Hock et al., 2009; Figure 7.22).

Applying a regression-based temperature index model to 42 glaciers located north of 60° N, de Woul and Hock (2005) found that the mass balance sensitivity to a hypothetical temperature increase of +1 K ranged from -0.2 to -2.0 m/y. Sensitivity of the mass balance to a 10% increase in precipitation ranged from < +0.1 to +0.4 m/y. This offset the effect of a +1 K temperature increase by about 20% on average. Maritime glaciers had considerably higher mass balance sensitivities

than continental glaciers. Results were extrapolated to the entire Arctic by assuming area-weighted means of sensitivity to be representative for different Arctic regions. The estimated total contribution to sea level from Arctic glaciers (including the Greenland Ice Sheet) in response to a +1 K warming was about +0.6 mm/y. This compares to an estimate of +0.53 mm/y derived using a sensitivity approach that suggested about equal contributions from the Greenland Ice Sheet and the Arctic glaciers (Oerlemans et al., 2005).

Radić and Hock (2011) computed global mass balance and volume changes for more than 120 000 mountain glaciers and roughly 2600 ice caps around the world until 2100 using an elevation-dependent temperature-index mass balance model driven by output from ten GCMs forced by the IPCC A1B emissions scenario. Future volume changes were up-scaled to all glaciers using a regionally differentiated approach. For the Arctic glaciers (including those in Greenland surrounding the ice sheet), the projected volume loss due to melt (ablation by calving is not included) ranges from 51 to 136 mm sea-level equivalent (SLE) (i.e., 13% to 36% reduction in their current volume by 2100) depending on the choice of GCM (Figure 7.23). The multi-model mean is 88 ± 28 mm SLE, or 22% ± 7% volume reduction where the uncertainty range is ±1 standard deviation. Multi-model mean volume reductions by 2100 (relative to initial volumes) vary considerably among regions, with the smallest values in Greenland (8% ± 5%) and the largest values in Svalbard (54% ± 15%); however, the spread among models for most regions is large (Figure 7.23). Most models show the largest sea level contributions coming from the Canadian Arctic and Alaska (including the Yukon), followed by the Russian Arctic and Svalbard. Contributions from Arctic Canada show a very large range from less than 10 mm SLE contribution to the largest projected contribution for any of seven Arctic regions, indicating a large spread in the projected temperature and precipitation fields among the GCMs for this region (Figure 7.24).

7.6.3. Modeling ice dynamics and ice extent

Glacier size and geometry will change as a glacier responds to climate change. Hence, projections of glacier mass balance on timescales longer than a few decades need to take into account feedbacks between the mass balance and the changing glacier surface elevation.
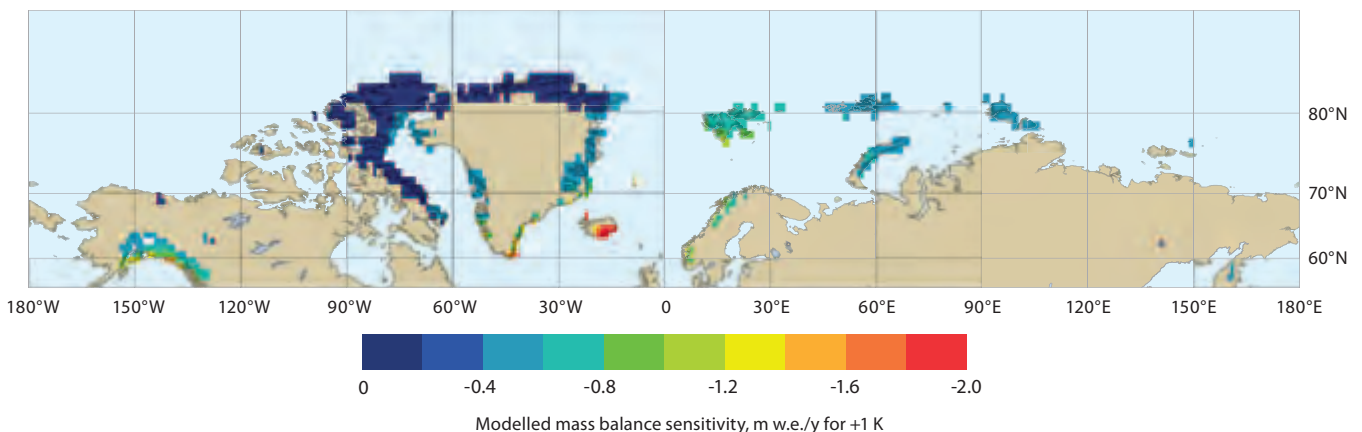
Figure 7.22. Modeled mass balance sensitivity, that is, the change in specific mass balance owing to a 1 K uniform warming. Source: Hock et al. (2009).
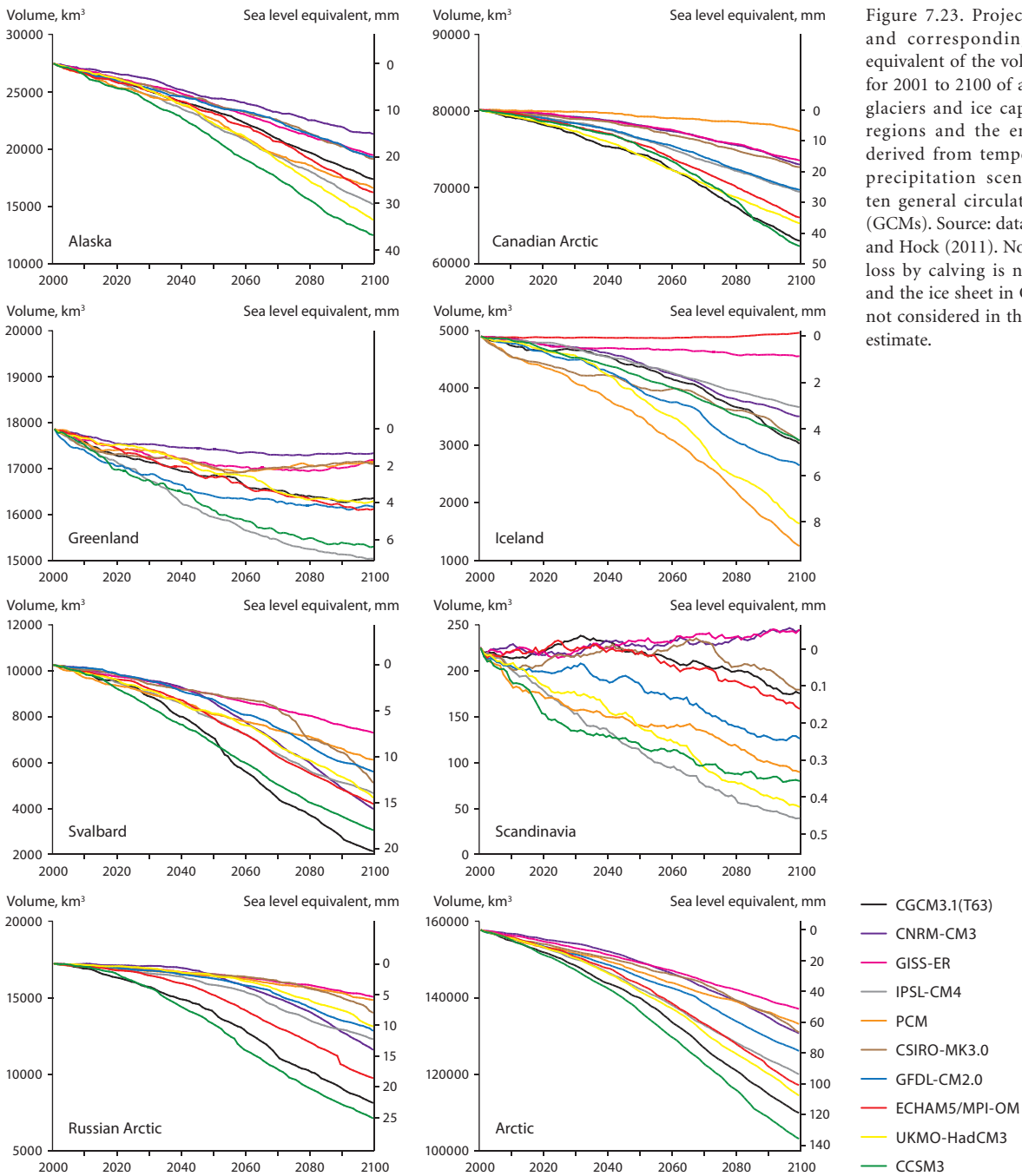
Figure 7.23. Projected volume and corresponding sea level equivalent of the volume change for 2001 to 2100 of all mountain glaciers and ice caps for seven regions and the entire Arctic derived from temperature and precipitation scenarios from ten general circulation models (GCMs). Source: data from Radić and Hock (2011). Note that mass loss by calving is not included and the ice sheet in Greenland is not considered in the Greenland estimate.

Ice flow models have been used to simulate area and thickness changes of individual glaciers (e.g., Schneeberger et al., 2003; Aðalgeirsdóttir et al., 2006). However, the detailed input data required to run such models are often not available. As a result, the effects of area changes are either simply neglected (e.g., Oerlemans et al., 2005; Zhang et al., 2007b) or approximated by scaling methods based on a functional relationship between glacier volume and glacier area (e.g., Radić and Hock, 2006). The total glacier volume is adjusted annually according to the volume changes computed from a climate-driven mass balance model. Annual area changes are then computed by inverting the volume-area relationship. Based on modeling of eleven glaciers, Schneeberger et al. (2003) showed that volume loss is overestimated by about 20% in a 100-year simulation if changes in glacier geometry are not included in the projections.

Detailed ice dynamic modeling is currently not possible on a regional scale due to a lack of the required data to run models for the vast majority of glaciers. Raper et al. (2000) developed a geometrical approach, in which the width, thickness, and length of a glacier are reduced as its volume and area decline. When applied statistically to the global population of glaciers and individually to ice caps, this approach shows that the reduction in glacier area strongly reduces ablation during the 21st century (Raper and Braithwaite, 2006). The reduction is about 45% under the IPCC A1B emissions scenario using output from two different GCMs.

A large number of glaciers in the Arctic terminate in the ocean and, therefore, lose mass through iceberg calving in addition to surface melting. Studies on a few Arctic marine-terminating ice caps (Dowdeswell et al., 2002, 2008; Burgess et al., 2005; Mair et al., 2009) indicate that calving may account for roughly 30% to 40% of total mass loss. However, model-

Figure 7.24. Projections of fractional volume change and sea-level equivalent of the volume change by 2100 for seven regions that include all Arctic glaciers. Projections are based on temperature and precipitation modeled by ten general circulation models (GCMs). Source: Radić and Hock (2011).

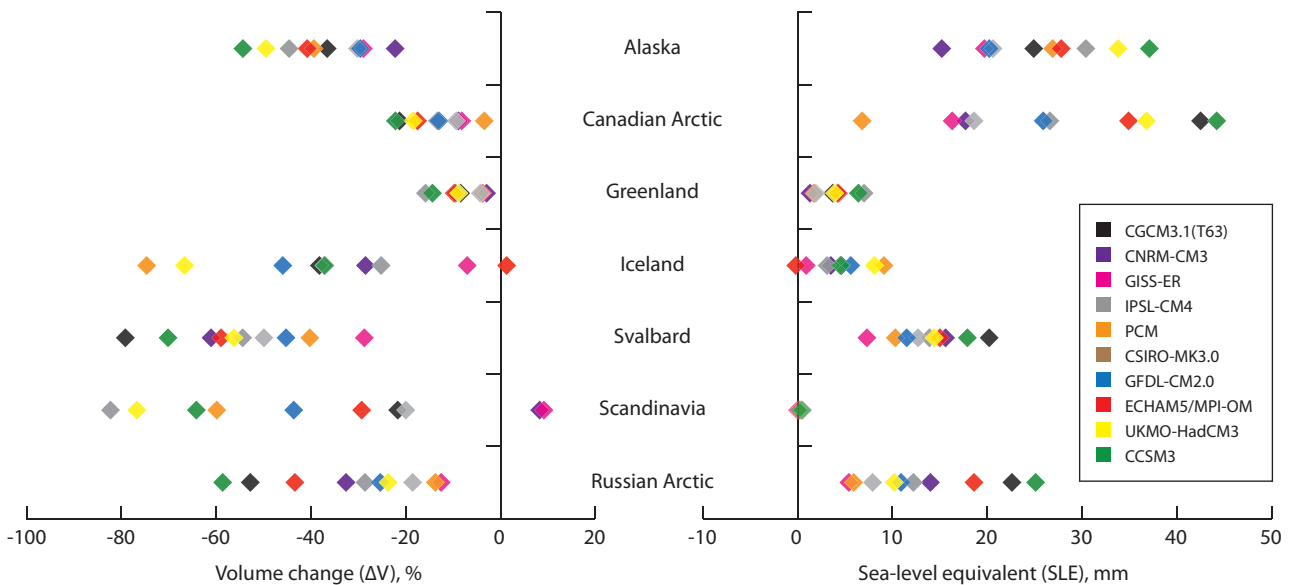based projections on regional scales neglect this effect. The current state-of-the-art in modeling does not allow meaningful projections of calving losses. However, calving rates are likely to increase as surface melting increases (Oerlemans et al., 2005).

7.7. Impacts of changes in mountain glaciers and ice caps

- The recent acceleration of mass loss from mountain glaciers and ice caps in the Arctic is such that, together with the Greenland Ice Sheet, they are likely to account for more than 60% of the current glacier wastage contribution to global sea level rise.

- Glacier wastage is likely to increase inputs of freshwater, sediment, and some nutrients to Arctic coastal waters, with potential impacts on water mass circulation, biological productivity, and ecosystem structure in affected fjord and nearshore marine environments.

- Retreat of tidewater glaciers onto land may be removing important feeding habitats that exist for seabirds and marine mammals in regions of upwelling in front of tidewater glacier termini and reducing the occurrence of grounded bergy bits that are used by seals for resting.

- Icebergs are a hazard to shipping and offshore platforms in the Arctic. The number and size distribution of bergs produced may change as tidewater glaciers retreat (e.g., more large bergs if floating ice tongues and ice shelves break up; more smaller bergs if calving from fast-flowing outlets increases). The way they circulate in Arctic waters is likely to change if sea-ice cover continues to decrease and surface water temperatures rise as a result. Bergs may become more mobile in summer as the restraining effect of sea ice declines, but disintegrate more rapidly in warmer surface waters.

- As climate warms, glacier runoff initially tends to increase due to higher melt rates, but ultimately declines as glacier

area shrinks. In most regions of the Arctic, the phase of declining runoff does not yet seem to have begun, but it may begin soon in the Russian Arctic mountains.

- Changes in glacier runoff will ultimately impact the viability of hydroelectric power operations in the Arctic. They will also be associated with changes in stream temperature, sediment load, and nutrient content that will initiate changes in the ecology of downstream river and lake environments.

- As rates of mass loss increase, 'legacy pollutants' stored in firn and glacier ice will be released back into the environment.

- As glaciers retreat, there are likely to be changes in the magnitude and frequency of a range of geomorphological hazards, including mass movements from deglaciated valley walls and outburst floods from ice-marginal and proglacial (moraine-dammed) lakes.

7.7.1. Impacts on sea level

In the most recent decade for which mass balance data were available for all regions (1996 to 2006), Dyurgerov and Meier (2005) reported that Arctic glaciers collectively lost mass at a rate of 165 Gt/y (0.45 mm/y sea-level equivalent; SLE). The dataset compiled by Cogley (2009a) yields a loss rate of 180 Gt/y (0.5 mm/y SLE) for the decade 1995 to 2005, and a loss rate of as much as 282 Gt/y (0.75 mm/y SLE) for the 2000 to 2005 pentad (see Section 7.9 for a more extended discussion). For comparison, Rignot et al. (2008) calculated a mass loss from Greenland of 267 Gt/y (0.71 mm/y SLE) for 2007. These results suggest that, together, Arctic glaciers and the Greenland Ice Sheet may account for more than 60% of the current glacier wastage contribution to global sea level rise (2.1 mm/y; Cazenave et al., 2009).

The acceleration of mass loss from Arctic glaciers and ice caps for the period 1985 to 2003 calculated using data from Dyurgerov and Meier (2005) is -7.3 km^3/y^2 ice equivalent, or 6.5 Gt/y^2 (0.02 mm/y^2 SLE) (Figure 7.25). Note that at this rate
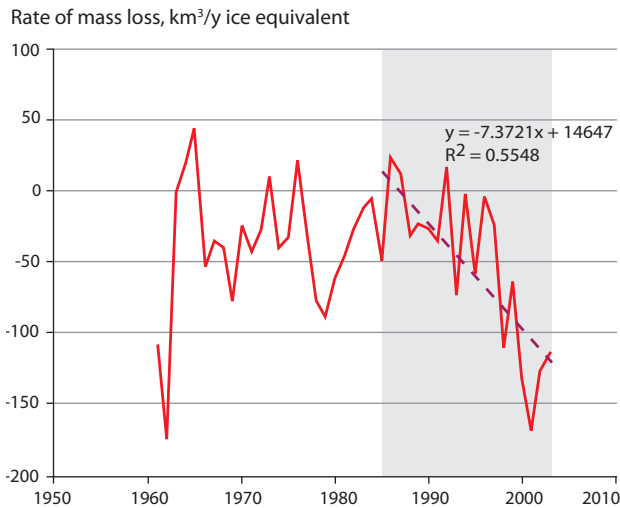
Rate of mass loss, km³/y ice equivalent

Figure 7.25. Rate of volume loss from Arctic glaciers and ice caps, derived from Dyurgerov and Meier, (2005). Source: AMAP.

of acceleration the estimated reservoir of mountain glacier and ice cap sea-level equivalent of 0.41 m (Radić and Hock, 2010) is being drawn down by only about 0.2% per year. While the current trend cannot be reliably extrapolated any significant distance into the future, depleting the reservoir is unlikely to result in a decline in annual total balance for at least the next several decades.

Analysis of the Cogley (2009a) dataset yields an estimate of mass losses from the glaciers surrounding the Greenland Ice Sheet of about 35 to 45 Gt/y for the 2000 to 2005 pentad. Some methods used to assess the mass balance of the Greenland Ice Sheet, such as most GRACE gravimetric observations, account for these losses, while other methods, such as satellite altimetric and mass flux observations, do not. It is therefore important, when combining mass loss estimates from this chapter with those in Chapter 8 on the Greenland Ice Sheet, to identify which glaciers are included in each particular method to avoid double counting.

7.7.2. Impacts on the marine environment

7.7.2.1. Freshwater and associated chemical fluxes

Glacier runoff is projected to increase in many parts of the Arctic for a period of decades or longer before declining (see also Section 7.7.3). This increased freshwater flux will influence conditions in coastal and marine waters. In fjords, freshwater discharge in summer produces a stable surface layer with low salinity (e.g., Svendsen et al., 2002), an effect that would be intensified with accelerated glacier melt. In the Gulf of Alaska, the majority (78%) of freshwater discharge occurs via a large number of smaller rivers and streams rather than a small number of point sources (Neal et al., 2010), producing baroclinic transport along the coast that is modified by wind stress (Royer, 1982). This along-coast transport is a major source of freshwater to the Bering Sea shelf and the Arctic Ocean (Weingartner et al., 2005). Runoff from glaciers and icefields contributes about half of the freshwater discharge to the Gulf of Alaska, with 10% currently from wastage (Neal et

al., 2010). Because the magnitude of baroclinic transport varies with freshwater flux, future changes in glacier runoff volume and timing will have significant implications for the strength of the Alaska Coastal Current, associated marine ecosystems, and economically valuable fisheries.

In addition to changes in freshwater flux, changes in material export may influence coastal and marine ecosystems. Hood and Scott (2008) monitored three streams in southeastern Alaska and found that decreasing glacier cover was associated with decreased phosphorus export and increased export of nitrogen and dissolved organic matter, although with a decrease in the more labile forms of organic matter. Nitrogen export increases in the early stages of vegetation succession on deglaciated forelands due to the establishment of nitrogen-fixing species such as alder, but is likely to decrease with the later establishment of other species such as conifers.

Increased glacier melt may be associated with increased sediment loads, which would increase turbidity in coastal waters. Increased turbidity would reduce the penetration of solar radiation, including harmful UV-B radiation, with implications for primary production and ecosystem functioning (Hanelt et al., 2001; Erga et al., 2005).

7.7.2.2. Seabird and marine mammal habitat

Tidewater glacier fronts are known to be 'hot spots' for seabirds and marine mammals in many Arctic areas. In Svalbard, a strong preference for these habitats has been identified for ringed seals (*Phoca hispida*) and beluga (*Delphinapterus leucas*) during summer and autumn (Lydersen et al., 2001; Freitas et al., 2008). Individuals of both species spend significant amounts of their time in proximity to tidewater glacier fronts, and their diving behavior in these areas suggests that they are feeding in these locations, which are known to be areas of high productivity. Enhanced production in these areas is probably due to freshwater outflows that drive upwelling of deep water close to glacier fronts, bringing nutrients to the surface and stimulating phytoplankton blooms and zooplankton growth. In addition, the cold-water outflows from the glaciers may stun or kill invertebrates, thus attracting predators such as polar cod (*Boreogadus saida*) and capelin (*Mallotus villosus*) that are an important part of the diet of beluga and ringed seal (Dahl et al., 2000; Labansen et al., 2007), as well as a variety of seabirds and other marine mammals. In addition, ringed seals, bearded seals (*Erignathus barbatus*), and other pinnipeds use bergy bits as resting platforms. Retreat of tidewater glacier fronts onto land would decrease the availability of this preferred habitat at a time when decreases in late summer sea-ice extent are making the other preferred summer feeding habitats of Arctic pinnipeds less available. In addition, changes during glacier retreat in the lability of organic matter and the amount and types of nutrients that are delivered from land to the marine environment in meltwater runoff may reduce the productivity of these environments and their attractiveness to marine mammals and seabirds.

7.7.2.3. Iceberg production and drift

In some Arctic and sub-Arctic offshore areas, icebergs present a significant danger for marine operations including shipping

and commercial fisheries, geological exploration surveys, and production of hydrocarbons from sea surface structures such as oil rigs (Figure 7.26). The Grand Banks of Newfoundland, the Labrador Sea, offshore of West Greenland, Davis Strait, and the Barents Sea are typical examples of such areas. Damage to or even destruction of a stationary platform or its communication lines by an iceberg will result in enormous economic losses and possibly significant environmental damage. Both iceberg plan form and keel depth are important quantities to measure in this context (Woodworth-Lynas et al., 1991). Non-stationary platforms and other floating structures (e.g., floating production, storage, and offloading systems), capable of leaving their operation localities, will limit economic losses owing to interruption of hydrocarbon production for the period of shutdown due to iceberg threat.

The general direction of the drift of icebergs near the Labrador coast and the Grand Banks of Newfoundland coincides with the direction of a strong Labrador Current. Iceberg trajectories and the most dangerous drift directions are known. Over the past few decades, Canada has gained significant experience in minimizing iceberg threats and has introduced a complex system of operational practices known as the Iceberg Management System.

Currents in the Barents Sea carry icebergs from the north to the southwest (Johannessen et al., 1999; Dmitriev and Nesterov, 2007). Some may enter the central and southern regions of the area, threatening navigation and operations in the *Shtokman* gas-condensate field. An analysis of atmospheric conditions at the time of the southernmost spreading of icebergs (e.g., 1881, 1929, 1989, and 2003) has revealed that all cases were characterized by prolonged periods (three to five months) of winds from the north (Buzin et al., 2008). Under such conditions, large numbers of icebergs concentrated near Spitsbergen and Franz Josef Land (their source) are exported southward. In some years (1989, 2003), icebergs reached the *Shtokman* gas-condensate field area and in particularly extreme cases (1881, 1929) reached the coast of northern Norway

and the Kola Peninsula. Sightings of an anomalously large concentration of icebergs and bergy bits (around 100 pieces) in the *Shtokman* gas-condensate field area in May 2003 (Naumov et al., 2003) significantly influenced the design concept for the production complex. The boundary of iceberg drift in the Barents Sea has been displaced to the south over the past 60 years (Zubakin et al., 2006; Buzin et al., 2008), probably related to climate warming and a corresponding increase in the production of icebergs (Figure 7.27). The tendency for the ice edge in the Barents Sea to be displaced toward the north in summer during the past 10 to 15 years (Divine and Dick, 2006) may result in earlier and easier release of icebergs stuck in the straits of Franz Josef Land. Consequently, a larger number of icebergs may be in free drift and represent a threat to shipping and platforms. However, the life expectancy of the icebergs also may be reduced in a warmer climate (Weeks and Campbell, 1973).

The iceberg discharge from the Greenland Ice Sheet is known to have increased by more than 40% from 1995 to 2005 (Rignot and Kanagaratnam, 2006). This has not, however, increased iceberg numbers on the Grand Banks or in the Labrador Sea. The number of icebergs reaching the Grand Banks has actually decreased, probably owing to increased surface water temperatures in the region and decreased ice cover (which protects icebergs from the dynamic impact of ocean waves) (McClintock et al., 2007). However, potential iceberg damage to sea-floor engineering structures is a function of iceberg keel depth as well as iceberg number. Icebergs with keel depths greater than 500 m have been observed in the Scoresby Sund fjord system of East Greenland, but few other systematic measurements of the keel depth of Arctic icebergs exist (Dowdeswell et al., 1992). The occurrence of recent plow marks on the sea floor, produced where drifting icebergs go aground, is evidence of this natural hazard, although many plow marks on Arctic continental shelves are not produced by modern icebergs (Dowdeswell et al., 1993; Syvitski et al., 2001).

Figure 7.26. Large iceberg (100 m length, 30 m width, and 15 m height above sea surface) drifting in the Barents Sea to the southwest of Franz Josef Land on 30 May 2009. Source: Andrey Masanov, Arctic and Antarctic Research Institute, Russian Federation.
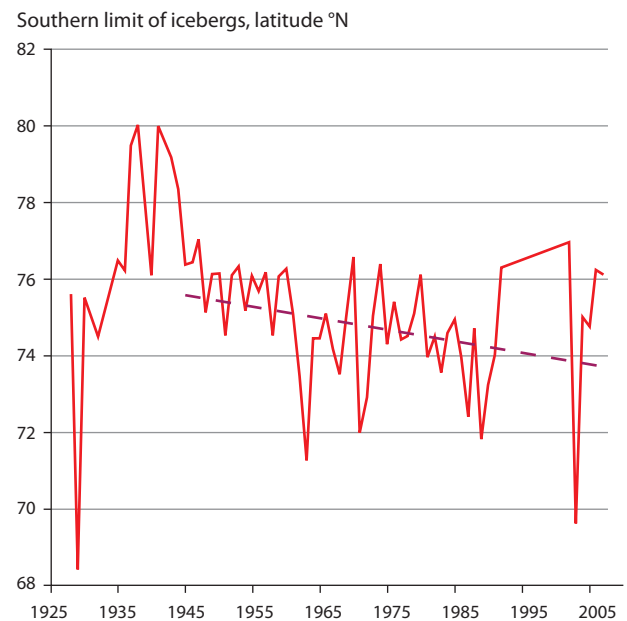

Figure 7.27. Multiyear variability of the southern limit of icebergs in the Barents Sea for the period 1928 to 2008. Source: Buzin et al. (2008).

7.7.3. Impacts on water resources

7.7.3.1. Glacier runoff

Streamflow in glacier-fed streams tends to peak later in the summer and to continue to exhibit diurnal fluctuations later in the melt season than streamflow in nival (snowmelt-dominated) streams (Church, 1974). However, the contrast between nival and proglacial regimes can be obscured by the weather patterns dominating the particular melt season, particularly when melt begins later than normal (Marsh and Woo, 1981). Research in temperate areas and in the Arctic indicates that detectable glacier melt contributions to streamflow occur in catchments with more than about 5% glacier cover, while contributions to summer flow are detectable even in catchments with as little as 2% glacier cover (e.g., Marsh and Woo, 1981; Stahl and Moore, 2006). Based on these criteria, glaciers are an important water resource throughout much of the Arctic.

Over annual or longer timescales, the influence of glaciers on streamflow can be assessed through the catchment water balance, which can be expressed as follows:

$$[1] \quad Q = P - E - f_g B + GW_{net} - \Delta S$$

where Q is streamflow, P is precipitation, E is evapotranspiration (including sublimation), f_g is the fractional glacier area within the catchment, B is the net mass balance for the glaciers, GW_{net} is the net flux of groundwater across the catchment boundaries, and ΔS represents changes in storage of seasonal snow, subsurface water, and surface water. All terms in [1] are typically expressed as a depth of water averaged over the catchment area. Negative net mass balance (glacier loss) augments streamflow relative to a basin lacking glacier cover. This contribution to streamflow ($f_g B$) is commonly referred to as 'wastage runoff'.

Table 7.3 highlights the contributions of glacier wastage (negative mass balance) to mean annual streamflow over recent decades, based on published water balance calculations and hydrological modeling. The significance of glacier melt contributions to streamflow tends to be greater than average in years with low precipitation and warm summers. At De Geerdalen, Svalbard, for example, glacier wastage contributed up to 20% of annual runoff, roughly double the average, in a year when annual precipitation was 40% below average (Killingtveit et al., 2003).

Assuming no further climate change, glaciers currently experiencing negative mass balance would recede until they achieved a hypsometry for which $B = 0$, as long as the ELA is below the top of the glacier. If the ELA lies above the maximum elevation of the glacier, the glacier would ultimately disappear. Under this assumption of constant climate, average annual runoff would ultimately decrease by an amount equal to the current contribution due to glacier wastage. Decreases in runoff would be particularly severe in years with low precipitation. Flowers et al. (2005) simulated the future evolution of Vatnajökull, Iceland, assuming a continuation of the 1961 to 1990 climatology, and found that glacier runoff would decrease by about 20% after 200 years.

Climatic warming will result in a longer melt period and more intense melt, particularly as firn cover is depleted, exposing lower-albedo glacier ice to solar radiation. As a consequence, glacier runoff will increase with the onset of a warming trend. However, increased melt is accompanied by glacier thinning and retreat, and at some point the decrease in glacier area will cause glacier runoff to decrease. While this conceptual model is widely recognized (e.g., Jansson et al., 2005; Moore et al., 2009), the timescales associated with the transition from increasing to decreasing flow are uncertain. Available studies focused on southern Alaska, northeastern British Columbia, and southwestern Yukon have reported positive trends in streamflow for glacier-fed rivers, suggesting that the glaciers in those regions are still in the initial phase of their response to warming (Fleming and Clarke, 2003; Brabets and Walvoord, 2009).

Runoff from glaciers in the temperate and sub-Arctic regions of Russia was calculated from the USSR Glacier Inventory for the period 1966 to 1980 based on the assumption that glacier systems are in equilibrium with climate. The total glacier runoff over the period 1966 to 1980 was estimated to be no more than 3 Gt (Ananicheva and Krenke, 2007). To project future changes in glacier runoff in Russia, the morphology and regime of glacier systems were adjusted on the basis of predicted changes in ELA under a climate-warming scenario derived from the ECHAM4 GCM driven by the IPCC A2 emissions scenario. Despite a significant reduction of the glacierized area, the glacier runoff from southeastern Siberia and the Sredinny Range (Kamchatka) was predicted to increase through the 21st century owing to increased ablation. Glaciers in the Kronotsky Range (Kamchatka) were predicted to almost disappear, resulting in a sharp reduction in glacier-melt runoff (Ananicheva and Krenke, 2007).

Transient responses of glacier geometry and runoff to climate warming have been simulated for Hofsjökull and

Table 7.3. Runoff contributions associated with glacier wastage.

Catchment	Location	Period	Catchment area, km²	Glacier coverage, %	Wastage contribution to streamflow, %	Q, kg/m²	Source
Zackenberg	Northeast Greenland	1997 – 2005	512	20	63	428	Mernild et al., 2008a
Mittivakkat	Ammassalik Island, Greenland	1993 – 2004	14.2	78	23	1970	Mernild et al., 2008b
Bayelva	Svalbard	1990 – 2001	30.9	55	23	1050	Killingtveit et al., 2003
De Geerdalen	Svalbard	1990 – 2001	34.4	17	9	539	Killingtveit et al., 2003
Endalen	Svalbard	1990 – 2001	28.8	20	19	545	Killingtveit et al., 2003

Vatnajökull, Iceland (Flowers et al., 2005; Aðalgeirsdóttir et al., 2006), and the results are consistent with the conceptual model presented above. Figure 7.28 illustrates that, as the rate of warming increases, the interval with maximum glacier runoff for Hofsjökull moves forward in time and the increase in runoff is greater. For simulations involving a warming-related increase in precipitation, the effect is to increase the magnitude of runoff increase and delay the timing of peak runoff. Flowers et al. (2005) also reported that changes in the surface topography of the ice cap would cause shifts in the hydrological divides, effectively redistributing outflow among the outlet glaciers.

Climate warming and glacier response will modify seasonal patterns of runoff in addition to changing annual totals. Jónsdóttir (2008) simulated runoff in Iceland for climate projections based on the IPCC A2 and B2 emissions scenarios and projections of glacier geometry using a dynamic glacier model (Aðalgeirsdóttir et al., 2006; Jóhannesson et al., 2006). For the period 2071 to 2100, glacier-melt contributions to streamflow are projected to increase in all seasons, while runoff from unglacierized areas increases from October to April and decreases from May to September, jointly producing a significant shift in the seasonal pattern of runoff in partially glacier-fed catchments.

The highest peak flows in proglacial streams are often associated with outburst floods (Ng et al., 2007), in which water stored on, under, or within the glacier is released suddenly, for example, by the rapid growth of a meltwater channel (e.g., Mernild et al., 2008a). There is some evidence that subglacially derived outburst events are more extreme in warmer summers when there is greater meltwater generation (Skidmore and Sharp, 1999), suggesting that flood risk in glacier-fed Arctic catchments may increase as a result of climate warming. When outburst floods are derived from ice-dammed marginal lakes, thinning of the ice dams due to increased melting may result in floods that occur earlier in the melt season and are smaller and more frequent than they are at present. Lakes will fill sooner and more rapidly than in the past and the accumulated lake water may breach the ice dam more easily than before. Earlier onset of lake drainage may allow multiple filling/drainage cycles during a melt season, especially because melt seasons will be longer in a warmer climate.

7.7.3.2. **Water quality**

Climate warming will influence proglacial water quality as well as streamflow. Increasing meltwater runoff in the initial phases of warming should tend to moderate diurnal increases in stream temperature. However, as glaciers retreat, the length of stream exposed to solar radiation and other heat inputs will increase, potentially counteracting the effect of increased flow. In addition, the emergence of proglacial lakes can increase warming. Milner et al. (2008) found that maximum stream temperatures at Wolf Point Creek, which drains into Glacier Bay, Alaska, were 2 °C in 1977, roughly 30 years after the creek emerged due to glacier retreat. From 1996 onward, maximum temperatures have consistently exceeded 15 °C, at least in part due to the increase in size of the upstream lake, which initially formed in the 1970s. Once glaciers reach the stage of declining flow, there should be a trend to increasing stream temperature. At that point, the main factor mitigating this increase would be development of
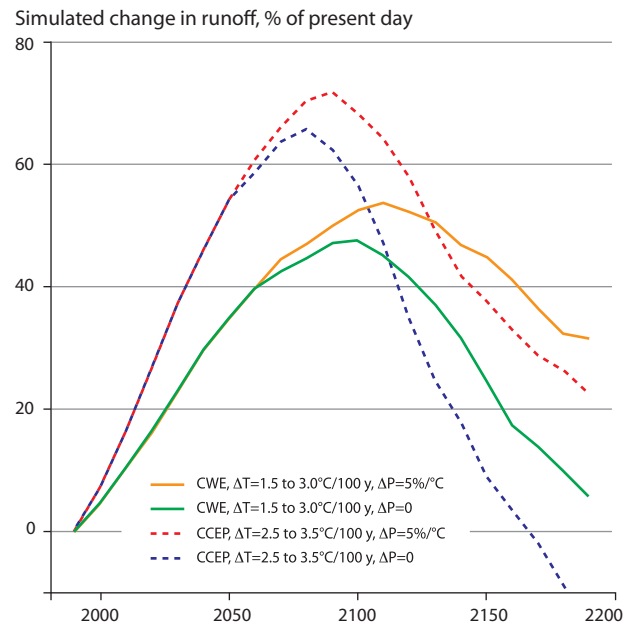
Figure 7.28. Simulated variations in glacier runoff from Hofsjökull, Iceland, under four climate scenarios. Source: Aðalgeirsdóttir et al. (2006).

riparian forests where they do not currently exist, which would increase stream shading and enhance bank stability, resulting in a narrower, deeper stream. Both effects would tend to reduce stream warming, but would be relevant only in sub-Arctic and temperate areas such as the Gulf of Alaska.

Focusing on temperate regions, Moore et al. (2009) concluded that the onset of negative mass balance and glacier retreat would be likely to result in an initial increase in suspended sediment concentrations in proglacial streams, followed by a longer-term decrease. This long-term projection is consistent with studies that show that glacier-fed streams in the Arctic typically yield orders of magnitude more sediment than those without glacier contributions (e.g., Hasholt and Mernild, 2008). However, temporal changes in sediment concentration could vary substantially among basins due to site-specific factors, particularly the emergence of proglacial lakes, which trap sediment and thus reduce downstream transport (Hasholt et al., 2008). Future patterns of suspended sediment concentration would also be complicated in cases where glaciers shift from being cold-based or polythermal to temperate, as well as by thawing of permafrost, which could influence the erosion of stream channel beds and banks (McNamara and Kane, 2009).

Glacier retreat may be accompanied by changes in streamwater chemistry. Glacial forefields are subject to different weathering and nutrient cycling processes than those that dominate in subglacial environments, particularly as soils and vegetation develop (Anderson, 2007; Hood and Scott, 2008). Available studies, not all from Arctic areas, suggest that concentrations of nutrients and dissolved organic carbon should increase as glaciers retreat, although the forms of some nutrients and organic matter might change, possibly influencing their availability to aquatic organisms (Lafrenière and Sharp, 2005; Filippelli et al., 2006; Hood and Scott, 2008). In southeastern Alaska, decreasing basin glacier cover was found to be associated with higher concentrations of soluble reactive phosphorus and higher proportions of bioavailable (labile)

dissolved organic carbon (Hood and Berner, 2009; Hood et al., 2009). Fluvial transport of dissolved organic carbon could be influenced by changes in suspended sediment concentration, because high turbidity reduces photodegradation of dissolved organic carbon in streamwater.

7.7.3.3. Water resources and hydroelectric power generation

Most studies suggest that Arctic glaciers are currently in the initial phase of their response to warming, in which runoff increases, and that glacier runoff should continue to increase over the next few decades or so, providing an increase in water availability. However, some Russian results indicate that some areas (especially those where the glaciers are predominantly small) are likely to experience flow decreases over the next few decades as a result of glacier retreat (Ananicheva and Krenke, 2007). These contrasting behaviors could either reflect fundamental differences among glaciers in the timescale of response to warming, or suggest that glaciers are currently at different stages of their response. Further site-specific assessments are clearly warranted.

Glacier runoff is significant for hydroelectric power generation in the upper Yukon basin (at Whitehorse, Yukon, Canada), Iceland, Greenland, and in many river systems in Norway. A trend to increasing flows would also tend to increase the hydroelectric power resource. In Iceland, Jónsdóttir (2008) projected a substantial increase in hydroelectric power potential for the period 2071 to 2100, but a substantial decrease over the following century as the glaciers disappeared under the IPCC A2 and B2 emissions scenarios employed.

Reservoir operations could be influenced not only by the effects of changes in streamflow, but also by changes in sediment transport and stream temperature. Increased sediment transport would reduce the operational lifetime of a reservoir by infilling, while suspended sediment concentration and inflow temperature together would influence mixing processes and temperature profiles, in turn influencing habitat conditions for aquatic organisms as well as the temperature of water downstream.

7.7.3.4. Aquatic ecosystems

Glacier-fed rivers typically support a distinctive flora and fauna determined by physicochemical conditions, including the dominant influences of water temperature and channel stability (Milner and Petts, 1994; Milner et al., 2001). While bedload and suspended sediment transport typically peak during maximum glacial melt in summer (Østrem, 1975b), lower water levels during spring and autumn result in greater water clarity and channel stability, creating 'windows of opportunity' when algal and macroinvertebrate productivity is higher. Some of the largest runs of salmon in south-central Alaska occur in glacially influenced systems (Dorava and Milner, 2000) as glacial runoff enhances summer flow relative to rainfall- or snowmelt-dominated catchments, facilitating the migration of salmon from the ocean to their spawning grounds in tributary streams via the main channel corridor.

Since increases in glacial runoff will enhance summer flow and increase the duration of meltwater runoff, they will probably also facilitate the migration of adult salmon to their spawning grounds. Increased glacial runoff will also increase the amount of available spawning habitat and would potentially enhance winter baseflow through recharge of groundwater during the summer (Smith et al., 2001; Brown et al., 2003). The survival of young salmonids during the winter is significantly correlated with the amount of winter discharge or winter groundwater inputs (Fleming, 2005). Any changes in water source contributions that enhance winter flow should be beneficial to salmonid populations and mitigate the effect of overwintering mortality on salmonids.

In the phase of declining glacier runoff, bedload and suspended sediment transport will potentially decrease in the long term as glacial and paraglacial erosion (erosion by non-glacial processes of previously deposited glacial and fluvioglacial sediments) decrease and the stream's transport capacity decreases (Fleming and Clarke, 2005). In addition, newly formed proglacial lakes trap sediment, reducing sediment loads downstream. Fleming (2005) emphasized the potentially favorable effects of glaciers and lakes in combination on salmonid habitat: glaciers maintain late summer flow while lakes reduce sediment transport. Reduced bedload transport could allow streams to become more stable, and thus narrower and deeper. The projected reduction in suspended sediment load with ongoing glacier retreat should allow greater light penetration into the water column and increase algal productivity, particularly during certain periods of the year. The effect of increased light would be augmented by an increase in nutrient concentrations as vegetation and soils develop following glacier retreat. However, the effects of decreased turbidity on ecosystem productivity need to be considered in the context of possible decreases in late summer streamflow and increasing late summer water temperature.

There is clear evidence of a strong relationship between catchment glacierization and macroinvertebrate taxon richness in glacially influenced streams. In a stream fed by a receding glacier in southeastern Alaska, Milner et al. (2008) found that the benthic macroinvertebrate community supported only five taxa when 70% of the catchment was glacierized but 24 taxa after the glacier had disappeared two decades later. Füreder (2007) found similar trends in alpine streams with differing degrees of glacierization. This evidence suggests that changes in glacial runoff and fluvial sediment transport due to glacier shrinkage may cause an increase in the diversity of the benthic macroinvertebrate community. An increase in algal productivity will potentially enhance macroinvertebrate abundance. With a change in benthic community structure, dominant species traits will reflect a reduction in glacial contribution to flow. Macroinvertebrate body size will become larger and life cycles longer, with the benthic community dominated by habitat generalists with less omnivory and flattened forms (Ilg and Castella, 2006; Füreder, 2007). The fauna would also be less cold-adapted with reduced mobility of the adults. Enhanced abundance and increased body size will favor food availability for fish. Although an overall increase in taxonomic richness and abundance in benthic macroinvertebrate communities is predicted with decreased glacial inputs, some cold-specialist benthic species will be lost, including some endemic species. Some species will not adapt rapidly enough to survive

the anticipated rate of deglaciation in future decades (Hari et al., 2006).

The effect of shrinking glaciers on fish populations in glacially influenced rivers will depend upon whether the system is an alpine or a more lowland system. However, over time reduced glacial input into these rivers will reduce summer flow and may influence salmon migration upstream. Lower flow will also reduce the amount of habitat available adjacent to the main glacier-fed channel, including side channels and side sloughs. Side channels and sloughs are sustained by meltwaters of the main channel (Richardson and Milner, 2005). For many river systems in northerly regions, winter is a critical period when overwintering mortality for salmonids can be high if they are unable to migrate to suitable refugia, such as side channels. Water temperature is a critical variable influencing fish distribution and each species has an optimum range for different life stages. Matulla et al. (2007) predicted that the ranges of alpine stream salmonids (e.g., brown trout *Salmo trutta* and grayling *Thymallus thymallus*) would contract upstream due to changes in water temperature regimes under climate change scenarios in which the cold water input from glacial runoff decreases. Projected changes in water temperature in cold rivers can also favor non-indigenous species that have wider thermal tolerance and can out-compete indigenous taxa at higher water temperature (Ficke et al., 2007). In the Rocky Mountains, cutthroat trout (*Oncorhynchus clarkii*) have been displaced by non-indigenous brook trout (*Salvelinus fontinalis*) and brown trout that are more efficient at securing food resources at higher temperatures (Hauer et al., 1997). However, in streams that are currently colder than is optimal for salmonid growth, some degree of warming might increase salmonid production.

Large-scale glacial recession creates new habitat that can potentially be colonized by organisms including fish. The creation of new stream habitat due to glacial recession has been extensive along the coasts of southeastern and south-central Alaska due both to climate change and changing local environmental conditions (Milner et al., 2008). Although many of the recently formed streams are relatively short, the creation of new salmonid habitat is significant at a time when human activity threatens salmonid stocks in other parts of the Pacific Northwest region of North America (Nehlsen et al., 1991). Pink salmon (*Oncorhynchus gorbuscha*) can establish substantial populations from a small colonizing population of spawners within a few generations, and other species (including non-anadromous fish such as sticklebacks) colonize rapidly (Milner et al., 2008). As migrant salmon die soon after spawning, rotting of their carcasses results in a significant input of nutrients and organic matter to affected catchments, and this can promote productivity in other parts of the ecosystem.

7.7.3.5. **Contaminant release from melting glaciers**

Many contaminants originating from lower latitudes are carried to the Arctic by long-range atmospheric transport and stored in the firn and ice of glaciers and ice caps (Chapter 11, Section 11.3). These contaminants include sulfate and nitrate, which have increased the acidity of snow and ice above pre-industrial levels (Koerner and Fisher, 1982), metals such as lead (Zheng et al., 2003), persistent organic pollutants such as PCBs (polychlorinated biphenyls), DDT and other organochlorine pesticides (Gregor and Gummer, 1989; Hermanson et al., 2005), as well as radionuclides associated with the Chernobyl reactor failure in 1986 and with nuclear testing dating back to 1961 to 1962 (Pinglot et al., 1999). Stored contaminants may be released back into the environment by glacier melt and the process can be accelerated if melt rates increase over time (Blais et al., 2001).

Contaminants stored in seasonal snow away from glaciers interact with soil and aquifer materials prior to their discharge to receiving waterbodies such as lakes and streams. During these interactions, a proportion of the contaminants (especially persistent organic pollutants) released via melting will be retained through sorption onto organic matter and cation exchange on clay particles in the soil. In contrast, glacier meltwater has much less interaction with organic matter and clays in soils before leaving the glacier, leading to less retention and higher rates of contaminant export to lakes and rivers downstream, as has been found in the Canadian Rockies (Blais et al., 2001; Lafreniére et al., 2006).

All the contaminants stored in glaciers would eventually be discharged in meltwater runoff regardless of climate change, but there would be significant lag times associated with the flow of ice from the accumulation to ablation zones. Climate warming accelerates melting of snow, firn, and ice, potentially releasing stored contaminants to receiving waters in a more concentrated pulse. However, thus far there has been no research in the Arctic directly addressing this impact.

7.7.4. **Glacier ecosystems**

Distinctive microbial ecosystems are found on and under glaciers (Hodson et al., 2008). On the glacier surface, the supraglacial ecosystem exists in the snowpack, supraglacial streams, melt pools, and cryoconites (water-filled holes on glacier surfaces that contain dust and microbial communities). It comprises a diverse array of bacteria, algae, fungi, viruses, rotifers, and tardigrades. *In situ* rates of primary production and respiration on glacier ice surfaces in summer are comparable to those encountered in soils in warmer and more nutrient-rich regions. Measurements of net production suggest that glacier ice surfaces may be largely autotrophic systems (Anesio et al., 2009), unlike most lakes and rivers, which tend to be heterotrophic systems.

At the glacier bed, a subglacial ecosystem is found in basal ice, subglacial sediments, meltwaters, and volcanic subglacial lakes (Gaidos et al., 2004, 2009; Hodson et al., 2008). It includes a mixture of aerobic and anaerobic (chemoautotrophic and heterotrophic) bacteria and probably also fungi and viruses (D'Elia et al., 2008). These microorganisms have metabolisms based on the oxidation of organic carbon, sulfide, sulfur, or hydrogen, and they use oxygen, nitrate, sulfate, iron(III), manganese(IV), or carbon as terminal electron acceptors. They play a critical role in subglacial chemical weathering (Tranter et al., 2002, 2005) and in controlling the biogeochemistry, nutrient loading, and properties of dissolved organic matter in glacial runoff (Hodson et al., 2008; Gaidos et al., 2009). They may also be involved in the production of greenhouse gases (carbon dioxide and methane), with possible feedbacks to the climate system if these gases are stored in basal ice or subglacial sediments and released to the atmosphere during deglaciation (Sharp et al., 1999; Skidmore et al., 2000; Wadham et al., 2008; Boyd et al., 2010).

The impact of climate change on these ecosystems is likely to be complicated as there will be a number of competing influences (Hodson et al., 2008). In the long term, changes in global glacier area will limit their geographical extent, and total deglaciation would be likely to eradicate these ecosystems completely. In the shorter term, increased surface melt would probably enhance supraglacial ecosystem productivity. As end-of-summer snowline elevations rise, the extent of supraglacial stream and meltwater habitats will increase at the expense of snowpack habitats. Higher ice melt rates may swamp these systems with water, continually redistributing them across the glacier surface. Increased melting of glacier ice may increase the rate of release of glacially entombed microorganisms into the supraglacial environment. This may increase the transfer of carbon, nutrients, and biota into downstream (subglacial and proglacial) aquatic environments. Increased water fluxes would be likely to increase the oxygenation of glacier beds in summer, favoring aerobic over anaerobic bacterial processes. They might also result in more rapid transit of meltwater through the subglacial ecosystem, especially as glaciers shrink, reducing the efficiency of utilization of supplies of nutrients and organic matter by subglacial microbial communities. Alternatively, increased delivery of carbon and nutrients could stimulate microbial activity, and create anoxic conditions at glacier beds in winter when surface water inputs cease.

Paradoxically, however, negative surface mass balances and glacier thinning have resulted in cooling and net freezing at the beds of some Arctic glaciers (Dowdeswell et al., 1995; Hodgkins et al., 1999; Glasser and Hambrey, 2001). Such changes would lead to isolation and cryostasis (the reversible cryopreservation of live organisms) of the subglacial ecosystem. Whether this might ultimately eliminate the subglacial ecosystem from individual glaciers or drive it deeper into unfrozen subglacial sediments is currently unknown. Elimination of the ecosystem would substantially alter the biogeochemistry of glacial runoff, with potential implications for downstream ecosystems (Hodson et al., 2004, 2005).

7.7.5. Geomorphological hazards

Glacier changes, particularly glacier retreat, can produce a range of hazards involving slope instability, increased proglacial fluvial sediment transport, and changes in water flow, particularly outburst floods. To augment the available research conducted in the Arctic, the review below draws heavily on work outside the Arctic. However, the general principles should apply throughout the Arctic region as broadly defined in this chapter.

7.7.5.1. Slope instability

Glacier down-wasting and retreat appear to be partly responsible for destabilizing adjacent terrain, leading to landslides, debris flows, rock avalanches and falls, and some catastrophic rock-slope failures in high mountains (Evans and Clague, 1994; Holm et al., 2004). Many marginally stable slopes that were buttressed by glacier ice during the Little Ice Age failed after they became deglaciated in the 20th century. A factor that has possibly contributed to such failures is steepening of rock slopes by cirque and valley glaciers during the Little Ice Age. These effects are most pronounced in mountain ranges with the largest ice cover; because it is there that ice losses in the 20th century have been greatest. An extreme example is Glacier Bay, which until the end of the 18th century was covered by glacier ice. Since then, Glacier Bay has become deglaciated, with the loss of over 1000 km² of ice in 200 years. The amount of ice lost is so great that the uplift of the land due to isostatic rebound is measurable (Larsen et al., 2005).

7.7.5.2. Outburst floods from moraine- and ice-dammed lakes

Lakes dammed by moraines and glaciers in high mountains have drained suddenly to produce floods orders of magnitude larger than normal nival or rainfall floods (Costa and Schuster, 1988; Clague and Evans, 1994). Lakes dammed by Neoglacial end and lateral moraines are susceptible to failure because they are steep-sided and consist of loose, poorly sorted sediment that in some cases is ice-rich (Clague and Evans, 2000). Irreversible rapid incision of a moraine dam may be caused by a large overflow triggered by an ice or snow avalanche or rockfall. Other failure mechanisms include earthquakes, slow melt of ice within the moraine, and slow removal of fine-grained sediment from the moraine by groundwater flow (piping). As climate warms, lakes impounded by glaciers may drain suddenly and unexpectedly following a long period of stability due to progressive wastage of the glacier dam and the formation of subglacial, supraglacial, or ice-marginal channels. Drainage may be initiated by hydrostatic lifting of that part of the base of the glacier facing the lake. Lakes may also form during glacier surges; they drain, often catastrophically, soon after they form.

Most outburst floods display an exponential increase in discharge, followed by a gradual or abrupt decrease to background levels as the water supply is exhausted (Clarke, 2003; Ng et al., 2007). Peak discharges are controlled by lake volume, dam height and width, the material properties of the dam, failure mechanism, and downstream topography and sediment availability. Floods from glacier-dammed lakes tend to have lower peak discharges than those from moraine-dammed lakes of similar size because enlargement of tunnels within ice is a slower process than overtopping and incision of sediment dams.

Floods resulting from failures of natural dams may transform into debris flows as they travel down steep valleys. Such flows can only form and be sustained on slopes greater than 10 to 15 degrees and only where there is an abundant supply of sediment in the valley below the dam. Entrainment of sediment and woody plant debris by floodwaters may cause peak discharge to increase down valley, which has important implications for hazard appraisal.

Outburst floods from lakes dammed by moraines and glaciers commonly alter river floodplains tens of kilometres from the flood source. The floodwaters erode, transport, and deposit huge amounts of sediment. They commonly broaden floodplains, destroy pre-flood channels, and create a new multi-channel, braided to anastomosing (channels that repeatedly branch and reconnect) plan form. The changes can persist for decades after the flood, although rivers quickly reestablish their pre-flood grades by incising the flood deposits.

Climate is an important determinant of the stability of moraine and glacier dams (O'Connor and Costa, 1993; Evans and Clague, 1994). Most moraine-dammed lakes formed

during the past century as glaciers retreated from bulky end moraines constructed during the Little Ice Age. The lake dams soon began to fail as climate warmed. With continued warming and glacier retreat, the supply of moraine-dammed lakes that are susceptible to failure will be exhausted, and the threat they pose will diminish. Glacier-dammed lakes have typically gone through a period of cyclic or sporadic outburst activity, lasting up to several decades, since climate began to warm in the late 19th century. The outburst floods from any one lake ended when the glacier dam weakened to the point that it could no longer trap water behind it. However, with continued glacier retreat, the locus of outburst activity may, in some cases, shift up-glacier to sites where new lakes develop in areas that are becoming deglaciated. Landslide damming is less clearly linked to climate.

7.7.5.3. **Erosion and sedimentation**

Fluctuations of glacier margins on timescales of decades and centuries can remobilize glacial sediment. During glacier advance, initial incision due to increased competence of meltwater streams is quickly followed by deposition as sediment supply increases (Maizels, 1979). Sediment stored within and beneath glaciers is delivered at an increasing rate to fluvial systems as glaciers advance (Karlén, 1976; Maizels, 1979; Leonard, 1986, 1997; Karlén and Matthews, 1992; Lamoureux, 2000). Similarly, subglacial erosion increases during glacier advance, and meltwater may carry more sediment into river valleys than at times when glaciers are more restricted. Sediment delivery to streams in the Coast Mountains of British Columbia, for example, increased during the Little Ice Age, and the streams responded by filling in their channels and braiding over distances up to tens of kilometres down valley from glaciers (Church, 1983; Gottesfeld and Johnson-Gottesfeld, 1990). Glacier retreat typically exposes large areas of unstable, unvegetated sediment that is easily entrained by meltwater (Church, 1983; Desloges and Church, 1987; Gottesfeld and Johnson-Gottesfeld, 1990; Brooks, 1994; Ashmore and Church 2001; Clague et al., 2003).

7.7.5.4. **Glacier-volcano interactions and related hazards**

Glaciers are found in regions of active volcanism in Alaska, Iceland, and Kamchatka, where high seismic and volcanic activity can generate significant glacier-related hazards. Consequences of volcanic activity include glacier burial by ash falls, melt-induced formation of craters and lakes on glaciers, significant mass loss by subglacial melting during subglacial eruptions, and accelerated glacier flow. Several volcanogenic *jökulhlaups* (outburst floods from glaciers) have been documented in Iceland (Russell et al., 2006).

Lahars (fluidized volcanic debris flows produced when hot pyroclastic material falls onto glaciers) can be extremely destructive owing to their high density and low viscosity while in motion, and their tendency to over-consolidate (stiffen) upon halting. Lahars are especially destructive because their mode of origin allows large volumes of material to be activated quickly, and because the topographic relief of the volcano provides abundant potential energy. In Kamchatka, lahars have caused significant damage to infrastructure and the environment. For example, the 2001 eruption of Shiveluch Volcano produced lahars which

traveled about 20 km and damaged the road to Ust-Kamchatsk. The 2005 eruption of Kluchevskaya Volcano created lahars that carried blocks 3 to 4 m in size a distance of over 30 km.

It is unclear from the available literature how glacier-volcano interactions and the associated hazards will be influenced by future changes in climate and resulting changes in glaciers and ice caps. However, there is some evidence that volcanic activity may increase as ice thins (Tuffen, 2010).

7.7.5.5. **Glacier surges**

Surging glaciers are found in Alaska, the Yukon Territory, Arctic Canada, Greenland, Iceland, Svalbard, and Novaya Zemlya. Those in Novaya Zemlya tend to terminate in water (Grant et al., 2009), and surges of these glaciers are a potential danger to local populations and infrastructure. Some surging glaciers in Alaska, such as the Black Rapids Glacier, are a potential hazard to the Trans-Alaska (oil) Pipeline (Heinrichs et al., 1995). Surge behavior has been documented for 26 glaciers in Iceland (Björnsson et al., 2003). Surges of these glaciers are associated with an increase in proglacial stream turbidity, but not discharge. The geographical variability of surge behavior, coupled with the incomplete understanding of glacier surge mechanisms, makes it difficult to predict whether climate change will alter the frequency of surging or the impact of the phenomenon on human society.

7.7.6. **Glacier-related tourism**

More than 1.5 million tourists currently visit the Arctic each year, up from one million in the early 1990s. Longer and warmer summers keep Arctic seas freer of ice floes, so cruise ships can visit places that were once inaccessible. The annual number of visitors to Svalbard has surged by 33% in the period 2002 to 2006 to about 80 000 (Naik, 2007). At the beginning of the 21st century, tourism has clearly become a major engine of growth for the Alaskan economy. According to the Alaska Tourism Satellite Account (Anon, 2004), travel and tourism's economic contribution in Alaska reached USD 1.6 billion in 2002. This amount – representing sales net of related imports into the state – contributed 5.6% to Alaska Gross State Product (GSP). This includes the direct and indirect effects of all travel and tourism expenditure, but not induced (multiplier) effects. Travel and tourism's core industry – that is, only the direct impact of end-providers of goods and services to travelers – generated USD 856 million in local value added in 2002: 3.0% of Alaska's GSP. Using the core industry definition, travel and tourism was the third largest private sector employer – fourth overall – in the State, with 26 158 direct full-time-equivalent jobs (9.1% of Alaska's total employment) in 2002. Travel and tourism-generated jobs provided USD 579 million in core labor income (benefits and salaries) to Alaska. Adding public sector employment, travel and tourism's economic contribution to employment reached 39 420 full-time-equivalent jobs. Those jobs provided Alaskan workers with USD 1.15 billion in income. Glacier-related tourism is an important part of the economy in several regions of the Arctic, especially in Alaska. Every year hundreds of thousands of people visit coastal Alaska to see the glaciers. Melting of the glaciers in the warming Arctic may be a threat to this sector of the economy.

7.8. New information expected from International Polar Year projects

- Eleven projects focusing on the behavior of Arctic mountain glaciers and ice caps were initiated during the International Polar Year (IPY). They contributed to several IPY projects including GLACIODYN and 'State and Fate of the Cryosphere'.

- New regional surveys of glacier changes, which contribute to the IPY program 'State and Fate of the Arctic Cryosphere', have been conducted in Alaska, the Yukon, Labrador, Iceland, the Russian Arctic Islands, and northeastern Siberia. Surveys in Alaska and Iceland involve new mapping of glacier surface topography with LIDAR and laser altimetry. Satellite laser altimeter data from ICESat have also been used extensively in these surveys. The Yukon surveys reveal a 22% reduction in glacier area since the 1957/1958 International Geophysical Year.

- GLACIODYN projects focus on understanding and modeling the role of changing ice dynamics in the response to climate change of polythermal and calving tidewater glaciers. Target glaciers for this work are Columbia and McCall Glaciers in Alaska, Belcher Glacier on Devon Island, Canada, and Austfonna and Vestfonna in Nordaustlandet, Svalbard.

- A Russian IPY project has focused on the formation, dynamics, and decay of icebergs in the western sector of the Russian Arctic, a topic of particular interest given the increasing amount of shipping and offshore resource exploration activity in the region.

7.8.1. Alaska

Three U.S. IPY projects will contribute to knowledge of Arctic glacier changes and mechanisms of change. As part of IPY-37 GLACIODYN, there is a continuation of the 30-year record of observations on the retreat of Columbia Glacier, Alaska, which at present is the largest single Alaskan contributor to sea level change (at 3.3 Gt/y from 1982 to 2001, and 6.6 Gt/y around 2001). The ongoing work at Columbia Glacier includes re-analysis and archival of the 145-mission, 30-year sequence of aerial photography of the glacier (the longest and most detailed record of its kind ever acquired), a continuation of this aerial photogrammetric record, and a variety of additional measurements and analyses including photogrammetric feature-tracking, terrestrial time-lapse photogrammetry, airborne radar, GPS surveying, seismic observations of calving, and meteorological monitoring. Collaborative work on Svalbard with Polish colleagues is included in this project. The results will add to knowledge of the physics of calving and tidewater glaciers, and improve the ability to predict future marine-ending glacier behavior, not only on Arctic glaciers and ice caps, but on marine-ending glaciers everywhere, including the outlet glaciers of Greenland and Antarctica. Work under this funding is in progress and research supported in part by this funding is presented by O'Neel and Pfeffer (2007).

A second GLACIODYN project aims to investigate the mass balance and ice dynamics of polythermal land-terminating glaciers using McCall Glacier in Arctic Alaska as the primary field site. During the IPY, ice cores to bedrock were extracted at three locations. One complete core and selected parts of the other two cores have been returned for analyses designed to elucidate past climate conditions in the region and establish a history of pollutant deposition. Thermistor strings were installed to bedrock in the drill holes as part of a study of internal accumulation within the glacier and its effects on the glacier's mass balance and thermal regime. A program of stream hydrology measurements was initiated both at the glacier terminus and in the river that drains from the region to the Arctic Ocean (these rivers have never previously been gauged and glaciers currently contribute roughly 90% of their discharge). A full 3D higher-order ice flow model of the glacier is being run using high-resolution bedrock and surface maps. To help understand how representative the McCall Glacier measurements are of a broader region, a new LIDAR digital elevation model covering about 500 glaciers in the eastern Brooks Range was created based on surveys conducted in 2008. It will be used to measure changes in glacier volume in the region.

As part of IPY-105 'State and Fate of the Cryosphere', glacier mass loss across Alaska is assessed using airborne laser altimetry along glacier centerline profiles. Repeat profiles are compared to previous mass change estimates to address the primary question of continued acceleration of glacial wastage.

7.8.2. Arctic Canada

Two Canadian IPY projects will contribute new knowledge about the glaciers in Arctic Canada. One component of the project 'Variability and Change in the Canadian Cryosphere' (a contribution to IPY-105 'State and Fate of the Cryosphere') involves producing a multi-temporal glacier inventory and mapping changes in the extent of glaciers in the Yukon and Labrador since the International Geophysical Year (1958/1959) from aerial photography and satellite imagery. Initial results suggest a reduction of 22% in Yukon glacier area over that period (Barrand and Sharp, 2010) and a 25% reduction in the extent of Labrador glaciers between 2003 and 2007. A second goal of the project is to undertake annual mapping of melt extent, melt duration, melt onset and freeze-up dates, and the end-of-summer distribution of snow and ice facies on all large glaciers and ice caps in the Arctic (including Greenland) using enhanced resolution data from the SeaWinds scatterometer on QuikSCAT. The records extend over ten years, to the end of the 2009 melt season, and will provide new insight into the evolution of summer climate over the ice-covered areas of the Arctic land mass.

A second project is a Canadian contribution to IPY-37 GLACIODYN, which aims to investigate the role of ice dynamics in the response of Arctic glaciers and ice caps to global warming, and improve the ability to predict future changes and their impact on global sea level and fluxes of freshwater to the ocean. The Canadian contribution to GLACIODYN focuses on the Belcher Glacier in the northeastern sector of the Devon Island Ice Cap, Nunavut. This is the fastest flowing outlet from the ice cap and a major source of icebergs. The project involves an intensive field and remote sensing study of the hydrology and dynamics of the glacier linked to the development and

validation of a state-of-the-art, high-order coupled model of ice flow dynamics and glacier hydrology. The model will be used to test hypotheses about the effects of climate warming on meltwater inputs to Arctic outlet glaciers and their impact on ice flow, and to perform simulations of the response of the Belcher Glacier to recent and projected future climate change.

7.8.3. Iceland

A major Icelandic IPY project involves mapping the surface topography of Icelandic ice caps using LIDAR. The results will be used to correct digital photogrammetric maps from the 1990s, allowing an analysis of surface elevation changes of the ice caps over the past one to two decades. The LIDAR mapping is a collaboration between the Icelandic Meteorological Office and the Institute of Earth Sciences of the University of Iceland and is funded by RANNIS (the Icelandic Centre for Research), Landsvirkjun (the National Power Company of Iceland), the Icelandic Public Road Administration, the Reykjavík Energy Environmental and Energy Research Fund, and the National Land Survey of Iceland.

High-resolution LIDAR digital terrain models (DTMs) of most of Hofsjökull, Langjökull, Eiríksjökull, and Snæfellsjökull have been made from surveys conducted during the 2007/2008 IPY. DTMs of Mýrdalsjökull, Eyjafjallajökull, Hofsjökull, and southeastern Vatnajökull were completed in 2010. Mapping will be continued in 2011 to 2013 to produce DTMs of the southern, western, and northern drainage basins of the Vatnajökull ice cap.

The new LIDAR DTMs of proglacial areas and nunataks (isolated peaks protruding through the surface of the ice caps) will be used as a reference for re-analysis of the digital photogrammetric maps of glacier ablation areas in the 1990s, which should reduce systematic bias in these maps. This will make it possible to compare the photogrammetric maps with the LIDAR DTMs and quantify volume changes of the Icelandic ice caps during the past 10 to 20 years.

It is important that the glaciers are accurately mapped now when rapid changes are occurring in response to a warming climate. Accurate DTMs make it possible to pursue various lines of glaciological research, but they also have considerable economic value for the assessment and monitoring of water resources for hydroelectric power generation. Potential uses of the DTMs include modeling of surface mass balance and ice flow, mapping the boundaries of surface and subglacial watersheds, mapping subglacial water flow pathways, and quantifying rates of glacier change and their impact on global sea level. The DTMs are also important for investigations of glacier surges and isostatic uplift due to reductions in ice mass.

7.8.4. Svalbard

During the IPY, two international project consortia worked on glaciology in Svalbard: the IPY Kinnvika project and GLACIODYN. Both studied mass balance and dynamics with a focus on large ice caps. The glaciological part of the Kinnvika project focused on Vestfonna (2500 km²) and involved studies of accumulation rates, ablation rates, ice temperatures, and ice geometry and measured ice velocity fields. Fixed mass balance stations and automatic weather stations were established across the ice cap. Shallow cores were taken to find the average point

mass balance over a few decades. The data provide input to and validation of modeling and remote sensing studies.

The Norwegian part of GLACIODYN focused on Austfonna in Nordaustlandet and on Kongsvegen / Kronebreen in northwestern Spitsbergen. GLACIODYN monitored the current status of more than a dozen Arctic ice caps and glaciers with the aim of predicting the future fate of these terrestrial Arctic ice systems. The Austfonna investigations focused on (i) direct surface mass balance measurements; (ii) elevation and volume changes using satellite data, airborne laser profiles, and ground-based GPS measurements; (iii) dynamics of surge, calving and subglacial processes; and (iv) modeling of mass balance and dynamics. Five years of net surface mass balance studies on Austfonna show a slightly negative average balance (-0.1 m w.e./y), but with large interannual variations both in snow accumulation and total ablation. The calving is important (-0.30 m w.e./y) and represents 30% to 40% of the total ablation, similar to values observed on other Arctic ice caps. However, the elevation change measurements on Austfonna show a thickening in the interior of about 0.5 m/y and an increasing thinning closer to the coast of 1 to 2 m/y. A total of 230 km, or about 70%, of the ice cap margin is a calving front, at which all the ice is grounded. The calving front has been retreating at an average rate of 50 m/y since 1990 (Dowdeswell et al., 2008). There seems to be a large dynamic instability, as the measured mass flux is much less than the flux that would be expected given the observed surface mass balance.

On Spitsbergen, a Russian IPY project has used surface radio echo sounding surveys and the analysis of available multi-temporal cartographic, aerial, and satellite-derived data to determine changes in the surface elevation, thickness, ice volume, and internal structure of Fridtjovbreen resulting from a surge event in the 1990s.

7.8.5. Russian Arctic and mountains

The Russian National Program and Implementation Plan for the IPY 2007/2008 includes three projects relevant to glaciers in the Russian Arctic. The project 'Current state of glaciers and ice caps in the Eurasian Arctic' is a Russian contribution to GLACIODYN. The main goal of the project is to study the area changes, mass balance, hydrothermal state, and potential instability of glaciers and ice caps in the Russian Arctic islands and Svalbard. The main fieldwork programs in 2007 and 2008 included airborne and surface radio echo sounding surveys of ice thickness and bedrock and surface topography of ice caps and glaciers, which have been supported by the analysis of spaceborne remote sensing data. On the basis of radio echo sounding surveys of glaciers on Franz Josef Land and Novaya Zemlya and satellite altimetry data, characteristic heights and thicknesses of glacier fronts producing icebergs have been determined, including Glacier No. 1 and the Moscow Ice Cap on Hall Island, the northern part of the glacier complex on George Land (Franz Josef Land), and the glaciers in the Inostrantsev Bay area, Novaya Zemlya. New criteria for the estimation of iceberg hazards from the glaciers of Novaya Zemlya and Franz Josef Land were developed. Franz Josef Land has the greatest potential for regular formation of large icebergs (with thicknesses of up to 150 to 200 m and extents of more than 1 to 2 km) (Kubyshkin et al., 2009).

A second project, 'Formation, dynamics and decay of icebergs in the western sector of the Russian Arctic', aimed to collect new data on the formation, distribution, and properties of icebergs in the Barents and Kara Seas and to estimate the current state of outlet glacier fronts in the Russian Arctic archipelagos. During a cruise of R/V *Mikhail Somov* in September 2007, iceberg-producing glaciers on Franz Josef Land, Novaya Zemlya, and some other islands were surveyed. The survey included a wide range of iceberg studies, following their life history from calving to melting and disintegration. The study continued the survey of 2003, when an abnormally large group of icebergs (over 40, the largest of which was over 400 m long and weighed about 3 million tonnes) was discovered in the northeastern part of the Barents Sea, in the vicinity of the *Shtokman* gas-condensate field. The unusual ice conditions allowed R/V *Mikhail Somov* to circumnavigate Franz Josef Land. Helicopter radio echo sounding and aerial photography of glaciers on Prince George Land, Salisbury Island, Luigi and Champ Islands, Hall Island, and Wilczek Land allowed determination of ice thickness in potential iceberg-producing areas. Photogrammetry was used to reconstruct the geometry of glacier fronts and the above-water parts of icebergs. On some glaciers and icebergs, the vertical distribution of ice temperature (down to 20 m depth) and energy balance components of the upper 3 m layer were measured. Observations were made on the surface and in the ice (down to 3 m depth). On some glaciers, markers were installed for measurement of glacier flow. Glaciological studies were supplemented with oceanographic temperature and salinity profiling in the Franz Josef Land straits. Several groups of large tabular icebergs (more than 1 million tonnes) were found not far from their calving areas (Elena Guld Bay on Wilczek Land; the straits between Salisbury, Luigi, and Champ Islands; Geographers' Bay on Prince George Land). The majority of large icebergs were already drifting. Under favorable meteorological conditions, some may drift to the Barents Sea through the deep straits. A similar survey was undertaken of the glaciers of the northern end of the Northern Island of Novaya Zemlya (Buzin et al., 2008). This survey was repeated in autumn 2008.

The project 'Climatic factors in the contemporary evolution of North-Eastern Siberia glaciation' is another Russian contribution to GLACIODYN. This project continued studies of climate–glacier interactions in the poorly explored northeastern Siberia region. This region is of special interest because its climate is influenced both by Atlantic and Pacific air masses. It includes a part of the Northern Hemisphere where climate change (weakening of the Siberian High and warming in recent years) and cryospheric change have been detected. Meteorological parameters were measured along a transect from Magadan to Oymyakon, and in the northern massif of Suntar-Khayata. A study has been completed of glacier change in the region based on modern satellite images and data from the USSR Glacier Inventory. Infrared, visual, and aerial photo surveys were conducted for the Suntar-Khayata glaciers in order to update the Glacier Inventory (Ananichava and Kapustin, 2010).

7.9. Synthesis of the current state of Arctic mountain glaciers and ice caps

- Mass balance assessments based on field measurements and geodetic surveys indicate a marked increase in the rate of mass loss from Arctic glaciers (especially those in Alaska, the Canadian Arctic, and Greenland) since the mid-1990s. From 1961 to 2004, the mean rate of loss was in the range -91 to -141 Gt/y, but pentadal estimates for the period since 1995 range from -170 to -280 Gt/y.

- The recent increase in mass loss coincides with increases in summer air temperature and reductions in glacier area.

- Recent rates of mass loss are comparable to those from the Greenland Ice Sheet, emphasizing the importance of the Arctic as a whole as a first-order control on the rate of global sea level rise.

- Iceberg calving is an important component of mass loss in Alaska, Arctic Canada, the Russian Arctic, and Svalbard (where it accounts for about 32% of the net mass loss). Individual tidewater glaciers can discharge very large amounts of ice into the ocean (e.g., Columbia Glacier discharged 6.6 Gt/y in 2001). However, temporal changes in this mass loss term are not well known and are not included in projections of future mass and volume change.

- Model-based projections suggest a 13% to 36% reduction in Arctic glacier volume by 2100 due to changes in surface mass balance alone. While some individual glaciers will disappear during this period (and others have already done so), mountain glaciers and ice caps in the Arctic will continue to influence change in global sea level well beyond 2100.

- Meltwater runoff from Arctic glaciers is expected to increase for a few decades at least, but will eventually decrease as the influence on melt production of decreasing glacier area becomes more significant than that of increasing melt rate.

7.9.1. Mountain glacier and ice cap mass balance

Two global mass balance datasets, one based on glaciological mass balance measurements (DM05) (Dyurgerov and Meier, 2005) and the other a combination of both glaciological and geodetic measurements (C09) (Cogley, 2009a), are sub-setted to determine a pan-Arctic glacier mass balance. DM05 is a weighted arithmetic mean of single-glacier time series within a set of climatically homogeneous regions, based on data from the World Glacier Monitoring Service (WGMS, which collects standardized observations of change in the mass, volume, area, and length of glaciers worldwide; www. wgms.ch), individual published sources and, in some cases, unpublished measurements. Each time series is weighted by the area of its glacier, and then extrapolated to the climatic region based on estimates of the total glacierized area in each region. Nearly all the glaciological measurements in DM05 are derived from land-terminating glaciers (with larger glaciers underrepresented). C09 uses a nearly identical glaciological dataset to DM05, but also includes geodetic datasets that sample several tidewater glaciers. For the annual glaciological dataset,

C09 uses a spatial interpolation algorithm that fits a polynomial in geographical coordinates (centered successively on each cell of a 1° by 1° grid) to the complete (glaciological plus geodetic) set of single-glacier balances. The measured mass balances are inverse-distance weighted to the center of the grid cell, and the resulting estimate is multiplied by the total glacierized area within that cell. Estimates are averaged over five-year (pentadal) time periods. For the geodetic measurements, which usually span several years, the time-averaged mass balance during the entire period is assumed to represent the annual balances within that period. Interannual variations in the geodetic balances are accounted for by calculating a standard deviation that is a function of the warmest month's air temperature in a given year, interpolated to the glacier equilibrium line. The temperatures used are climatological means from the NCEP/NCAR Reanalysis.

While all the mass balance measurements included in the DM05 and C09 datasets are total mass balances (surface mass balance plus iceberg calving), very few are from calving glaciers. Circumstantial evidence presented by Cogley (2009a) suggests that calving glaciers may currently have more negative mass balances than land-terminating glaciers. If so, underrepresentation of calving glaciers in these datasets may result in underestimation of rates of mass loss in both global mass balance assessments. As yet, it is not understood why rates of mass loss from calving glaciers might be higher than those from land-terminating glaciers and it is not possible to properly quantify the magnitude of the bias in global mass balance assessments that arises from underrepresentation of calving glaciers in the datasets on which these assessments are based. Equally, it is not yet possible to predict how the rate of mass loss by iceberg calving will evolve under different climate change scenarios.

Both estimates derived from measurements (DM05 and C09) show that mass balances were predominantly negative during the period of record, with large interannual variability (Figure 7.29). During the period 1961 to 2004, the average mass balance of Arctic glaciers was -91 Gt/y using the DM05 data and -138.4 Gt/y using the C09 data. Estimates from DM05, based only on glaciological measurements, are more positive than the C09 estimates, possibly because they underrepresent calving glaciers. A third global mass balance estimate, derived from a surface mass balance model driven by climate station data (Hock et al., 2009), suggests that the surface mass balance of Arctic glaciers was -141 ± 65 Gt/y during the period 1961 to 2004 (Figure 7.29). Mass balances were significantly more negative in the period 1995 to 2005 than previously (of the order of -170 to -280 Gt/y according to the C09 dataset). These recent rates of mass loss are comparable in magnitude to recent estimates of mass loss from the Greenland Ice Sheet (see Chapter 8). When the mass loss rates from Arctic mountain glaciers and ice caps are combined with those from the Greenland Ice Sheet, it becomes clear that wastage of glacier ice in the Arctic is a first-order control on the rate of global sea level rise.

The regional contributions that comprise the pan-Arctic mass balance estimate derived using the combined geodetic-glaciological method (C09) are shown in Figure 7.30. This figure also shows the pentadal mean summer (June to August) 700 hPa air temperature for each region from the NCEP/NCAR Reanalysis. The time series of estimated contributions from regions with observed records (such as southern Alaska, the
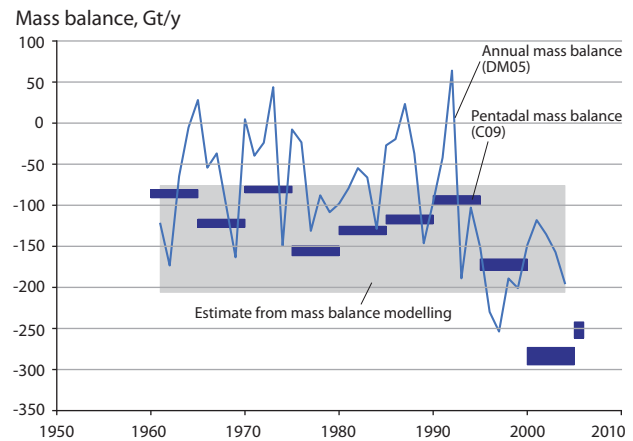

Mass balance, Gt/y

Figure 7.29. Annual mass balance (Dyurgerov and Meier, 2005) and pentadal mass balance (from data compiled by Cogley, 2009a) of Arctic glaciers as outlined in Figure 7.9. Annual estimates are extrapolated from glaciological measurements. Pentadal estimates are derived from a combination of geodetic and glaciological data. The grey box is an estimate from mass balance modeling (Hock et al., 2009) marking the error bounds around the mean estimate. Source: AMAP.

Canadian Arctic, Scandinavia, and Iceland) are inversely related to regional summer air temperature variations. However, the time series for some regions with fewer observations (such as the Russian Arctic) are not consistent with local climate fluctuations. In these regions, the reconstructed mass balance records may be overly influenced by regional interpolation from areas with more mass balance data and different temperature histories. In general, the degree of confidence in the regional mass balance reconstructions decreases as the amount of smoothing of observed data required to generate them increases. This suggests that records from such regions need to be treated with caution and that deriving more reliable records for such regions should be a high priority.

Estimates of regional mass balance based on local measurements also illustrate the relative contribution of each region to rising sea level (Figure 7.31) over specific intervals of time. All areas with regional syntheses show mass losses during their periods of measurement, with Alaska and the Canadian Arctic being the dominant contributors to rising sea level. It is important to note that these estimates for different regions are not directly comparable owing to the differences in measurement periods.

7.9.2. Ablation due to iceberg calving

The relative proportions of ablation due to calving and surface mass balance are not well known, but available data suggest that calving plays an important role in the mass balance of glaciers in Alaska, the Canadian Arctic, Svalbard, and the Russian Arctic. There are 54 tidewater glaciers in northwestern North America, some of which are continuing their rapid retreat that commenced about 200 years ago. Among these, Columbia Glacier is undergoing the most rapid retreat exceeding 1 km/y since 2000, discharging an average of 3.3 Gt/y of calved ice during the period 1982 to 2001, with 6.6 Gt/y calved ice in 2001 (O'Neel et al., 2005). In Svalbard, calving losses accounted for about 32% of the overall glacier ablation (Błaszczyk et al., 2009). In Arctic Canada, Svalbard, and Severnaya Zemlya, calving comprises a significant fraction of the total ablation from several large ice caps including Devon Ice Cap and Prince of Wales Icefield, Austfonna, and Academy of Sciences Ice Cap
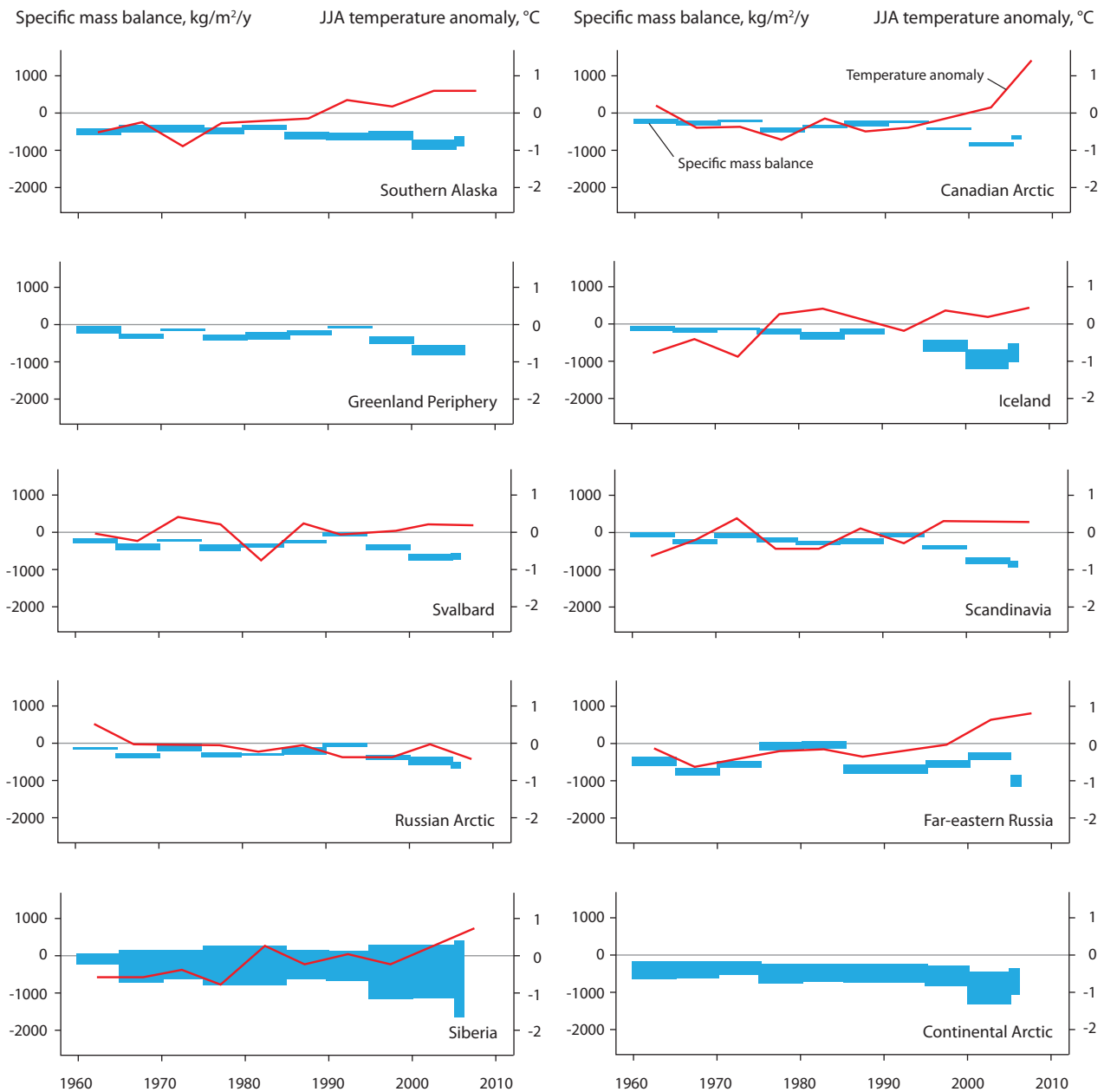
Figure 7.30. Estimates of pentadal mean annual specific mass balance for glacierized regions of the Arctic derived using the combined geodetic-glaciological dataset of Cogley (2009a) and the pentadal mean of annual anomalies in summer (June – August) air temperature at a geopotential height of 700 hPa over each region from the NCEP/NCAR Reanalysis. Temperature data are not plotted for the Greenland Periphery and Continental Arctic regions due to the non-contiguity of these regions. Source: Dan Moore, University of British Columbia.

(Dowdeswell et al., 2002, 2008; Burgess et al., 2005; Williamson et al., 2008). However, ongoing monitoring of calving fluxes is almost non-existent and, as a result, the magnitude of temporal variability and the existence and sign of any trend in calving fluxes are essentially unknown.

7.9.3. **Observed trends in ice extent**

Observed changes in ice extent over the past half-century are consistent with changes in mass. In nearly all regions, mass losses occurred coincidentally with decreases in glacier area. Although a small number of glaciers have advanced during this time, nearly all of these are tidewater or surge-type glaciers. The area evolution of tidewater and surge-type glaciers depends more on dynamic cycles than on direct climatic forcing, although it may be influenced by regional climate patterns.

7.9.4. **Relationship to climate and circulation changes**

Arctic air temperatures have increased during the past half-century (Hinzman et al., 2005), which probably explains much of the increase in rates of mass loss observed on Arctic glaciers. Increases in summer air temperature explain the accelerated losses from glaciers in northwestern North America (Arendt et al., 2009) and Arctic Canada (Gardner and Sharp, 2007), but uncertainties remain large due to the difficulties in modeling dynamic losses owing to calving.

In general, changes in synoptic-scale atmospheric circulation patterns, resulting from variations in sea-surface temperature, polar jet stream position, and other factors, correlate well with Arctic glacier mass balance time series. In northwestern North

Figure 7.31. Estimates of the mass balance of Arctic glacier regions for which regional assessments have been conducted. The vertical extent of each box describes the range of uncertainty in the mass balance estimates (uncertainty estimates are not available for Iceland or the Canadian or Russian Arctic). Glaciological estimates relate only to the surface mass balance. Source: Anthony Arendt, University of Alaska, and W. Tad Pfeffer, INSTAAR, University of Colorado.

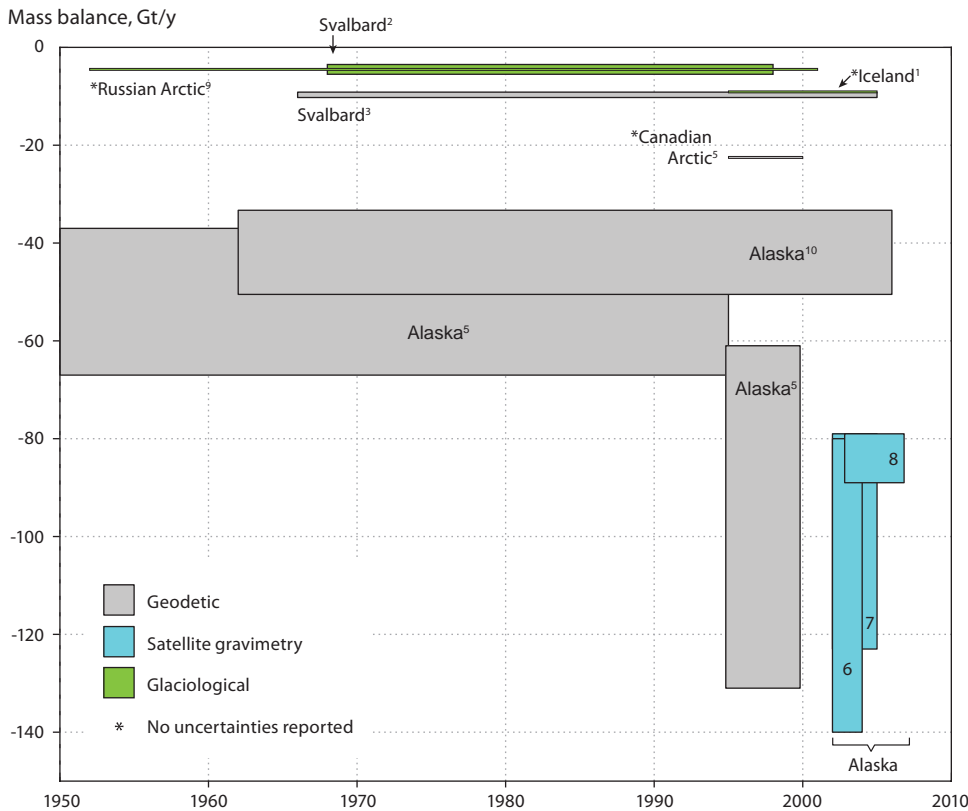[1]Björnsson and Palsson (2008): data from two ice caps comprising 81% of the ice-covered area of Iceland; [2]Hagen et al. (2003a): surface mass balance component, Hagen et al. (2003b) calving component; [3]Nuth et al. (2010): date ranges are 1965/90 to 2003/07; [4]Abdalati et al. (2004); [5]Arendt et al. (2002); [6]Tamisiea et al. (2005); [7]Chen et al. (2006); [8]Luthcke et al. (2008); [9]Glazovsky and Macheret (2006); [10]Berthier et al. (2010).

America, two well-documented 'regime shifts' occurred in 1977 and 1989, during which temperatures abruptly shifted to more positive values and storm tracks changed direction. These shifts are evident in the mass balance time series of glaciers in Alaska (Bitz and Battisti, 1999; Josberger et al., 2007). In the Canadian Arctic, accelerated surface mass loss after 1987 was associated with changes in the position of the July circumpolar vortex. In the Russian Arctic, mass balance changes have been linked to sea-surface temperature variations in the Barents Sea.

7.9.5. Projections of future change

Modeling studies indicate that many individual glaciers may disappear by the end of the 21st century if air temperatures in the Arctic continue to rise as projected in the scenarios used by the suite of climate models employed in the Fourth Assessment of the IPCC (Solomon et al., 2007). The total volume of Arctic glaciers is projected to decline substantially by 2100. Projected reductions are of the order of 13% to 36% of current glacier volume. However, these projections vary greatly from model to model, are sensitive to the choice of climate scenario, and generally do not include mass losses by iceberg calving, which is known to be an important mode of ablation. Nonetheless, it is clear that glaciers in the Arctic are likely to continue to influence changes in global sea level well beyond the end of the 21st century.

7.9.6. Impacts

In many parts of the Arctic, climate warming should cause glacier runoff to increase for a few decades or longer, but glacier retreat will ultimately cause glacier runoff to decline. Runoff decline may already have started in some regions where most glaciers are small.

Changes in glacier runoff will influence freshwater habitat, water supply, hydroelectric power generation, flood and avalanche hazard, estuarine and coastal habitat, and patterns of ocean circulation. Glacier retreat may result in the destabilization and failure of valley-side slopes. Outburst floods from moraine-dammed lakes are likely to decrease in frequency as glaciers retreat further from moraines formed during the Little Ice Age. The frequency and magnitude of outbursts from existing glacier-dammed lakes may also change as ice dams are thinned by ice melting. However, new lakes may form further up-glacier as glacier margins recede down valley-side slopes and melt rates increase.

7.10. Knowledge gaps and recommendations

- The major gaps in knowledge that limit the ability to quantify current rates of glacier wastage in the Arctic and predict future trends are (i) the lack of a complete glacier inventory for the Arctic (including information about glacier location, area, surface topography, and ice thickness); (ii) the limited number of *in situ* mass balance measurements, the strong spatial bias in the distribution of available measurements, and the complete lack of measurements from areas such as the Russian Arctic; and (iii) the lack of ongoing monitoring programs for the iceberg calving component of mass balance, and limited knowledge of the major controls on calving fluxes.

- A coordinated effort to fill these knowledge gaps is needed in order to predict more reliably the current and future contribution of Arctic glaciers to global sea level change and better assess how glacier change will affect regional water resources.

7.10.1. **Key gaps in knowledge**

Despite progress since the Arctic Climate Impact Assessment (ACIA, 2005), major gaps remain in knowledge about Arctic glaciers and ice caps. The keys gaps are as follows:

1. There is a severe lack of basic information about glaciers in the Arctic. Knowledge gaps include the geographical distribution, total area and size distribution of glaciers in some regions, the distribution of area with elevation (hypsometry), and glacier thicknesses and, therefore, volume (Figure 7.32).

2. Glaciological mass balance measurements are limited to a small number of Arctic glaciers (50 in the 2000 to 2004 pentad, of which 30 were in Scandinavia), with larger glaciers and calving glaciers underrepresented. There have been no mass balance measurements conducted in Arctic Russia since 1990, and all but one measurement series from that region ended before 1980. In Arctic Canada, the number of measurement sites peaked at 13 in the early 1970s and is now only five. Given that the uncertainties associated with estimates of regional mass balance increase substantially as the number of local measurements decreases, there is a pressing need to continue (and expand) *in situ* monitoring programs and to develop alternative well-validated methods for assessing regional-scale glacier mass changes, including remote sensing observations (although a problem with these is the lack of temporal and methodological continuity between sensors and satellite missions) and numerical modeling.

3. Observations, process understanding, and modeling capabilities related to ablation by calving are severely limited, introducing large uncertainties into current and future mass balance estimates.

7.10.2. **Basic observations**

Basic information about the geometry and distribution of glaciers is required to assess and model current and future mass changes and contributions of Arctic glaciers to changes in sea level, but this information is currently incomplete for the Arctic. As it is unlikely that direct volume measurements on regional scales will become available in the near future, volume estimates will continue to rely on volume-area scaling techniques. Nonetheless, more direct volume measurements, for example by airborne radio echo sounding, are required to validate scaling techniques.

Completion of high-quality Arctic-wide inventories of glacier areas and size distributions from recent optical satellite imagery is also essential to improve Arctic glacier mass balance assessments. It is important to have regional inventories based on glacier regions; regions should not be bisected by political borders.

There is need for an accurate pan-Arctic inventory of surge-type glaciers and a better understanding of surge recurrence intervals, to provide better estimates of the drawdown of ice by these unusual glaciers. The fraction of mountain glacier and ice cap area drained by marine outlets, and the size and discharge of the marine outlets (including the temporal variability in discharge), must be known if present-day estimates and future forecasts of mass loss by iceberg calving are to be improved.

7.10.3. **Mass balance measurements, proxies, and modeling**

Most of the uncertainties associated with mass balance estimates arise from the limited number of *in situ* observations. Because the number and distribution of *in situ* observations are unlikely to increase in the short term, there is a need to use remote sensing data to monitor ongoing changes in glacier geometry, surface mass balance, dynamics, and iceberg production. In particular, the potential for estimating mass change of Arctic ice caps using satellite gravimetry (GRACE) and repeat-track altimetry data from the Geoscience Laser Altimeter System on ICESat and its successor needs to be explored more fully. The successful launch in April 2010 of the European Space Agency's CryoSat-2 satellite will create opportunities for monitoring changes in the thickness of the larger ice masses in the Arctic.

Most indices of glacier change at regional scales can only be derived from remote sensing sources, so it is important to continue to develop and maintain long and continuous time series of remote sensing data. The development of regional-scale indicators of mass balance is currently limited by the low spatial resolution of much readily available remote sensing data. This seriously impedes the ability to map the properties of smaller ice bodies containing a significant fraction of the total glacier mass, resolve local features, and properly interpret and validate interpretations using sub-pixel *in situ* data. This is particularly the case with mapping melt duration and identifying the limits of snow and ice facies (especially the superimposed ice zone) from active microwave data that are freely available at the pan-Arctic scale. The accuracy of the results of facies mapping affects the accuracy of secondary indices (e.g., equilibrium-line altitude and accumulation area ratio) derived from facies maps. The ability to monitor regional-scale mass balance proxies at high spatial and temporal resolution is dependent on the
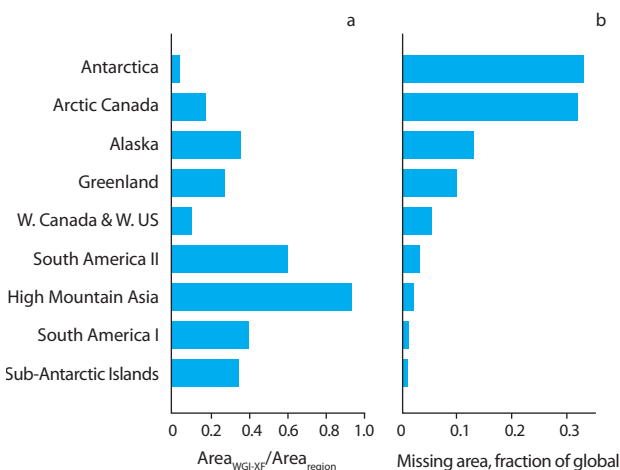
Figure 7.32. (a) Ratio between the area of Extended World Glacier Inventory (WGI-XF) glaciers (Cogley, 2009b), A_{WGI-XF}, and regional glacier area, A_{region}, for the nine glacierized regions that have incomplete glacier inventories. (b) Missing area in WGI-XF as a fraction of missing global glacierized area. While 'a' shows the degree of completeness of each regional inventory, 'b' indicates how much the different regions contribute to the global missing area. South America I refers to Bolivia, Ecuador, and Peru, while South America II refers to Argentina and Chile. Source: after Radić and Hock (2010).

availability of microwave scatterometer data and enhanced resolution products (e.g., QuikSCAT SIR). It is crucial for the construction of long-term records that these data sources be replaced following the recent demise of QuikSCAT.

Uncertainties in nearly all mass balance methods are poorly quantified and rarely reported, although the latter problem is abating steadily. Often it is not clear whether reported mass balances include internal accumulation or calving, both of which can be major components of the mass balance; this makes direct comparison of different measurements problematic. The effects of internal accumulation and superimposed ice formation on glaciological mass balance measurements need to be quantified and these processes need to be included in mass balance modeling. Rigorous identification of sources and magnitudes of error is necessary through intercomparison of methods. Methods for combining mass balance datasets derived from different sampling techniques and representing different time periods and regions need to be expanded and improved to deliver firmer estimates of mass balance and contributions to sea level rise for all Arctic ice masses. The number of glaciologists with expertise in mass balance compilation and integration is also declining and a new generation of glaciologists with these skills must be trained soon.

Given the limited number of *in situ* observations, modeling will continue to play an important and growing role in the assessment of ongoing change in surface mass balance as well as in the projection of future changes. Regional-scale modeling remains a challenge because of the difficulties involved in downscaling climate re-analysis data and climate model results to the complex topography and the heterogeneous climatic conditions of mountain glacier and ice cap environments. These problems are exacerbated by the limited availability of ground measurements made on glaciers to validate the results of downscaling. Projections of future mass balance are also challenging owing to the magnitude of the uncertainty in climate projections for the Arctic that results both from the choice of forcing scenarios used to drive the climate models and large inter-model differences in climate response to the same forcing scenario.

7.10.4. Ice dynamics and ablation by iceberg calving

Although, with the possible exception of internal accumulation (but see Reeh et al., 2005 and Reeh, 2008 for recent progress), processes related to surface mass balance are relatively well understood, there is a severe gap in understanding and modeling of mass loss by calving. Glaciers that terminate in water can have dynamic instabilities that result in rapid mass losses. However, for glacier calving and glacier-marine interactions, there is a lack of both the physical understanding of critical processes and the observations necessary to validate the mathematical representation of these processes. There is also a lack of direct measurements of calving flux and its variability, and of the roles that glacier-marine interactions, bathymetry, ocean temperature, and terminus geometry and dynamics play in determining rates and patterns of glacier retreat and advance. Another area where understanding is limited involves the relationships between summer melt, the development of meltwater drainage systems on, in, and under glaciers, the delivery of water to glacier beds, and the subsequent effects on glacier sliding.

Such information would provide constraints on estimates of total dynamic mass losses and would improve modeling results. There appear to be fundamental differences between calving from tidewater glaciers that terminate in the ocean and calving from those that terminate in freshwater, but these remain poorly understood. Ablation by calving has generally been neglected in predictions of future mass balances. Hence, there is an urgent need to incorporate suitable parameterizations in Arctic-wide mass balance modeling. Since approaches developed for modeling individual glaciers are not yet suitable for application at regional scales due to the paucity of constraining observational data and the largely unresolved nature and complexity of the processes involved, there is a need to explore simpler empirical parameterizations. Given that iceberg keel plowing of sea-floor sediments is a major hazard to underwater engineering structures, there is very little knowledge of the size-frequency distribution of iceberg keel depths or the sources of large icebergs.

7.10.5. Impacts

There is a general understanding that climate warming should cause an initial increase in glacier runoff, followed by a decreasing trend. However, the magnitude of peak runoff and the timing of the shift from increasing to decreasing runoff are not known for most watersheds. Indeed, it is possible that some glaciers are currently in the phase of declining runoff, although the total stored ice volume is large enough that an Arctic-wide decline in loss will not occur in the near future. This knowledge gap is particularly critical because changes in glacier runoff control many other impacts, including those on water quality, stream channel dynamics, and circulation in estuaries and coastal waters. There is a need for further simulations of the transient response of glacier geometry and runoff, and their application to a range of drainage basins and climate scenarios, to understand better the sensitivity of runoff changes to climate forcing and its regional variability.

There is a reasonable consensus regarding the direction of stream temperature response to changing glacier runoff and glacier retreat. However, the magnitudes are uncertain. There is less certainty regarding the changes in biogeochemistry and sediment yields. Further process-based research and model development are required to address these gaps. Without predictions of these water quality responses to future climate-glacier change it is not possible to make reasonable projections of the impacts on aquatic ecosystems.

Finally, the impact of climate change on the magnitude and frequency of glacier hazards needs to be understood with a considerably greater degree of confidence to be useful.

Appendix 7.1 Glossary of glaciological terms

Definitions largely follow those proposed by Cogley et al. (2011).
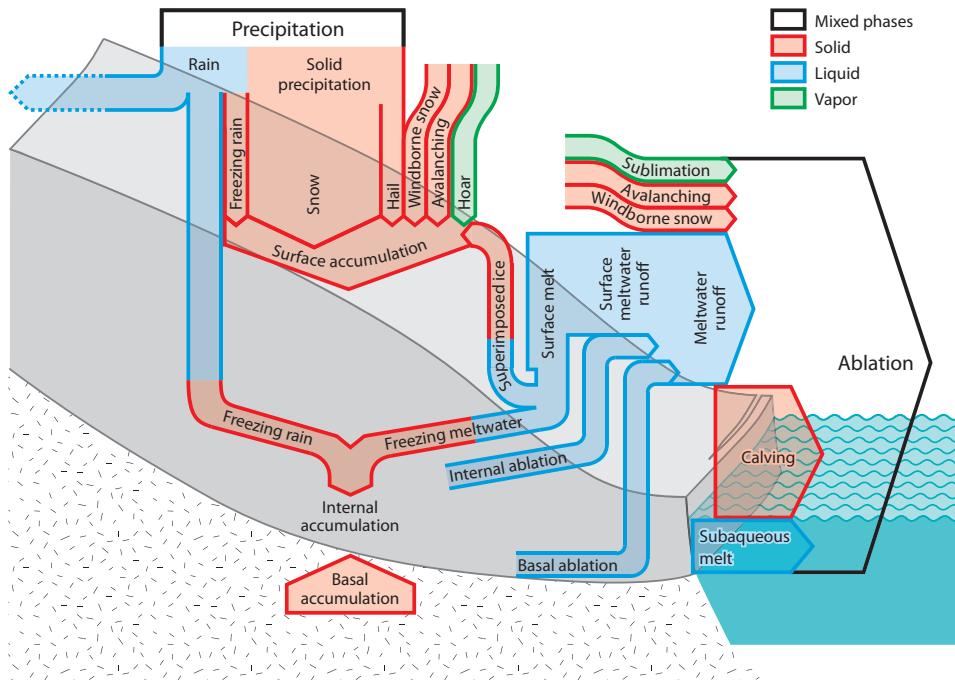
Components of the mass balance of a glacier. The arrows have arbitrary widths and do not indicate physical pathways of mass transfer. Source: after Cogley et al. (2011).

Accumulation area ratio

The *accumulation area ratio* (AAR), often expressed as a percentage, is the ratio of the area of the accumulation area to the area of the glacier. The AAR is bounded between 0.0 and 1.0 and its value often correlates well with the surface mass balance. The likelihood that mass balance will be positive increases as the AAR approaches 1.0.

Calving

Calving involves the breaking off of discrete pieces of ice from a glacier terminus into a lake or marine waters, producing icebergs. Calving of outlet glaciers is responsible for roughly half of the mass loss of the Greenland Ice Sheet (see Chapter 8), and can account for a significant part of the mass lost from glaciers in the Arctic.

Calving flux

Calving flux is the *mass flux by calving from a glacier terminus*, with dimension mass per unit time [M/T]. The calving flux can be determined from ice thickness and velocity measurements close to the terminus and the change in ice mass due to terminus advance or retreat.

Equilibrium line

The equilibrium line is defined by the set of points on the surface of the glacier where the surface mass balance is zero at the end of the melt season. The equilibrium line separates the *accumulation area* (where the annual mass balance is positive) from the *ablation area* (where the annual mass balance

is negative). It coincides with the snowline (defined as the set of points on a glacier forming the lower boundary of the snow-covered area) only if all mass exchange occurs at the surface of the glacier and there is no superimposed ice. Where superimposed ice forms, the equilibrium line marks the lower limit of superimposed ice formation and occurs at a lower elevation than the snowline.

Equilibrium-line altitude

The spatially averaged altitude of the *equilibrium line* is known as the equilibrium-line altitude (ELA). The ELA is generally determined by fitting a curve to data representing *surface mass balance* as a function of altitude, although it may also be defined by direct visual observation. The ELA often correlates well with the surface mass balance.

Glacier surge

A glacier surge represents anomalously fast glacier motion (10 to 100 times non-surging speeds) sustained for a period of months to years. Surges commonly alternate with slower motion lasting for periods of years to centuries during which the glacier thickens in its upper reaches and thins in its lower reaches. Surging transfers most of the accumulated reservoir of ice from high on the glacier to its lowest reaches, and often extends the glacier terminus by significant distances (kilometres). Surges recur at fairly consistent, glacier-specific intervals. Surge-type glaciers may end on land or in the ocean and are clustered geographically for reasons that are not well understood. Surges seem to be related to changes in the subglacial hydrological regime and not primarily to climatic fluctuations. They are

controlled in some cases by thermal processes ('Svalbard-type surging', characterized by long quiescent periods) and in other cases by hydraulic processes ('Alaskan-type surging' characterized by shorter quiescent periods and higher velocities relative to Svalbard-type surges). While the processes involved in surging are understood in broad terms, many details of their behavior remain elusive (Raymond, 1987).

Internal accumulation

Internal accumulation is the refreezing of surface meltwater (or freezing of rain) that is in transit through the glacier that otherwise would have left the glacier as runoff. Water that freezes within the current year's snowpack is not counted as internal accumulation. Internal accumulation may be regarded as simply redistributing mass within the glacier. On many Arctic glaciers, internal accumulation is a major component in the mass balance and failure to account for it results in a measurement bias towards overestimation of mass loss. Typically, internal accumulation is not detected by traditional surface mass balance measurements.

Mass balance

The mass balance is the change in the mass of a glacier or ice body over a stated span of time, often expressed as a rate. Mass gain (accumulation) mainly occurs by snowfall. On some glaciers deposition of hoar, gain of blowing and drifting snow, avalanching, and basal freeze-on (usually beneath floating ice) are significant processes of accumulation. The glacier loses mass (ablation) mainly by melting and iceberg calving. On some glaciers, sublimation, evaporation, loss of blowing and drifting snow, and ice avalanching are significant processes of ablation. In some cases, a fraction of the meltwater produced (plus rainfall) can refreeze, either within snow or firn (see *internal accumulation*) or on the ice surface (see *superimposed ice*), and does not contribute to mass loss from the glacier. These processes are summarized in Figure 7.8. There are several ways to measure mass balance; these are described in Section 7.3.2.1.

The principal mass balance components are surface accumulation (C_{sfc}, usually mostly snowfall and superimposed ice formation), surface ablation (A_{sfc}, usually mostly melt), internal accumulation C_i, internal ablation A_i, basal accumulation C_b, and basal ablation A_b, and (for ocean- or lake-terminating glaciers) iceberg calving D, so the glacier-wide mass balance $\dot{M}$ expressed as a rate is given by:

$$\dot{M} = \dot{C}_{sfc} + \dot{A}_{sfc} + \dot{C}_i + \dot{A}_i + \dot{C}_b + \dot{A}_b + \dot{D}$$

All terms have the dimension mass per unit time, [M/T]. Surface accumulation and ablation tend to be major components on most land-terminating Arctic glaciers. Internal accumulation is also important on the Greenland Ice Sheet and many polythermal/cold glaciers. Basal ablation can be a dominant component of the mass balance of floating tongues of marine-terminating glaciers, while basal accumulation and internal ablation are generally negligible.

The sum $C_{sfc} + A_{sfc}$ defines the **surface mass balance,** B_{sfc} or SMB, the latter symbol often being used in ice sheet studies. In estimates of ice sheet mass balance, the term has often been extended to include internal accumulation. In addition to snowfall and superimposed ice, surface accumulation also includes deposition of hoar, freezing rain, solid precipitation in forms other than snow (e.g., hail), and snow transported by wind or avalanches.

Mass balance units

The principal dimension for mass balance is mass per unit time [M/T]. Ice sheet balances are often expressed in Gt/y (1 Gt = 10^{12} kg). However, the mass balance, especially of smaller ice bodies, is usually treated as a rate of change of mass per unit area, in which case it is referred to as *specific mass balance* and its dimension becomes [M/L²/T], for example, kg/m²/y, which is numerically identical to the millimetre water equivalent, mm w.e./y. While mass units (e.g., kg or Gt) are useful for hydrological or sea level calculations, specific units (e.g., kg/m², mm w.e.) are needed for comparisons of glaciers of different sizes, for example, when investigating glacier-climate interactions. Mass balances, in particular for larger ice bodies, are often expressed in **sea level equivalent** (SLE), which is the change in mean global sea level that would result if glacier mass were added or removed from the ocean. SLE in mm/y is often simply obtained by

$$SLE = \frac{0.001 \dot{M} A_{glacier}}{\rho_w A_{ocean}}$$

where the glacier mass change $\dot{M}$ is in kg/m²/y and the areas of the glacier ($A_{glacier}$) and the ocean (A_{ocean}) are in m²; ρ_w is the density of freshwater (1000 kg/m³).

Superimposed ice

Ice accumulated on top of glacier ice by the refreezing of rain or meltwater produced during the current mass balance year is known as superimposed ice. It is different from internal accumulation, which represents refreezing within firn below the summer surface formed at the end of the summer in the previous mass balance year. Measurement of superimposed ice accumulation requires special attention in conventional mass balance programs.

Tidewater glacier

A tidewater glacier terminates in the sea, with its terminus either floating or grounded below sea level. Tidewater glaciers can undergo a periodic growth/shrinkage instability with long periods (centuries) of slower motion and slow advance alternating with shorter periods (decades) of rapid motion, high calving flux, and rapid (~1 km/y) retreat (Meier and Post, 1987).

Termed *tidewater instability*, this cyclic behavior has been documented most extensively in Alaska (Post, 1975), although it has been observed in many other regions, and the present retreat of certain Greenland Ice Sheet outlet glaciers appears to operate in a nearly identical fashion (Howat et al., 2005). During rapid retreat, a glacier can thin far faster (by loss of the marine-grounded tongue plus drawdown of an upstream basin by rapid flow) than can be accomplished by surface mass

balance alone. In its advancing phase, a tidewater terminus is stabilized by back-stress against a push moraine at its advancing margin. A tidewater glacier in an advanced position can switch abruptly (~years) to an unstable phase of thinning in its ocean-grounded reach, fast flow, and rapid calving which results in terminus retreat. Once initiated, fast flow and retreat can continue until the glacier's terminus retreats to a point where the glacier bed rises above sea level.

The causes of unstable retreat are not completely understood, but involve high basal water pressures tied to the depth of the glacier bed below sea level, rapid sliding, vertical thinning, and positive feedback between thinning and fast flow. The processes governing the size and rate of production of icebergs are extremely poorly known. Climate-induced thinning can play a critical role in initiating unstable retreat, but once initiated, unstable retreat is modulated by channel geometry and englacial and basal hydrology and is essentially independent of climate and surface mass balance.

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8. The Greenland Ice Sheet in a Changing Climate

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Contents

Key Findings

- At present, Greenland is warming at more than twice the global average due to the polar amplification at high Arctic latitudes. Annual Arctic temperatures are projected to be 2.8 to 7.8 °C higher in 2100.

- Mass balance estimates reveal that the Greenland Ice Sheet is losing ice at an increasing rate. The mass loss has increased from 50 (±50) Gt/y in 1995–2000 to 205 (±50) Gt/y in 2005–2006.

- A slight increment of accumulated mass is more than counterbalanced by the increased loss of mass by melting and ice discharge to the ocean. Runoff and ice discharge have both increased, each with more than 100 Gt of mass lost each year since 1990.

- The discharge occurs mainly through an increase in the velocity of the fast-flowing glaciers that terminate on ocean water in the fjords. Many fast-flowing glaciers have increased their velocity – some as much as doubling their velocity – and the glaciers reacting are observed further and further northward along the west coast of Greenland with global warming.

- The increased velocity from the fast-flowing glaciers appears to be related to warming of the ocean water.

- The present loss of mass (205 Gt/y) from the Greenland Ice Sheet if unchanged will cause a 5 cm global mean sea level rise by 2100. Models predict further loss of mass and increased discharge with the ongoing warming and 10 to 19 cm of eustatic sea level rise from Greenland Ice Sheet can be expected by 2100. However, the models still suffer from a very incomplete understanding of atmosphere ocean ice interactions. The maximum projected loss of mass from the Greenland Ice Sheet by 2100 is believed to result in a 40 cm eustatic sea level rise.

- Since 2003, the global contribution to sea level rise from the ice sheets and glaciers has increased from 50% to 75–85%. In the Arctic, mass loss from the Greenland Ice Sheet is very likely to exceed the contributions from the Arctic mountain glaciers and ice caps and from thermal expansion of the sea water during the next century.

- The increased meltwater runoff and iceberg discharge are reducing the salinity and density of the recipient marine surface water. These changes result in stronger gradients between the fresh surface water and the underlying saline water, as a result boosting circulation and increasing the transport of nutrients, which has a strong impact on the primary production of the affected marine areas.

- Planned hydroelectric power plants rely solely on runoff from the Greenland Ice Sheet and retreat of the ice sheet needs to be understood to be able to construct reliable and economic hydropower plants.

- Retreat of the Greenland Ice Sheet will open areas for mining, but will result in longer walking distances for hunters.

- Tourism in relation to the accelerating and retreating ice streams, especially at the UNESCO World Heritage site at Ilulissat Ice Fjord, is increasing and plays an important role in witnessing climate change first-hand.

Summary

The Greenland Ice Sheet is a massive ice cap nearly 3000 m thick. It is the largest body of freshwater ice in the Northern Hemisphere. There are 2.93 million km^3 of ice locked up in the Greenland Ice Sheet. Ice is lost from the ice sheet either at the surface, where it is melted, or from the edge, where it breaks off as chunks of solid ice. Measured rates of both types of loss from the Greenland Ice Sheet have increased.

Most ice comes off the Greenland Ice Sheet in a series of fast-flowing glaciers that discharge to the ocean through fjords along the west coast of Greenland. Studies show that these have increased their rate of flow – some doubled their speed between 1995 and 2000. The trigger for these changes is thought to be warming of the ocean water that is in contact with the outflowing end of these glaciers. One well-known glacier, the one terminating in the Ilulissat Ice Fjord (Kangia in Greenlandic), which drains ice from around 7% of the entire ice sheet, has doubled its flow rate and the ice front has retreated by 15 km in the past eight years. These changes are dramatic and the glacier has become a popular tourist destination. Estimates of the amount of ice leaving the Greenland Ice Sheet as solid icebergs have risen from a pre-1990 value of 300 gigatonnes per year (Gt/y) to 400 Gt/y in 2005.

Results published in 2007 indicated that the area of the Greenland Ice Sheet's surface that experiences some melting each year had been steadily increasing since satellite observations began. The area of surface melting is not a direct measure of the volume of water being lost from the ice sheet (called 'runoff'), but both the area and the number of days of surface melting are strongly related to the amount of ice lost.

To calculate the net loss of mass from the Greenland Ice Sheet, losses by melt and discharge are offset against mass gains caused by snow falling onto the ice sheet's surface. The elevation of the Greenland Ice Sheet at its centre, measured from satellites, has risen slightly recently – probably reflecting an increase in snowfall – but this increase is more than outweighed by the increase in loss of ice due to melting and ice discharge.

When mass budgets were first calculated for the Greenland Ice Sheet in the 1960s, the ice sheet appeared to be in balance. Those early measurements were very uncertain, but they showed similar amounts of ice being added and lost each year. The latest measurements tell a very different story, with large net losses of ice every year. The net loss of ice has increased from an average estimate of 50 (± 50) Gt/y from 1995–2000 to 200 (± 50) Gt/y from 2004–2008. The current rate of loss (~200 Gt/y) represents enough water to supply more than one billion city-dwellers.

8.1. **Introduction**

- The Greenland Ice Sheet is the second largest body of ice on Earth and its mass is equivalent to a global mean sea level rise of about 7.5 m.

- During the last million years, sea level has ranged from being over 100 m lower than at present during glacial periods to up to 20 m higher during interglacial periods.

- The warming during the past decade in the Arctic is more than twice the global average.

- Annual temperatures over the Arctic (i.e., all areas north of 60° N) are projected to be 2.8 to 7.8 °C higher in 2100 and the mass loss from the Greenland Ice Sheet is projected to increase significantly.

8.1.1. **Evolution of the Greenland Ice Sheet**

The Greenland Ice Sheet (Figure 8.1) is the largest body of ice in the Northern Hemisphere and, globally, is dwarfed only by the Antarctic Ice Sheet. The entire mass of the Greenland Ice Sheet has a volume of 2.93 million km³ of ice, equivalent to a global mean sea level rise of 7.5 m (Table 8.1).

This section presents recent scientific results, focusing specifically on the impacts of contemporary changes in the Greenland Ice Sheet. Although many reports on climate change have addressed impacts resulting from the evolution of the ice sheet, none provide a thorough and up-to-date assessment of the importance of the Greenland Ice Sheet in a changing climate. The importance of the ice sheet as a component of the Arctic cryosphere and the enormous amount of new observations and data provide the background for a better understanding of the evolution of the Greenland Ice Sheet and its potential impacts.

Table 8.1. Total volume of the ice bodies on Earth and the equivalent global sea level change if the ice bodies were to melt completely.

	Volume, million km³	Equivalent sea-level rise, m	Source
Antarctic Ice Sheet	25.4	57	Lythe et al., 2001
Greenland Ice Sheet	2.93	7.5	Bamber et al., 2001
All remaining glaciers and ice caps	0.051 – 0.133	0.15 – 0.37	Lemke et al., 2007

The high latitudes of the Northern Hemisphere are the regions that have experienced the strongest changes in climate since the 1951–1980 reference period (Figure 8.2). While the global annual temperature anomaly for 2009 in relation to the 1951–1980 reference period is 0.7 °C, the anomaly over Greenland is 1.5 °C, which is twice the global value. Average surface temperature changes observed in the Arctic are greater than those observed on average over the Antarctic Ice Sheet.

The trend of strongly enhanced changes in surface

a

Surface temperature anomaly (2009 vs 1951-1980), °C

b

Zonal mean temperature anomaly, °C
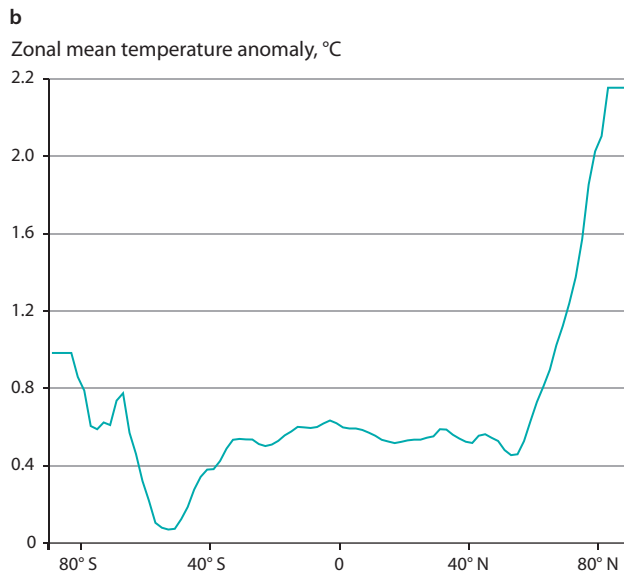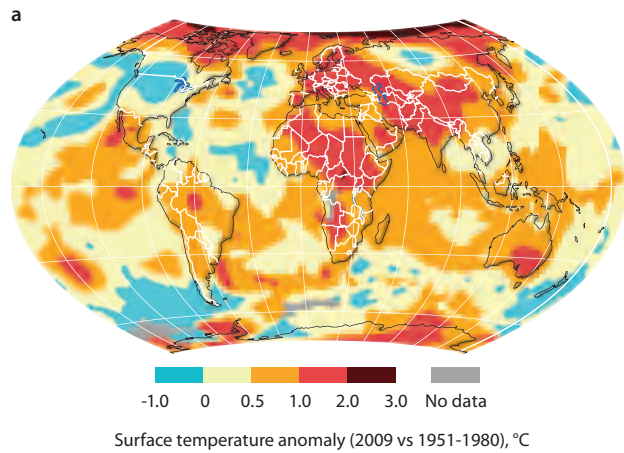
Figure 8.2. Mean surface temperature changes between 2009 and the reference period 1951–1980: (a) map averaged over 1200 km, (b) zonal means of the data shown on the main map. Source: data.giss.nasa.gov/gistemp.
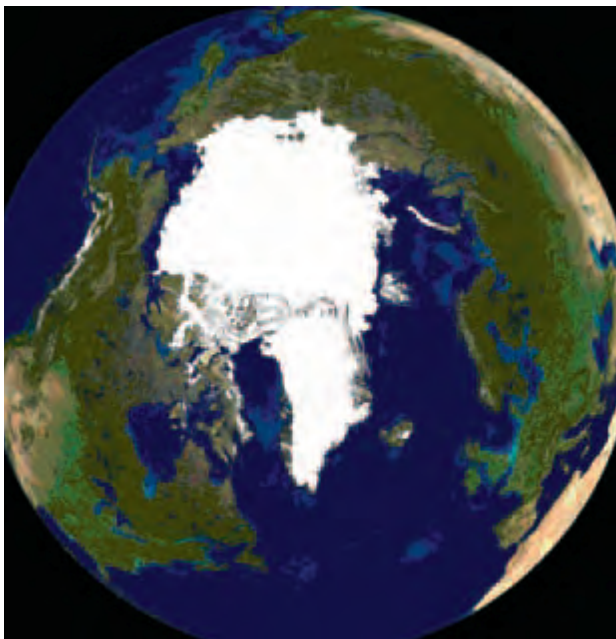

Figure 8.1. The high latitudes of the Northern Hemisphere are covered with ice and snow. The high reflectance (albedo) of the white areas reduces the absorbance of incoming solar energy. As the white areas diminish in a warming climate more solar energy will be absorbed by the ice-free ocean and land surfaces. Source: Earth View image.

temperature in the high northern latitudes is not only an issue in terms of the changes observed to date but is also a cause for concern in relation to the predicted increase in future temperatures. The IPCC A1B emissions scenario is projected to result in annual temperatures over the Arctic (i.e., all areas north of 60° N) 2.8 to 7.8 °C higher in 2100 than they were during the 1951–1980 reference period, with winters warming at a greater rate than summers (Meehl et al., 2007). A temperature increase of this magnitude is serious and would be the largest that the Greenland Ice Sheet would have experienced since the previous interglacial period, the Eemian, 130 000 to 115 000 BP (North Greenland Ice Core Project members, 2004).

8.1.2. History of the Greenland Ice Sheet

The global climate has changed enormously over time; there have been periods with glaciations and periods with no ice on Earth. While the Northern Hemisphere experienced ephemeral glaciations from 38 to 4 Myr BP, the onset of extensive glaciations did not occur until 3.3 Myr BP (Lunt et al., 2008).

Over the past three million years, the Greenland Ice Sheet is believed to have waxed and waned in accordance with glacial and interglacial periods consistent with orbital timescale variations in benthic $^{18}O/^{16}O$ ratios (Weidick, 1993; Lisiecki and Raymo, 2005; Weidick and Bennike, 2007) (Figure 8.3). Glacial deposits (i.e., moraines) 20 to 50 km from the present shoreline of Greenland can be used to reconstruct the Greenland Ice Sheet in earlier climatic periods. While the maximum extent of the ice sheet during glacial periods can be observed through studies of moraines and marine sediment cores, the degree to which the Greenland Ice Sheet retreated in earlier warmer periods is not clear.

Observations by Imbrie et al. (1984) indicate that sea level has been higher than at present several times during the past 400 000 years. 130 000 to 115 000 years ago during the Eemian climatic period (MIS5) sea level was 5 to 8 m above present (Kopp et al., 2009; Rohling et al, 2009) and 400 000 years ago
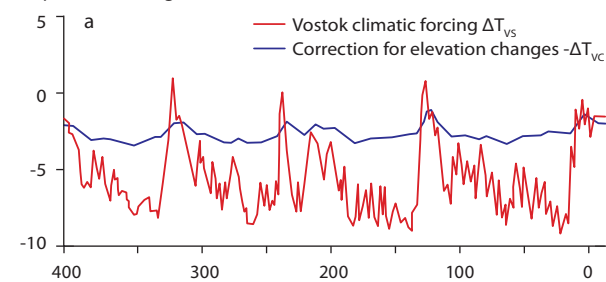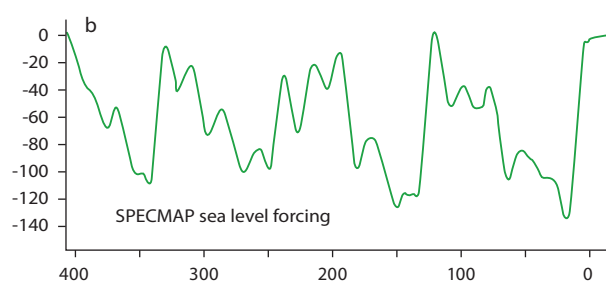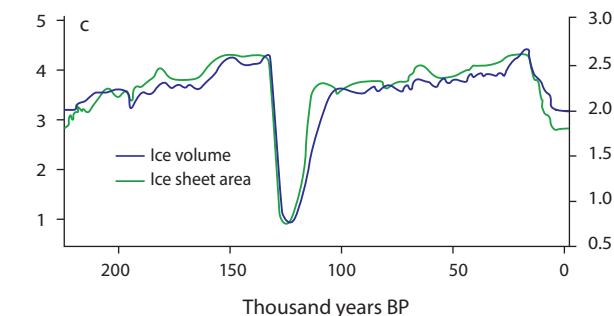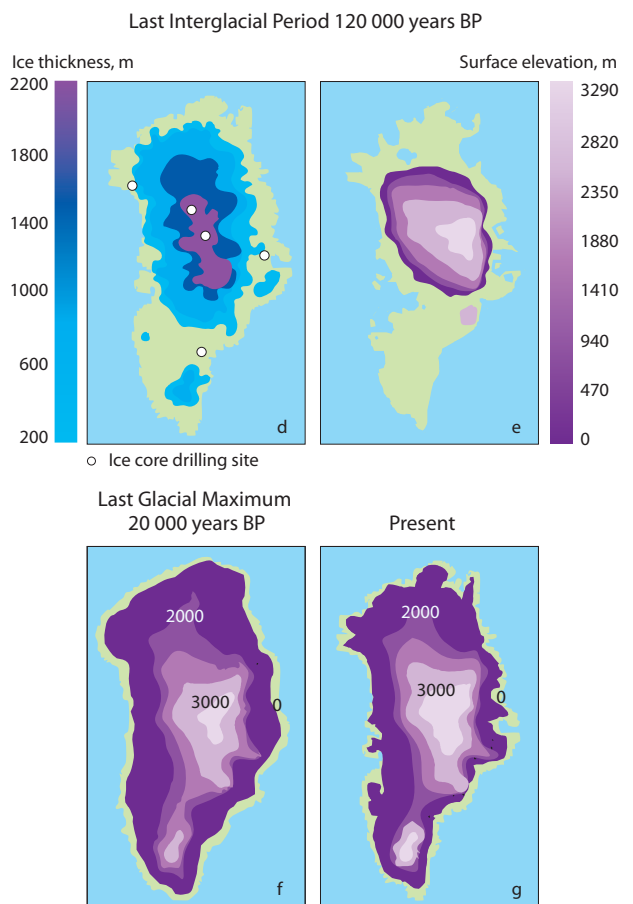
Figure 8.3. The volume and area of the Greenland Ice Sheet have changed as the climate has evolved through glacial and interglacial periods. The reconstructed temperatures, based on the Vostok ice core data (a), are compared with reconstructed sea level changes (b) over the past 400 000 years. The volume and area of the ice sheet for the past 200 000 years are shown in (c). Snapshots, 120 000 years ago, of the modeled Greenland Ice Sheet from the Last Interglacial Period, 130 000 to 115 000 BP (d-e), show the very reduced volume during the previous warm period. During the cold Last Glacial Maximum, 20 000 BP, the ice sheet volume increased and model simulations (f) estimate that the added volume of the ice sheet was equivalent to a reduced average global sea level lowering of 2.9 m. For comparison, the model simulation of the present Greenland Ice Sheet is shown in (g). Source: (a, b, c, f, g) Huybrechts (2002); (d) Otto-Bliesner et al. (2006); (e) Cuffey and Marshall (2000).

during the warm interglacial (MIS20) the highest stand is reconstructed at 20 m above present (Alley et al., 2010).

The 5 to 8 m of additional sea level during the previous interglacial period, the Eemian (MIS5), is of particular relevance for the future evolution of the ice sheet (Stirling et al., 1998; Lambeck et al., 2002; Kopp et al., 2009). The average air temperature is believed to have been 5 °C warmer than during the 1951–1980 reference period and to have remained so for several thousand years (North Greenland Ice Core Project members, 2004). The reaction of the Greenland Ice Sheet during the Eemian can to some degree be seen as an analogue to the evolution of the ice sheet in future warming scenarios.

Model reconstructions of the ice sheet during the Eemian show very reduced ice volumes (Figure 8.3), especially in southern Greenland (Letréguilly et al., 1991; Cuffey and Marshall, 2000; Otto-Bliesner et al., 2006), predicting a sea level contribution of 1 to 5 m from the Greenland Ice Sheet. Findings of Eemian ice in several Greenland ice cores and the presence of older ice at the base of the southern Greenland ice core drill site, DYE 3 (North Greenland Ice Core Project members, 2004; Willerslev et al., 2007; de Vernal and Hillaire-Marcel, 2008) support volume changes equivalent to a 1 to 3 m rise in sea level.

During the last glacial maximum, 25 000 to 14 000 BP, annual temperatures over Greenland are believed to have been 25 °C colder than at present (Cuffey et al., 1994; Cuffey and Clow, 1997; Dahl-Jensen et al., 1998), sea level to have been 120 m lower than at present (Lambeck et al., 2002), and the volume of the Greenland Ice Sheet to have been 140% higher than at present (Huybrechts, 2002; Lambeck et al., 2002), corresponding to a global average sea level lowering contribution of 3 m (Fleming and Lambeck, 2004).

Throughout the last glacial period, the climate was very unstable with 25 rapid climate changes, termed 'interstadials' or 'Dansgaard-Oeschger events'. The North Atlantic region and the Greenland Ice Sheet experienced rapid temperature increases of 10 to 15 °C occurring over a few decades (North Greenland Ice Core Project members, 2004; Landais et al., 2005; Steffensen et al., 2008) together with sea level increases of 5 to 20 m (Siddall et al., 2003). The rapid changes are believed to have been connected to enormous surges of ice into the ocean, especially from the Laurentide Ice Sheet over North America and changing strength of the ocean circulation in the Atlantic Ocean (Flückiger et al., 2006). After the abrupt warmings during the Dansgaard-Oeschger events, surface temperatures gradually cooled over the subsequent 1000 to 5000 years, before the next rapid warming occurred. The contribution of the Greenland Ice Sheet to sea level during these rapid warming events is not known. Ice core data show that glacial accumulation over the central parts of the Greenland Ice Sheet was less than 50% of the present accumulation, showing a strong correlation between surface temperature and precipitation (Cuffey and Clow, 1997; Johnsen et al., 2001). The 25 rapid climate changes were also followed by strong changes in precipitation rates, apparent from changing annual layer thicknesses in the ice cores (Alley et al., 1993; Andersen et al., 2006a; Svensson et al., 2006).

After the glacial period, the climate warmed into the present interglacial period, the Holocene, and the coastal regions around Greenland experienced an isostatic uplift of up to 100 m in response to the retreating ice (Letréguilly et al., 1991; Weidick, 1993; Huybrechts, 2002).

The onset of the Holocene brought the climate directly into a warm period named the 'Climatic Optimum' 5000 to 8000 years ago with Greenland temperatures 2 °C warmer than in the 1951–1980 reference period (Dahl-Jensen et al., 1998). The onset of the present interglacial period with abrupt warming changed the shape and volume of the Greenland Ice Sheet. The modeled volume changes of the ice sheet entering the warm Holocene climatic period differ depending on the modeled extent of the ice sheet during the last glacial maximum, the mass balance applied, and the evolution of the temperature in the ice (Letréguilly et al., 1991; Huybrechts, 1996; Marshall and Cuffey, 2000; Clarke and Marshall, 2002; Vinther et al., 2009). Investigations in the region around the Jakobshavn Isbræ glacier on the west coast of Greenland show that 5000 to 8000 years ago the margin had retreated 10 km further east than at present (Weidick et al., 1990; Weidick and Bennike, 2007).

Following the Climatic Optimum, surface temperature gradually cooled until The Little Ice Age (occurring between 1500 and 1800) with temperatures 1 °C colder than during the 1951–1980 reference period (Dahl-Jensen et al., 1998). Temperatures then increased and the years around 1940 were about 1 °C warmer than in the reference period. In general, temperature change during the present Holocene climate (the past 12 000 years) has been moderate and of the order of 1 to 2 °C, compared to changes of the order of 10 to 20 °C during the last glacial period. The warming of 1.5 °C over the past decades is of the same order as other warming events recorded during the Holocene interglacial period. Due to the relatively small temperature changes recorded during the Holocene the variability in the proxy paleo-data leading to temperature and precipitation records is of the same order as the changes themselves (Bales et al., 2001, 2009; Mosley-Thompson et al., 2001; Andersen et al., 2006a; Vinther et al., 2010). The relation between temperature change and change in mass balance during the Holocene is not well established and is further complicated by regional patterns. Decadal-scale trends over the past century are also complicated by the large degree of natural variability (Chylek et al., 2006).

8.1.3. Evolution of the Greenland Ice Sheet in the future

Experience from modeling the past evolution of the Greenland Ice Sheet indicates the need for improvements in order to accurately model the evolution of the ice sheet in a changing climate. Not enough is known about many fundamental parameters, such as ice sheet mass balance, basal processes, and ice-ocean interactions. Observations are made difficult by the vast extent of the ice sheet and the inaccessibility of the ice sheet base and submerged front.

This inadequate knowledge of processes is affecting the ability to predict the evolution of ice sheets and led the Intergovernmental Panel on Climate Change (IPCC) to conclude that:

Taken together, the ice sheets of Greenland and Antarctica are very likely shrinking, with Greenland contributing about 0.2 ± 0.1 mm yr^{-1} and Antarctica contributing 0.2 ± 0.35 mm yr^{-1} to sea level rise over the period 1993 to 2003. There is evidence of accelerated loss through 2005. Thickening of high-altitude, cold regions of Greenland and East Antarctica, perhaps from

coastal regions of Greenland and West Antarctica in response to increased ice outflow and increased Greenland surface melting. (Lemke et al., 2007:376).

Since the latest IPCC assessment was published in 2007 (IPCC, 2007), several other compilations have raised concern about the evolution of the big ice masses in response to global warming (Rignot et al., 2008; Steffen et al., 2008; Velicogna, 2009).

There is consensus from observations that the mass loss from the Greenland Ice Sheet has accelerated since 2000. While the contribution from the ice sheet is currently 10–20% of the observed global sea level rise of 3 mm/y, the Greenland Ice Sheet is believed to be capable of reacting more strongly to the warming over the next 100 years with a mass loss that could increase sea level by several tens of centimetres (Rignot et al., 2008; Steffen et al., 2008).

8.2. The Greenland Ice Sheet today

8.2.1. Overview of the Greenland Ice Sheet

- The Greenland Ice Sheet presents a major orographic barrier to Arctic weather systems.

- Long-term temperature records from coastal stations show a warming over Greenland since 1950. Since 1985 in West Greenland, the warming has primarily been driven by higher winter temperatures.

- The basal hydrological system plays an important role in the flow of the Greenland Ice Sheet and is also affected by increasing amounts of surface water drained to the base.

8.2.1.1. Arctic climate and large-scale synoptic observations

Greenland is the largest island in the world with a surface area of 2.2 million km², stretching 2600 km from 59.8 to 83.6° N, where – for several months of the year – it is either polar night or continuous daylight. Over 80% of Greenland is covered by a dome of inland ice (Figure 8.4) rising along an average gradient of about 1% from sea level in various regions to over 3300 m along the central spine.

Greenland plays a pivotal role in determining the climate of the Northern Hemisphere because of its size, location, elevation gradient, and mass of water stored in the ice sheet. During summer, Greenland is ideal for monitoring changes in meridional energy transport into the Arctic because it is located in the predominant direction of cyclone flow. Greenland's extreme North Atlantic location and the ice-filled ocean that surrounds it are the principal factors influencing the climate. The northern branch of the Gulf Stream, known as the North Atlantic Current, runs northward along the Norwegian coast into the Arctic Ocean, mixes with cold polar water and returns southward along the east coast of Greenland. Nearly all the water that leaves the Arctic Ocean passes through Fram Strait (Figure 8.4) via the East Greenland Current, eventually wrapping around Cape Farewell and continuing northward along the west coast of Greenland for several hundred kilometres. A similar, counterclockwise gyre is active along the coast of West Greenland with warm water moving northward

Figure 8.4. Greenland Ice Sheet elevation map and location of WMO climate stations along the coast and Greenland Climate Network (GC-Net) automatic weather stations (blue) in the ice sheet interior and the climate stations from the Geological Survey of Denmark and Greenland (red). Source: Konrad Steffen, CIRES, University of Colorado.

until it meets colder polar water flowing through Kennedy Channel. The resultant cold current flows southward along northeastern Canada.

Wind in the lower troposphere is forced to flow along the coasts of Greenland because of the height and size of the ice sheet. As a result, Greenland is an important participant in the exchange of air mass between the southern and northern latitudes of the Northern Hemisphere. Northerly and southerly air flow are about evenly distributed during summer. However, the predominant flow during winter is northerly due to high pressure in the colder regions to the west or northwest. Airflow in the free atmosphere at the 500 hPa level plays a key role in the Greenland climate because it governs the North Atlantic storm track. Generally, airflow at 500 hPa is from the southwest during winter and from the west during summer.

A typical North Atlantic cyclone develops as a wave in the polar front and propagates along the front. Therefore, winter cyclones generally travel along the east coast of the United States tracking along the edge of the Gulf Stream heading northeast. The cyclones often pass south of Greenland continuing toward Iceland and into the Norwegian Sea. However, they can also track more to the north through Davis Strait and into Baffin Bay. Occasionally, a cyclone will split near the southern tip of Greenland producing separate lows that track along both

coasts. During summer, lows tend to be less intense but are frequently more northerly and, therefore, often influence summer conditions in western Greenland.

Passing cyclones are generally accompanied by strong winds. However, during periods with no cyclone activity, the wind regimes are determined by local conditions, which usually relate to katabatic flow on the Greenland Ice Sheet, and sea breezes during summer or land breezes during winter near the coasts. Whether the high velocity katabatic winds on the ice sheet propagate down the fjords to the coastal areas largely depends on the temperature of the air mass as it reaches the head of the fjords. If the katabatic air mass is warmer than the air in the fjord due to adiabatic warming it will not be able to replace the air near the top of the fjord where it will be experienced as a warm 'foehn wind'. As a result, it will ride above the colder and denser surface air mass. If the katabatic flow is colder than the air in the fjord, it will replace the local air mass and probably travel all the way to the coast where it will be experienced as an icy downhill wind.

Climatic conditions in the fjords in the absence of cyclonic or katabatic influences are usually characterized by sea breezes during summer and week land breezes during winter due to ocean-land temperature differences. The winds associated with passing cyclones are, generally, very strong and heavily influenced by local topography and direction of the wind relative to the coast. Winds blowing directly toward the coast will often lead to precipitation due to orographic lifting. This scenario can often lead to very strong barrier winds, which blow clockwise relative to the land mass. Strong winds can cause re-deposition of snow accumulation and even wind-generated mass loss to the sea. Another special feature of the Greenland wind regime is that of very rapid change from calm to gale force conditions. These situations often develop as a result of cold Canadian air reaching the eastern coast of Greenland via the ice sheet behind a cyclone tracking to the northeast. The ice sheet topography determines the direction of the cold outflow from the ice sheet, focusing the wind into the low-lying coastal regions. The Greenlandic term for such an event is *pitoraq* (*piteraq* in East Greenland) which is closely related to the word for 'being attacked'. Piteraqs are most common during autumn and winter. Wind speeds typically reach 50 to 80 m/s. On 6 February 1970, the community Tasiilaq, also known as Ammassalik and the most populous community in East Greenland, was hit by a very strong piteraq causing severe damage. Since the beginning of the 1970s, special piteraq warnings have been issued by the Danish Meteorological Institute.

Surface air temperature in Greenland during summer is dominated by radiation effects as the sunlight returns to the Arctic. As a result, the mean temperature in July in the northernmost part of the country is only about 2 °C colder than in the southernmost part. However, temperatures at the coast are strongly influenced by the ocean and sea-ice variability; inland temperatures in ice-free regions can, therefore, be 5 °C warmer than at the coast. The proximity of the ice sheet does not seem to cool neighboring areas because the air flow off the ice during summer is usually warmer than the low lying local air mass due to adiabatic warming.

The latitudinal temperature gradient is much greater during winter than during summer with average sea level February

air temperatures in the north of -36 °C and in the south of -4 °C. Variability is also much greater without the moderating influences of the open ocean in the north and the melting ice sheet. Katabatic storms often raise winter temperatures above freezing via the combined effects of adiabatic warming and mixing of the inversion layer, particularly in the southern regions, where winter temperatures in the fjords can reach 10 °C (Cappelen, 2003).

Analysis of the long-term temperature record (1840 to 2001) from coastal stations in Greenland (Vinther et al., 2006, 2010) puts the more recent observations from satellites and weather stations on the Greenland Ice Sheet into a broader context (Box, 2002). The warmest decades in the long-term Greenland temperature record (1840 to 2000) are the 1930s and 1940s. Generally, the west coast experienced warming from 1873 to 1930 followed by cooling until 1988. There has been a warming trend since 1985 of 2 to 4 °C in West Greenland, primarily driven by winter temperature anomalies. The current warm period, beginning in 1988, is not unprecedented because the record warm decades occurred during the 1930s and 1940s.

The recent warmth occurred in spite of a persistently positive North Atlantic Oscillation (NAO) since around 1970. The NAO has a significant influence on the coastal temperature record particularly in the western and southern regions of Greenland during winter. The spectra of the temperature records display peak power at 3.7 years which Box (2002) attributed to the NAO, which has a spectral peak at ~4.0 years. The positive trend in coastal temperatures in southwestern Greenland appears to be responsible for the ice-sheet thinning observed in the region (Ohmura et al., 1999).

Volcanism has a significant cooling effect on temperatures in southern and western Greenland with a peak lag of 5 to 10 months following major eruptions (Box, 2002). The warming trend in Greenland since the early 1990s, in spite of major volcanic eruptions in 1982 and 1991, stands in contrast to the warmth of the 1930s and 1940s during a period of anomalously low volcanic activity (Overland et al., 2004).

Meteorological station records and regional climate model output were combined to develop a continuous 168-year (1840 to 2007) spatial reconstruction of monthly, seasonal, and annual mean Greenland Ice Sheet near-surface air temperatures (Box et al., 2009). Box and co-workers found that volcanic cooling episodes were concentrated in winter and along the western ice sheet slope, that inter-decadal warming trends coincided with an absence of major volcanic eruptions, and that 2003 was the only year in the 168-year series with a warm anomaly that exceeded three standard deviations from the 1951–1980 base period. The magnitude of the annual whole ice sheet 1919–1932 warming trend is 33% greater than the 1994–2007 warming. Box et al. (2009) also found that the recent warming was stronger along western Greenland in autumn and southern Greenland in winter. Spring trends marked the 1920s warming onset while autumn leads the 1994–2007 warming.

The climate on the Greenland Ice Sheet has been studied intensively in an effort to quantify the surface mass balance and to estimate its contribution to global sea level. However, *in situ* observations are limited due to the large area and remoteness of the ice sheet. The notable exception is the climate record from the Greenland Climate Network (GC-Net) of automatic weather stations, operating since 1995 and distributed across the

ice sheet (Figure 8.4). Scientists are, therefore, forced to rely on remotely sensed observations and estimations from downscaled general circulation models (GCMs) or high-resolution regional climate models (RCMs). However, systematic biases in the models relative to the *in situ* record illustrate that the climate of the ice sheet is more or less decoupled from the coastal regions where most of the direct observations that feed the models are made. Nevertheless, a consensus picture of change due to a warming climate since 1995 is emerging.

8.2.1.2. Climate observations

The Danish Meteorological Institute (DMI) data are the most comprehensive meteorological records available for Greenland's coastal region and have been homogenized, most notably the original observations have been checked and the data compared with time series of related climatic elements for the same stations, with the specific purpose of studying long-term climatological trends (Cappelen et al., 2001).

Monthly air temperature records from eight DMI synoptic stations at the coast of Greenland (located mainly in the southwest but including Tasiilaq in the southeast) show a pronounced warming since the early 1990s. The warming follows an overall trend of regional cooling between 1958 and 1992, concentrated in the winters of the 1960s to 1980s (Hanna and Cappelen, 2003). Summer trends for the 7-station DMI

average (i.e., excluding Kangerlussuaq because of its relatively short record) indicate a significant warming for 1961 to 2006 of 0.9 °C.

For the DMI station average, 2003 had the warmest summer (i.e., June to August) on record, with a mean temperature of 8.1 °C at 3.3 standard deviations above the most recent climatological 'normal' period (1971–2000) mean. The second warmest summer occurred in 2005 (7.6 °C) at 2.5 standard deviations above the 1971–2000 mean. 2005 was more than 0.5 °C warmer than the third warmest summer, 2006 (7.07 °C) which was closely followed by 2001 (7.02 °C), 1965 (7.00 °C), and 2004 (6.97 °C) (Hanna et al., 2008).

The record for the GC-Net automatic weather station 'Swiss Camp' (1170 m above sea level in 1991 and since then decreasing 0.32 m annually; Stober and Hepperle, 2006; Figure 8.4) (Steffen and Box, 2001) was used to gauge temperature changes on the western flank of the Greenland Ice Sheet, where extensive seasonal melt and relatively high runoff from this relatively low elevation zone contribute a large proportion of the total runoff (Hanna et al., 2005; Box et al., 2006). This record, by far the longest GC-Net series, spans 19 years (1991 to 2009), and its interannual variability is significantly correlated with that of the mean of the seven DMI coastal stations (detrended series r = 0.65, $p < 0.01$; Figure 8.5).

Similar to the positive (7-station mean) DMI temperature
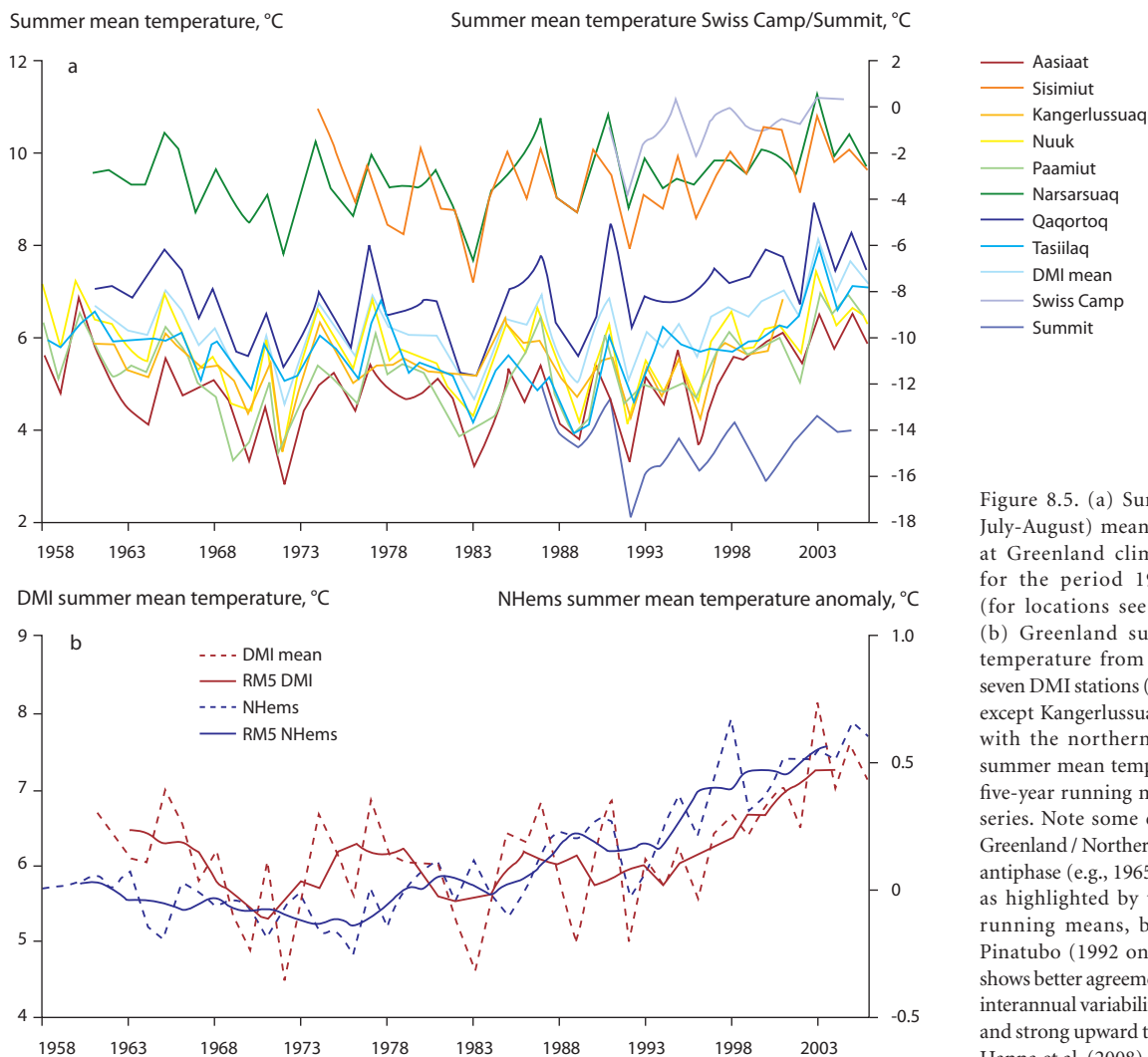
Figure 8.5. (a) Summer (June-July-August) mean temperature at Greenland climate stations for the period 1958 to 2006 (for locations see Figure 8.4). (b) Greenland summer mean temperature from the mean of seven DMI stations (see Figure 8.4, except Kangerlussuaq) compared with the northern-hemisphere summer mean temperature, with five-year running means of both series. Note some early years in Greenland / Northern Hemisphere antiphase (e.g., 1965, 1971, 1983), as highlighted by the disparate running means, but the post-Pinatubo (1992 onward) period shows better agreement in terms of interannual variability, correlation, and strong upward trends. Source: Hanna et al. (2008).

trend, Swiss Camp summer mean temperature increased significantly by 2.4 °C from 1991 to 2008, or 2.2 °C for the period 1993 to 2008, excluding the global cooling effect of the Mount Pinatubo eruption in 1992. The three summers of 2003/2004/2005 were almost equally record warm (mean temperatures 0.3, 0.3, and 0.2 °C, respectively) at Swiss Camp, alongside 1995 (0.5 °C) and the record year 2007 (1.2 °C). The latter season has been previously noted for its relatively high modeled runoff compared with most other years between 1958 and 2003 (Hanna et al., 2005). The latest data from the Swiss Camp GC-Net record show a continued warming with another record mean summer temperature of 1.2 °C in 2007; 0.7 °C above the previous maximum in 1995.

A reprocessed and updated near-surface air temperature series for Summit (1987–2005, 3205 m above sea level) (Shuman et al., 2001) shows a slight overall -0.3 °C cooling in summer, in contrast to all the other Greenland temperature records (all from much lower elevations and generally around the ice sheet margin; Figure 8.5). This new Summit series is a re-analyzed composite, primarily of University of Wisconsin and ongoing GC-Net automatic weather station data, supported by SSM/I (Special Sensor Microwave / Imager) brightness temperature data (Shuman et al., 1996, 2001). There is a highly significant correlation between individual years' fluctuations in the de-trended DMI and the Summit series (Figure 8.5).

Three hypothesized possible causes for the cooling trend observed at Summit compared to the warming trend along the coast are: (i) continued relative suppression of more regional climatic change by thermal inertia of the huge central Greenland ice mass as noted for Arctic Ocean sea ice (Serreze and Francis, 2006); (ii) a differential response of the high elevation zones of the Greenland Ice Sheet in accordance with the well-known lower tropospheric warming / higher atmospheric cooling response to increased greenhouse gases (Stott et al., 2006), which has been demonstrated specifically for Greenland in a recent analysis of radiosonde data spanning the period 1964 to 2006 (Box and Cohen, 2006); and/or (iii) regional changes in wind, cloud cover, or radiation patterns over the Greenland Ice Sheet.

This observed pattern (coastal, i.e., marginal, warming combined with little change or slight cooling in the high interior) is opposite to the output of simulations from Atmosphere-Ocean General Circulation Models (AOGCMs). In all high-resolution temperature changes studied in the present century, Huybrechts et al., (2004) and Gregory and Huybrechts (2006) found that modeled summer warming is actually greatest over the interior of Greenland and least along the coast.

High-resolution (5 km × 5 km) surface air temperature (SAT) data were bilinearly interpolated from the 0.5° resolution 40-year European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and corrected for ECMWF terrain errors using empirically derived ice sheet surface lapse rates, as explained by Hanna et al. (2005). These SAT data provide corroborating evidence for anomalously (3 °C) high summer temperatures around the Greenland margins during recent warm summers with a noted high melt, which occurred most widely within the southern and western marginal ablation zones of the ice sheet.

The accumulation zone of the Greenland Ice Sheet (above 2000 m elevation) apparently had cold summer anomalies in 2003 and 2006. This is in contrast to warm anomalies in more outlying areas but in line with the Summit temperature results discussed previously. Summer temperatures both in Greenland and across the Northern Hemisphere exhibit strongly rising trends since the early 1990s, although the earliest part of this period (1992–1993) characterizes the general temperature recovery following the cooling after the 1991 Mount Pinatubo volcanic eruption (Robock and Mao, 1995).

Observations and models both indicate the occurrence of recent high snow accumulation events in the winter of 2004/2005, concentrated in western Greenland (Nghiem et al., 2007), and winter–spring 2002/2003 in southeastern Greenland (Krabill et al., 2004; Hanna et al., 2006). Huybrechts et al. (2004) hypothesized that such events may become more frequent in Greenland as storm tracks intensify or shift position with climate change.

8.2.1.3. **Surface and basal properties and processes**

Surface melting in the ablation zone (Figure 8.6) is highly sensitive to changes in surface albedo. Melting of ice in the ablation zone exposes a surface layer of dust that was originally deposited with snowfall high on the ice sheet. Dust in snow and on ice reduces the solar reflection (albedo) and, thus, increases surface melt further. Surface albedo decreased dramatically at Swiss Camp during 1993 to 2008 and this led to increased surface melt and ablation at Swiss Camp (Figure 8.7).

Basal melt and meltwater exert a strong influence on ice flow (Fahnestock et al., 2001). The limited knowledge of basal melt is derived from models and sparse observations of bed properties. A basal melt rate of 0.1 m/y over a substantial area requires a regional geothermal heat flux of 970 MW/m^2, which is 17 times higher than the continental background of 56.5 MW/m^2 (Sclater et al., 1980). This does not account for heat conducted through the ice. For most of the ice sheet, the heat conducted through the ice is close to the background geothermal heat flux. It can be lower in low-accumulation regions because the thermal gradient in the basal ice is lower. The volume of meltwater being produced in the areas of high basal melt is of the order of 1 km^3/y. This water and warm basal ice are responsible for the onset of rapid ice flow in the large ice stream that drains the north side of the summit dome in Greenland (Fahnestock et al., 2001). Strong melt also occurs due to deformation heat under the fast-flowing ice streams thereby further increasing their flow. Melt rates of metres per day are observed where the base of the ice streams is in contact with the ocean (Rignot et al., 2010).

Basal flow can transport ice at velocities exceeding rates of internal deformation (i.e., from hundreds to more than 10 000 m annually) and glacier surges, tidewater glacier flow, and ice stream motion are governed by basal flow dynamics (Clarke, 1987). Glaciers and ice sheets that are susceptible to basal flow can move quickly and erratically, making them intrinsically less predictable than those governed by internal deformation. They are more sensitive to climate change because of their high rates of ice turnover which gives them a shorter response time to climate (or ice margin) perturbations. In addition, they may be directly responsive to increased amounts of surface meltwater production associated with climate
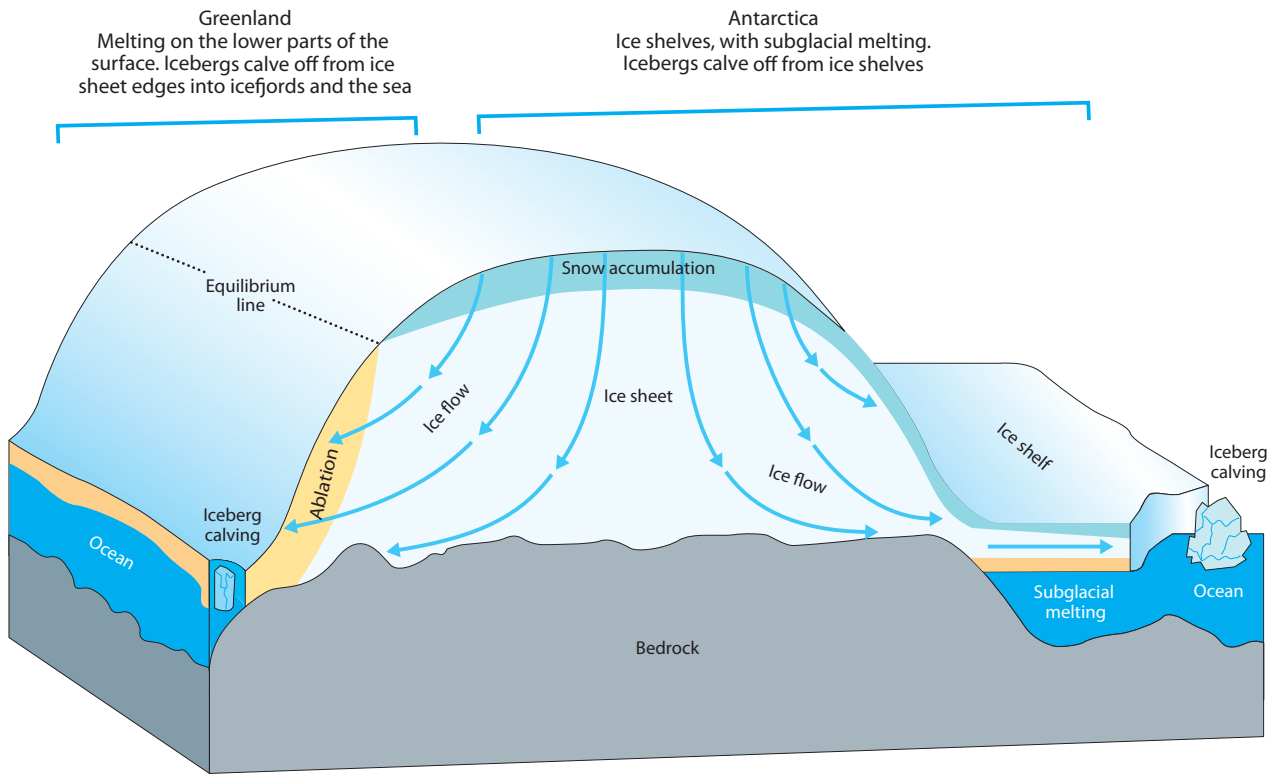
Figure 8.6. The ice cover in Greenland and Antarctica has two components – thick, grounded, inland ice that rests on a more or less solid bed, and thinner floating ice shelves and glacier tongues. An ice sheet is actually a giant glacier, and like most glaciers is nourished by the continual accumulation of snow on its surface. As successive layers of snow build up, the layers beneath are gradually compressed into solid ice. Snow input is balanced by glacial outflow, so the height of the ice sheet stays approximately constant through time. The ice is driven by gravity to slide and to flow downhill from the highest points of the interior to the coast. There it either melts or is carried away as icebergs that eventually melt, thus returning the water to the ocean from where it came. Outflow from the inland ice is organized into a series of drainage basins separated by ice divides that concentrate the flow of ice into either narrow mountain bounded outlet glaciers or fast-moving ice streams surrounded by slow-moving ice rather than rock walls. In Antarctica, much of this flowing ice has reached the coast and has spread over the surface of the ocean to form ice shelves that are floating on the sea but are attached to ice on land. There are ice shelves along more than half of Antarctica's coast, but very few in Greenland. Sources: UNEP Maps and Graphs; Konrad Steffen, CIRES, University of Colorado.
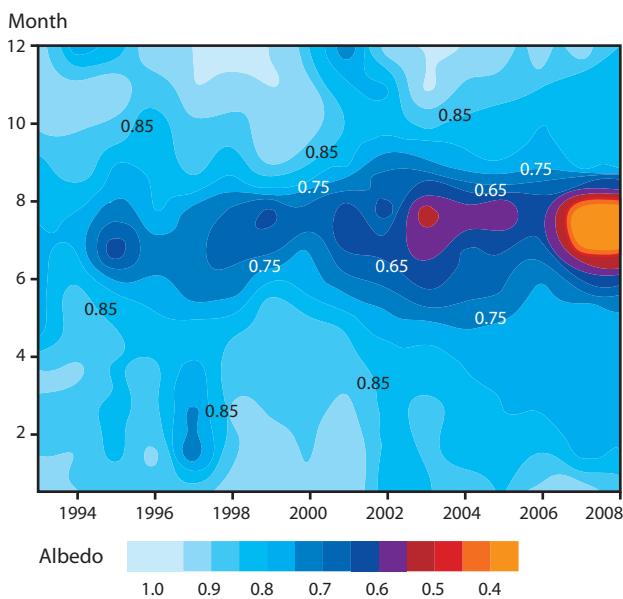

Figure 8.7. Interannual variability of monthly mean albedo at Swiss Camp (1993 to 2008). Note the decrease in surface reflectivity (albedo) for the summer months (June–August) with values as low as 0.4 in 2007 and 2008. Increase in summer air temperature at this elevation enhances snow and ice melt as well as reducing albedo. Hence, less solar radiation is reflected and this further increases surface melt (positive albedo feedback). Source: Konrad Steffen, CIRES, University of Colorado.

warming. This latter process is crucial for predicting dynamic feedbacks to the expanding ablation area, longer melt season, and higher rates of surface meltwater production that are predicted for most ice masses.

Although basal meltwater is most likely to be the primary source of subglacial water, models have shown that supraglacial streams with discharges of over 0.15 m^3/s can penetrate down through 300 m of ice to reach bedrock, via self-propagation of water-filled crevasses (Arnold and Sharp, 2002). There are several possible subglacial hydrological configurations: ice-walled conduits, bedrock conduits, water film, linked cavities, soft sediment channels, porous sediment sheets, and ordinary aquifers (Mair et al., 2001; Flowers and Clarke, 2002).

Modern interest in water flow through glaciers originated in a pair of theoretical papers published in 1972. In one, Shreve (1972) discussed the influence of ice pressure on the direction of water flow through and under glaciers, while Röthlisberger (1972) presented a theoretical model for calculating water pressures in subglacial conduits in the other. Through a combination of these theoretical considerations and field observations, Röthlisberger concluded that the englacial drainage system probably comprises an arborescent network of passages. The millimetre-sized fingertip tributaries of this network join downward into ever larger conduits.

Locally, moulins provide large direct connections between the glacier surface and the bed. Beneath a valley glacier, the

subglacial drainage is likely to occur as a tortuous system of linked cavities transected by a few relatively large and comparatively straight conduits. The average flow direction in the combined system is controlled by a combination of ice overburden pressure and bed topography, and, generally, is not normal to contours of equal elevation on the bed.

Luthje et al. (2006) studied the summer evolution of supraglacial lakes on the Greenland ice margin using a one-dimensional model to calculate the surface ablation for a bare ice surface and beneath supraglacial lakes for 30 days in the summers of 1999 and 2001. The surface ablation beneath the lake was enhanced by 110–170% for the two years 1999 and 2001 compared with the ablation for bare ice. Within the region of the ice sheet where supraglacial lakes presently occur, the area covered by supraglacial lakes was found to be more or less independent of the summer melt rate but controlled by topography. Luthje and co-workers predicted that, inside the ablation region, the area covered by supraglacial lakes will remain constant even in a warmer climate.

Although theoretical studies usually assume that subglacial conduits are semicircular in cross section, there are reasons to believe that this ideal is rarely realized in nature. Much of the progress in subglacial hydrology has been theoretical, as experimental techniques for studying the englacial hydraulic system are few and, as yet, not fully exploited and observational evidence is difficult to obtain. How directly and permanently do these effects influence ice dynamics? It is not clear at this time. The process is well known in valley glaciers where surface meltwater that reaches the bed in the summer melt season induces seasonal or episodic speed-ups (Iken and Bindschadler, 1986). Speed-ups have also been observed in response to large rainfall events (e.g., O'Neel et al., 2005).

8.2.1.4. Summary

Is the climate in Greenland changing? Coastal records since 1840 show the warmest decades to be the 1930s and 1940s and a warming trend since 1988 of 2 to 4 °C in West Greenland, mainly during the winter months. Recent, short-term temperature records on the ice sheet in West Greenland (Swiss Camp) show a 2.2 °C warming in summer since 1991 with the highest air temperatures in 2007. Air temperatures at the top of the Greenland Ice Sheet (Summit) revealed a slight cooling of -0.3 °C between 1987 and 2005. Decadal-scale trends over the past century, however, are complicated by the large degree of natural variability (Chylek et al, 2006). The observation period for the recent climate warming over Greenland is short.

8.2.2. Surface mass balance

- The surface mass balance of the Greenland Ice Sheet has decreased since 1990. The mean of three model reconstructions of the surface mass balance show a 40% decrease from 350 Gt/y (1970–2000) to 200 Gt/y in 2007.

- The runoff has increased significantly due to higher temperatures, increasing ablation zone area and a longer melt season.

- These trends in surface mass balance have been linked to a global warming trend that exceeds the local natural variability in climate.

8.2.2.1. Introduction

The total mass balance of the Greenland Ice Sheet (Section 8.2.4) determines its overall state of health and is the sum of two terms: the surface mass balance (SMB), discussed here, and solid ice discharge from marine terminating glaciers, discussed in Section 8.2.3. The SMB is the net mass added or removed from the surface of an ice sheet by a range of processes. The aim of this section is to introduce the factors that control SMB, and how they respond to climate change. This is achieved by reviewing recent estimates of the components that make up the SMB, discussing the degree of consistency in these components, both from observations and modeling, and considering areas of uncertainty requiring further development.

The annual SMB is defined as the sum of mass fluxes to and from the ice sheet surface, integrated over a year and over the area of the ice sheet. Mathematically, this is expressed as:

$$SMB = P + TMT - R - Q$$

where P is total precipitation, comprising the solid fraction (snow, hail, freezing rain) as well as the liquid fraction (rain); TMT is turbulent moisture transport, representing the net effect of surface evaporation, sublimation/deposition and snowdrift sublimation; R is runoff, representing the net effect of condensation, melt, refreezing and retention; and Q is the snowdrift blown off the ice sheet integrated along the entire margin. All fluxes are in Gt/y and integrated over the area of the ice sheet. Snowdrift plumes being blown into the ocean have been occasionally observed, but Q is believed to impact SMB significantly only locally and is not further considered here (Box et al., 2006). The accumulation zone is the region where SMB > 0 and the ablation zone is the region where SMB < 0. The equilibrium-line altitude (ELA) is the elevation where SMB = 0. It is important to note, however, that precipitation takes place in the ablation zone and runoff can take place in the lower part of the accumulation zone. Above the ELA, however, accumulation (via snowfall) exceeds ablation (via runoff) and below the ELA the opposite is true. For an ice sheet as a whole to be in a state of balance, SMB must match the solid ice flux crossing the grounding line and flowing into the ocean. For the Greenland Ice Sheet, this was roughly the case prior to the 1990s but there has since been a marked increase both in solid ice flux (Rignot and Kanagaratnam, 2006) and runoff (e.g., Hanna et al., 2008; Ettema et al., 2009). Accumulation includes all processes by which mass is added to the ice sheet, which, in this case is almost entirely through precipitation in the form of snowfall. The accumulation rate is what is typically measured with *in situ* observations from snow pits or shallow ice cores in the high elevation zones of the ice sheet where there is basically no runoff. It is usually expressed as a water equivalent (w.e.) thickness change per unit area per unit time. Ablation includes all process by which mass is lost from the ice sheet. This section, however, only considers ablation via surface melting (called runoff). The other major component of ablation due to solid ice discharge into the ocean is dealt with in Section 8.2.3. The ablation rate is identical and has the same units but is a measure of net mass loss at the surface rather than mass gain. Ablation rates can be measured during the summer melt season using, for example, stakes drilled into the ice surface.

These observations of surface accumulation/ablation rate are equivalent to the local SMB.

There are a number of approaches for estimating SMB. The first, and the only tractable approach until quite recently, involves the interpolation of *in situ* observations of point SMB (accumulation/ablation) estimates from snowpits, ice cores, stake measurements and automatic weather stations to estimate accumulation rates (Figure 8.8). The data were combined with simplified models of runoff, R, to determine SMB (Ohmura and Reeh, 1991; Reeh, 1991; Ohmura et al., 1999; Bales et al., 2001; Steffen and Box, 2001). The more recent interpolations to determine accumulation rates also included a formal assessment of the error associated with field data and their interpolation (Cogley, 2004; Bales et al., 2009) as discussed in Section 8.2.2.2.

Commonly, runoff has been estimated using an empirical model of melt, M (see Section 8.2.2.3), combined with simplified assumptions about refreezing, RF, calibrated using stake measurements (Reeh, 1991; Braithwaite et al., 1992; Braithwaite, 1995). The differences in the estimations of runoff from the different models are of the order 25% (Table 8.2). The individual components contributing to runoff (M and RF) vary by far more than this. Thus, the agreement in R is largely based on compensating differences. The different runoff models have very different sensitivities to forcing and, as a consequence, the ability to predict the true response of SMB to a future change in climate also contains significant uncertainties (Bougamont et al., 2007).

Recently, alternative approaches for estimating SMB have been developed, especially following the release of a high-quality, consistent ~50 year global climatology produced by ECMWF. The data set is known as ERA-40 (i.e., ECMWF Re-Analysis covering the ~40 year period 1957-2002). ERA-40 is a hind-cast generated using the ECMWF atmospheric forecast model driven by surface and satellite meteorological observations (Uppala et al., 2005). ERA-40 produces 6-hourly and daily fields of key atmospheric parameters at 1.125° resolution, which equates to a grid cell size of ~125 km in latitude. This resolution is too coarse to be used unadjusted for SMB calculation as the ablation zone, for example, is mostly less than a single grid cell in width. As a result, various approaches, commonly termed downscaling, have been used to interpolate ERA-40 to higher resolution. The simplest approach has been to downscale surface air temperature using seasonally adjusted lapse rates and high-resolution surface topography as well as a simple interpolation of precipitation (Hanna et al., 2002, 2005) (Table 8.2 and Figure 8.9).

Another approach is to use an atmospheric RCM as a physically-based 'interpolator' of the reanalysis data. Generally, an RCM is run at high resolution for a limited area and forced at its boundaries by climate data from a coarser-resolution GCM. To date, three RCMs have been used to produce a high-resolution time series of the Greenland Ice Sheet SMB (Box et al., 2004, 2006; Fettweis, 2007; Ettema et al., 2009) although others are under development.

It is interesting to note how SMB responds to changes in climate. Increased near-surface air temperatures bring about increased moisture and, therefore, precipitation. Simulations with coupled atmosphere-ocean GCMs suggest a 5%/K sensitivity of P to surface air temperature (Gregory and Huybrechts, 2006). This value can be combined with SMB 'climate sensitivities' (especially to changes in temperature and precipitation) estimated from an energy balance model that suggests that the area-averaged SMB sensitivity, CT, to temperature is -49 mm/K and to precipitation, CP, 3.8 mm/% (Oerlemans et al., 2005). Using the sensitivity of precipitation
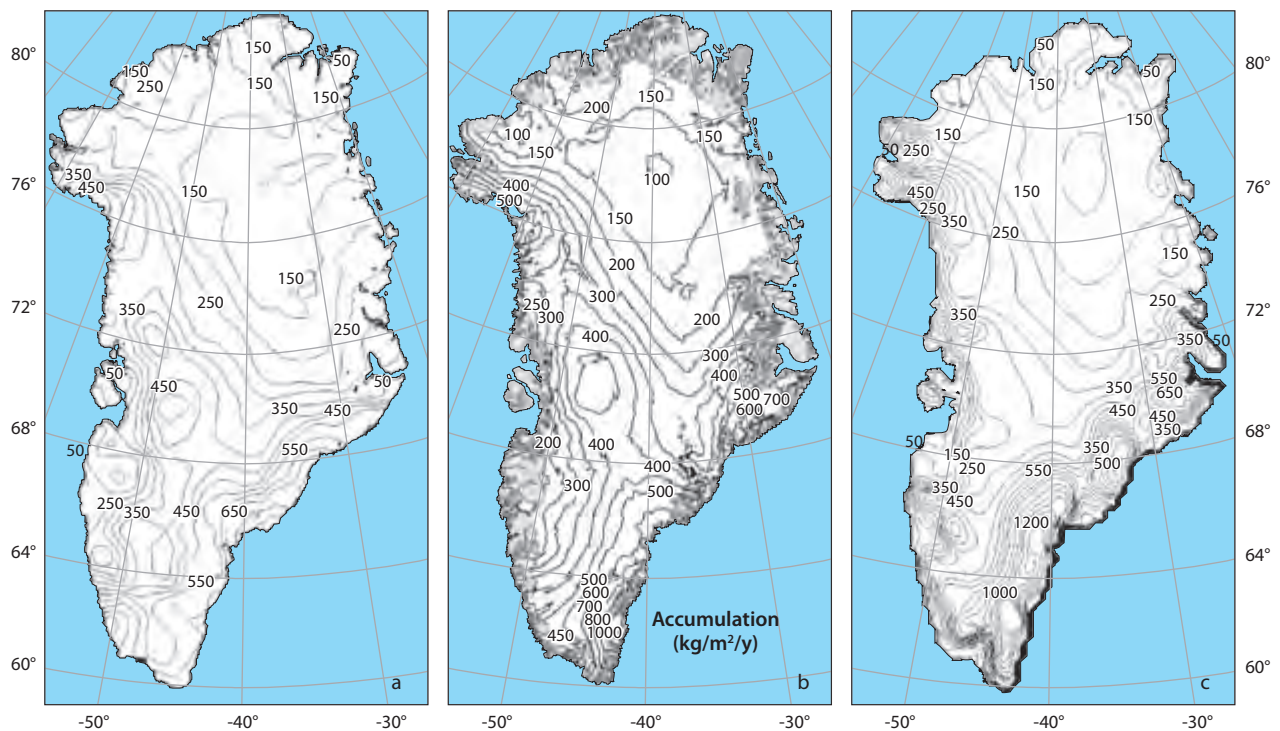

Figure 8.8. Net accumulation (snowfall minus turbulent moisture transport) from three recent estimates. Figures a and b are from observations, while c is from a regional climate model. Sources: (a) Bales et al. (2001, 2009); (b) Cogley (2004); (c) MAR (Fettweis, 2007).

Table 8.2. Estimates of the SMB components in Gt/y for Greenland from satellite (Mote, 2003) and modeling studies for the range of time periods specified. The regional climate models, PMM5, RACMO2/GR and MAR are discussed in Section 8.2.2.2. The sum of RF and R does not necessarily equal M, depending on how rain is treated.

Model	Period	Area × 10⁷ km²	P	TMT	Rain	M	R	RF	SMB
MAR[a]	1958–2007	1.701	575	-5	20	532	282	250	264
PMM5[b]	1958–2006	1.691	555	-63	16	228	213	48	326
Hanna[c]	1958–2007	1.678	559	-35	26	313	261	77	264
RACMO2/GR[d]	1958–2007	1.69	743	-26	46	404	248	199	469
Mote[e]	1988–1999	1.648	591	-69	23		255		239
Reeh[f]	1990s	1.707	552[g]				279		273

P: Precipitation, comprising the solid fraction (snow, hail, freezing rain) as well as the liquid fraction (rain); TMT: Turbulent Moisture Transport, comprising surface evaporation, sublimation/deposition and snowdrift sublimation; Rain: liquid fraction of precipitation; M: total volume of Meltwater; R: Runoff, comprising the liquid water fraction from melt and rain that leaves the firn/ice column; RF: Re-Freezing; SMB: Surface Mass Balance.
[a] Modèle Atmosphérique Régional (Fettweis, 2007); [b] Polar Mesoscale Model version 5 (Box et al., 2004); [c] Hanna et al. (2002, 2005); [d] Ettema et al. (2009); [e] Mote (2003); [f] Reeh et al. (1999); [g] Reeh and co-workers (1999) did not estimate the individual terms in accumulation and the tabulated value represents, therefore, P-TMT-Rain.

to temperature, given above, makes it possible to express CP with respect to temperature, which gives CP = 19 mm/K. Thus the increase in runoff greatly outweighs the increase in accumulation (49 *vs* 19) for a fixed temperature change. Assuming the area of the ice sheet to be 1.7 million km² then the increased mass loss from the ice sheet is 47 Gt/K, excluding any dynamic response. Thus, simulations of future SMB for various greenhouse gas emissions scenarios generally predict a decreasing SMB (Gregory and Huybrechts, 2006). Most importantly, this is what has been observed to have taken place over the past 15 years, since the temperature increase over Greenland started (Hanna et al., 2008).

8.2.2.2. **Accumulation**

8.2.2.2.1. **Observations**

Shallow ice cores and snowpit measurements have been collected and used to determine accumulation rates, going back to the 1920s (Ohmura and Reeh, 1991). In this compilation, 251 measurements of accumulation rate were used to derive a map of precipitation for the whole of Greenland. Since the 1990s, the Program for Arctic Regional Climate Assessment (PARCA), a NASA-funded, multi-decadal, coordinated project, has been supplementing these records and collecting many other types of data with the aim of improving estimates of
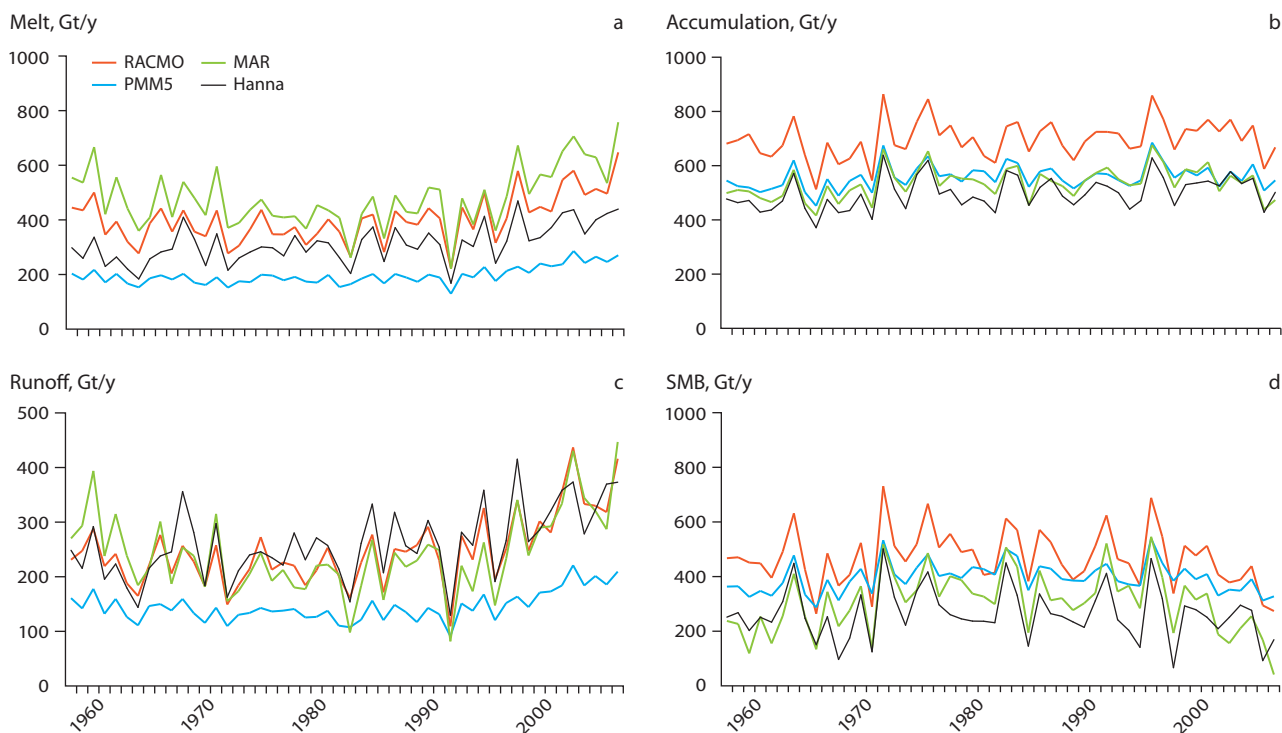

Figure 8.9. Time series of surface mass balance components from four different models: (a) melt, (b) accumulation, (c) runoff, (d) SMB. Note that the y-axis scales are not the same for each plot. Source: Jonathan Bamber, University of Bristol.

a wide range of key parameters for the ice sheet (Abdalati, 2001). Among the measurements taken were over 100 shallow ice core estimates of accumulation. These were combined with a subset of the historical records to produce the most comprehensive observation-based map of accumulation across the ice sheet to date. The estimated mean error for this compilation was 24% with regional variations of 15–30% (Bales et al., 2001). This compilation has recently been updated with a number of new cores and coastal station data (Figure 8.10) (Bales et al., 2009). The recent update resulted in regional changes, compared with the earlier study, of as much as ±50%, suggesting that uncertainties in accumulation rates were significantly underestimated in previous studies: a conclusion that is corroborated by the substantial differences in modeled accumulation shown in Figure 8.9.

The average accumulation rate of 300 kg/m^2 over the ice sheet agrees well with an earlier estimate obtained from different sources and covering a different time period (Ohmura and Reeh, 1991), and with a further estimate using a different interpolation approach (Cogley, 2004). Although there is agreement between the average values of the estimated accumulation, examination of the spatial distribution demonstrates marked and significant differences (Figure 8.8).
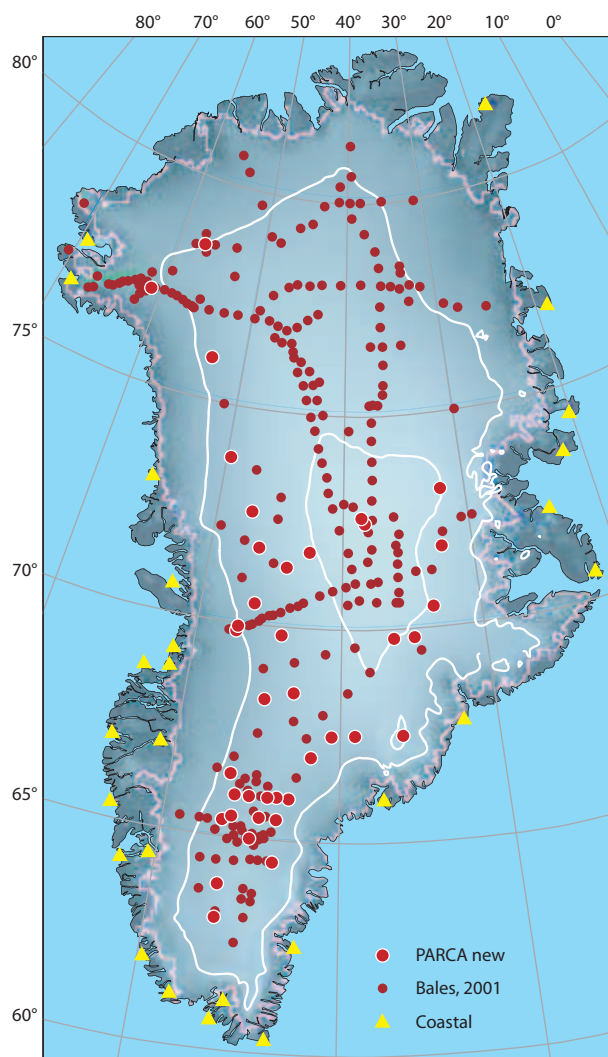

Figure 8.10. Locations of ice cores (red symbols) and coastal meteorological station (yellow symbols) data used to reconstruct accumulation rates over the ice sheet. Source: Bales et al. (2009).

An error analysis of *in situ* snowpit and ice core data suggested an uncertainty of 7.7% in the averaged 30-year mean. This is in agreement with the error estimate from the most recent study (Bales et al., 2009). It does not, however, reflect the interannual and spatial uncertainties which are very much greater (Bales et al., 2009). Uncertainties in the spatial pattern can have a significant impact on modeled estimates of ablation due to the influence of snowfall on surface albedo and refreezing (see discussion in Section 8.2.2.3). In addition, rain is a small but important term of the accumulation rate. Although its volume is low (~4% of total precipitation), its impact can be important because upon refreezing it raises the heat content of the snowpack, and in doing so increases total melt and runoff volume. The increased percentage of precipitation falling as rain in a warming climate will have a stronger influence on runoff and, hence, SMB.

8.2.2.2.2. Regional climate models

One big advance in assessing the SMB of the Greenland Ice Sheet, its temporal evolution and the relative importance of different components and their sensitivity to climate has been the development of RCMs. The first high-resolution RCM for Greenland was based on a modified version of the fifth generation mesoscale model known as PMM5 (Polar MM5; Box et al., 2004). This model ran at 24 km resolution and was calibrated using *in situ* data to correct for biases in melt energy and water vapor fluxes (Box et al., 2004). It was run, initially, for the period 1957 to 2004 forced by ERA-40 data (Box et al., 2006) (see Table 8.2 and Figure 8.9).

More recently, a 25 km resolution RCM was developed by Gallée and Schayes (1994) and applied to the Greenland Ice Sheet by Fettweis (2007) for the period 1958 to 2006. Unlike PMM5, the MAR (Modèle Atmosphérique Régional) has not been calibrated or corrected using *in situ* data. Comparison of modeled water vapor fluxes from MAR with automatic weather station data suggests that TMT is underestimated in MAR (Fettweis, 2007). Precipitation and runoff, however, are broadly comparable with other estimates from modeling and observations. The individual terms that comprise the runoff (M and RF) are both much greater than, for example, those of Hanna et al. (2008) (Table 8.2, Figure 8.9).

The third, a 50-year RCM reconstruction driven by ERA-40 data, is known as RACMO-2/GR. This is an adapted version of the ECMWF regional forecast model and was initially developed to run over Antarctica (van de Berg et al., 2006). More recently it has been coupled to a snow metamorphism and melt model (Bougamont et al., 2005) and run at 11 km resolution over a domain covering the whole of Greenland (Ettema et al., 2009). Like MAR, it was not calibrated using any *in situ* observations. The total accumulation from the RACMO-2/GR RCM is 7–24%, higher than the estimates from the other RCMs (Figure 8.9c). The greatest differences between the accumulation estimates from the RCMs are found in the wet southeast of Greenland where precipitation is estimated to exceed 4 m w.e /y locally. This unusually high value is consistent with a single ice core obtained from this region (Ettema et al., 2009).

It should be noted that RCMs are the only approach where all the components of SMB can be separated. *In situ* data provide the net SMB: that is, precipitation minus TMT terms

and runoff. Also, measuring melt volume and refreezing, in the field, is challenging (Bøggild et al., 2005). The models suggest that TMT is generally negative, indicating net sublimation/evaporation, with a magnitude of about 10% of P, except in MAR, where it is an order of magnitude smaller.

8.2.2.3. Ablation

8.2.2.3.1. Observations

Direct observations of ablation rates around the margins of the Greenland Ice Sheet are sparse both in space and time (Figure 8.11). Ablation has been measured using stakes at a number of sites in Greenland (Braithwaite et al., 1992) and, more recently, a technique has been developed for making automated measurements (Bøggild et al., 2004a). However, the spatial coverage is far from representative. One of the more comprehensive and uninterrupted *in situ* data sets has been collected since 1990 by Utrecht University along a 150-km transect near Kangerlussuaq, central western Greenland, known as the K-transect (Greuell et al., 2001; van de Wal et al., 2005). Stake measurements of mass balance have been collected at eight sites along the K-transect since 1990 and automatic weather station data are available since 1997 at one site and since 2003 at two further sites.

This represents the longest existing continuous record of *in situ* mass balance and ablation data for the Greenland Ice Sheet, but how representative the data are of other parts of the ablation zone is uncertain. This region of the ablation zone is, generally, smoother and less rugged compared with areas to the north and south. Small-scale topography affects the microclimate and therefore local ablation variability. A modeled time series used to assess the spatial autocorrelation of SMB showed little autocorrelation in the ablation zone in the southeast, a negative correlation in the northeast, and a strong positive correlation along the western margins suggesting that ablation depends on local conditions (as would be expected) as well as larger-scale climate forcing (Box et al., 2006). For example, in northeastern Greenland at the Storstrømmen glacier at 77° N, the observed large spatial variability in ablation is mainly controlled by the katabatic windfield producing a highly uneven winter snow cover distribution and higher melt rates locally at convex surface undulations.

Although the *in situ* data cannot be used to provide direct estimates of ice-sheet integrated ablation or the individual ablation components, they can be used to validate and calibrate satellite and model-based estimates. The spatial coverage of the quite sparse observational points, however, is currently not adequate to provide representative sampling to validate and calibrate satellite and model-based estimates of ablation.

The runoff component of the ablation has also been estimated using calibrated satellite observations of passive microwave melt extent and duration (Mote, 2003) and satellite-derived albedo data (Greuell and Oerlemans, 2005). Both approaches rely on an empirically derived relationship between melt duration or albedo and runoff. Estimating runoff from routinely acquired satellite observations is a promising technique but is empirically based and it remains to be shown whether the relationships derived are robust under current and future (changing) climate conditions. Recent results have
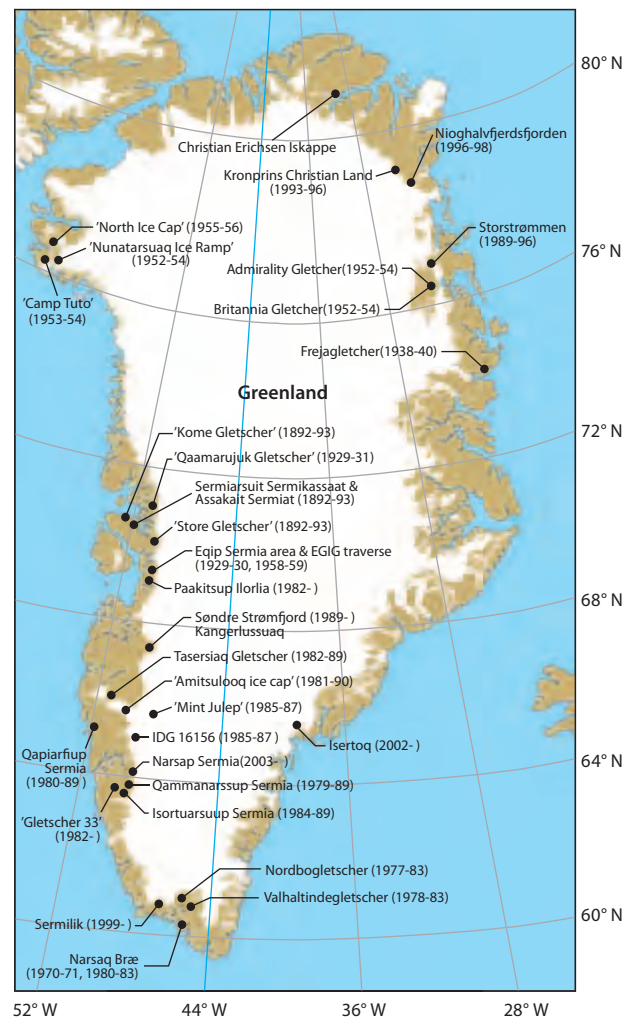
Figure 8.11. Location of ice sheet ablation measurements up to 2004. Source: Bøggild et al. (2004b).

shown, for example, that for areas in Antarctica where melt occurs, passive microwave-derived values underestimate SMB (Magand et al., 2008).

In addition to estimating runoff, passive microwave data have also been used to measure the area of the Greenland Ice Sheet subject to surface melt (Abdalati and Steffen, 2001; Mote, 2007). Results indicate that the melt area has been steadily increasing since satellite observations began (Figure 8.12). Although this is not a direct measure of runoff (because it does not estimate the volume of melt or refreezing) there is a good correlation between melt area, number of melt days, and runoff (Mote, 2003).

In this approach, the satellite-based melt duration data are used as a proxy for surface air temperature. Although this method has some merit, the development of a consistent, homogeneous temperature data set derived from ERA-40 and extending back to 1958, provides a longer time series with roughly the same correlation with runoff (Hanna et al., 2002, 2005). Its advantage, however, is that it represents a direct observation of changing surface properties as opposed to a model reconstruction and can, therefore, be used to corroborate the estimates derived from re-analysis data. In addition, it provides observational evidence for changes in the length of the melt season and its onset date (Figure 8.12).

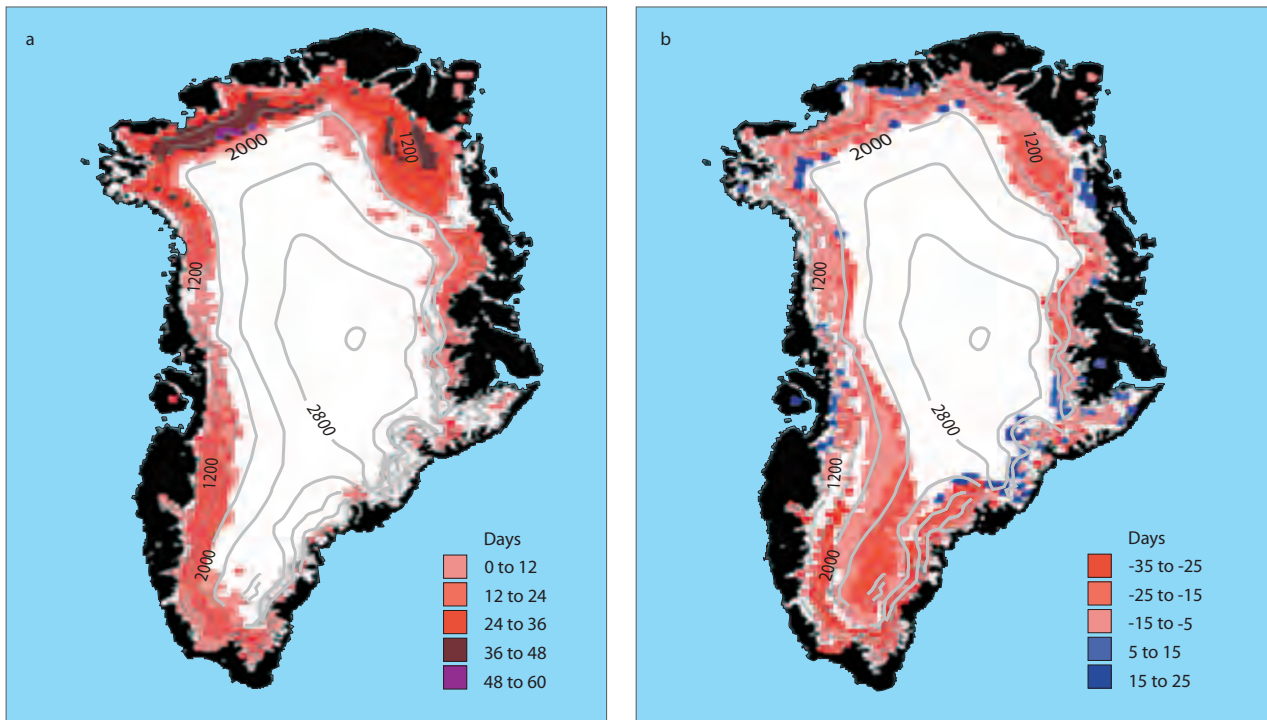Figure 8.12. (a) Departure of the 2008 melt frequency from the mean melt frequency for 1973–2008, and (b) departure of the date of melt onset in 2008 from the mean date of melt onset for 1973–2008. Elevation contours are shown in grey, and black indicates land area. Source: Mote (2007); T.L. Mote, pers. comm.

8.2.2.3.2. **Runoff modeling**

The first reasonable, quantitative attempts to model runoff from the Greenland Ice Sheet used a positive degree day (PDD) approach (Reeh, 1991). Here, the melt volume is related to the number of days where the near surface air temperature (usually taken at 2 m above the surface) exceeds 0 °C, and by how much. Melt volume is calculated by multiplying the positive temperature by a degree day factor (typically ~ 8 mm/°C per day for ice). This approach has been, and still is, used extensively for modeling the present-day runoff and SMB of the Greenland Ice Sheet (Braithwaite, 1995; Abdalati et al., 2001; Mote, 2003; Hanna et al., 2006; Rignot and Kanagaratnam, 2006) as well as its response in the future (Huybrechts et al., 2002, 2004).

The compelling advantage of the PDD approach is that it relies on a single, smoothly varying and easily interpolated parameter, i.e., near surface air temperature. Climatological maps of this parameter have been produced from a combination of coastal station and inland automatic weather station data (e.g., Ohmura, 1987) and more recently downscaled from ERA-40 (Hanna et al., 2002, 2005). The weakness of the approach is that it relies on an empirically based relationship, calibrated using a limited range of data for present-day conditions only. As a result, it is unclear how well a PDD model can predict melt in the future (Bougamont et al., 2007). For example, a PDD model has no sensitivity to changes in cloud cover or type unless they impact near-surface air temperature. Modifications to the approach have been suggested to incorporate more of the physical processes involved (Pellicciotti et al., 2005) but remain, fundamentally, empirical.

A more physically rigorous approach is to calculate the energy balance at the surface and, from this, the quantity of energy available for melting. Most recently, energy balance models have been used to estimate melt volume over Greenland by downscaling meteorological hindcast data such as ERA-40 (Box et al., 2004, 2006; Fettweis et al., 2005). Generally, these types of model incorporate all the key processes that affect melt but, as a result, require a large number of meteorological inputs that may not be constrained particularly well. For example, to estimate the turbulent, sensible and latent heat fluxes, requires prescribing a surface roughness as well as atmospheric stability corrections. It also requires near-surface humidity and the vertical wind profile in the boundary layer, all of which may not be particularly well known.

The single most important surface property for controlling runoff in the energy balance, however, is surface albedo, $\propto$. This can be estimated by the model based on snow age and/or density (Greuell and Konzelmann, 1994; Konzelmann et al., 1994) or prescribed from satellite observations (Stroeve, 2001). Model-based estimates, however, are generally only reliable for snow. Ice in the ablation zone can have an albedo of 0.4 or less (see Figure 8.7), which is substantially less than that of snow. Ice albedo depends on aerosol content and this is highly spatially heterogeneous. The advantage of an energy balance model compared to a PDD approach is that it incorporates the physical processes responsible for melt explicitly and should, therefore, respond appropriately to changing climate forcing. To do this, however, requires the 'correct' forcing. For long integrations (e.g., multi-centennial) such input data typically do not exist or may not be reliable enough to justify the use of a more complex and more 'data hungry' energy balance model. Thus, although an energy balance model may provide a more physical representation of the processes, it will not always be the best tool to use.

Preconditioning of the snowpack and the firn is very important for subsequent melt and runoff. In the SMB time

series of Hanna et al. (2005), high runoff years (except 2003) are generally synchronous with low precipitation / accumulation and vice versa. More accumulation results in a higher albedo for a longer time, which, in turn, reduces absorbed energy available for melt; the available surface energy needs to melt any snow first before ice can melt in the ablation region. Also, higher volumes of meltwater are retained in the thicker snowpack, which tends to reduce net runoff. Thus, accurate reconstruction of the timing, pattern and volume of accumulation is important for reliable estimation of runoff.

Another important consideration in modeling runoff is refreezing of meltwater within the snowpack. A variety of approaches exist from the simplest, using a fixed fraction of the annual accumulation (Reeh, 1991) to full snow metamorphism models (Bougamont et al., 2005; Mernild et al., 2006; Fettweis, 2007). The latter contain explicit terms for the metamorphism of the snowpack as meltwater percolates into it but can, in general, be driven only by an energy balance model because they require the energy balance of the snowpack to be calculated, which is not possible using a simpler PDD approach. A study of three relatively simple refreezing schemes for the Greenland Ice Sheet found that they accounted for about 20% of the total melt production, produced similar estimates, and had reasonably similar sensitivities to changing climate (Janssens and Huybrechts, 2000).

More recent analyses incorporating these simple approaches as well as snow metamorphism models reached a different conclusion, however, and found that the different models produced different volumes of refreezing and had different sensitivities (Bougamont et al., 2007; Wright et al., 2007). The MAR RCM, for instance, estimates that refreezing is about 50% of the total melt production (Table 8.2). These studies suggest that further work on understanding and modeling refreezing is required. They also indicate that modeling refreezing remains a significant uncertainty in present-day and future estimates of SMB. A major hurdle for improving refreezing estimates is both the difficulty in obtaining reliable *in situ* validation data (Bøggild, 2000) and the difficulty of incorporating preferential water flow deep into the snowpack with the models, which is known to take place (Bøggild, 2000). Instead of refreezing in the cold snow, water channels into vertical flow fingers followed by horizontal conduits and eventually results in runoff (Bøggild, 2000).

8.2.2.4. Differences and trends in measured and modeled SMB

The average SMB for the four reconstructions (RACMO2/GR, MAR, PMM5, Hanna; see Table 8.2) covering the past 50 years is 354 Gt/y with a range of 206 Gt between the lowest and highest estimates. This range is 58% of the mean value and provides some measure of the uncertainty in the SMB. This compares with a mean estimate of precipitation of 599 Gt/y. Thus, roughly half the precipitation is lost as runoff during this period, although estimates vary substantially (Table 8.2). It should be noted that runoff is not equal to the total volume of meltwater, M, because a proportion of this percolates into the underlying snowpack and refreezes. Considerable uncertainty exists in estimating this refreezing term, RF (see Section 8.2.2.4).

Table 8.2 summarizes some recent estimates of Greenland SMB, derived from a variety of methods. It is important to note that the values presented in Table 8.2 are not entirely independent. The models labeled Mote and PMM5 derived accumulation rates from almost identical downscaling approaches. Hanna, MAR, RACMO2/GR and PMM5 all use the same source: ECMWF climate re-analysis (known as ERA-40) and operational analysis data as the input for their downscaling approach. Thus, biases in ERA-40 may affect all of these estimates in similar ways.

Estimates of SMB from downscaling of ERA-40 data agree reasonably well when averaged over several years but there are substantial differences in spatial patterns, individual components of SMB, and interannual variability. Figure 8.9 compares the components of SMB from several different approaches: A simple downscaling of ERA-40 derived by Hanna et al. (2005) and the RCMs, MAR (Fettweis, 2007), RACMO2/GR (Ettema et al., 2009) and PMM5 (Box et al., 2006). It is not surprising that the trends are in agreement and that there is reasonable correlation between minima and maxima as this is largely constrained by the common forcing used in the simulations (i.e., ERA-40). There are, however, substantial departures in absolute values. For 2003, for example, the Fettweis (2007) SMB anomaly is about 180 Gt less than Hanna and PMM5, while RACM2/GR is consistently even higher due to its higher estimate of accumulation (Figure 8.9d). These differences are about as large as the SMB for that year.

The difference between the estimates is thus larger than the loss of mass from the ice sheet over the past decade (Section 8.2.5).

In addition, the refreezing component, RF, is highly uncertain. The range for RF between the four time series that use ERA-40 data as input is from 39 to 238 Gt/y: a difference that is 58% of the mean SMB. Refreezing is, therefore, an important, yet poorly known, component of SMB, which has received relatively little attention compared with other terms, partly due to the difficulty in making direct measurements of its magnitude. Although the fundamental physical processes that influence SMB are, in the main, relatively well understood, Table 8.2 and Figure 8.9 clearly show that much uncertainty still remains in constraining it.

Figure 8.13 summarizes the main trend in SMB over the past 50 years. Hanna et al. (2008) estimated a statistically significant
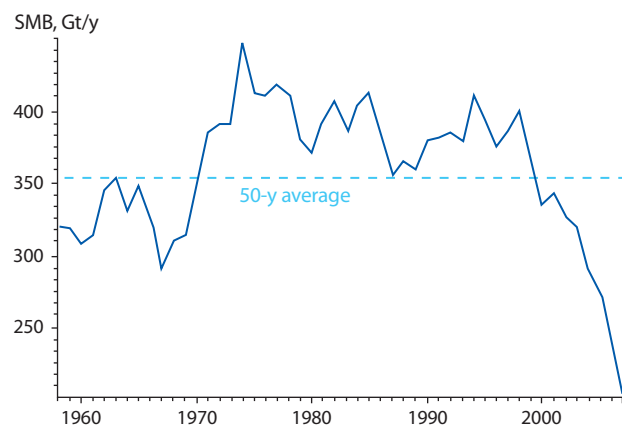

Figure 8.13. Mean SMB of the three reconstructions (MAR, PMM5, Hanna) shown in Figure 8.9. The data have been smoothed over a running 5-year period to reduce interannual variation. Source: Jonathan Bamber, University of Bristol.

positive trend in precipitation from 1958 to 2006 of 84 Gt/y, equivalent to 1.7 Gt/y². Runoff, however, has increased at a greater rate, particularly since the early 1990s (Figure 8.9) resulting in a net reduction in SMB during this time (Figure 8.13) as expected in a warming climate. The modest increase in precipitation is supported by *in situ* measurements from an ice core in North East Greenland (Figure 8.14), which show an increase since about 1990 once interannual variability is reduced by smoothing. However, observational data from one site is not enough owing to climate variability and significant noise in the data (Fisher, 1985; Andersen et al., 2006b). There are, however, few studies extending to the most recent years and the data illustrate the need for many and long records before changes in accumulation can be statistically verified. The interannual variability in precipitation from the RCMs is around 20% and, as can be seen in Figure 8.9, means that short-term signals over, for example, the past decade are strongly influenced by this variability.

The SMB has a 50-year mean average value of 354 Gt/y. During the 1960s and early 1970s, SMB was 10–20% below the 50-year mean. Since the mid-1990s, however, SMB has decreased steadily to its lowest value over the entire reconstructed period, such that the most recent years are more than 100 Gt below the 50-year mean and 200 Gt below the value in the mid-1990s. This is the most striking trend in the time series and has been linked to a global, rather than regional, warming trend that exceeds the background natural variability (Hanna et al., 2008). This is an important conclusion, which suggests that the trend is likely to continue into the future. Also, crucially, the increased melt during this period has warmed the surviving snowpack at lower elevations by some 5 to 10 °C, priming it for even greater losses in the near future (van den Broeke et al., 2009).

8.2.2.5. Summary and recommendations

The mean SMB, based on the four ~50-year records, is 354 Gt/y with a range between estimates that is 54% of the mean, providing a qualitative indication of the uncertainty in the SMB. This is of the order of magnitude of the standard deviation of the interannual variability, which ranges from 62 Gt/y for PMM5 to 124 Gt/y for MAR. Since the mid-1990s, runoff has increased significantly with only a modest change in accumulation, resulting in a reduction in the SMB of around ~200 Gt over the past 13 years (Figure 8.13). This trend in SMB appears to be a response to global climate change and above the level expected due to natural variability (Hanna et al., 2008). It should be noted, however, that the gradient of the trend is dependent on the epoch, due to the large interannual variability discussed above. Also of note, is that 2007 had the lowest SMB of any year in the 50-year time series. Significant differences exist between the various estimates of the components of the SMB derived from numerical modeling and downscaling of climate re-analysis data. These differences can be as large as the actual SMB estimate for a given year and the standard deviation of the differences is of the order of 92 Gt. This is of the same order of magnitude as the inferred increase in mass loss due to changes in ice dynamics over the past decade (Rignot and Kanagaratnam, 2006). The differences suggest that mass budget calculations (where the solid ice flux is subtracted from the SMB to estimate the overall net mass balance of the ice sheet) may have been seriously hindered by uncertainties in the SMB in the past. The most recent mass budget calculation, however, showed good agreement with results from independent satellite observations, giving confidence that progress is being made in constraining the components of mass loss from the ice sheet (van den Broeke et al., 2009). Nonetheless, resolving and explaining the cause of the large differences between reconstructions remains an important goal in improving the ability to model past, and predict future, changes in SMB.

To achieve this goal will require a concerted effort to collect more targeted *in situ* SMB data, especially from the percolation and ablation zone, with which to calibrate and validate the models. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow and subgrid-scale effects will also be valuable.

Accumulation, m ice-equivalent

δ¹⁸O, ‰
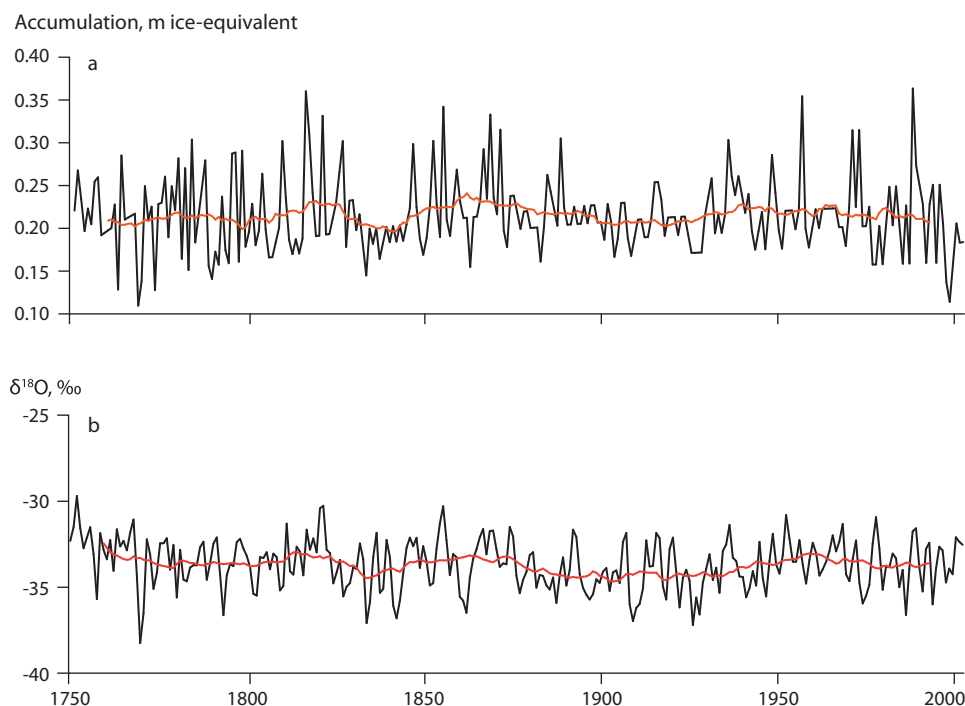
Figure 8.14. Results from a shallow ice core drilled at NEEM (77.5° N, 51° W) in 2007. The black curves represent annual values of accumulation (a) and stable water isotopes $\delta^{18}O$ (b) while the red curves represent 20-year sliding mean values for the period 1750 to 2003. The accumulation rates are determined from detailed $\delta^{18}O$ identification of annual layers based on 2.5 cm samples and a measured density record. Source: courtesy of the NEEM project.

8.2.3. **Ice discharge**

- Accelerating ice discharge from outlet glaciers since 1995–2002 is widespread and has gradually moved further northward along the west coast of Greenland with global warming.

- The ice discharge has increased from the pre-1990 value of 300 Gt/y to 400 Gt/y in 2005.

- The acceleration of fast-flowing marine terminating glaciers coincides with terminus retreat.

- The increase of glacier velocities is believed to be related to warming ocean water and the basal melt under the floating tongues can be several metres per day.

The Greenland Ice Sheet loses mass by melting as well as by discharge of ice into the ocean. The discharge comprises calving of icebergs as well as submarine melting. This section addresses the mechanisms of fast ice flow leading to transport of ice from the inland to the margin and the mechanisms of iceberg calving and submarine melting.

The margin of the Greenland Ice Sheet is characterized by spatially variable flow; areas of slow flow are separated by fast-flowing outlet glaciers and ice streams (Figure 8.15; Weidick, 1995; Rignot and Kanagaratnam, 2006). Parts of the ice are in direct contact with the surrounding ocean. In southern Greenland, ice-ocean contact is restricted to fast-flowing outlet glaciers that are topographically constrained by fjords, which often extend subglacially into the present-day ice sheet. The slower flowing ice usually terminates on dry land inland from the coast. However, in southwestern Greenland ice marginal lakes are quite common, and ice can be lost there by freshwater calving (Warren, 1991). The role of calving into ice marginal lakes has not been investigated extensively. It is likely to become increasingly important as the Greenland Ice Sheet retreats into the bedrock depression created by its own weight. In the north, along a significant section of the ice margin, both slowly and rapidly flowing ice is in direct contact with the ocean.

Around the Greenland Ice Sheet, the highest rates of ice discharge and the highest variability of discharge are observed at ocean-terminating outlet glaciers (Figure 8.15). The major mass loss mechanism for these glaciers is calving into the ocean (Reeh, 1968, 1994) and melting from ocean water (Rignot, 1996; Rignot et al., 1997). Rignot and Steffen (2008) showed that the floating ice shelf in front of Petermann Glacier, in northwestern Greenland, experiences massive bottom melting that removes 80% of its ice before calving into the Arctic Ocean. Commonly, estimates of ice discharge into the ocean implicitly include mechanical calving as well as melting by the ocean, since the ice contributes to sea level as soon as it crosses the grounding line and begins to float. Many of the outlet glaciers with high discharge rates flow through deep and narrow fjords. The largest and best studied of these is Jakobshavn Isbræ on the west coast of Greenland, for which the mechanisms of fast flow have been examined in detail (Iken et al., 1993; Funk et al., 1994; Lüthi et al., 2002, 2003). The findings are summarized here to illustrate the mechanism of fast outlet glacier flow.

Jakobshavn Isbræ is characterized by a deep subglacial trough that extends to 1500 m below sea level about 50 km inland from the current ice front (Clarke and Echelmeyer, 1996). This fjord (i.e., the trough) is relatively narrow (~ 5 km) but channels the ice from a drainage area that comprises
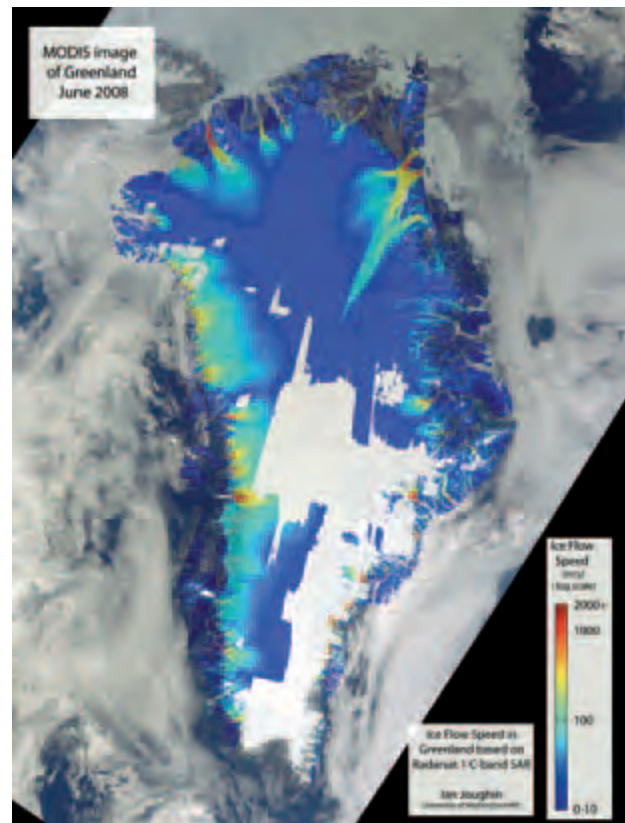

Figure 8.15. Velocity mosaic of the Greenland Ice Sheet derived from radar satellite data. The mosaic shows how the slow-flowing inland ice transitions into the fast-moving outlet glaciers, and the large variability of flow along the margin. White ice sheet areas are lacking radar satellite data. Source: Ian Joughin, University of Washington and Mark Fahnestock, University of New Hampshire, unpubl. (2008).

about 6.5% of the Greenland Ice Sheet (Bindschadler, 1984; Echelmeyer et al., 1991). Ice flow in such a channel is inherently complicated. The various commonly-used ice flow models that are simplified by approximations based on shallow ice (i.e., where ice thickness is much less than horizontal extent) are not adequate to fully describe the flow in this type of channel. Such models include the Shallow Ice Approximation (SIA, Hutter, 1983), and the Shallow Shelf Approximation (SSA, MacAyeal and Thomas, 1982), which can also be applied to shallow ice streams with low basal drag (MacAyeal, 1989). Instead, a fully 3-D thermo-mechanical treatment of ice flow, without approximation of the stress terms in the momentum balance, is necessary (Hutter, 1983) to describe outlet glaciers whose depths are comparable to their widths. No such model has yet been applied to either the Greenland or Antarctic ice sheets or even to an entire drainage basin of one of these ice sheets. (The hierarchy of ice flow models is discussed in Section 8.3.2.)

The relatively steep and very thick ice of Jakobshavn Isbræ creates high driving stresses that are balanced by high basal drag, particularly at the flanks of the subglacial trough (Lüthi et al., 2003; Truffer and Echelmeyer, 2003). The high driving stress influences the stress field far into the surrounding ice sheet. Using a numerical model, Lüthi et al. (2003) found differences to sheet flow at lateral distances from the ice stream of seven times the ice sheet thickness. The high stresses lead to high deformation rates, because they are combined with soft ice at depth, due to a temperate layer and lower viscosity Wisconsian ice (Funk et al., 1994; Lüthi et al., 2002; Truffer and Echelmeyer, 2003).

The presence of this temperate ice adds an additional modeling challenge, because it introduces a free boundary: the cold-temperate surface (Hutter, 1982) and it requires the treatment of waterflow in the temperate layer (Aschwanden and Blatter, 2009). Also, the existence of a temperate basal layer implies water at the bed and a certain amount of basal motion. This is a boundary layer process that needs to be parameterized through a 'sliding law'. The most commonly used sliding laws (see also Section 8.3.2) relate the basal motion to the basal shear stress and the subglacial water pressure; the latter is generally unknown but thought to be relatively high, that is, comparable to the ice overburden pressure.

This type of sliding law has well-known limitations, which are due to spatially and temporally variable parameters and to the limited knowledge of spatial and temporal variability of basal water pressure, which is an important variable (e.g., Iken and Truffer, 1997). The present inability to predict the nature and the amount of basal motion under large changes

in geometry, water supply, and stresses is a major limitation to credibly predictive models of outlet glaciers.

Reeh (1994) used mass balance considerations to estimate a total ice discharge into the ocean of 316 Gt/y. This is higher than a subsequent estimate of 170 to 270 Gt/y, which is based on calving rate estimates from parameters such as ice thickness at the front and water depth (Biggs, 1999). Both estimates applied for some time before the 1990s. Reeh's (1994) estimate is very close to one by Rignot and Kanagaratnam (2006) for 1995, which was derived from measurements of ice surface velocity and thickness at flux gates, as well as an estimate of surface ablation downstream from these gates. Rignot and Kanagaratnam (2006) also estimated discharge for 2000 and 2005, which showed a progressive increase in discharge within just a few years (Figure 8.16). The various estimates are summarized in Table 8.3. It is clear that ice calving into the ocean is a significant ice loss mechanism for the Greenland Ice Sheet, and that it can be highly variable (Bamber et al., 2007).
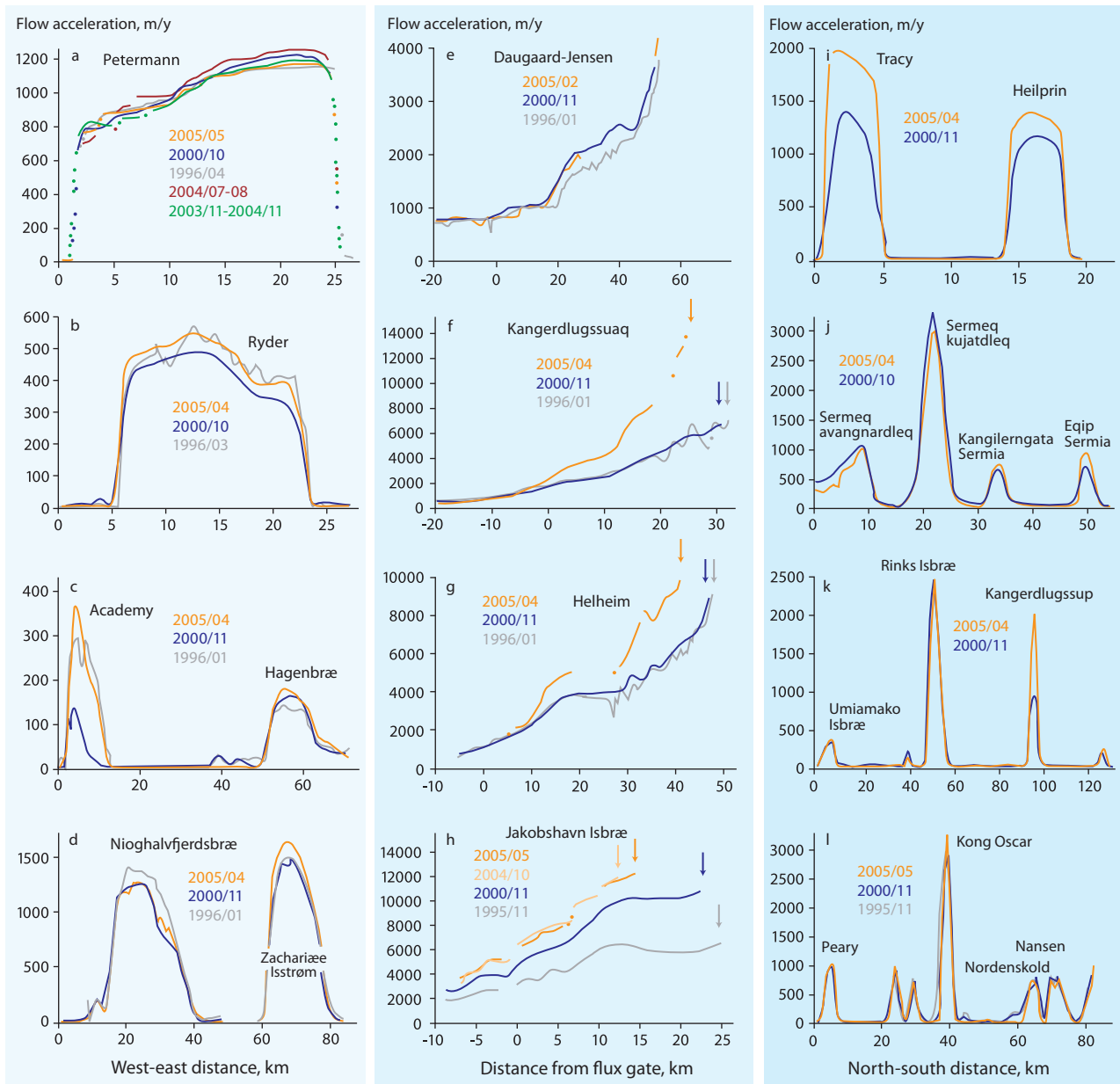

Figure 8.16. Flow acceleration (in m/y) at a selection of outlet glaciers. Cross profiles of velocity (a-d and i-l) were used for ice discharge calculations. e-h show along-flow velocity profiles. Note that acceleration is accompanied by glacier retreat. Source: Rignot and Kanagaratnam (2006).

Table 8.3. Calving flux estimates for the Greenland Ice Sheet.

Calving flux estimates, Gt/y	pre-1990	1995	2000	2005
Reeh (1994)[a]	316			
Biggs (1999)[b]	170-270			
Rignot and Kanagaratnam (2006)[c]		321	354	421

[a]Estimate is based on mass balance considerations; [b]estimate based on empirical calving rate relationships derived in Greenland and elsewhere; [c]estimate derived from measurements of surface velocity and ice thickness at flux gates and a correction for surface ablation downstream of these gates. The measurements have been converted to Gt/y assuming a density of 900 kg/m³.

Ice calving is a major factor accounting for the recent negative mass balance of the Greenland Ice Sheet.

The variability of ice discharge is confirmed by observations of individual outlet glaciers, which have shown dramatic changes in recent years. In the period 2002 to 2007, Jakobshavn Isbræ went through a period of thinning, retreat by loss of its floating tongue, and acceleration to twice its former speed (Thomas et al., 2003; Joughin et al., 2004, 2008c; Podlech and Weidick, 2004; Luckman and Murray, 2005). There is no evidence of similar increases in discharge in the 20th century at Jakobshavn Isbræ (Weidick and Bennike, 2007), but old moraines and glacier trim lines do indicate that the glacier has lost its floating tongue at least once since the Little Ice Age (15th to 19th centuries), while it was retreating to a position that remained stable for the next 50 years (Csatho et al., 2008). Large changes have also been observed at two big glaciers on Greenland's east coast: Kangerdlugssuaq Glacier and Helheim Glacier (Howat et al., 2005, 2007; Luckman et al., 2006; Stearns and Hamilton, 2007).

Rignot and Kanagaratnam (2006) showed that the acceleration of ice discharge is widespread. Between 1995 and 2000 many of the outlet glaciers south of 66° N accelerated, often doubling their discharge. This pattern spread to 70° N by 2005, and appears to have been related to regional warming. The detailed pattern is complicated, however. For example, Howat et al. (2007) reported that the Helheim and Kangerdlugssuaq glaciers decreased their discharge again to near former levels in 2006. This was partially due to a slow-down connected with a re-advance of the terminus, and partly due to thinning, so that at similar ice velocity less volume can be discharged. Other glaciers, such as Jakobshavn Isbræ, have maintained their high rates of ice discharge (Dietrich et al., 2007; Amundson et al., 2008; Joughin et al., 2008c).

The out-of-balance flux of ice into the ocean from many of the outlet glaciers in southern Greenland has led to large thinning of near-margin ice (Pritchard et al., 2009, Figure 8.17). This thinning is affecting large parts of the drainage basins (Stearns and Hamilton, 2007; Joughin et al., 2008c). For Jakobshavn Isbræ, the thinning is closely related to increased ice flow that can be accounted for by an inland-propagating change in ice surface slope (Joughin et al., 2008c).

Coincident with ice-flow acceleration is a widespread retreat of outlet glacier terminus positions (e.g., Figure 8.18). Moon and Joughin (2008) recorded terminus positions of 203 outlet glaciers and concluded that retreat accelerated during 2000 to 2006 compared to 1992 to 2000. Again, the detailed picture is
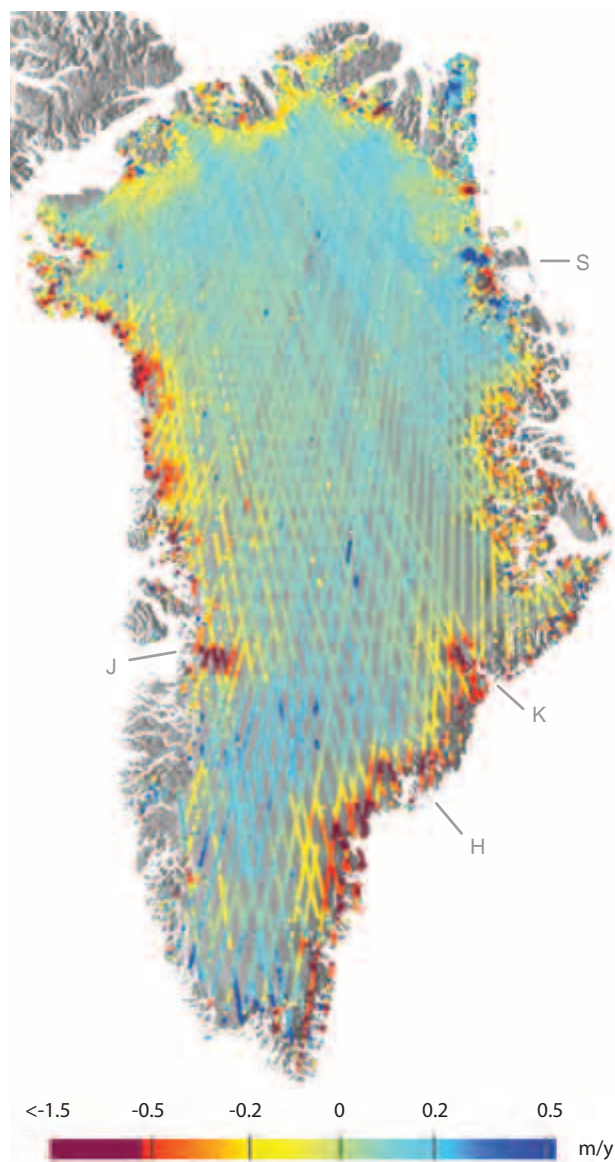
Figure 8.17. Thinning rates across the Greenland Ice Sheet from ICESat repeat laser measurements. Elevation changes over time intervals larger than 365 days in the period 2003 to 2007. The thinning is concentrated at the margin of the ice sheet. (Part of Figure 2, Pritchard et al., 2009).

more complicated and Moon and Joughin suggested that there is a close relationship between air temperature and ice front advance / retreat, without suggesting that there is a cause and effect relationship between the two.

The observed acceleration of outlet glaciers coincides with an increase in teleseismically discovered glacial earthquakes (Ekström et al., 2003, 2006). The earthquakes occur seasonally and are strongly related to the occurrence of large calving events (Joughin et al., 2008b). Ekström et al. (2003) and Tsai and Ekström (2007) attributed the seismicity to large slip events on outlet glaciers. Recent ice displacement measurements do not confirm this hypothesis (Amundson et al., 2008; Nettles et al., 2008), and the former authors favor a model in which the glacial earthquakes are generated by the scraping of large overturning icebergs. Only a few seismic studies have been carried out on the ice and they point to a rich variety of seismic sources, many of which are presently not well understood (Rial et al., 2009).

Various hypotheses have been proposed for the causes of the speed-up of outlet glaciers. Thomas (2004) and Johnson
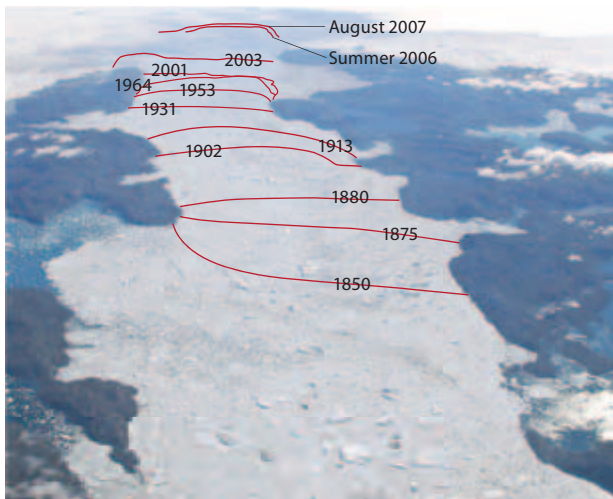
Figure 8.18. Changes in terminus position at Jakobshavn Isbræ (see Figure 8.35). Photo: Henning Thing.

et al. (2004) presented a model of Jakobshavn Isbræ in which the floating tongue provides resistance (i.e., backstress) to ice flow. The loss of the floating tongue then leads to ice flow acceleration. This model is also invoked by Joughin et al. (2008c) who observed a striking correlation between the speed of Jakobshavn Isbræ and the seasonally varying position of the calving front. Howat et al. (2005) used a force balance method to show how the loss of grounded ice near the terminus of Helheim Glacier leads to flow acceleration. The recently observed increases in ice discharge were always coincident with the loss of a floating tongue. But once the process is initiated, a pattern of fast flow, thinning and retreat can become established. This is a common observation at temperate tidewater glaciers, which only rarely develop floating tongues (Meier and Post, 1987). The idea of rapid tidewater glacier retreat is directly linked to that of a reverse bed slope, that is, the glacier bed deepens in the up-glacier direction. Nick et al. (2009) suggested that, in the case of Helheim Glacier, the phase of accelerated flow may be relatively short-lived, but it remains to be seen how universally applicable this model is.

While the exact form of a calving law remains unknown (e.g., Benn et al., 2007), it is clear from observations that a retreat into deeper water will lead to higher rates of calving, thinning, and further retreat. Vieli et al. (2001) and Schoof (2007) showed that no stable terminus position can exist on a reverse bedrock slope. This holds in advance as well as in retreat. The glacier acceleration that accompanies retreat can then be due to the effect of thinning, as suggested by Meier and Post (1987) and used by Pfeffer (2007).

Thinning influences glacier velocity in several ways. First, spatially-uniform thinning leads to a reduction in driving stress and thus in horizontal velocity. Second, thinner ice leads to a reduction in effective stress (ice overburden minus basal water pressure). This is because the basal water pressure has to exceed a level that is given by sea level for subglacial drainage to occur. As a tidewater glacier thins, basal water pressure cannot decrease due to this set downstream level, causing an overall drop in effective pressure. The reduction of effective pressure, as the lower part of the glacier approaches floatation, is expected to lead to higher rates of basal motion in any reasonable sliding law. Finally, enhanced thinning near the

terminus leads to increases in surface slope and hence driving stress and ice velocity.

Pfeffer (2007) derived a stability index for tidewater glaciers by considering the first and second of the three effects, which have opposite effects on ice flow. Pfeffer showed that an ice thickness above a certain threshold:

$$h_{th} = C \cdot \rho_w / \rho_i \, h_w$$

is stable against thinning in that thinning leads to a reduction in ice flow, whereas an ice thickness below that threshold reacts to thinning by ice acceleration. Here, h_w is the depth of the water and ρ_w and ρ_i are the densities of ocean water and glacier ice, respectively. C is a constant that depends on the choice of the ice flow law and the sliding law. Pfeffer (2007) suggested that C = 4/3, given flow parameters in the literature.

The analyses by Vieli et al. (2001), Pfeffer (2007) and Schoof (2007) all apply to flowline models and do not consider the embedding of an outlet glacier in its surrounding ice sheet. This could be an important factor contributing to the stability of outlet glaciers. As an outlet glacier retreats and thins, the surface slope increases in a lateral direction (i.e., toward the ice sheet), and more convergent flow should be expected, which could stabilize the glacier and delay further thinning and retreat (Howat et al., 2007).

Measurements from outside the fast-flowing ice in West Greenland have also indicated a seasonality in ice flow (Lüthi et al., 2002; Zwally et al., 2002b; Joughin et al., 2008b; van de Wal et al., 2008). This seasonality is related to the seasonality of surface melt and was attributed to meltwater reaching the base of the ice (Zwally et al., 2002b). Das et al. (2008) observed the draining of a supraglacial lake to the bed and a corresponding local speed-up of the ice sheet. They concluded that the integrated effect of multiple lake drainages could explain some of the observed acceleration of the ice sheet. However, for several reasons, it is unlikely that increased melt directly leads to the observed speed-up of outlet glaciers.

First, the ice-stream acceleration was first observed near the termini of these glaciers and then propagated inland (Joughin et al., 2004; Howat et al., 2007). Second, at least in the case of the main channel of Jakobshavn Isbræ, there is no observed distinct seasonal reaction to the large variability of meltwater supply. Rather, the seasonality of ice flow is closely related to the position of the glacier terminus, which varies seasonally, but with a distinctly different phase than the meltwater production (Joughin et al., 2008c). Third, van de Wal et al. (2008) did find a large seasonal velocity signal that correlates well with meltwater production, but does not translate into a correlation between annual velocities and melt. In fact, the mean annual velocity of the ice sheet has a negative correlation with melt. This has also been reported for mountain glaciers (e.g., Truffer et al., 2005) and is presumably due to the theoretically expected increase in effectiveness of water drainage as discharge increases.

Reeh et al. (2001) and Joughin et al. (2008c) observed a correlation between sea-ice cover and the stability of glacier fronts. While it is perhaps difficult to imagine a sea-ice cover that provides substantial resistance to glacier flow, it is possible that icebergs held together by sea ice inhibit calving, which then leads to glacier advance because ice is not lost at the terminus (Joughin et al., 2008c). This can then have an influence on ice velocities by the increased resistance due to lateral shear, and

perhaps by decreased effective pressures near the grounding line due to ice thickening. In practice, it is often difficult to tell whether the influence of sea ice is direct, or whether the sea ice reacts to the same forcing as the outlet glaciers, for example, warmer ocean water.

That all large observed changes in ice-stream flow initially occurred near the ocean and then propagated inland, points to the importance of understanding ice-ocean interactions. Holland et al. (2008) reported that the acceleration of Jakobshavn Isbræ is temporally coincident with a pulse of deep warm ocean water that was observed along the west coast of Greenland. Typical fjord circulation patterns that bring in warm saline water at depth and return colder and fresher water at the surface could bring this pulse to the glacier front and contribute to the very high melting rates that were reported on Jakobshavn Isbræ before the disintegration of its floating tongue (Thomas et al., 2003). Straneo et al. (2010) and Rignot et al. (2010) also pointed to the importance of warm water incursions for regulating glacial flow. Howat et al. (2008) noted a correlation between sea surface temperature and outlet glacier acceleration and retreat in southeastern Greenland. Murray et al. (2010) observed simultaneous acceleration and deceleration of many southeastern glaciers and proposed that increased ice and freshwater discharge led to a strengthening of a fresh coastal current that kept warm ocean water from entering glacial fjords, thus providing a negative and self-regulating feedback between ice discharge and melting by ocean water. Ocean conditions can also favor the formation of a strong melange (mix of sea ice and icebergs) that can inhibit calving and lead to a strong seasonality in terminus position (Amundson et al., 2010). The length of the winter season determines the amount of time a grounded terminus is exposed to calving. This provides a link between seasonality and longer term glacier behavior.

In a different setting, Motyka et al. (2003) showed that melting of submerged ice at a tidewater glacier front could account for about half of the total ice flux into the ocean. The important factors for melting were the temperature of the deep fjord water and the strength of the fjord circulation. Motyka et al. (2003) hypothesized that the buoyant freshwater emerging at the glacier's base plays an important role in the fjord water circulation. This provides an interesting physical link between atmospheric temperatures, surface melt, and fjord circulation.

The fast flow of Jakobshavn Isbræ and other outlet glaciers can be explained as a consequence of subglacial topography. This is not, however, necessarily the case for all outlet glaciers. One that stands out in particular is the northeastern Greenland ice stream (Fahnestock et al., 1993; Figure 8.15), which extends far into the ice sheet (Joughin et al., 2000). The fast flow of this ice stream might be controlled by geothermal heat. Fahnestock et al. (2001) used the spacing of deep radar layers to derive the geothermal heat flux under this ice stream and discovered unusually high melt rates.

Some of Greenland's glaciers also show surge-type behavior: long periods of quiescent flow interrupted by short periods of very rapid flow (Meier and Post, 1969). Glacier surges are a non-equilibrium phenomenon that is believed to be due to a positive feedback between basal water storage and basal motion (Kamb et al., 1985; Harrison and Post, 2003) that can

last several months. Slower surges at polythermal glaciers might also be thermally controlled (Clarke et al., 1984). In Greenland, surging has been reported for the northeastern ice stream Storstrømmen (Reeh et al., 1994) and several glaciers in central East Greenland (Murray et al., 2002; Jiskoot et al., 2003). A short surge-like flow acceleration was also seen at Ryder Glacier (Joughin et al., 1996). It is important to remember the temporary nature of such changes in ice discharge. The mechanisms (and perhaps duration) are quite different from outlet glacier changes.

8.2.4. Total mass balance

- The Greenland Ice Sheet is losing mass at an accelerating rate.

- The present (2005/06) loss of mass is 205 Gt/y corresponding to a 0.5 mm/y global average rise in sea level.

- Satellite and airborne observation in general reveal a slight increase in elevation in the regions above 2000 m and a strong decrease in elevation in the regions below 2000 m with very dominating regional patterns.

8.2.4.1. Total net balance of the Greenland Ice Sheet

The total mass balance of an ice sheet is the sum of the surface mass balance and ice discharge. It is the net result of the mass changes due to the addition and loss processes discussed in the previous sections. Owing to the seasonal nature of these processes, the mass of an ice sheet varies throughout the year, between years, and over many years as longer-term trends in climate become established. There are strong spatial gradients in mass balance that must be taken into account to determine variations in the total mass of the ice sheet.

For the Greenland Ice Sheet, which is subject to substantial surface melt and the resulting runoff of meltwater, mass exchange through the year has a large seasonal component at lower elevations caused by winter accumulation giving way to summer surface melt. These changes around the ice sheet margins are accompanied in outlet glaciers by changes in ice flow and iceberg production that may be linked in part to summer warming. At high latitudes and high elevations in the interior of the Greenland Ice Sheet, the relatively low input of snow and relatively small loss of snow to sublimation and wind transport lead to smaller amplitude annual variations in mass.

The time it takes for significant changes in accumulation or rates of ice loss around the margins of the Greenland Ice Sheet to impact on the ice flow in the bulk of the ice sheet interior determines the response of the ice sheet to forcings on ice-age time scales. Changes in the flow of the Greenland Ice Sheet over millennia are driven by slowly changing temperature profiles and ice characteristics. The rate of these changes is controlled by thermal diffusion, slow flow of ice downward and outward as new snow is added to the ice sheet surface, and the slow evolution of the shape of the ice sheet as changing accumulation and melt patterns add and remove mass.

Prior to recent measurements of changes in the rates of outlet glacier flow and melt in Greenland, the ice sheet appeared to be roughly in balance. The mass turnover in the Greenland Ice Sheet was estimated at 500 Gt/y by Benson (1962) with about half the mass added through snowfall each year lost by

surface melt and subsequent runoff, and the other half returned to the ocean through outlet glacier discharge leading to iceberg calving. But, as discussed in Sections 8.2.2 and 8.2.3, recent observations show quite large and rapid changes in surface melting and ice discharge. At the same time, different analyses of the total mass balance show the Greenland Ice Sheet overall to have lost mass since the early 1990s.

8.2.4.2. Techniques employed to observe or estimate total mass balance

Current techniques for estimating or measuring total mass balance include: (i) the mass budget approach, comparing total net snow accumulation with losses by ice discharge and meltwater runoff; (ii) repeated altimetry, to estimate volume changes; and (iii) tracking temporal changes in gravity, to infer mass changes.

The first two provide estimates of mass balance for regions included within survey boundaries, whereas the third provides an estimate for large regions with less distinct boundaries. All three techniques can be applied to the Greenland Ice Sheet as a whole, within limitations specific to each technique.

8.2.4.2.1. Mass budget approach

Changes in the Greenland Ice Sheet mass can be inferred from the difference between estimates of accumulation (i.e., snowfall) and mass loss through ablation (i.e., melt and subsequent runoff and sublimation) or ice discharge and calving. Each of these quantities involves a number of variables. Generally, snow accumulation is estimated from annual layering in ice cores, sometimes with interpolation between core sites using satellite microwave measurements or shallow radar sounding (e.g., Rotschky et al., 2004), or from atmospheric modeling (e.g., Box et al., 2004). Ice discharge is the product of ice flow velocity and ice thickness, with velocities measured *in situ* or remotely, preferably near the grounding line where velocity is almost depth independent. Ice thickness is measured by airborne radar, seismically, or from measured surface elevations assuming hydrostatic equilibrium for floating ice near grounding lines. Generally, meltwater runoff (large near the Greenland coast but small or zero elsewhere) is from model estimates calibrated against surface observations where available (e.g., Box et al., 2004; Hanna et al., 2005). Rignot and Kanagaratnam (2006) applied the mass balance approach around Greenland.

Mass budget calculations involve the often small difference of two large numbers, and small errors in either mass loss or gain terms can result in large errors in estimated total mass balance. These errors are difficult to assess for the Greenland Ice Sheet because of high temporal and spatial variability (see Section 8.2.2). Although broad interferometric SAR (InSAR) coverage and progressively improving estimates of grounding line ice thickness have substantially improved ice discharge estimates, incomplete data coverage and uncertainty regarding velocity-depth relationships and meltwater runoff limit accuracy of total discharge estimates. The mass budget uncertainty for the Greenland Ice Sheet is a result of all these factors.

Moreover, accumulation estimates are based on data from the past few decades, and there are indications that snowfall in Greenland may be increasing with time (e.g., Hanna et al., 2005). Similarly, it is becoming clear that glacier velocities can change substantially over quite short time periods (see Section 8.2.3). Both of these would add to estimated mass budget errors.

Thomas et al. (2000) gave a regional estimate for the mass budget for central Greenland above 2000 m elevation, using discharge velocities from repeat GPS measurements made at stake markers 30 to 40 km apart along a traverse at 2000 to 3000 m elevations, ice thickness from radar sounding, and snow accumulation rates from shallow ice cores. Results show that between the mid-1970s and mid-1990s, central Greenland was in overall mass balance, but with regions of thickening, particularly in the southwest, and of thinning, particularly in the southeast.

A more complete picture, including near-coastal regions and individual drainage basins, comes from balancing net SMB (Section 8.2.2) against ice discharge rates derived from SAR-derived ice velocities and ice thicknesses from radar or surface elevations near grounding lines, where the ice is floating (Rignot and Kanagaratnam, 2006). SMB for this work was derived from SMB anomalies calculated from Hanna et al. (2005) scaled and applied to modulate a surface accumulation pattern from shallow ice cores combined with ECMWF (re) analyses (Section 8.2.2).

Results show significant mass loss since the mid-1990s from most drainage basins, with very large losses after 2000 from Jakobshavn Isbræ, Kangerdlugssuaq Glacier and Helheim Glacier and from nearly all glaciers along the east coast south of Helheim (Joughin et al., 2004; Rignot and Kanagaratnam, 2006; Luckman et al., 2006; Howat et al., 2007, 2008). Velocity time series for these glaciers show up to a 100% increase after 2000, consistent with their negative mass balance.

8.2.4.2.2. Repeat altimetry

Rates of surface elevation change ($\delta S / \delta t$) reveal changes in ice sheet mass after correction for changes in depth / density profiles and bedrock elevation, or for hydrostatic equilibrium if the ice is floating. Satellite radar altimetry (SRALT) data have been widely used (e.g., Davis et al., 2000; Johannessen et al., 2005; Zwally et al., 2005), together with laser altimetry from airplanes (Krabill et al., 2000), and from NASA's ICESat (Zwally et al., 2002a; Thomas et al., 2006). Modeled corrections for isostatic changes in bedrock elevation (e.g., Peltier, 2004) are small (i.e., a few mm/y) but with errors comparable to the correction. Those for near-surface snow density changes (Arthern and Wingham, 1998; Li and Zwally, 2004; Helsen et al., 2008) are larger (i.e., a few cm/y) and also uncertain.

SRALT data provide the longest time series, with the first measurements in 1978. Resulting estimates of $\delta S / \delta t$ can be misleading, however, because derived elevations refer to the radar reflection horizon. The depth of this horizon below the actual ice sheet surface is affected by near-surface characteristics, such as snow wetness. Consequently, changes in these characteristics can affect SRALT-derived estimates of $\delta S / \delta t$ (Thomas et al., 2008), particularly in Greenland where the area affected by summer melting has increased substantially since the early 1990s. In addition, because of the broad radar beam, SRALT data are unreliable over the sloping and undulating surfaces near the coast, where rates of elevation change are greatest (e.g., Brenner et al., 2007).

Airborne Topographic Mapper (ATM) and satellite (ICESat) laser altimeters provide data that are easier to validate and

interpret because footprints are small (i.e., about 1 m for airborne laser, and 60 m for ICESat) and there is negligible laser penetration into the ice. However, clouds limit data acquisition, and accuracy is affected by atmospheric conditions and particularly by laser pointing errors. Moreover, existing laser data are comparatively sparse, being limited by practical limitations on the number of ATM surveys and the ICESat orbit separation of up to 40 km in southern Greenland.

ATM elevation estimates are accurate to ~10 cm along survey tracks (Krabill et al., 2002) with similar errors estimated for ICESat surveys, but increasing over steeper slopes (Brenner et al., 2007). Because there are large gaps both in ICESat and airborne coverage, $\delta S/\delta t$ values are generally supplemented by degree-day estimates of anomalous melting (Krabill et al., 2000, 2004). This increases overall errors and probably underestimates total losses, because it does not take into full account the dynamic thinning of outlet glaciers.

In summary, $\delta S/\delta t$ errors cannot be quantified precisely for SRALT data, because of the broad radar beam and time-variable penetration. For laser data the difficulty in quantification comes because of sparse coverage. If the SRALT limitations are real, they are difficult, if not impossible, to resolve. Laser limitations result primarily from poor coverage, and can be resolved by increasing coverage to include all regions of rapid change around the Greenland coast. All altimetry mass balance estimates include additional uncertainties due to changes in the density profile of near-surface snow and (small) rates of basal uplift. If elevation changes are caused by recent changes in snowfall, the appropriate density (ρ) may be as low as 300 kg/m^3; for long-term changes, it may be as high as 900 kg/m^3. This is of most concern for small rates of surface elevation change, where the safest assumption is that density $\rho = 600 \pm 300$ kg/m^3, implying a $\pm 50\%$ mass balance uncertainty. Slobbe et al. (2009) used this conservatively large range for the entire ice sheet for an ICESat-only analysis, finding results consistent with other laser analyses.

Large and rapid changes, commonly found near the coast, are almost certainly caused by changes in melt rates or glacier dynamics, and the appropriate density of the lost or added volume approaches that of pure ice (ρ ~900 kg/m^3). Moreover, densification rates are sensitive to snow accumulation rates, temperature, and wetness, with warm conditions favoring more rapid densification (Arthern and Wingham, 1998; Li and Zwally, 2004; Helsen et al., 2008). Thus, the recent Greenland warming probably caused ice sheet surface lowering simply from this effect alone. Finally, the rate of basal uplift is inferred from models and has uncertain errors.

An early estimate of surface elevation change in southern Greenland using SRALT showed either little change or that much of the variability could be attributed to decadal changes in accumulation rates as measured in ice cores (Davis et al., 2000; McConnell et al., 2000). Other estimates of the Greenland Ice Sheet volume change from SRALT altimetry (Johannessen et al., 2005; Zwally et al., 2005) show thickening at higher elevations, particularly in the south, comparatively modest thinning nearer the coast, and a small positive mass balance overall.

However, as previously discussed, time-variable radar penetration and 'blurring' of near-coastal results by the wide radar beam make these results suspect (Thomas et al., 2008). Estimates derived by comparing ATM surveys 1993/1994 and

1998/1999 (Krabill et al., 2000) show substantial thinning near the coast and an overall mass loss of about 55 Gt/y between surveys. Comparison of elevations from these surveys with later values from ICESat at locations where flight tracks crossed ICESat orbits show that mass losses have increased to about 80 Gt/y between 1998/1999 and 2004 (Thomas et al., 2006). These laser-based estimates probably represent lower bounds because the sparse coverage 'misses' outlet glaciers that are thinning dynamically. Annual thinning rates of tens of metres on outlet glaciers in southeastern Greenland are shown clearly from repeat stereo satellite coverage (Howat et al., 2007, 2008; Stearns and Hamilton, 2007).

8.2.4.2.3. **Temporal variations in Earth's gravity field**

Of all the techniques that are used to monitor the changing mass of the Greenland Ice Sheet, only gravity measurements are directly sensitive to changes in that mass; all other techniques convert changes in other quantities, such as melt, ice flux or surface elevation change, into changes in mass. Since 2002, the GRACE (Gravity Recovery and Climate Experiment) mission, joint between NASA and Deutsche Forschungsanstalt für Luft und Raumfahrt (DLR), has measured Earth's gravity field and its temporal variability. GRACE comprises two satellites, following each other in the same orbit 200 km apart, that carefully monitor the distance between themselves (Tapley et al., 2004a,b).

As the first satellite approaches a more massive object on the surface, it accelerates away from the trailing satellite, which then accelerates as it approaches the same object, closing the increased distance between the two orbiters. A detailed record of the changes in this inter-satellite distance over time provides a direct indication of changes in mass at the Earth's surface. Other accelerations are monitored with on-board accelerometers. GRACE has produced time series of these small changes in gravitational acceleration that start in 2002 and still continue (Tapley et al., 2004a,b). The spatial scale of the changes in gravity that GRACE can detect is determined by the height of the orbit, the spacing of the two satellites, and the accuracy with which the range rate between the two satellites can be determined.

After removing the effects of other loading, high latitude data, like those for Greenland, contain information on temporal changes in the mass distribution of the ice sheet and the underlying bedrock. Because of its high altitude, GRACE makes coarse-resolution measurements of the gravity field and its changes with time and the resulting mass balance estimates are also at coarse resolution – hundreds of kilometres. Despite this limitation, this technique has the advantage of covering the entire Greenland Ice Sheet, which is extremely difficult using other techniques. GRACE estimates include mass changes on the small ice caps and isolated glaciers that surround the Greenland Ice Sheet, regions not included in mass balance estimates from other techniques.

There are two different analysis schemes for GRACE data that have been applied to determining changes in the mass of the Greenland Ice Sheet. The first uses products derived by the GRACE mission that represent the global gravity field, expressed as a spherical harmonic expansion to a high order, differencing fields from each 30-day period to detect changes in

ice sheet mass over time (Velicogna and Wahr, 2005, 2006; Chen et al., 2006). This technique requires convolving the harmonic expansion of the global gravity field with a filter representing either part or all of the Greenland Ice Sheet to isolate the changes in gravity that are due to changing ice mass. The gravity fields produced by the GRACE team have been updated since the time of publication of these papers; results may change with this revised input data. Ramillien et al. (2006) used a similar technique, but began with a different input dataset, which is a version of the Earth's geoid (i.e., another representation of the gravity field) derived for 10-day periods by a group at Centre National d'Etudes Spatiales (CNES). Slobbe et al. (2009) used a common analysis scheme on gravity fields produced by four different centers, and reported a significant range of resulting net balance rates.

Corrections applied in these analyses include compensation for changes in global patterns of water storage on land, ocean mass, and atmospheric mass. In addition, a correction must be made for longer-term changes in crustal mass distribution due to glacial isostatic adjustment (GIA) (crustal uplift that is a continued response to past deglaciation, or post-glacial rebound). Each of these analyses uses different input data sets and modeling strategies to make the corrections. Chen et al. (2006) used an optimized filtering technique to fit mass change time series on a $1° \times 1°$ grid with high resolution aided by use of knowledge about where changes were occurring (i.e., in outlets near the ice margin identified by Rignot and Kanagaratnam, 2006). Ramillien et al. (2006) discussed the impacts of errors from each of these corrections. Barletta et al. (2008) conducted an extensive analysis of the impact of different factors on the GIA corrections to GRACE estimates in Greenland and Antarctica. They did make a somewhat simplified estimate of the rate of mass loss in Greenland from GRACE data to illustrate these impacts. This result is not included in Figure 8.19 or Table 8.4, but is similar to the others in the table.

Wu et al. (2010) have presented a different analysis also based on the glacial isostatic adjustment and the redistribution of ice and water masses. It is suggested that the GIA correction is large and negative (-0.72 ±0.006 mm/y) resulting in a mass loss of 104 Gt annually during the period 2003 to 2008 from the Greenland Ice Sheet; this mass loss is smaller than those published by other authors.

The second type of GRACE analysis scheme uses the range rate measurements between the two satellites directly, modeling the local changes in mass required to produce the observed changes in these rates during different overpasses of Greenland (Luthcke et al., 2006). This work also makes corrections for GIA using published paleo ice sheet reconstructions, but not the smaller effects from atmospheric and oceanic mass variations.

Error sources can include measurement uncertainty, leakage of gravity change signal both from regions surrounding the Greenland Ice Sheet, and for some analyses, globally, and causes of gravity changes other than ice sheet changes. Of these, the most serious are the gravity changes associated with vertical bedrock motion (i.e., GIA). These are inferred from models of deglaciation history and its impact on vertical crustal motion, primarily post-glacial crustal uplift; the errors are also modeled and remain highly uncertain. Results from all gravity-based analyses show mass loss over their periods of observation, with rates ranging from -100 Gt/y to more than -200 Gt/y. The wide

spread in results requires further investigation, but several studies using different analysis techniques show continued and accelerated change over the GRACE time period.

8.2.4.3. Synthesis and state-of-the-art

The results for each of the techniques are shown in Figure 8.19, plotted as boxes that encompass the length of the measurement period along the time axis, and the published range of the mass change rate over that interval. Data used for Figure 8.19 are summarized in Table 8.4.

With the exception of measurements from SRALT, estimates of the Greenland Ice Sheet total mass balance discussed here are negative, and are becoming more negative with time, with the caveat that many of these records are relatively short. The trend and timing are largely consistent with increased ice discharge from outlet glaciers around the edge of the Greenland Ice Sheet. Where these measurements do not agree within published errors, the sources of the discrepancies may be attributable to shortcomings in spatial sampling, complex sets of corrections, and input data that continue to evolve as satellite missions mature; limitations in measurement techniques are better characterized, and surface conditions on the Greenland Ice Sheet respond to a changing climate.

The complexity of measuring or estimating total balance has led to the development and exploitation of a number of measurement strategies. With respect to the Greenland Ice Sheet, these strategies have been developed and are maturing into a system that will one day be capable of tracking well constrained and internally consistent mass changes on a sub-annual basis. The extension of present measurement time series, combined with follow-on instruments and new airborne and satellite campaigns (e.g., Cryosat II), will help solve problems associated with limited coverage, slope and surface change induced error, and disparities in analysis of common data sets.

This represents a rapid evolution in capability that is presently matched by the rapid evolution of the Greenland Ice Sheet mass itself. One lesson that may be taken from the present state of knowledge is that multiple approaches and observations of mass input and mass loss processes have been required to make the progress that has been made to date.

8.2.5. Summary and outlook

Compared with processes that take place at the base of the ice sheet, those that influence SMB are relatively well understood and considerably easier to observe. Despite this, published reconstructions of the individual components that make up SMB differ significantly. The differences can be as large as the inferred increase in mass loss due to changes in ice dynamics over the past decade. The uncertainties in SMB arise, largely, as a result of the paucity of relevant, spatially extensive *in situ* observations. To reduce uncertainties to a level useful for mass budget calculations and reliable SMB predictions will require a concerted effort to collect more targeted *in situ* data, especially from the percolation and ablation zone. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow and sub-grid scale effects will also be needed. Although there are differences between reconstructions, they all indicate a negative trend over the past ~15 years. This is largely due to an increase in runoff since about 1995, resulting
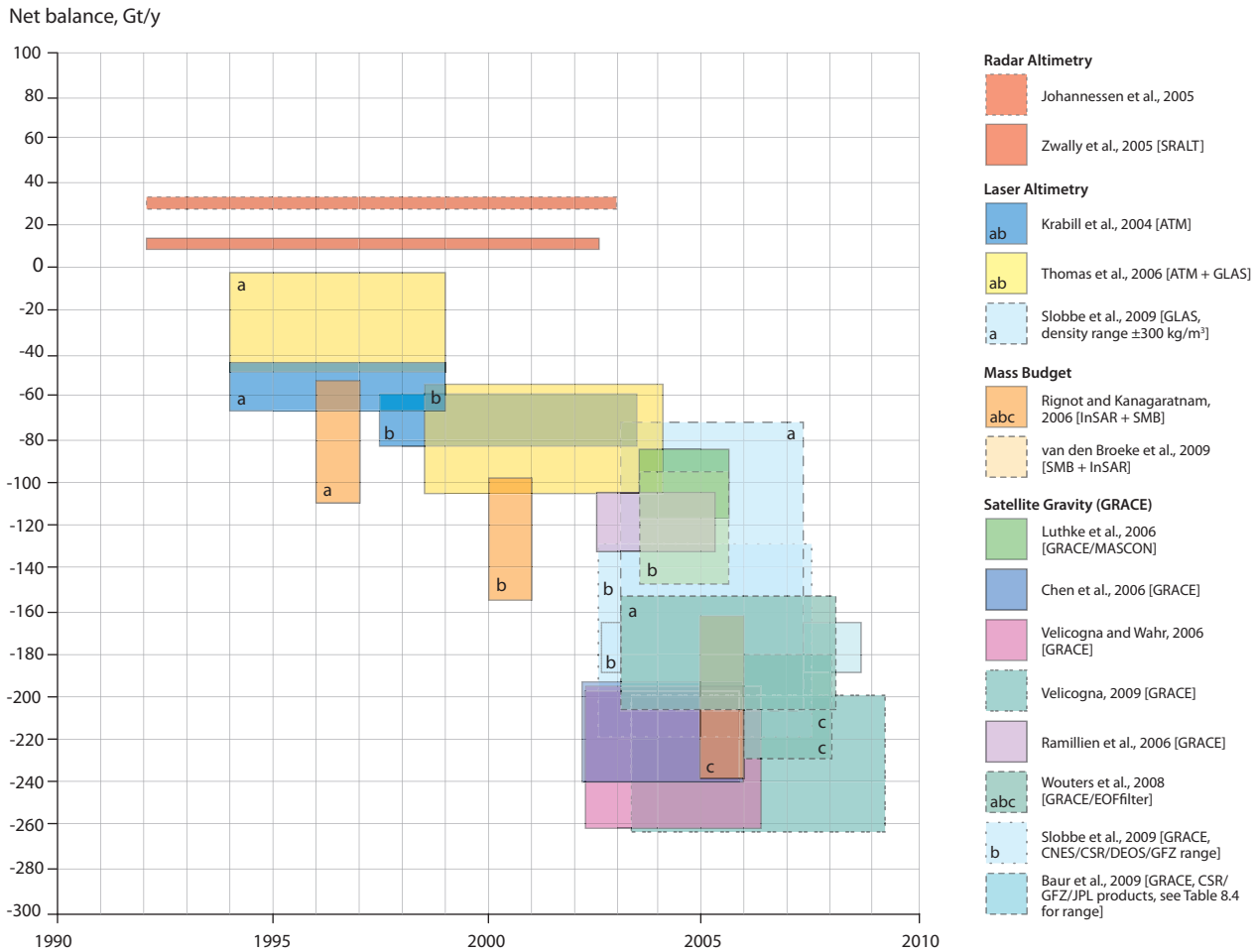
Net balance, Gt/y

Figure 8.19. Results from the Recent Large Area Total Balance Measurements discussed in the text, placed into common units and displayed versus the time intervals of the observations. The heights of the boxes cover the published error bars or ranges in mass change rate over those intervals. Source: Mark Fahnestock, University of New Hampshire. See Table 8.4 for details.

in a marked reduction in SMB since then. Also of note is that 2007 had the lowest SMB of any year over the 50-year record for which reliable estimates exist.

Much of the ice loss from the Greenland Ice Sheet occurs by ice discharge into the surrounding ocean. This has been estimated at 50% of the mass loss on average over the past ~50 years, but recently large variability in the flow of many large outlet glaciers has been documented. The temporal variability of Greenland's ice mass balance has a large contribution from variability of flow in its outlet glaciers. Observations during the past decade have shown that ice discharge can increase by a factor of two within a few years and, in some cases at least, that this can also be reversed. These changes are coincident with observations of warming in the ocean and atmosphere, as well as the disappearance of near-coastal sea ice. While the general mechanisms of fast flow are reasonably well understood, the necessary tools for predicting future behavior have not been developed. This is primarily due to a lack of understanding of the ice-ocean coupling and the role of surface water in determining the rates of ice flow.

The total mass balance of the ice sheet (i.e., the sum of SMB and ice discharge) was exceptionally difficult to determine prior to the development of the present suite of observational and analysis techniques. While it is clear from the spread of published results that estimation remains a challenge for

these techniques, it is important to note that nearly all the approaches show very similar trends, indicating clearly that the Greenland Ice Sheet is losing significant mass, and has been doing so at an accelerating rate over the past ten years. This conclusion is supported by all approaches sensitive to the ice margins: (i) the change in ice discharge from outlet glaciers combined with increasingly negative SMB; (ii) surface lowering in outlet glaciers measured directly by laser altimeters; and (iii) the reduction in the mass of the ice sheet measured by the GRACE satellite gravity mission. This reduction of mass is consistent with what would be expected to happen in a warming Arctic and the diverse, independent suite of approaches used to document this change in total mass balance provides confidence in the result.

The challenge of more accurately determining the present and future rate of mass change of the Greenland Ice Sheet, by reducing the uncertainties inherent in the different approaches, will require improvements in measurements, more consistent and complete observational time series, better analysis schemes, and an improved understanding of the physical processes involved in recent rapid changes.

This chapter has attempted to show the state-of-the-art with respect to observations and understanding of present-day, and recent past, behavior of the Greenland Ice Sheet. It is clear that, although the state-of-the-art has greatly advanced over the past

Table 8.4. Rate of total mass change in the Greenland Ice Sheet.

Source	Technique	Time period	Rate of total mass change, GT/y	Range (or stated error), ± Gt/y	Notes
	Radar Altimetry				Satellite radar altimetry is affected by surface slope near ice margins and changes in penetration
Zwally et al., 2005	SRALT (ERS 1-2) + limited ATM	1992 – 2002 (10.5 y)	11	± 3	
Johannessen et al., 2005	SRALT (ERS 1-2) low elevations sparsely sampled	1992 – 2003 (11 y)	30	± 3	Densities of 0.9, 0.5, 0.4, 0.3, 0.3 Gt/km³ assumed to convert area-weighted height change in elevation intervals (<1.5, 1.5–2, 2–2.5, 2.5–3, >3 km) into mass
	Laser Altimetry				
Krabill et al., 2004	ATM (airborne laser altimetry)	1993/4 – 1998/9	-55	± 3	
		1997 – 2003	-73	± 11	
Thomas et al., 2006	ATM	1993/4 – 1998/9	-4 to -50		High end of range likely due to limited coverage at low elevations
	ATM/ICESat	1998/9 – 2004	-57 to -105		High end of range likely due to limited coverage at low elevations
Slobbe et al., 2009	ICESat (GLAS)	2003/2 – 2007/4	-139 ± 68		Used large density range (± 300 kg/m³) for error bounds
	Mass Budget				
Rignot and Kanagaratnam, 2006	InSAR (ice motion) 1996 + balance anomaly estimate (Hanna et al., 2005)	1996	-83	± 28	
	InSAR (ice motion) 2000 balance anomaly estimate (Hanna et al., 2005)	2000	-127	± 28	
	InSAR (ice motion) 2005 + balance anomaly estimate (Hanna et al., 2005) (extrapolated)	2005	-205	± 38	
van den Broeke et al., 2009	Mass budget validated with data from GRACE.	2003–2008	-237	± 20	
	Satellite Gravity (GRACE)				
Luthcke et al., 2006	GRACE MASCON – (change in gravitational acceleration on each orbital pass fit to forward model of mass history of local basins)	July 2003 – July 2005	-101	± 16	Gain of 54 GT/y above 2000 m and loss of 155 GT/y below 2000 m
Chen et al., 2006	GRACE – averaging filter applied to monthly spherical harmonic gravity field product	April 2002 – Nov 2005	-219	± 21	
Velicogna and Wahr, 2006	GRACE – averaging filter applied to monthly spherical harmonic gravity field product	April 2002 – April 2006	-227	± 33	
Velicogna, 2009	Monthly measurements of time-variable gravity from the GRACE.	April 2002 – February 2009	-230	± 33	
Ramillien et al., 2006	GRACE – averaging filter applied to 10-day gravity solutions	July 2002 – March 2005	-118	± 14	
Wouters et al., 2008	GRACE EOF decomposition of monthly spherical harmonics iteratively fit with forward model of mass change of basins (based initially on R&K and iterated)	February 2003 – January 2008	-179	± 25	

Source	Technique	Time period	Rate of total mass change, GT/y	Range (or stated error), ± Gt/y	Notes
		July 2003 – July 2005	-121	± 27	Same period as Luthcke et al. (2006) Similar elevation dependence seen
		January 2006 – January 2008	-204	± 25	
Slobbe et al., 2009	GRACE using products from CNES (Centre National d'Etudes Spatiales), CSR (Center for Space Research, University of Texas-Austin), DEOS (Delft Institute of Earth Observation and Space Systems, Delft University of Technology, Delft), and GFZ (GeoForschungsZentrum, Helmholtz-Zentrum Potsdam)	July 2002 – June 2007	-128 to -218		Range in net balance is range of results using gravity products from four different centers, rather than estimated error
Baur et al., 2009	Analysing time-variable GRACE gravity fields	August 2002 – July 2008	-177	± 12	Value is the average derived from monthly gravity field models provided by CSR, GFZ and JPL, with individual contributions of 242 ± 14 km^3/y, 194 ± 24 km^3/y and 96 ± 23 km^3/y, respectively

decade or so, major gaps and challenges remain. During this period, unpredicted and striking variability in ice dynamics has been detected from satellite observations as well as a marked decrease in SMB coincident with a decreasing trend in total mass balance. Serious limitations in the ability to model and monitor these processes, however, are also apparent from the disparity between published results.

Progress will need to come from many sources, including:

- Improving acquisition of more comprehensive and complete time series suitable for ice velocity determination, including seasonal variability. This will be predominantly from satellite and airborne remote sensing.

- Acquisition of improved measurements of surface elevation change by satellite and aircraft.

- Better measurement of ice thickness in rapidly flowing glaciers for improved ice discharge estimation.

- Improvements in analysis strategies and measurement capabilities for all processes which determine SMB.

- More, and better, *in situ* measurements of refreezing and ablation.

- Better understanding of links between outlet glaciers, surface melt, and marine conditions.

- Perhaps, most importantly of all, the need to ensure that the time series of observations is secure into the future, and extended back in time with the aid of historic and paleo-proxy data.

8.3. **Predictions and sensitivity**

- The Greenland Ice Sheet is likely to lose mass at an accelerating rate in the 21st century in response to ongoing warming.

- Increases in snow accumulation in central Greenland may partially offset the increased melting and runoff from the ice sheet periphery, but are unlikely to compensate for the effects of higher temperatures, which increase the area of

the ablation zone as well as the duration and intensity of the summer melt season.

- Between 5 and 10 cm of global sea level rise can be expected by 2100 as a result of declining SMB. Adding estimates of increasing ice discharge means a total ice mass loss of 10 to 19 cm can be expected by 2100.

- Greenland Ice Sheet retreat is projected to accelerate beyond 2100 under most IPCC emissions scenarios, as the regional climate continues to warm and the dynamical feedbacks associated with surface lowering become more important over time scales of several centuries.

- A regional warming of 3 to 4.5 °C (equivalent to a global warming of 2 to 3 °C) would move the Greenland Ice Sheet into a state of negative SMB, which would cause large-scale ice sheet collapse if maintained for several centuries to millennia.

The Greenland Ice Sheet response to climate change is too complex to be predicted through simple theory or analytical equations. SMB depends on the interaction of weather systems with the ice sheet orographic influences and surface energy balance. Spatial and temporal patterns of ice sheet evolution therefore feed back onto the meteorological fields over the ice sheet. The regional- and synoptic-scale weather patterns that affect the ice sheet are also evolving as part of global-scale climate change and accompanying shifts in sea ice, ocean conditions, the hydrological cycle, and atmospheric dynamics. These systems are all interlinked and the interactions can be described only by numerical weather or climate models (Box 8.1).

Similarly, the evolution of the ice sheet depends on the climate fields that dictate SMB (see Section 8.2.2), on the oceanic influences that affect the rates of iceberg calving and basal melting at the ice-ocean interface, and on the dynamic ice sheet processes that govern the flux of ice to the ice sheet margin. Numerical models have been developed to describe

Box 8.1. Numerical models of the oceans, atmosphere and ice sheets

Numerical models of the oceans, atmosphere, and ice sheets take similar approaches to simulating the large-scale evolution of each system. All are based on the principles of mass, momentum, and energy conservation, with these conservation equations governing mass and energy transfer within or between these systems. The models dissemble the oceans, atmosphere, and ice sheets into 3-D grid-cell networks and solve the governing equations to predict the system state in each grid cell. Neighboring cells interact to simulate the fluxes of water, air, ice, energy, and climatically active variables such as salt and water vapor.

The whole system is driven by boundary conditions such as global topography and bathymetry, the Earth's rate of rotation, and temporal and spatial patterns of solar input. The resulting fluid motions and energy exchanges provide a representation of the climate system. Integrating the system forward in time, system interactions can give internal weather/climate variability or can simulate climate system response to a change in boundary conditions, for example, increases in solar activity or greenhouse gas concentrations.

The resolution of the simulation is determined by the size of the 3-D grid cells used to discretize the oceans, atmosphere, and ice sheets. Although computational limitations are continually receding, simulating climate dynamics still taxes the world's most advanced computational facilities. This places limits on the available simulation capabilities. For century-scale climate change scenarios, typical grid resolutions are 1° latitude-longitude for the oceans, 2° to 3° for the atmosphere, and 0.2° (~20 km) for ice sheets.

the ice sheet. However, some processes are poorly understood and unresolved in the models, particularly those associated with basal sliding of ice over the bedrock. Furthermore, SMB and ocean-ice interactions need to be prescribed to simulate ice sheet response to climate change. Although efforts to couple ice sheet, atmosphere, and ocean models to describe the co-evolution of these systems are still in early stages, this is the best available route for physically-based forecasts of how the Greenland Ice Sheet will change in the coming decades and centuries.

8.3.1. Aspects of modeling climate change for Greenland

The IPCC AR4 models incorporate many improvements compared to their predecessors, but there are still shortcomings (see Stroeve et al., 2007; as well as Chapter 2). While some studies suggest anthropogenic forcing may favor a positive Northern Annular Mode (NAM), there is evidence that climate models underestimate NAM-like variability (e.g., Stenchikov et al., 2006). Maslanik et al. (2007) analyzed three types of atmospheric circulation pattern that appear most significant in terms of Arctic Basin winds and ice transport. The 'light ice' phases of these patterns include decreased mean sea level pressure (SLP) in the North Atlantic (an 'NAO-like' pattern resembling the positive phase of the NAO), a low pressure cell within the Arctic Basin (a 'central Arctic' pattern), and a dipole pattern of high pressure over the Canadian Arctic paired with low pressure

over the Siberian Arctic. Winds and ice transport patterns that favor reduced ice cover in the western and central Arctic have continued since the late 1980s, but the Arctic Oscillation (AO) index is not a reliable indicator of these patterns.

Atmospheric modeling over complex terrain is a great challenge. The major reason for this is the difficulty represented by the topographical relief. If not properly resolved, the atmospheric circulation will not, for an otherwise perfect model, be able to depict the real local circulation patterns. This, in turn, will lead to systematic deficiencies in the models' ability to capture regional / local climatological features.

The most obvious example is that of topographically lifted precipitation that causes enhanced precipitation on the upstream slopes of a mountain, while producing less precipitation on the lee side downstream. Only when the appropriate mountain ridges are captured by the resolution, will the precipitation patterns be accurately simulated. The importance of this effect at larger scales is not well understood.

High resolution is offered by regional climate models (RCMs) (see Chapter 2). Results from RCMs are sensitive to the choice of integration domain, the horizontal and vertical resolution and parameterizations, and all contribute to the uncertainties of RCM simulations. There is no exact choice of domain any more than there is for resolution or parameterization, which means that domain must be treated as yet another element contributing to the uncertainty of RCM simulations. RCM simulations are also sensitive to lateral boundary conditions and to the surface conditions prescribed. The climate scenario projections come from interpolated atmosphere-ocean general circulation model (AOGCM) simulations and contribute to the overall uncertainties of RCM simulations and projections. The RCM results can vary depending on the AOGCMs used, in terms of large-scale atmospheric circulation patterns and their teleconnections. RCM results are often based on only one specific AOGCM and are for just one selected scenario. A multi-model ensemble approach could increase the confidence of future projections of extreme climate. Accurate simulation of surface mass accumulation over the ice sheets requires a spatial resolution which is currently not available from AOGCM simulations. Therefore high-resolution RCMs (i.e., PMM5 and MAR) have been run for shorter time slices driven by AOGCM boundary conditions and combined with an ice sheet mass balance model as described by Gregory and Huybrechts (2006) and Fettweis et al. (2008), see also Chapter 2.2.

One of the remaining challenges is that sequential modeling (i.e., climate modeling followed by ice sheet modeling) lacks any possible feedback between the climate and ice sheet. Thus, a poor climate model performance over the ice sheet will result in a possibly erroneous climate change signal over the ice sheet as well. This in turn will possibly result in an erroneous estimate of the change in SMB, even if the ice sheet model is excellent. A consistent development and coupling of models of atmospheric processes with ice sheet models is therefore in demand (see also Chapter 2).

According to the recent IPCC AR4 GCM studies, the Arctic is projected to be generally warmer and wetter (Kattsov et al., 2007). Using an RCM, Rinke and Dethloff (2008) projected future changes over the Arctic by the end of the 21st century according to the IPCC A1B emissions scenario. The results showed distinct regional patterns. Over the Arctic Ocean

these are mainly associated with sea-ice retreat and over land are mainly associated with surface characteristics, such as orography, and the agreement with the large-scale temperature and precipitation projections within the driving GCM is striking. Regional detail is evident along the North Atlantic storm track and, during summer, close to orographic obstacles. This is attributed to the higher resolution of the RCM.

Modeling sea ice in a realistic manner is still a great challenge, particularly with respect to the minimum ice extent at the end of summer. Dorn et al. (2007, 2008, 2009) investigated modified descriptions of ice growth, snow and ice albedo, and snow cover on ice used in the coupled regional atmosphere-ocean-ice model HIRHAM-NAOSIM. A series of sensitivity experiments were performed in order to assess the need for more sophisticated parameterizations of these processes in coupled regional and global models. These showed that the simulation of Arctic summer sea ice is very sensitive to the parameterization of snow and ice albedo as well as to the treatment of ice growth. The parameterization of the snow cover fraction on ice plays an important role in the onset of summer ice melt. This has a crucial impact on summer ice decay when more sophisticated schemes for ice growth and ice albedo are used. The work by Dorn and co-workers illustrates that when using a harmonized combination of more sophisticated parameterizations, simulation of the summer minimum in ice extent can be considerably improved due to the more realistic representation of interactions between the atmosphere and sea ice influencing Greenland and the Arctic as a whole.

Current understanding of model limitations as briefly described here, suggest that as a necessity for achieving improved simulations of the climate over Greenland, model development efforts should be directed towards the following:

- Use of an accurate ice sheet topographic dataset
- Accurate albedo parameterization
- More realistic surface layer model, including thermal conductivity for snow/firn
- Accurate planetary boundary layer momentum exchange.

These recommendations would improve the simulation of regional effects of Greenland's climate due to finer-resolved orography and land-sea contrasts, better-resolved non-linear interactions between the large-scale and mesoscales, better simulation of hydrodynamic instabilities and synoptic cyclones, and better description of hydrological and precipitation processes. Better representation of the northward development of the storm track, and invasion of the Arctic Basin by cyclonic systems from the lee of Greenland, which varies greatly with the NAO, would have a major influence on the Arctic sea-ice movement and melting in a coupled regional climate system. At higher resolution, the air-sea interaction changes, the air-sea heat flux is enhanced and localized in the lee of Greenland, and the wind stress increases.

8.3.1.2. **Model limitations and challenges**

In order to capture even the large-scale snow accumulation rates, it is essential for the atmospheric models to include a good representation of the mountains around the Greenland Ice Sheet. The detailed deposition of moisture in the mountains as well as over the ablation zone of the ice sheet will influence

the amount of moisture transported further into the interior of the ice sheet. It is also of great importance to use an accurate elevation model for the ice sheet itself. Box and Rinke (2003) looked at how the ice sheet topography used in the standard version of the HIRHAM regional climate model compared to a more recent digital elevation model (e.g., Bamber et al., 2001), and found substantial differences. They considered these to be responsible for several degrees of systematic temperature bias, particularly in the ablation zone.

Figure 8.20 illustrates how the topography of Greenland is depicted with increasing resolution (based on Bamber et al., 2001). At the coarsest grid resolution (~150 km), which is typical for most advanced state-of-the-art GCMs, the resolution does not allow for discrimination between the outermost part of the ice sheet and the coastal mountains. This basically implies that the atmosphere 'sees' all of Greenland as one big lump of ice. Increasing the grid resolution to 25 km improves the situation, but the gradients over the ice margins are not resolved even at this scale. Typically, the incremental rise in topography within the ablation zone is 300 to 500 m for a 25 km grid, which is potentially in conflict with the formulation of the vertical structure of the atmospheric model and clearly introduces systematic surface temperature

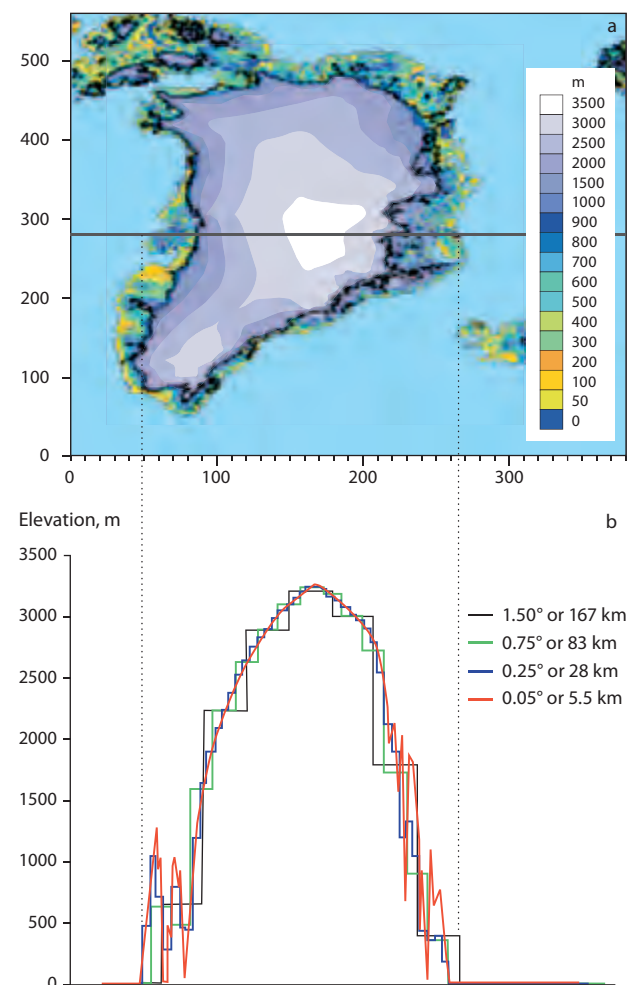Figure 8.20. (a) Topography of Greenland, with the ice sheet extent indicated by the black contour, (b) elevation as seen by models along a W-E transect (indicated in 'a') at GCM scale (black line) (~150 km), high-resolution GCM scale (green line) (~75 km), high-resolution RCM (blue line) (~25 km), and very high-resolution (red line) (~5 km). Source: Courtesy of Philippe Lucas-Picher, Danish Meteorological Institute.

errors that may exceed 3 °C. At a grid resolution of 12 km, this feature is almost resolved. But it is not until a resolution of about 5 km is applied that the surface temperature error no longer introduces serious inconsistencies, in cases where the modeled temperature field is used without any adjustment.

At present, it is still speculative as to whether a grid resolution of 5 km will also improve the precipitation climatology. Figure 8.21, however, compares the precipitation patterns from three simulations with the RCM HIRHAM with increasing resolution (Stendel et al., 2008; Philippe Lucas-Picher, Danish Meteorological Institute, pers. comm. 2008). The three HIRHAM realizations have a grid distance of ~80 km, 25 km and 5 km, respectively. The large-scale patterns are similar, but whether the precipitation is falling in the coastal margins, in the ablation zone or even in the accumulation zone does not appear to be resolved unless using a fine grid resolution.

In order to provide realistic driving conditions for an ice sheet model, a necessary requirement for the atmospheric model must, therefore, be that it resolves the features that are of most relevance in controlling the mass balance of the ice sheet. If this cannot be provided directly, additional adjustments to the data are necessary. This, in turn, raises questions about internal consistency of the energy and water cycle balances.

Representing the detailed regional conditions of Greenland in a climate model is also affected by the ability of the model to simulate realistically the energy and moisture cycles in the surrounding seas. Clearly, part of the moisture supply for the precipitation systems influencing the ice sheet originates from the regional seas surrounding Greenland. Because the oceanic circulation around Greenland is complex and poorly resolved by coarse-resolution coupled climate models, it is not surprising that detailed sea-ice conditions are often poorly represented.

At best, only large-scale sea-ice conditions can be depicted. Stendel et al. (2008) gave an example of a climate change downscaling experiment, where the driving GCM does have a relatively good representation of present-day sea-ice conditions, both with respect to seasonal variability and to interannual variability. However, details are still not matching observed

conditions to a level providing confidence that the changes are indeed a result of a realistic timing of changes in circulation and moisture support.

When ice sheet models are forced with atmospheric model output from GCMs, it is always an issue as to how realistic the simulated sea-ice temperatures and sea-surface temperatures actually are.

The IPCC concluded in its Fourth Assessment Report (IPCC, 2007) that mean temperature within the Arctic is likely to increase at a greater rate than mean global temperature, confirming results from previous IPCC assessments and the Arctic Climate Impact Assessment (ACIA, 2005). For the IPCC A1B emissions scenario, the annual mean Arctic temperature increase by the end of the 21st century is projected to be about twice that of the global mean increase (5 to 7 °C *vs* 2.5 to 3.5 °C) (IPCC, 2007; Christensen et al., 2007). Figure 8.22 shows an extract from an evaluation of 21 model simulations of global change under the A1B emissions scenario (Christensen et al., 2007), highlighting three models, the NCAR PCM, GFDL CM2.0, and MPI ECHAM5 as well as the 21-model mean. It is clear that the three individual models qualitatively show the same climate change response, but that the magnitude of the change differs by several degrees with the PCM model showing the lowest and the ECHAM5 model the highest degree of warming, while the GFDL CM2.0 model is close to the ensemble mean.

It is interesting to note, however, that the projected climate signals are to some degree caused by quite different mechanisms. Figure 8.23 shows an extract of an analysis of the performance of 14 model simulations for the period 1958 to 2000 (Walsh et al., 2008). Again, the ensemble mean behavior is shown with the same three individual models. A common feature for most of the models, reflected by the ensemble mean, is a clear cold bias in the Barents Sea due to a tendency to simulate too much sea ice, with the MPI model a clear exception. The implication is that only the MPI model simulates sea-ice coverage in this region reasonably well. At the same time, the greatest warming by the end of the study period is simulated exactly over this region in the ensemble mean as well as by the individual models.
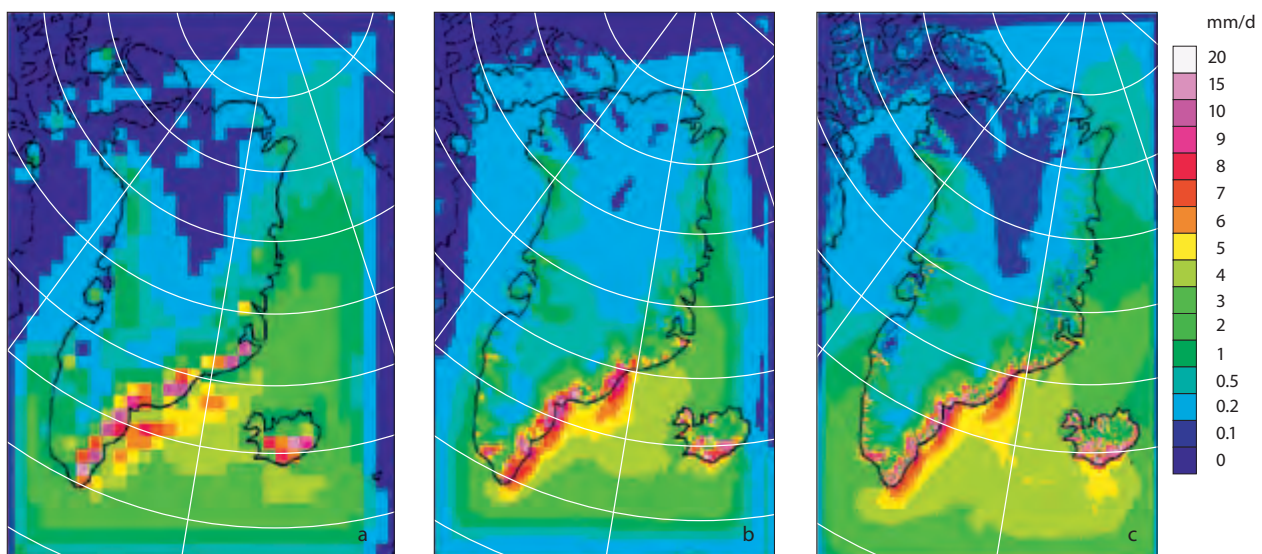

Figure 8.21. Resolution dependency of simulated winter precipitation over Greenland. (a) RCM resolution (~80 km), (b) RCM resolution (~25 km), (c) RCM very high resolution (~5 km). Source: Courtesy of Philippe Lucas-Picher, Danish Meteorological Institute.
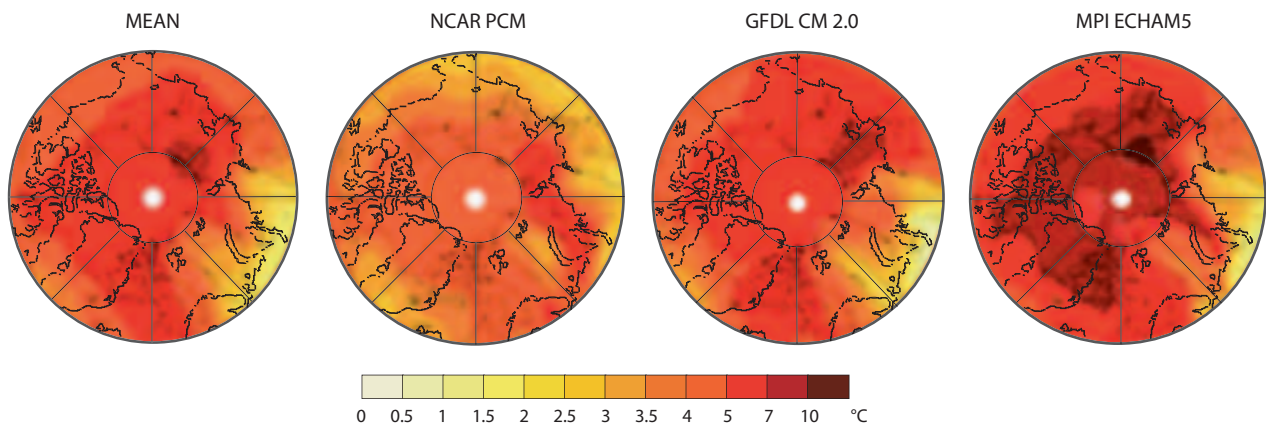
Figure 8.22. Maps of composite (based on 21 models) and three individual model-simulated annual mean temperature changes (2080–2099 *vs* 1980–1999) for the IPCC A1B SRES scenario. Source: IPCC (2007).

In the NCAR and GFDL models, this partly reflects the bias in present-day sea-ice conditions, while this apparently cannot be the case in the MPI model because the present-day sea ice appears to be captured with some realism. It should also be noted that, in general, the greatest warming occurs in the area with too much ice (strong cold bias) under current conditions, and in the NCAR model in particular, even though here winter data are compared with the annual mean.

This section has shown that, to varying extents, results at the regional scale are clearly subject to systematic errors in present-day simulations. Using an ensemble of models masks this deficiency. The role of the ocean as a driver of the climate in the Arctic is obvious, and there should be concern about interpreting a climate change signal, when most of the signal is apparently due to systematic error in simulating present-day conditions.

As the models discussed above have indicated, maps of warming must be carefully analyzed and, without further analysis, data cannot be used in a region with non-linear feedbacks such as the presence and absence of sea ice. Thus, it is also likely that the simulated warming and precipitation change signals are strongly influenced by these systematic errors, not only in the vicinity of the imperfections themselves, but also further away due to non-linear interactions in the atmospheric and oceanic circulations.

Figure 8.23. Ensemble mean temperature (based on 14 models) and three individual model temperature biases for winter (1958–2000). Source: Walsh et al. (2008).

8.3.2. Modeling Greenland Ice Sheet dynamics

8.3.2.1. Objectives of numerical ice sheet modeling
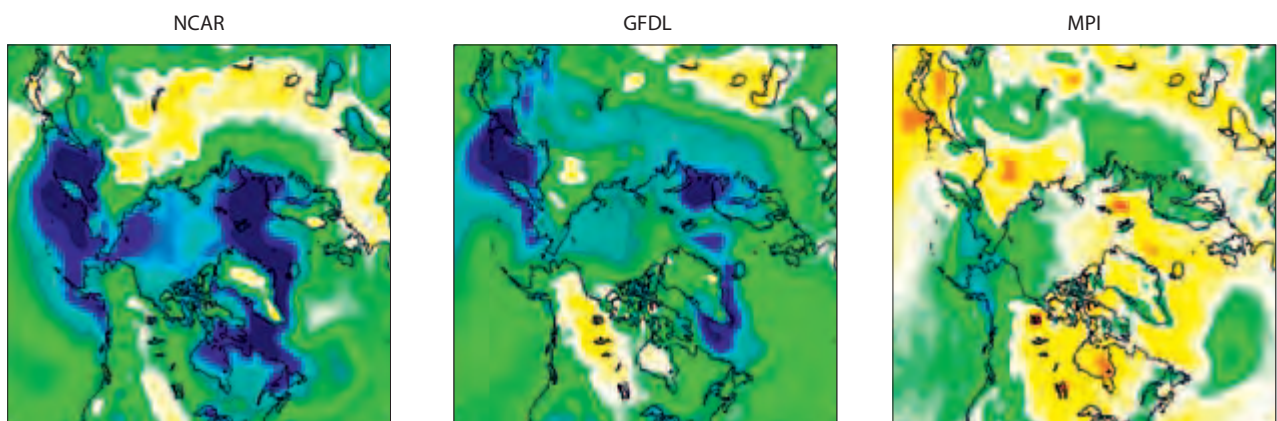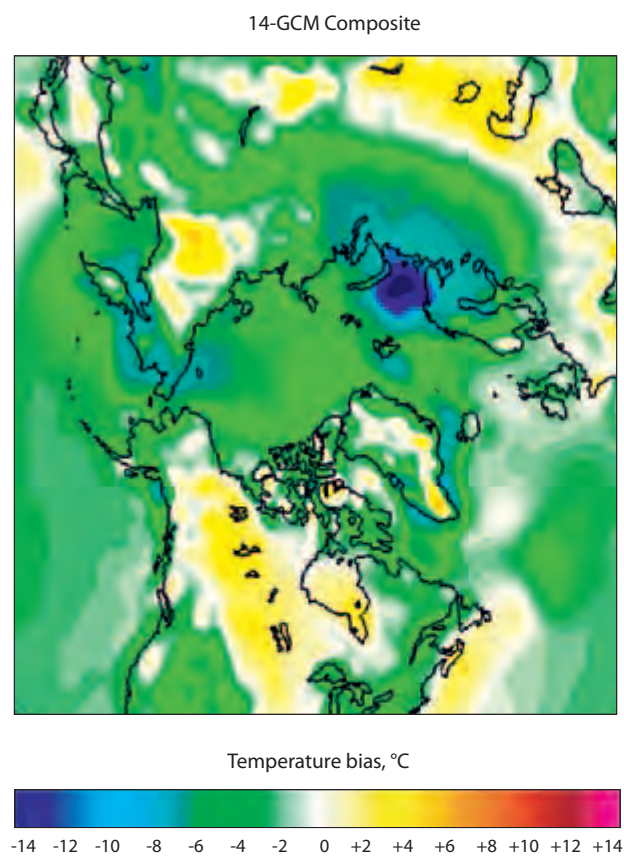
The future evolution of the Greenland Ice Sheet depends on the extent of climate change over the ice sheet, in particular the specific impacts on SMB as well as past climate forcing. As discussed in Sections 8.2.3 and 8.2.4, overall ice sheet mass balance depends strongly on the rate of ice discharge. In particular, dynamical mass loss is governed by the flux of ice to the ablation area. This includes terrestrial margins, where ice is removed through melting and runoff, and marine margins, where ice loss occurs through iceberg calving and melting of ice that is in contact with the ocean. High ice velocities lead to greater transport of ice to the ice sheet margins and higher rates of mass loss through each of these processes.

Ice sheet dynamics are difficult to predict or extrapolate from present-day observations because ice sheet velocities do not remain constant on decadal and century timescales. There are unanswered questions as to how ice flux will respond to changes in ice sheet geometry, oceanic boundary conditions, and surface meltwater reaching the bed. All of these changes are expected in the coming decades and centuries.

Numerical ice sheet models have been designed to describe first order ice sheet physics and to provide prognostic solutions for ice sheet evolution in response to such changes. Numerical models are required because the flow and deformation of ice at a particular point depends on a complex array of physical parameters, including ice sheet thickness, surface and bed slopes, ice temperature (which influences the deformation of the ice), ice fabric, valley walls, regional ice flow, and basal friction which is a function of bed geology, basal ice temperature, and subglacial hydrological conditions. By simulating these processes and features of the ice sheet system, physically-based estimates of ice flow and deformation and thus ice sheet response to climate change are possible.

8.3.2.2. The physical basis of ice sheet models

Detailed descriptions of ice sheet physics are provided by van der Veen (1999) and Paterson (1994). Box 8.2 provides a simplified overview of the theoretical basis of ice sheet models. The ice sheet is discretized into a 3-D array of grid cells and the equations that govern ice thickness, velocity, and temperature evolution are solved in each grid cell. Atmospheric and oceanic forcing of the ice sheet are introduced as boundary conditions.

The Greenland Ice Sheet is still adjusting to the last cycle of glaciation and deglaciation, particularly with respect to ice temperature at depth, where most of the ice deformation occurs. Simulations of future evolution of the ice sheet therefore require a 'spin up' simulation that takes Greenland through one or more glacial cycles (Huybrechts et al., 1991; Letréguilly et al., 1991). The climate forcing for these simulations is taken from Greenland ice core records. Climate forecasts are then imposed for studies of future ice sheet changes, generally as perturbations from present-day climatology.

When considering longer time scales, adjustment of the Earth's crust to ice loading and unloading must be included in the model as this represents a potentially important

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Box 8.2. Theoretical basis of ice sheet models

Three-dimensional models applied to the Greenland Ice Sheet to date all make use of the shallow-ice approximation, in which the gravitational driving stress is locally balanced by drag at the glacier base (Nye, 1957; Hutter, 1983). The gravitational driving stress is $\tau d\,(z) = \rho\,g(hS-z)\,\nabla S$, where ρ is the ice density, g is gravitational acceleration, hS is the glacier surface elevation, and ∇hS is the surface slope. At the glacier bed, $\tau d\,(hb) = \rho\,gH\,\nabla hS$, for ice thickness H. Glen's flow law relates ice deformation rates to the stress field in the ice. In the shallow-ice approximation, the ice velocity (u) associated with vertical shear deformation (d), i.e., ud, follows $ud \propto \nabla S\,n\,H\,n+1$, where n = 3 is the exponent in Glen's flow law.

Ice deformation rates are sensitive to ice temperature, with the effective viscosity of ice varying by a factor of about 1000 over the range of temperatures found in the Greenland Ice Sheet (Marshall, 2005). Ice sheet models therefore simulate the 3-D ice thermodynamics – advection and diffusion of energy and strain heating due to ice deformation – to model the temperature distribution in the ice sheet.

In addition to internal deformation, ice can flow via basal motion where the bed is at the pressure-melting point, through some combination of subglacial sediment deformation and decoupled sliding over the bed. Basal flow rates, ub, often exceed the ice motion associated with deformation by one or several orders of magnitude. The vertically-averaged velocity, ū, is calculated from the sum of these two contributions: ū = ūd + ub. Large-scale basal flow generally requires a layer of pressurized subglacial meltwater at the bed. Models make some allowance for basal flow, usually through a local sliding 'law' relating basal flow rates to gravitational shear stress at the bed. There is little knowledge of subglacial hydrological conditions, and the physical processes that determine basal water pressure are not yet modeled or parameterized in ice sheet models. For this reason, sliding rates and the spatial-temporal variability in basal flow are highly uncertain in current models.

Given an estimate of the vertically-averaged ice velocity, the equation for conservation of mass describes the rate of change of ice thickness at each point on the ice sheet:

$$\frac{\partial H}{\partial t} = -\nabla\cdot(\overline{u}H) + b$$

The first term on the right describes the divergence of the ice flux and b is the local mass balance rate (accumulation minus ablation; basal melting can be included in this term but is, generally, negligible compared to surface accumulation and ablation). This is usually considered over one year, that is, expressed as metres per year of ice-equivalent gain or loss of mass. Given measurements or climate model predictions of b, this equation can be integrated forward to simulate the evolution of ice thickness at all locations on the ice sheet. This is the basis of ice sheet modeling.

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feedback through the coupling between SMB and surface elevation (e.g., Oerlemans, 1980). Different models have been adopted for estimating isostatic adjustment, with for example, Marshall et al. (2000) assuming a local, damped return to

isostatic equilibrium, while Le Meur and Huybrechts (1998) incorporated a self-gravitating spherical visco-elastic Earth model.

For whole-ice sheet simulations over one or more glacial cycles the horizontal grid spacing is typically 20 km or more, with some 30 layers in the vertical (e.g., Huybrechts, 2002). Grid sizes of 5 to 10 km are tractable, but each time the grid is reduced by a factor of two the computational demand increases by an order of magnitude. There are also theoretical and pragmatic limits to resolution. The shallow-ice approximation that is used in most whole-ice sheet models is based on an assumption that horizontal grid dimensions are greater than the ice thickness. Further, available bedrock maps for the ice sheet, derived from airborne radar surveys, are currently available only at 5-km resolution. Issues associated with model resolution are discussed below.

Parameterization of SMB is a key parameterization in the models. Most ice sheet modeling efforts to date prescribe the SMB from monthly or annual surface air temperatures and precipitation rates. These can be based on observed (present-day) meteorological patterns or meteorological fields downscaled from a climate model. In general, the SMB fields simulated by climate models are not accurate enough for coupled ice sheet-climate modeling, in part because GCMs are too coarse to adequately resolve the ice sheet ablation areas (see Section 8.3.1). This makes it difficult to estimate ablation through a direct surface energy balance, so degree-day methods, driven solely by temperature fields, are generally adopted for melt and runoff modeling, following Reeh (1991) and Huybrechts et al. (1991).

Interest in decadal-scale ice sheet changes and the development of improved regional-scale meteorological models (e.g., Box et al., 2006) should soon permit direct estimates of SMB from meteorological models for future predictions, rather than parameterizations based on temperature and precipitation fields. This is desirable because such models are more physically-based and can capture the spatial variability in climate fields as ice sheet geometry and surface properties change; parameterizations tuned to present-day conditions become less reliable as both the climate and the ice sheet depart from their present-day state. This is discussed in more depth in Sections 8.2.2, 8.3.1, and 8.3.3.

8.3.2.3. Modeling of the Greenland Ice Sheet

To date, Greenland Ice Sheet simulations are all based on the shallow-ice approximation and the physics described in Box 8.2 (Huybrechts et al., 1991, 2004; Letréguilly et al., 1991; Greve, 1997, 2000; Ritz et al., 1997; Huybrechts and de Wolde, 1999; Marshall and Cuffey, 2000; Tarasov and Peltier, 2002; Gregory and Huybrechts, 2006). Basal flow is typically assumed to be active wherever the basal ice is at the pressure-melting point and is prescribed as a function of the gravitational driving stress.

Figure 8.24 gives an example of modeled and observed topography for the Greenland Ice Sheet on a 5-km grid. Present-day ice sheet area and volume can be simulated to within a few percent, but there are often systematic regional biases in modeled ice thickness (Figure 8.24c). The Northeast Greenland Ice Stream (see Figures 8.15 and 8.16) is present in

the model (Figure 8.25b), but the modeled flow rates in the ice stream are too slow, contributing to the ice thickness anomaly in this region. The ice margin position is poorly simulated in northern Greenland, leading to thickness anomalies of several hundred metres. The model is anomalously thin in southern-central Greenland, for reasons that are unclear. It may be related to a bias in the climate forcing or modeled rates of basal flow that are excessive, driving a thinner, lower-sloping ice dome.

The velocity field over the whole ice sheet indicates high flow rates on the ice sheet flanks and areas of concentrated ice flux where they are expected (Figure 8.16). These regions correspond
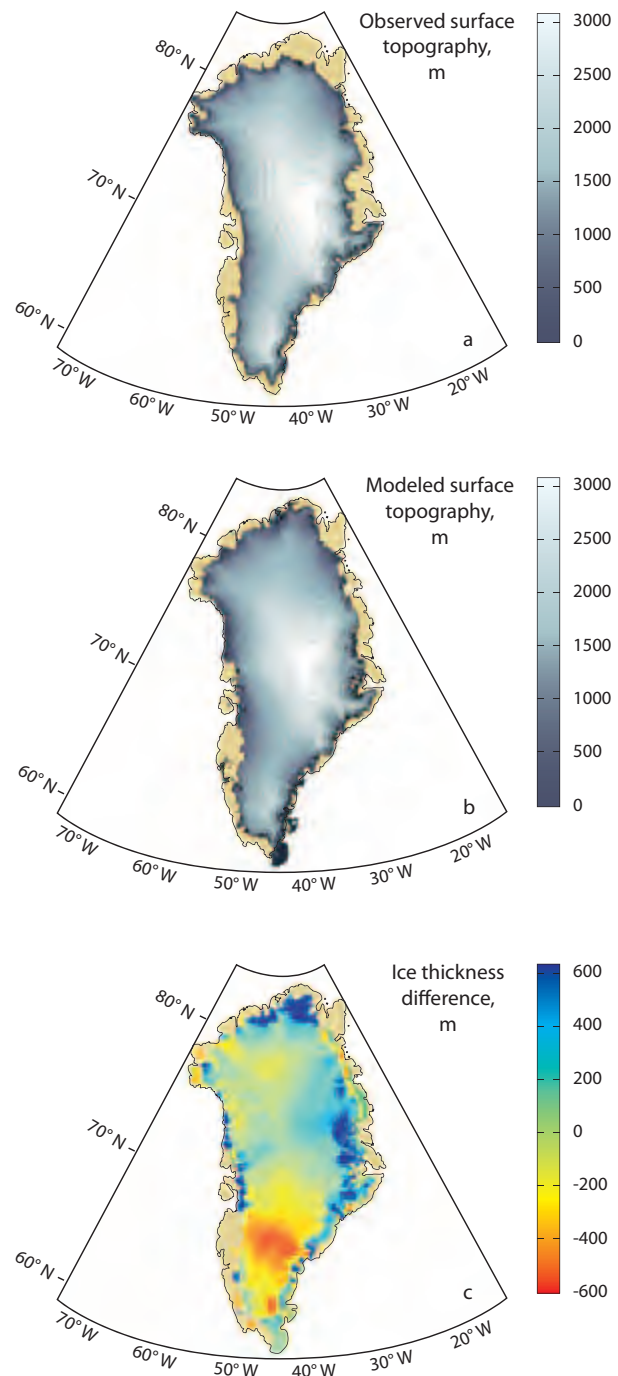

Figure 8.24. Model reconstructions of the present-day Greenland Ice Sheet on a 5-km grid: (a) observed and (b) modeled ice sheet surface topography. (c) Modeled minus observed ice sheet thickness. Source: Unpublished results, based on the Greenland Ice Sheet model described by Marshall and Cuffey (2000).

to major topographic channels and drainage outlets, which are resolved in this simulation. However, the model fails to capture the high rates of discharge in major outlet glaciers such as Jakobshavn Isbræ and Helheim Glacier; simulated velocities in some of these systems are an order of magnitude too low. This is most likely to be related to inadequate resolution of the fjord geometry, discussed further below.

Basal flow in this model is parameterized as a function of gravitational driving stress and is enabled wherever the ice sheet bed is at the pressure-melting point. Figure 8.25 plots modeled present-day basal ice temperatures and ice sheet velocities, including contributions from basal flow in warm-based sectors of the ice sheet. Few data are available to validate or test these predictions from the model.

Deep borehole data confirm that central and eastern Greenland are generally cold-based, due to the relatively thin ice that drapes the underlying mountains and the advection of cold ice to the bed along the ice divide. On the ice sheet flanks, deformational heating warms the ice to produce large areas that are at the melting point. In the major topographic channels, this can produce a thick temperate ice layer which has a low effective viscosity, introducing a positive feedback

that accelerates deformation and ice flow in these channels.

These models have limitations, elaborated below, but they also capture many large-scale features of the ice sheet dynamics in Greenland. Bearing the model limitations in mind Figure 8.26 illustrates the modeled response time of the Greenland Ice Sheet to simple warming scenarios. A regional warming of 2 to 8 °C is introduced as a linearly increasing temperature perturbation over the period 2000 to 2100, relative to late 20th century baseline temperature fields over the Greenland Ice Sheet as compiled by Ohmura (1987). The temperature anomaly is held constant from 2100 to 3000. Figure 8.27 plots ice sheet configurations at 3000 for the 4 and 8 °C scenarios.

Most of the ice volume loss in these scenarios comes from southwestern Greenland, with the ice dome in southern Greenland collapsing under sufficient warming. This result has been well established in the modeling by Huybrechts and others, as illustrated in Figure 8.28 from Alley et al. (2005). These results correspond to three different scenarios of future atmospheric carbon dioxide (CO_2), with values in excess of 750 ppm inducing summer warming of more than 5 °C in Greenland, sufficient to excite collapse of Greenland's southern dome on a timescale of several centuries.

These results reflect the plausible impact of atmospheric warming on increased melt and drawdown of the Greenland Ice Sheet. There is an accelerating ice loss in late stages of the retreat, when the interior of the Greenland Ice Sheet falls below the equilibrium line altitude. Due to simplified fast-flow and ice-marginal physics in the model, as well as one-way climate forcings (i.e., missing climatic feedbacks of ice sheet retreat), the time-scale of modeled Greenland Ice Sheet retreat in Figures 8.26, 8.27 and 8.28 is uncertain. Many of the missing feedbacks and processes give a systematic bias towards underprediction of ice sheet sensitivity to climate warming; it is expected that Greenland Ice Sheet retreat will proceed more quickly than forecast for a given warming scenario. However, how quickly this could transpire is not well-constrained, because no models with the requisite physics are available to assess this. This is picked up again in Section 8.3.3. The next section addresses several of the outstanding challenges for the current ice sheet models.
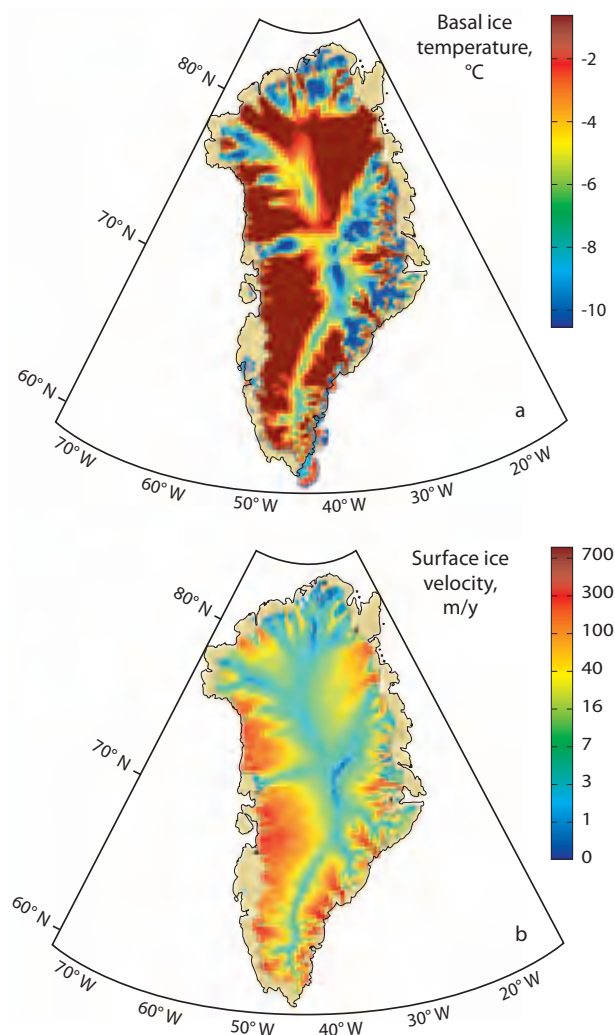

Figure 8.25. Model reconstructions of the present-day Greenland Ice Sheet (a) basal ice temperatures below the pressure-melting point, and (b) surface ice velocity. Source: Unpublished results, based on the Greenland Ice Sheet model described by Marshall and Cuffey (2000).
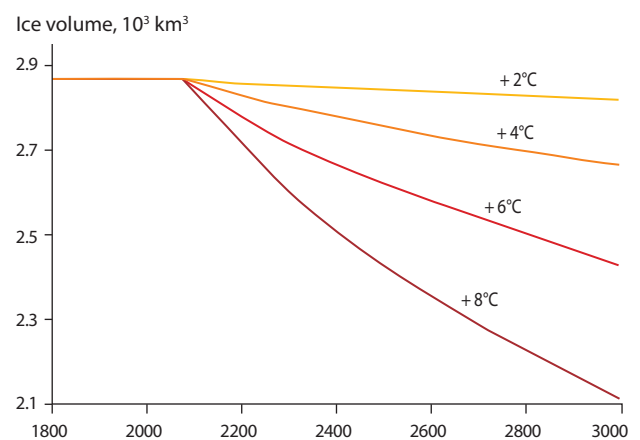
Figure 8.26. Modeled Greenland Ice Sheet volume response to different climate warming scenarios. Warming scenarios are introduced as a linear temperature perturbation from 2000 to 2100, with warming held constant at that level (2, 4, 6, or 8 °C) from 2100 to 3000. Source: Unpublished model results from Shawn Marshall, University of Calgary. For illustrative purposes only; not intended as a forecast for the ice sheet.
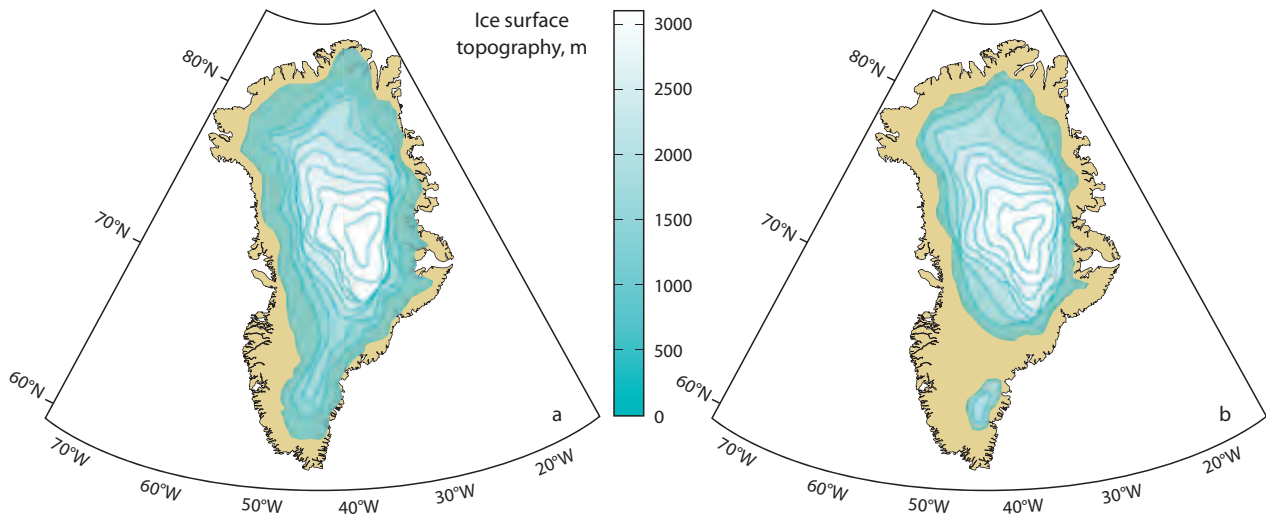
Figure 8.27. Modeled Greenland Ice Sheet surface topography in 3000 AD for (a) the 4 °C and (b) 8 °C warming scenarios of Figure 8.26. Source: Shawn Marshall, University of Calgary.
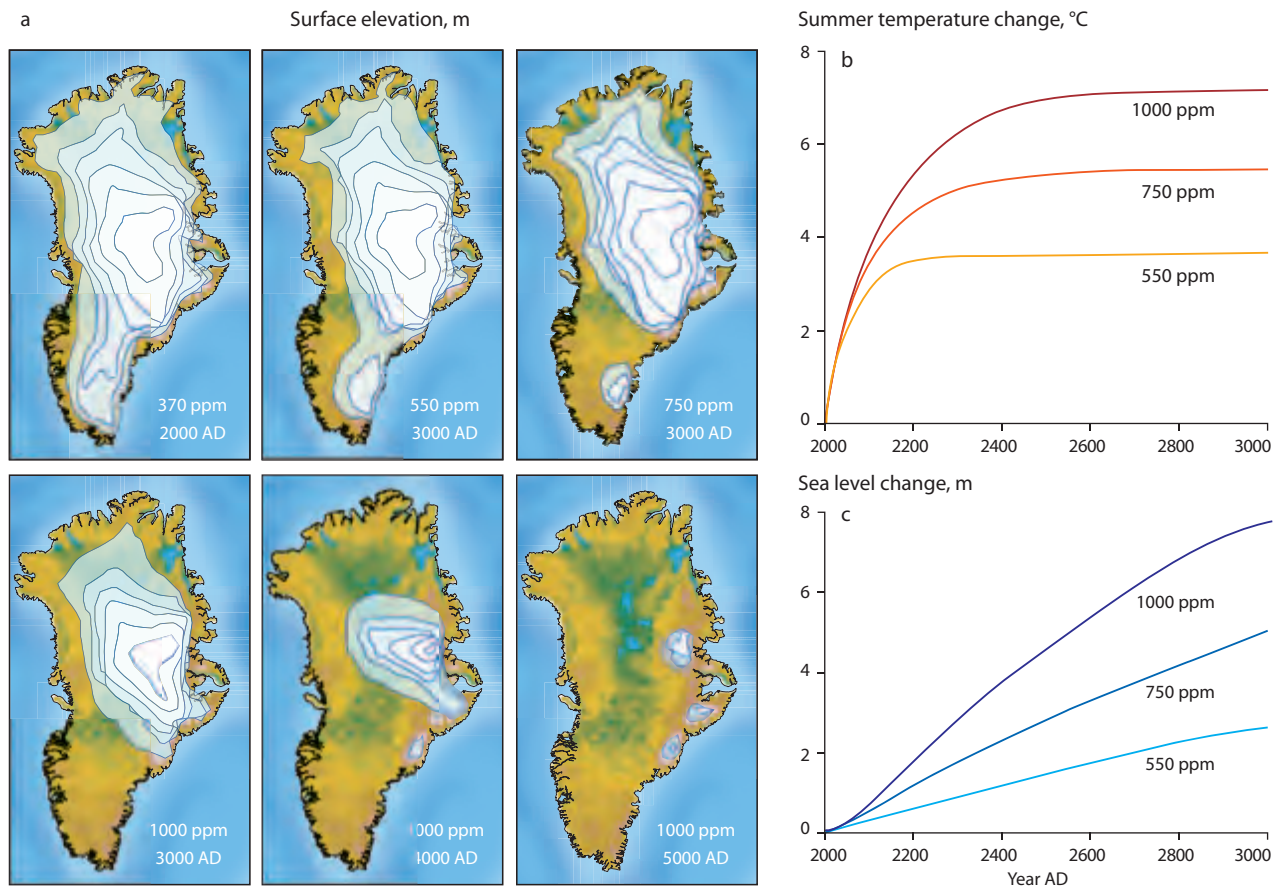
Figure 8.28. Greenland Ice Sheet simulations for 3000 to 5000 AD under different greenhouse gas scenarios: (a) surface elevation, (b) surface temperature change, and (c) sea level change. Source: Alley et al. (2005).

8.3.2.4. Modeling challenges

The modeling studies cited above have a general skill in simulating Greenland Ice Sheet dynamics, but some features and processes are difficult to capture. These include ice sheet-ocean interactions, ice stream dynamics, basal flow processes (sliding and sediment deformation at the glacier bed), and coupling of ice marginal dynamics and inland ice dynamics. These features are critical to questions of ice sheet sensitivity and response time to climate change.

The IPCC AR4 summarizes this well: *Dynamical processes related to iceflow not included in current models but suggested by recent observations could increase the vulnerability of the ice sheets to warming, increasing future sea level rise. Understanding of these processes is limited and there is no consensus on their magnitude.* (IPCC, 2007:17).

An expanded discussion of shortcomings of existing ice sheet models was provided by van der Veen and ISMASS (2007). Shortcomings with existing Greenland Ice Sheet models

fall into three categories: those related to physical processes; those associated with spatial resolution; and uncertainties in specification of boundary conditions, including ice-ocean and ice-atmosphere interactions. These issues are elaborated in the following sections, ranked according to a subjective judgment of their relative importance to modeling Greenland Ice Sheet ice dynamics. These processes and considerations have different levels of importance in different parts of the ice sheet, and in some cases the process is not understood well enough to assess how important it will prove to Greenland Ice Sheet response to climate change.

8.3.2.4.1. Uncertainties associated with physical processes

Basal flow and hydrology

The extent and physical controls of glacier motion due to basal flow remain poorly known and difficult to quantify (Clarke, 2005). This is true across a wide spectrum of ice masses, including mountain glaciers, outlet glaciers in Greenland and other polar icefields, and Antarctic ice streams. Basal flow is generally the main mechanism for fast flow (except for enhanced creep rates in deep fjords), so must be well quantified to provide a realistic model of ice sheet dynamics.

Elaborate theories have been constructed to describe the sliding of ice over variously shaped obstacles and glacier beds with different roughness characteristics, but it is unclear whether the models are relevant to large-scale basal flow. The models describe salient physical processes in the subglacial environment, including regelation and stress enhanced softening of glacier ice, which allow it to flow around basal obstacles, and the free slip of basal ice over air- and water-filled cavities in the lee of bedrock obstacles. While these processes are likely to contribute to integrated basal slip over large areas of the glacier bed, at least for hard beds (i.e., bedrock rather than deformable till), the governing physics operate at scales of centimetres to metres. Basal motion at scales of tens of metres to tens of kilometres, as observed in Antarctic ice streams and ice marginal regions of Greenland (Joughin et al., 2008a), is more relevant to overall ice flux.

Furthermore, motion associated with regelation and stress-softening around bedrock obstacles increases non-linearly with shear stress (Paterson, 1994:157), whereas basal flow in many settings (e.g., Antarctic ice streams) is associated with low values of gravitational driving stress: thin, low-sloping ice with subglacial water pressures near flotation, implying low values of basal shear stress over large regions (Bindschadler et al., 2000). Summer speed-ups in valley and outlet glaciers and the ice marginal region of the Greenland Ice Sheet provide another illustration of this. These speed-ups are associated with negligible change in surface geometry (i.e., gravitational driving stress), but are instead driven by reductions in basal friction due to the influx of surface meltwater to the bed (e.g., Iken and Bindschadler, 1986; Copland et al., 2003; Das et al., 2008; Joughin et al., 2008a).

The mechanics of large-scale basal motion therefore appear to be governed by the extent of the glacier bed that experiences ice-bed decoupling due to high subglacial water pressures. High water pressure promotes basal flow by effectively floating the glacier above the bed, reducing basal traction. This applies both to basal sliding and subglacial sediment deformation, although the rheological properties and supply of basal sediments also need to be considered for the latter.

The need to understand hydrological influences on basal flow and to incorporate these processes into models is pressing, considering recent evidence that surface meltwater can drain to the bed through hundreds of metres of cold ice in marginal areas of the Greenland Ice Sheet (Zwally et al., 2002b; Das et al., 2008) and smaller-scale polar icefields (Boon and Sharp, 2003). Similar to the long-standing observations from mountain glaciers, this influx of meltwater is capable of triggering transient speed-ups, increasing the discharge of ice (Zwally et al., 2002b; Joughin et al., 2008a). This presents a direct mechanism by which climate change can influence ice dynamics. Coupled with trends of increasing spatial extent and intensity of the melt season in Greenland (Tedesco, 2007), accelerated summertime flow is expected in future decades in the marginal zones of the Greenland Ice Sheet.

The magnitude of the speed-up and the extent to which ice sheet drawdown will propagate inland are less clear. Joughin et al. (2008a) presented observations over a 300-km wide region in southwestern Greenland which indicate that summer speed-ups of the major outlet glaciers, including Jakobshavn Isbræ, contribute only a few percent to the total annual discharge. Speed-ups on the slow-moving ice sheet flank (i.e., away from the major outlet glaciers) are relatively greater, increasing annual discharge by as much as 25%, but this region does not account for much of the total dynamic discharge of the Greenland Ice Sheet. Therefore, this mechanism represents a pathway for ice sheet response to climate forcing, but one which may be relatively stable (i.e., non-catastrophic).

These influences on ice dynamics are generally absent in the large-scale models that predict Greenland Ice Sheet response to climate change. Physically-based models of ice sheet hydrology are needed to simulate these processes. In a flowline model of the western flank of the Greenland Ice Sheet, Parizek and Alley (2004) simulated the supraglacial hydrology and englacial drainage to the bed, assuming that surface drainage induces speed-ups as seen by Zwally et al. (2002b). Extrapolating their results to the whole ice sheet, their most conservative scenarios gave 12% to 38% more volume loss from the Greenland Ice Sheet over the next several centuries due to surface-meltwater induced lubrication, with the effect increasing with warming. Parizek and Alley (2004) also suggested that the effect could be much higher, with the upper range of their modeling giving a tripling of the ice volume loss in Greenland, relative to models that exclude surface meltwater effects.

These models assume that basal flow is proportional to water supply at the bed, but subglacial water storage and basal water pressures are what really matter for ice-bed coupling and basal flow. The physics that governs the relationship between subglacial water pressure, bed geology, bed roughness, and basal traction still needs to be elucidated and parameterized in a way that is physically valid and testable in models.

A subglacial hydrological model is also required to provide a realistic estimate of spatial and temporal variations in basal water pressure. These have begun to be developed and coupled with models of glacier dynamics (Arnold and Sharp, 2002; Flowers and Clarke, 2002; Johnson and Fastook, 2002; Flowers et al., 2005; Marshall et al., 2005), but subglacial hydrological

conditions are notoriously difficult to observe, so models are difficult to test and constrain at large scales. To date, most models of subglacial hydrology treat the subglacial water system as a distributed water sheet, with water fluxes driven by hydraulic potential gradients. Conduit physics and the nuances of regime shifts between distributed and channelized drainage systems need to be better understood and included in the models in order to capture spatio-temporal patterns of basal water pressure and their influence on glacier dynamics. This has yet to be attempted in Greenland.

High-order stresses (horizontal stress coupling)

The shallow-ice approximation estimates ice motion as a function of the local gravitational driving stress, with the explicit assumption that non-local effects are negligible. This is known to be a poor approximation in certain parts of the ice sheet, such as the ice divide, ice streams, ice shelves, and ice sheet margins. In these settings, higher-order stress terms such as horizontal shear stress and longitudinal stress gradients, sometimes known as non-local stress coupling effects, become important to ice deformation. For instance, horizontal shear stresses at ice stream margins and next to valley walls play an important role in holding back ice flow, while longitudinal strains (stretching) represent the primary mode of ice deformation in ice shelves and floating ice tongues.

Longitudinal stress coupling also provides a mechanism for the transmission of ice-marginal forcings to the interior of the ice sheet. Current thinning of Jakobshavn Isbræ is accompanied by rapid retreat of the calving terminus (Sohn et al., 1998; Joughin et al., 2004). Joughin et al. (2004) proposed that the velocity increase resulted from development of large rifts in the floating tongue, followed soon thereafter by complete collapse of this tongue. Associated reductions in buttressing would allow increased discharge from the grounded portion of the glacier (Thomas, 2004). On the other hand, Pfeffer (2007) proposed that thinning could have decreased the height above buoyancy, thereby reducing the effective basal water pressure. This, in turn, would lead to greater sliding speeds, especially if the ice thickness approached floatation. To adequately incorporate these processes into numerical models and to test the validity of these competing hypotheses, it is necessary to incorporate stress transmission across the grounding line.

Models of glacier dynamics with higher-order stresses have become more common in studies of valley glaciers as well as regional ice sheet applications (Albrecht et al., 2000; Schneeberger et al., 2001; Pattyn, 2002, 2003; Saito et al., 2003). These advances are relatively new, and have yet to be applied to whole-ice sheet simulations in Greenland. Higher-order models such as that of Payne et al. (2004) and Price et al. (2008) demonstrate the importance of longitudinal stress coupling for accurate simulation of ice sheet response to perturbations at the ice sheet margin, that is, inland propagation of changes due to ice-marginal thinning or acceleration.

Iceberg calving

Ice sheet grounding lines, floating ice tongues, and iceberg calving processes are poorly represented in current glaciological models. Existing models generally determine the position of the calving terminus (i.e., marine ice front) by either prescribing an ice limit based on a maximum water depth (e.g., the continental shelf break), or by freely simulating floating ice tongue advance based on a simple parameterization of iceberg calving, typically prescribing calving rates to be proportional to water depth. Zweck and Huybrechts (2005) examined several different treatments of calving in large-scale models; all are simple parameterizations that do not describe the actual process(es) and should be considered to be statistical-empirical models that aim to capture the bulk loss of ice through this mechanism.

The physical controls of iceberg calving are not well understood and may not be deterministic. Iceberg calving requires fractures to initiate and propagate at the ice front. The process has been well-observed, but the environmental controls of calving rates are difficult to isolate; ocean and air temperature, surface meltwater, sea-ice conditions, and tidal flexure at the ice front may all have an influence on calving, but there is no simple relationship. Sohn et al. (1998) observed an increase in calving rate following the onset of spring melting and found a strong correlation between surface melting on the Greenland Ice Sheet, the melting of fjord ice, and the rapid increase in calving rate during summer. Their observations are insufficient, however, to resolve conclusively whether the presence of abundant surface meltwater enhances the ability of surface crevasses to penetrate the ice thickness, or whether the break-up of confining fjord ice reduces the restraining force acting on the calving front.

Recent evidence indicates the likely role of regional ocean warming in destabilizing the floating ice tongue of Jakobshavn Isbræ (Holland et al., 2008). This is also consistent with tidewater-style retreat of Helheim Glacier on the southeast coast of Greenland (Nick et al., 2009). Ocean-ice sheet coupling is lacking in the current models, but clearly needs to be embraced for predictive models of ice sheet response to climate warming.

8.3.2.4.2. **Uncertainties associated with model resolution**

Whole-ice sheet models for Greenland now operate at scales of 5 to 20 km, but scale problems still arise because of the small spatial scale of most of the outlet glaciers where drainage to the sea is concentrated. For example, Jakobshavn Isbræ, which drains mass from 7% of the total area of the Greenland Ice Sheet, is less than 10 km wide over its entire ~100 km length (Echelmeyer et al., 1991). This glacier overlies a narrow (6 to 10 km) and deep (~1500 m below the surrounding basal landscape) subglacial trough (Clarke and Echelmeyer, 1996; Gogineni et al., 2006; Lohoefener et al., 2006) that cannot be resolved with a horizontal grid spacing of 10 km or more. This limits the capacity of models to simulate the enhanced creep and high rates of flow that are observed in Jakobshavn Isbræ and other major outlet glaciers that drain the Greenland Ice Sheet.

At the same time, detailed and accurate bed topography upstream of the grounding line or calving front is needed to assess the possible irreversibility of terminus retreat for Greenland's major outlet glaciers. Howat et al. (2007) suggested that the rapid retreat of the Helheim Glacier terminus occurred as the calving front retreated into deeper water and continued until reaching a reversed bed slope, similar to the behavior of tidewater glaciers (Nick et al., 2009).

Even if spatial resolution is improved sufficiently to capture

small-scale basal topography, narrow channels no more than a few kilometres wide and up to 1500 m deep challenge the assumptions in the shallow-ice approximation. As noted by Clarke and Echelmeyer (1996), these bedrock troughs are similar to valley walls in that the near-vertical faces produce lateral drag that partially opposes the driving stress – horizontal shear stresses become important in taking up the gravitational driving stress. Also, an overlying temperate layer of fast-moving ice is embedded within more slowly moving ice, which may provide additional resistance, similar to the situation on West Antarctic ice streams. As a start, an effective shape factor could be introduced (Clarke and Echelmeyer, 1996) to account for the additional flow resistance, but no rigorous tests have been conducted to evaluate the applicability of this approach.

Additional uncertainty is introduced by the well-known (if rarely discussed) sensitivity of modeled ice margin dynamics to grid resolution (Abe-Ouchi and Blatter, 1993). With a fixed grid and a free boundary, the standard treatment in continental ice sheet models, the advance and retreat of the ice sheet margin proceed by discrete 'jumps' to adjacent grid cells. This process is sensitive to grid resolution because the flux of ice into a neighboring cell is a function of the local surface slope. This sensitivity combines with mass balance elevation feedbacks to make ice margins more mobile under higher resolution; it is simpler for the ice sheet to advance as grid size $\Delta x \to 0$. Strategies involving adaptive grids (e.g., Price et al., 2008) or sub-grid tracking of the ice sheet margin are needed to give convergent behavior that is insensitive to model resolution.

8.3.2.4.3. Uncertainties associated with boundary conditions

Geothermal heat flux

Pronounced subglacial topography with a very deep basal valley such as that of Jakobshavn Isbræ intensifies the local geothermal heat flux by as much as 100% (van der Veen et al., 2007), potentially introducing an important feedback between subglacial topography and fast flow. On a broader scale, significant variations in crustal thickness have been inferred from airborne gravity, with the thinnest crust underlying the Northeast Greenland Ice Stream and other major drainage routes (Braun et al., 2007). Where the crust is relatively thin, geothermal heat flow may be expected to be greater than the continental average (Leftwich et al., 2007). High geothermal heat flux in northern Greenland is also believed to play a role in providing the temperate, lubricated bed that enables fast flow in this ice stream (Fahnestock et al., 2001). The extent to which spatial variations in heat supplied to the glacier base affect the regional flow regime remains to be resolved (Greve, 2005). This could explain discrepancies in areas where basal meltwater has been detected but where models incorrectly predict the basal ice to be frozen to the bed.

Climatic forcing

Ice sheet models require distributed (spatially-resolved) atmospheric fields to provide SMB fields. To examine climate change scenarios, temporal evolution of atmospheric conditions needs to be characterized. This appears as a boundary condition in ice sheet models, and is prescribed simplistically in most ice

sheet model experiments. Uncertainties in spatial-temporal climate evolution may well dominate uncertainties in ice dynamics modeling for prediction of Greenland Ice Sheet response to climate change. This is discussed in more detail in Section 8.3.3. Still striving to present a correct present-day mass balance, the challenge of reproducing past and future mass balances is large.

In addition to atmospheric fields, ocean and sea-ice conditions directly impact on marine outlet glaciers (e.g., Holland et al., 2008; Section 8.2.3). Marine influences are parameterized through the ocean heat flux that is prescribed as a basal boundary condition for floating ice; this is generally held fixed in current models, for example, a uniform flux of 2 W/m². Plus, ocean conditions can be incorporated in calving parameterizations, but this has rarely been done, in part because of unclear mechanistic links. Except for a small number of regional studies, ocean and ice sheet models are largely uncoupled and ocean variability is not considered in ice sheet modeling. Challenges in ice sheet modeling are revisited in the summary comments of Section 8.3.4, where the main sources of uncertainty for predicting Greenland Ice Sheet sensitivity to climate change are discussed.

8.3.3. Future climate and ice sheet scenarios: Response of the Greenland Ice Sheet to a warmer climate

8.3.3.1. Objectives of coupled ice sheet-climate modeling

Ice sheet models, as described in Section 8.3.2, can investigate the ice-dynamical response to idealized climate change scenarios, such as a uniform cooling or warming. However, patterns of change in climate fields (e.g., temperature, precipitation, radiation, winds) will not be uniform, as used in the models, over the area and elevation range of the Greenland Ice Sheet. Realistic climate forcing is required to provide more detail on ice sheet evolution in the coming decades and centuries. Ice sheet evolution over long timescales also introduces a number of feedbacks that need to be captured in future climate forecasts, such as changing elevation and albedo fields over the ice sheet. Coupled ice sheet-climate models are needed to capture these feedbacks.

The next section describes the coupling approaches adopted in ice sheet-climate modeling. Model predictions of Greenland Ice Sheet response to future warming are then summarized, followed by a discussion of some of the limitations of current models and priorities for future development.

8.3.3.2. Coupling approaches

8.3.3.2.1. One-way and two-way coupling

One-way and two-way approaches have both been adopted for coupling atmospheric climate and ice sheet models. In one-way coupling, output from the climate model provides input for the ice sheet model (e.g., SMB fields), but the ice sheet model output does not force the climate model. This allows existing climate change scenarios (e.g., the ensemble of IPCC climate model forecasts) to be applied to ice sheet simulations. To

increase realism of the exchange, relationships between surface elevation, latitude, and surface conditions have been prescribed to allow climate forcings to change as the ice sheet surface evolves (Huybrechts et al., 1991). This includes elevation-temperature (lapse rate) feedbacks, precipitation rates that increase with increasing temperatures, roughly in accord with a Clausius-Clapeyron relationship, and a modification of local precipitation rates as a function of local surface slope (Ritz et al., 1997).

In two-way coupling, climate model output forces the ice sheet model, while output from the ice sheet model (e.g., ice topography, albedo fields) serves as input to the climate model. Two-way coupling is desirable as the simulation more realistically represents the physical system and the climate feedbacks associated with temporal evolution of the ice sheet. However, as two-way coupling is more challenging than one-way coupling and there are, generally, discrepancies of resolution between climate and ice sheet models, most ice sheet modeling experiments to date have used one-way coupling.

The model results in Figures 8.25, 8.26, 8.27, and 8.28 are examples of ice sheet simulations with one-way climate forcing. SMB is estimated from temperature and precipitation fields, using a degree-day methodology to estimate surface melt and the fraction of precipitation to fall as snow (Reeh, 1991). Present-day precipitation and temperature fields are taken as the baseline and temporal variability is prescribed as a perturbation based on ice core reconstructions, climate model forecasts, or simple scenarios (e.g., a step warming of 2 or 4 °C). Many Greenland Ice Sheet modeling studies have adopted this approach (Huybrechts et al., 1991, 2004; Ritz et al., 1997; Huybrechts and de Wolde, 1999; Greve, 2000; Alley et al., 2005; Gregory and Huybrechts, 2006).

While several parameterized ice-climate feedbacks are included in these numerical experiments, the one-way approach may omit key physical processes that affect the circulation of the atmosphere and ocean, which in turn would modify temperature and accumulation patterns on the ice sheet. A small number of groups have begun to interactively couple ice sheet models with climate models of varying degrees of complexity and resolution. Huybrechts et al. (2002), Fichefet et al. (2003), and Driesschaert et al. (2007) forced an ice sheet model with an AOGCM and examined how Greenland Ice Sheet retreat may affect oceanic circulation in the coming decades. These studies use one-way ice-sheet forcing from the atmospheric model. Other studies focus on ice sheet-atmosphere feedbacks (Gregory et al., 2004; Toniazzo et al., 2004), while a small number of studies explore fully-coupled ice sheet and AOGCMs (Ridley et al., 2005; Mikolajewicz et al., 2007b; Vizcaíno et al., 2008).

A number of climate-system and ice sheet feedbacks identified in coupled model studies are discussed in Section 8.3.3.3. Specific predictions of Greenland Ice Sheet retreat and sea-level rise in response to climate change are summarized in Section 8.3.3.4.

8.3.3.2.2. Climate downscaling

The experiments described above require atmospheric fields to be downscaled to the ice sheet model grid for mass balance modeling. This is due to the mismatch that exists between even high-resolution AOGCMs running at 1.25° (~150 km) grid resolution and ice sheet models (10 to 20 km). Grid downscaling is thus required and commonly takes the form of a simple interpolation.

This step requires a good understanding of altitude-SMB relationships in different climate regions of the ice sheet. Constant and spatially-uniform temperature lapse rates (linear temperature decreases with elevation) are usually used to downscale temperatures, often with different summer and annual lapse rates (Huybrechts et al., 1991). This does not account for the effects of inversions, katabatic-wind induced cooling at low elevations on the ice sheet, or surface energy balance effects that govern near-surface (*vs* free-air) temperatures, so a uniform lapse rate is a simplification. Precipitation is more difficult to downscale, as it can vary over small spatial scales, particularly in association with orographic forcing of precipitation at the ice sheet margin. Most modeling studies to date are based on perturbations to the present-day spatial pattern of precipitation in Greenland.

Under interactive coupling, the influence of significant changes in ice sheet topography and boundary conditions are better accounted for. For example, processes such as orographic precipitation and the energy balance influence of increasing areas of open water and ice-free land can be included as feedbacks in the model, rather than an assumption that present-day climate patterns will hold.

Because of the uncertainty associated with interpolation of meteorological fields, the downscaled climate is typically parameterized into positive-degree-day (PDD) input in order to simulate snow/ice melt; a full surface energy balance is not generally attempted. PDD approaches have proven reasonable for melt modeling in Greenland, but downscaling or interpolation of temperature or PDD fields does not ensure conservation of energy between the applied surface climatology and the climate model. There is a similar concern for downscaled precipitation fields. Conservation of mass with respect to larger-scale climate model precipitation and moisture fields is not straightforward and is often disregarded. Conservation of energy and mass warrant serious attention, particularly in two-way coupling experiments where there is a possibility of cumulative biases leading to 'drift' in transient simulations.

8.3.3.2.3. Mass balance estimation in coupled models

For the reasons outlined above, most Greenland Ice Sheet modeling studies to date estimate SMB from degree-day calculations as a function of climate (i.e., temperature and precipitation) perturbations. As was discussed in detail in Section 8.2.2, SMB fields have also been simulated directly from climate model re-analyses (Hanna et al., 2002, 2005, 2008), remote sensing studies (Mote, 2003), and regional climate modeling (Box et al., 2006; Fettweiss et al., 2007, 2008). Like the reconstructions by Hanna and co-workers, regional climate models make use of re-analyzed climatology to provide the boundary conditions for atmospheric simulations. The main difference is the nature of accumulation and melt modeling in the two approaches; Hanna and co-workers use a degree-day approach, while regional climate models distribute precipitation based on atmospheric dynamics and estimate melt from a surface energy balance.

Table 8.2 gives a summary of simulated SMB terms from the historical period for several of these efforts. These studies

provide detailed estimates of Greenland Ice Sheet mass balance through the second half of the 20th century, but future forecasts cannot be carried out with re-analyzed climatology. RCMs offer a potential solution to this, as they can be trained/calibrated on the historical period and forced by global-model fields for future simulations. This has yet to be attempted for Greenland.

Regional climate models have the inherent capacity to represent the physical processes of how snowfall and melt relate to the local and regional surface characteristics and energy budgets, and these models are likely to form the basis for the next generation of Greenland Ice Sheet predictions. However, climate models exhibit systematic biases, for example, land surface temperature too high or too low, in response to biases in model surface energy budget. These errors have been attributed largely to cloud-radiation bias, such as too much or too little downward longwave radiation. Terrain biases are another typical source of systematic bias, commonly arising from inaccurate representation of the terrain or from the need to smooth the surface topography at the 'steep' ice sheet margin to avoid atmospheric dynamical instabilities, for example, gravity waves. Through comparison with *in situ* observations, calibration procedures can be developed and uncertainties can be quantified (e.g., Box et al., 2004, 2006). However, a careful assessment of the stationarity of model biases is needed before these models can be applied with confidence to future forecasts.

8.3.3.2.4. **Ice sheet-ocean coupling**

Regional-scale coastal and sub-ice shelf oceanographic models are also needed to improve ice-climate coupling, something that has been explored on a regional scale in Antarctica, but has not been considered in Greenland. Other than an influence of global sea level on calving rates, ice-ocean feedbacks have not been examined on the scale of the whole ice sheet; ocean temperatures and heat fluxes, sea-ice conditions, and other potential controls of iceberg calving and melting at the ice-ocean interface are invariant in current models. This point is discussed further in Section 8.3.4.

8.3.3.3. **Feedbacks in coupled models**

Simulations with two-way coupling of the ice-ocean-atmosphere system reveal positive and negative feedbacks on ice sheet retreat in a warmer world. Huybrechts et al. (2002) and Fichefet et

al. (2003) examined the coupled response of a 3-D thermo-mechanical model of the Greenland Ice Sheet and an AOGCM to future climate warming (IPCC B2 emissions scenario, mid-range greenhouse gas forcing). Simulated ice sheet melt is put into the ocean circulation model. Modeled melt rates are modest in the 21st century (global sea level rise of 4 cm in response to a warming of 4.5 °C over Greenland), but Fichefet et al. (2003) concluded that the freshwater addition to the ocean from ice sheet melt would contribute to a weakened Atlantic Meridional Overturning Circulation (AMOC) by the end of the 21st century; a feedback that lowers temperatures over East Greenland and partly counteracts warming in this region.

In a subsequent study, Driesschaert et al. (2007) concluded that Greenland Ice Sheet melting only significantly impacts on the AMOC in the 21st century under severe future warming scenarios. This is echoed in the results of Mikolajewicz et al. (2007b) and Vizcaíno et al. (2008), who used a coarse-resolution AOGCM interactively coupled with an ice-sheet model at 80-km resolution to explore millennial-scale future climate and ice sheet changes in Greenland. They also predicted that a weakened AMOC will reduce the degree of warming over southern Greenland (Figure 8.29), buffering Greenland Ice Sheet melt. However, changes in the hydrological cycle of the North Atlantic region are the main reason for this; Greenland Ice Sheet runoff is about 10% of the total freshwater forcing of the system (Mikolajewicz et al., 2007b).

Vizcaíno et al. (2008) predicted that a doubling or tripling of atmospheric CO_2 would result in 2 to 5 °C of summer warming in Greenland, with greater warming in northern Greenland (Figure 8.29). Simulations with $4 \times CO_2$ trigger a shutdown in North Atlantic deepwater formation, which creates a dramatic reversal in the projections and a cooling of up to 3 °C in southern Greenland.

Atmospheric feedbacks in response to Greenland meltback have been examined extensively through the productive collaboration of Philippe Huybrechts, Jonathan Gregory, and colleagues, using Huybrechts' ice sheet model and the Hadley Centre AOGCM. Toniazzo et al. (2004) examined the effect of removing the Greenland Ice Sheet on regional atmospheric circulation and climate. A lower (or absent) ice sheet reduces the persistent anticyclonic flow over the ice sheet, particularly in winter months, causing a net increase in precipitation in eastern and northern Greenland. Low elevation, reduced albedo, and
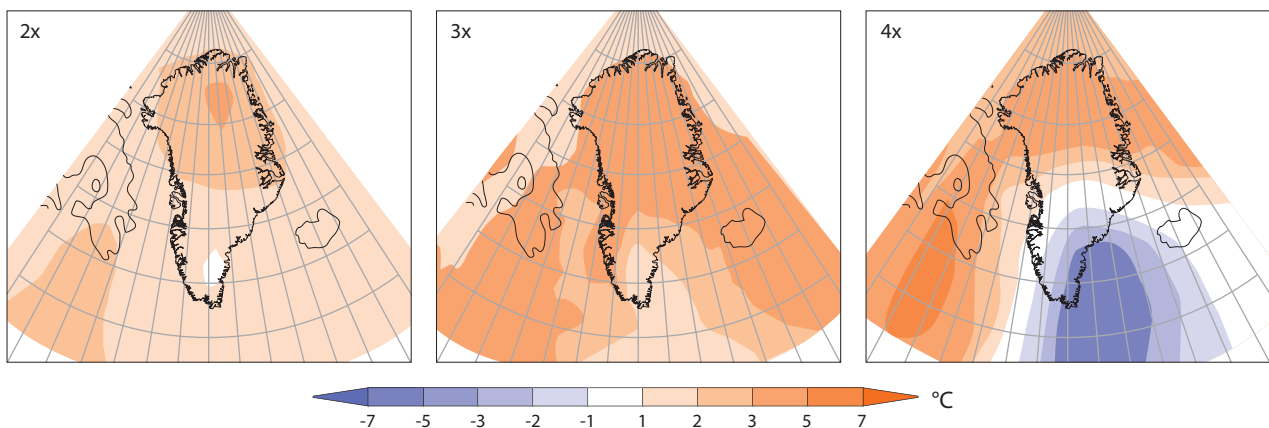

Figure 8.29. Modeled changes in summer temperature over Greenland as a result of 2×, 3× and 4× pre-industrial carbon dioxide concentration scenarios. These snapshots correspond to the period 2900–3000 AD and include ice-dynamical feedbacks. Source: Vizcaíno et al. (2008).

the development of vegetation (e.g., boreal forest) cause greatly intensified summer warming and seasonal melting. No perennial snow is predicted. This prevents the ice sheet from re-growing in today's climate; it is an artifact of the Pleistocene, preserved by its own elevation and regional cooling influence.

Ridley et al. (2005) examined two-way coupling for a $4 \times CO_2$ climate. In this analysis, the ice sheet model output forces the climate model through changes in ice topography, surface albedo, and freshwater runoff. Ridley et al. (2005) concluded that long-term trends are primarily sensitive to topographic feedbacks (i.e., elevation-temperature effects) that are reasonably well captured in the one-way forcing parameterizations of, for example, Huybrechts et al. (1991). There are some additional feedback effects that appear to slow down ice loss. Warming of the ice-free region at the ice sheet margin, which intensifies under climate warming (and with ice sheet retreat) generates a convective circulation that cycles warm air to the interior of the ice sheet, while enhancing the intensity of katabatic return flows at the surface. This cools the lower elevations, reducing melt rates at the ice margin and helping to preserve the ice there.

Ice dynamics introduce another negative feedback in the models during early stages of the ice sheet retreat. As described by Huybrechts and de Wolde (1999) and Ridley et al. (2005), higher ablation at the margins and increased snowfall at high elevations combine to give greater balance gradients and steeper ice sheet slopes. This produces an increased ice flux which buffers drawdown at the ice sheet margin, maintaining thicker ice and reducing the elevation-temperature feedbacks. This reduces the initial rate of ice sheet retreat, although it is not sustainable as drawdown in the interior regions eventually causes an expansion of the ablation area, with elevation and albedo feedbacks that hasten ice sheet retreat.

8.3.3.4. Predicted Greenland Ice Sheet response to warming

Bearing in mind the missing ice-dynamical processes and ice sheet-climate coupling challenges, a number of insights are available from simulations to date. It is clear that the Greenland Ice Sheet is very sensitive to climate warming. In steady-state, a regional warming of more than 3 °C is likely to be sufficient to precipitate ice sheet retreat in southern Greenland (Huybrechts et al., 1991; Alley et al., 2005). This is illustrated in Figure 8.28; the southern dome retreats by 3000 AD in these simulations at CO_2 levels between 550 and 750 ppm, which correspond to a summer warming of 4 to 5 °C in Greenland.

Based on an ensemble of models, including high-resolution (~1°) climate model representations of temperature and precipitation patterns over the ice sheet, Gregory and Huybrechts (2006) concluded that an average annual regional warming of 4.5 °C would push the Greenland Ice Sheet into a negative SMB. The threshold temperature is higher than in earlier studies (e.g., 2.7 °C in Huybrechts et al., 1991; 3 °C in Gregory et al., 2004), in part because summer warming over Greenland, ΔT_S, is forecast to be less than the mean annual warming, ΔT_A: Gregory and Huybrechts (2006) reported an average value $\Delta T_S / \Delta T_A = 0.81$. Summer temperature is the critical parameter for Greenland's mass balance, as this is the main driver of annual surface melt and runoff.

A warming of 4.5 °C in Greenland corresponds to an average global warming of 3.1 °C in the ensemble of models explored by Gregory and Huybrechts (2006). The relationship is plotted in Figure 8.30, along with estimates of rates of sea level rise due to SMB losses as a function of mean annual warming. There is an average polar amplification factor of ~1.5 over Greenland, less than the average value for the Arctic. Part of the moderate polar amplification factor here may be due to the cooling influence of weakened deepwater formation in the Nordic Seas that is predicted by many AOGCMs in the coming decades (see for example Figure 8.29). This is addressed in more detail in Section 8.4.1.

The critical warming value of 4 to 5 °C for Greenland to go into a state of negative SMB is often quoted as a threshold for 'irreversible' ice sheet decline. Dynamical ice losses would place the Greenland Ice Sheet into a negative total mass balance well before this SMB threshold was met – this threshold appears to have been reached in the 1990s and 2000s with a regional warming of close to 1 °C (Section 8.2.4). However, the temperature threshold for SMB to become negative is considered to be terminal because ice sheet thinning would provide a positive feedback to ice sheet decay. If this proceeds too far, ice sheet drawdown would make it difficult to reverse the decline, even with temperature stabilization. However, the timescale for such a decline is long – many centuries to millennia – so it is difficult to judge irreversibility on this timescale.

Charbit et al. (2008) argued that 3000 Gt of cumulative CO_2 emissions could lead past this threshold for an irreversible Greenland Ice Sheet retreat, such that anthropogenic activity in the 21st century could commit the world to several centuries of sea level rise. Ridley et al. (2010) examined this question through coupled modeling with climate restored to its pre-industrial state for several different Greenland Ice Sheet configurations. They argued for a potential two-stage destabilization of the ice sheet in Greenland, with reductions to 80–90% of the current ice sheet volume inducing ice loss in southern Greenland and an irreversible sea level rise of about 1.3 m over several centuries. If the ice sheet retreats further before climate is stabilized (i.e., under greater or sustained
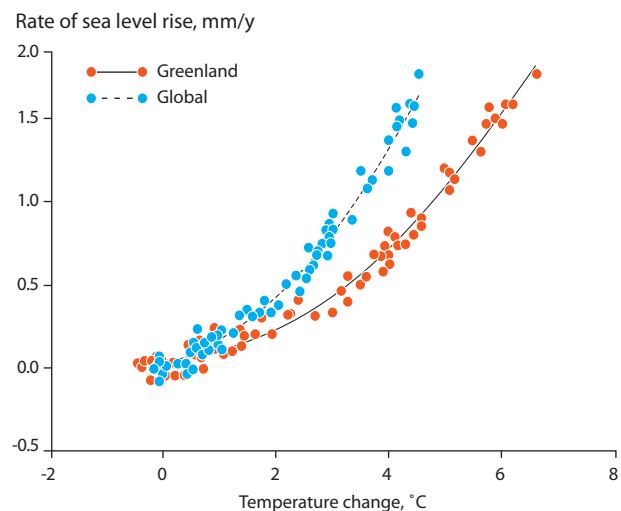

Figure 8.30. Rates of sea level rise as a function of mean global temperature change and the temperature change over Greenland for future climate simulations in the ensemble of AOGCMs examined by Gregory and Huybrechts (2006). Source: Gregory and Huybrechts (2006).

warming), such that the ice sheet falls below half of its current volume, Ridley et al. (2010) predicted an irreversible sea level rise of about 5 m.

Such a decline is consistent with reconstructions of a reduced Greenland Ice Sheet in the last interglacial period (Cuffey and Marshall, 2000), in response to orbitally induced spring and summer warming at high northern latitudes. Other ice sheet modeling studies predict a less severe ice sheet decline at this time (Huybrechts et al., 1991), due to different assumptions about the ice core isotope-temperature relationship. Miller et al. (2006) and Otto-Bliesner et al. (2006) estimated a summer warming of about 5 °C near Greenland at the peak of the Eemian warming. Ice sheet models predict an almost complete collapse of the ice sheet if warming of this magnitude persists for several thousand years, but the modeled retreat takes several millennia – changes over the first few centuries are modest.

For future warming, the extent of modeled Greenland Ice Sheet retreat by 2100 is primarily a function of the change in SMB, hence the climate scenario. Ice dynamics plays only a minor role over this time scale in the current models, for two main reasons: elevation-induced SMB and ice-dynamic feedbacks are minor over a one-century timescale, and ice dynamics models lack a direct connection with climate (e.g., ocean or surface-meltwater forcing). The latter is consistent with ice-sheet model simulations of the past century; there is little to no interannual variability in modeled ice dynamics. This is partly because ice dynamics in the current generation of ice sheet models is sensitive only to local gravitational driving stress (see Section 8.3.2), which changes slowly.

Table 8.5 provides a summary of future sea-level rise forecasts from model studies of Greenland Ice Sheet response to climate change. These studies are difficult to compare directly because they make different assumptions about future climate change, the climate downscaling strategy and SMB models differ in each study, and results are not always documented for 2100. These results all include ice sheet dynamical feedbacks.

Under CO_2 stabilization at concentrations of 2× pre-industrial, equivalent to the IPCC IS92a climate change scenario, Huybrechts and de Wolde (1999) simulated a contribution to sea level rise from the Greenland Ice Sheet

of 3.9 cm by 2100. Equivalent simulations for 4× and 8× pre-industrial CO_2 concentrations give ice sheet retreat equivalent to 10.6 and 20.8 cm of sea level rise. Using a more sophisticated climate treatment with the IS92a scenario, Huybrechts et al. (2004) estimated an average 21st century rate of sea level rise of 0.45 to 0.49 mm/y from an ensemble of climate models: a net sea level rise of 4.5 to 4.9 cm by 2100 (Figure 8.31). The range of model estimates for 2100 from Huybrechts et al. (2004) is 2 to 7 cm; these estimates are all for the IS92a emissions scenario, so this represents the variation in climate models. Alley et al. (2005) gave a similar estimate for 2100, 5 ± 4 cm, for a fuller range of emissions scenarios.

Greenland Ice Sheet retreat has also been studied by van de Wal and Oerlemans (1997), with a simpler (2-D, vertically-integrated) ice sheet model but using an energy balance model to calculate surface melt. They estimated 10.4 cm of sea level rise from Greenland between 1990 and 2100 due to changes in SMB under the IS92a scenario. Depending on assumptions about basal sliding and the initial state of the ice sheet, van de Wal and Oerlemans (1997) estimated a range from 5 to 16 cm of sea level rise. In a follow-up study, van de Wal et al. (2001) examined the impact of different emissions scenarios and reported a similar range of sea-level rise forecasts, from 6 to 14 cm by 2100. Sea level rise accelerates as the century progresses. The reference scenario, IS92a (10.4 cm), corresponds with 4.1 cm of sea level rise by 2070. With climate fields derived from an AOGCM for a similar scenario (2 × CO_2), van de Wal et al. (2001) estimated only 1.3 cm of sea level rise by 2070, which can be extrapolated to 3.3 cm by 2100. Increased snow accumulation in the AOGCM explains the lower total in this study, compared to the energy balance model results driven by the IS92a reference scenario.

The analysis by Gregory and Huybrechts (2006) is the most detailed examination to date of the expected future contributions to sea level rise from warming-driven changes in SMB. Figure 8.32 plots the phase space of simulated sea level rise as a function of change in temperature and precipitation. Decadal changes in temperature and precipitation from a suite of AOGCM simulations for the 21st century indicate projected temperature increases of up to 8 °C and precipitation

Table 8.5. Model projections of future sea level rise (SLR) associated with Greenland Ice Sheet retreat. Results are from ice sheet modeling studies with one-way climate forcing introduced as a perturbation to present-day temperature and precipitation fields, unless noted otherwise.

| Year	SLR, cm	Souce	Climate treatment
2100	4.8	Huybrechts et al., 1991	Warming anomaly increasing to +4.2 °C by 2100
2100	3.9	Huybrechts and de Wolde, 1999	Radiative forcing for stabilization at 2 × CO_2
2100	4.0	Huybrechts et al., 2002	IPCC SRESB2 scenario, single AOGCM
2100	2-7	Huybrechts et al., 2004	IPCC IS92a emissions scenario, ensemble of AOGCMs
2100	1-9	Alley et al., 2005	Range of AOGCM future climate scenarios
2100	10.4	van de Wal and Oerlemans, 1997	IS92a radiative forcing, energy balance model
2100	6-14	van de Wal et al., 2001	As above but a range of IS92 radiative forcings
2070	1.3	van de Wal et al., 2001	2 × CO_2, AOGCM forcing of energy balance model
3000	88-317	Huybrechts and de Wolde, 1999	Radiative forcing, stabilization at 2× and 4 × CO_2
3000	72-240	Greve, 2000	Temperature anomalies of +3 and +6 °C
3000	25	Vizcaíno et al., 2008	2 × CO_2, fully coupled ice model and AOGCM
2600	98	Vizcaíno et al., 2010	2 × CO_2, fully coupled ice model and AOGCM

Sea level change relative to 1975, cm
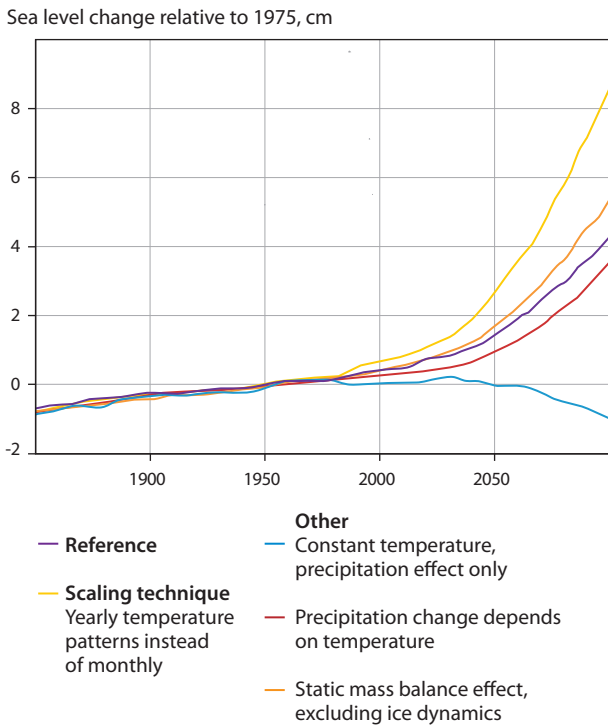
Figure 8.31. Modeled sea level rise from Greenland Ice Sheet retreat, 1860 to 2100, driven by modeled high-resolution (1.125°) temperature and precipitation patterns over Greenland, with temporal climate perturbations from the Hadley Centre AOGCM. The different lines denote different treatments of climate. Source: Huybrechts et al. (2004).

increases of up to 50%. The net effect on SMB is negative in the models; higher melt rates beat increases in accumulation. The corresponding rate of sea level rise due to declining SMB ranges from 0 to 2 mm/y for warmings of 0 to 8 °C. Examining Greenland Ice Sheet sensitivity to changes in temperature and precipitation, along with an ensemble of AOGCM forecasts for

the full spectrum of IPCC AR4 emissions scenarios, Meehl et al. (2007) reported a Greenland Ice Sheet contribution to global sea level of 1 to 12 cm by 2100.

Based on this collection of model studies, the direct SMB effect of expected warming in Greenland is likely to be a sea level rise of 5 to 10 cm by 2100, corresponding to 200 to 400 Gt/y loss of mass in average over the period to 2100. Within the range of anticipated climate change, ice volume losses in excess of 1 m of sea-level equivalent from the Greenland Ice Sheet require several centuries in all published modeling studies to date. The complete demise of the ice sheet requires 3000 years or more under most future climate change scenarios. The most severe climate warming simulated by Greve (2000), that is, 12 °C over Greenland, precipitates a full collapse (and a 7 m sea level rise) in 1000 years. Predictions by Greve (2000) for more moderate warming scenarios of 3 and 6 °C give a sea level rise of 0.72 m and 2.4 m after 1000 years. Contributions by 2100 are about 5% of this. Huybrechts and de Wolde (1999) found a similar rate of extreme ice sheet retreat in their most severe climate change scenario, $8 \times CO_2$, with an average rate of sea level rise of 60 cm per century and about 80% of the Greenland Ice Sheet melting within 1000 years under these conditions. Rates of ice loss increase non-linearly with time in these simulations, as elevation-climate feedbacks strengthen.

While the forecasts are reasonably well bounded for this century, models have a conservative bias for ice sheet response to climate change, due to missing processes in the glaciological models (see Section 8.3.2) and a number of missing climatic feedbacks. There is no sensitivity to changing conditions at the ocean interface in current models, that is, no capacity for outlet glacier acceleration.

Pfeffer et al. (2008) probed this by considering an extreme case where all of the marine-based outlet glaciers of the Greenland Ice Sheet double their discharge to the ocean. By taking the cross-sectional area of all marine outlet glaciers and

Temperature change, K
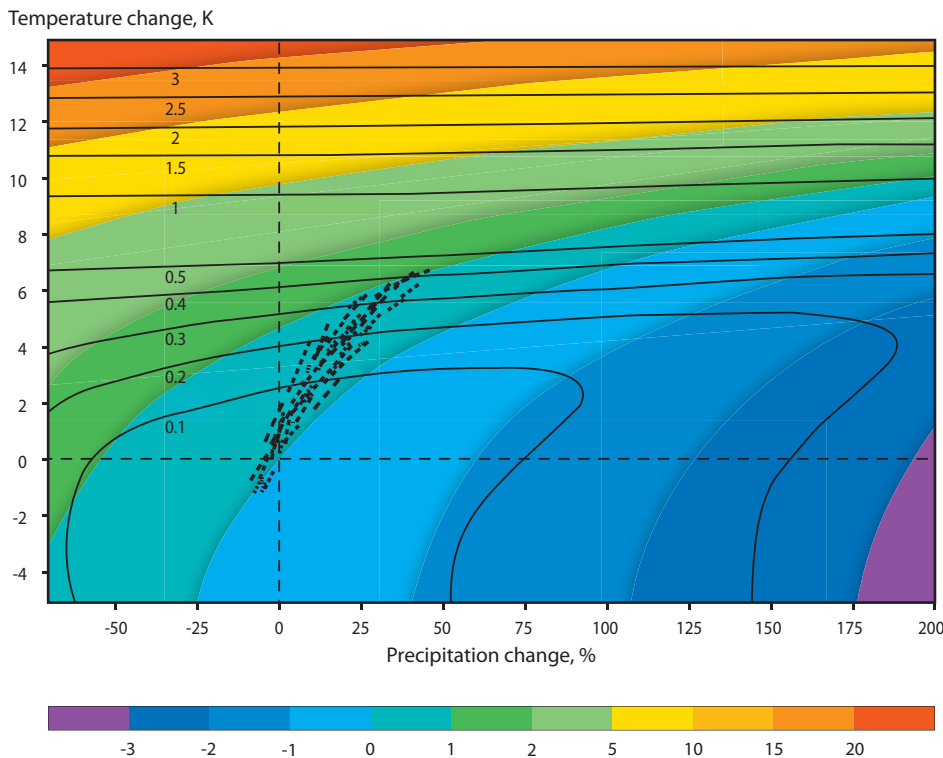
Figure 8.32. Modeled rates of sea level change as a function of changes in temperature and precipitation over Greenland based on a suite of AOGCM simulations for the 21st century. The rate of sea level rise is indicated by the colour, while the contour lines denote the standard deviation in an ensemble of AOGCM forecasts. The points indicate the results of decadal temperature and precipitation shifts in future simulations from the ensemble of models. Source: Gregory and Huybrechts (2006).

Sea level change, mm/y

assuming a constant ice thickness at these flux gates, Pfeffer et al. (2008) calculated that a two-fold increase in outlet glacier velocity would contribute 9.3 cm to sea level rise by 2100. This is an extreme case: all of the outlet glaciers would need to double their velocity immediately and remain at this higher rate of flow for the rest of the century, without significant thinning. This, therefore, provides a plausible upper bound on ice dynamical discharge in Greenland.

Pfeffer et al. (2008) added this term to the sea level rise associated with SMB losses to give an upper bound for Greenland mass loss. However, the Greenland Ice Sheet's 'baseline state' and current estimates of future sea level rise from Greenland (e.g., Huybrechts and de Wolde, 1999; Greve, 2000; Gregory et al., 2004; Huybrechts et al., 2004) include contributions from dynamical ice losses (i.e., iceberg calving). Current models have negligible 21st century variability in this term and muted sensitivity to climatic forcing, relative to recent observations, but there is still a background level of ablation from this mechanism. Hence, it might be appropriate to consider only the extra iceberg discharge associated with a two-fold speed-up of marine outlet glaciers in placing an upper bound on Greenland's contribution to sea level rise. Using the fluxes calculated by Pfeffer et al. (2008), this gives 4.7 cm of sea level rise by 2100. If this is based, instead, on the estimated baseline ice discharge from Greenland in the early 1990s, 350 km³/y (Table 8.3), the additional discharge from a doubling of the ice flux would give a 21st century sea level contribution of 8.8 cm. This can be added to the expected SMB-driven losses of up to 10 cm to give an upper bound of 19 cm of sea level rise from the Greenland Ice Sheet by 2100 corresponding to an average mass loss of 800 Gt/y in the period to 2100. This is not a true upper bound. The value could be higher, if climate change during this century is more severe than expected in Greenland, or as a result of additional climate feedbacks that are lacking in the current models.

8.3.4. Summary and outlook

The complexities of climate and ice sheet models are such that much work is still needed to refine model physics, improve model resolution, and assimilate observational data to better constrain and test simulations of the meteorology, mass balance, and dynamics of the Greenland Ice Sheet. Great strides have been made in coupling of ice sheet and climate models in recent years. RCMs now offer the possibility of grid-matching and a full surface energy balance calculation, to help reduce uncertainties associated with climate downscaling and some of the simplistic assumptions in current parameterizations of SMB.

A number of climate forcings and ice sheet-climate processes are still missing in current models, particularly at the ice-ocean interface. This is pressing in light of evidence that oceanic forcing is responsible for some of the recent, rapid changes in the Greenland Ice Sheet mass balance and ice dynamics (Holland et al., 2008). Most fast-flow processes in ice sheets are not well represented in current models, and potential climatic excitation of fast flow, through either the ocean or atmosphere, has not been explored. This means that the parameter space for Greenland Ice Sheet response to climate change is not fully explored; models cannot give reliable bounds on this.

The kinematic approach of Pfeffer et al. (2008) probably offers the best available upper bound on the ice-dynamical contributions to mass balance losses in Greenland. For a proper dynamical simulation of these effects, the ice-dynamical processes discussed in detail in Section 8.2.3 need to be captured by predictive models. This is important to improved estimation of the total ice sheet mass balance, given the potential for large-scale climate forcing of ice-dynamical ablation mechanisms. At present, ice sheet models are not sensitive to surface meltwater production or to ocean and sea-ice conditions, all of which have begun to change rapidly in the Arctic.

Further development of AOGCMs and regional equivalents is also needed. Little besides what has been reviewed here has been done to couple climate and ice dynamical models. Thus, more coupling experiments represent an obvious area of future work, particularly with respect to regional patterns of climate downscaling and atmospheric/mass balance feedbacks of changing ice sheet geometry.

Other climatic feedbacks on ice sheet mass balance are not fully explored, such as the impact of decreasing ice sheet albedo as the ablation zone of the ice sheet expands spatially and seasonally. Degree-day melt models typically use two melt factors, one for snow and one for ice, rather than the continuum of behavior that is evident in nature. The positive feedbacks of a wetter, lower-albedo snow surface may not be being captured in present model forecasts. More physically-based melt models (e.g., from energy balance calculations on high-resolution grids) will provide improved estimates of the threshold temperature at which the Greenland Ice Sheet switches into a state of negative SMB.

These limitations make it difficult to project the extent of sea level rise that can be expected from the Greenland Ice Sheet in this century and beyond. Current models suggest 5 to 10 cm by 2100, but conservative biases in the models mean that it may be in excess of this. If climate warming excites systematic, widespread acceleration of the outlet glaciers in Greenland, the ice sheet could contribute as much as 19 cm to global sea level rise this century. Extreme warming scenarios (e.g., 10 °C) are needed to extract more ice than this from Greenland by 2100, although long-term ice sheet losses may result from instabilities that are triggered in the coming decades.

The models are still in early stages of coupling and several key processes and feedbacks are absent or poorly parameterized, which compromises the ability of the models to simulate interannual variability in ice velocities (hence, ice discharge to the oceans), as well as some potentially important aspects of ice sheet sensitivity to climate forcing. Some of the key processes that need to be better understood and included in the models include:

- subglacial processes governing basal flow
- effects of surface meltwater on basal flow
- numerical representation of ice stream and outlet glacier dynamics
- oceanic influences on ice flow and ice-marginal ablation
- surface energy balance feedbacks as the ice sheet wanes (e.g., declining albedo).

These missing processes introduce a likely conservative bias in model studies to date, such that they may systematically underestimate Greenland Ice Sheet sensitivity to climate change. Recent observations and simplified (e.g., conceptual)

model studies indicate that increased ice discharge associated with a warmer climate could effectively double the ice sheet sensitivity to climate change, with ice dynamics contributing 5 to 9 cm to sea level rise by 2100. This is on top of the SMB-driven ice sheet losses of 5 to 10 cm, giving a total contribution to sea level rise of 10 to 19 cm.

8.4. Impacts caused by changes in the Greenland Ice Sheet

- The mass loss from the Greenland Ice Sheet will increase and cause a rise in sea level that will impact globally, especially the vast population living close to the coast.

- The increasing freshwater input to the fjords and ocean surrounding Greenland will change salinity and impact on the marine food web and thereby the fisheries.

- Freshwater input to the ocean will affect the thermohaline ocean circulation. But as the component from the Greenland Ice Sheet is less than 10% of the total freshwater input the effect of the increased melt is assumed to be small.

8.4.1. Ocean circulation

Freshwater from the Greenland Ice Sheet melt as well as iceberg calving into the sea have various effects on the waters of the Arctic region (i.e., Arctic Ocean with continental shelves, Nordic Seas, and Labrador Sea) as well as the World Ocean. The Greenland Ice Sheet runoff dilutes the marine surface water and reduces its salinity and density, thereby impacting global sea level rise and two types of ocean circulation: thermohaline circulation and estuarine circulation. All are important for the marine climate of the Arctic region although they are affected in different ways.

8.4.1.1. Sea level change

8.4.1.1.1. Global sea level change

The Greenland Ice Sheet represents, if entirely melted, a global average eustatic sea level rise of about 7 m. In a warmer climate with more precipitation, accumulation, ablation and ice discharge will increase but model studies suggest that the net balance of these factors will be toward a diminishing ice sheet (Gregory and Huybrechts, 2006). Depending on the future temperature evolution, the potential disappearance of the Greenland Ice Sheet or its transformation to a much reduced inland ice cap, will take of the order of millennia (Gregory et al., 2004).

The contribution to global sea level rise from the Greenland Ice Sheet has increased during the past decades (Cazenave et al., 2008) and global sea level has been observed to rise at a faster rate than projected by Meehl et al. (2007). Motivated by this apparent under-estimation of sea level rise from dynamical modeling, Rahmstorf (2007) suggested an alternative semi-empirical approach where a statistical model is fitted to historical time series of global sea level and temperature. By then driving the statistical model with temperature scenarios Rahmstorf obtained total global sea level changes of 0.5 to 1.4 m for the end of the 21st century; larger than by Meehl et al. (2007). This work has, however, been criticized by several authors (Holgate et al., 2007; Schmith et al., 2007; Rahmstorf,

2008; von Storch et al., 2008). An improved semi-empirical model by Grinsted et al. (2010) estimated the likely sea level rise to be in the range 0.7 to 1.6 m at the end of the 21st century.

8.4.1.1.2. Regional sea level change

The increased release of freshwater from the Greenland Ice Sheet will, together with the expected increase in net precipitation over the North Atlantic area, cause changes in the salinity and temperature distribution (Curry et al., 2003), which, through a dynamic response, will cause regional changes in sea surface height. This dynamic response can be either thermohaline in nature, such as weakening of the AMOC, or an adjustment of gyre circulations due to changes in the density distribution. Moving mass from the ice sheet to the ocean causes changes in the gravity field (Mitrovica et al., 2001; Milne et al., 2009). This has remained unnoticed to date, but causes noticeable regional differences in sea level rise. Diminishing the Greenland Ice Sheet is expected to produce the largest sea level increase in the South Atlantic and the North Pacific and a decrease in the North Atlantic and around Greenland. This does not take vertical isostatic land lift into account, which takes place on a longer time scale.

A quantitative assessment of the effect of freshwater release from the Greenland Ice Sheet does not exist at present. Landerer et al. (2007) analyzed a standard IPCC scenario run (i.e., without melting from the Greenland Ice Sheet) and discovered regional signals, most prominent a sea level increase of 0.3 m in the Arctic Ocean and a minor increase (i.e., up to 0.1 m) along the Greenland coast. In the center of the subpolar gyre, Landerer et al. (2007) found a decrease of the order of 0.1 m, which could be due either to changed density distribution or to changed wind forcing, or both.

8.4.1.2. North Atlantic circulation – deep convection and climate impact

Through thermohaline ventilation, the Arctic region is a significant source to the North Atlantic Deep Water as part of the global thermohaline circulation. Therefore, increased runoff from Greenland has the potential to affect the circulation of the Arctic region and the global ocean.

Off Greenland (Figure 8.33), the East Greenland Current flows southward along the shelf towards the Atlantic (Holfort et al., 2008). Together with the outflow through the Canadian Arctic Archipelago, this comprises a low-salinity surface outflow from the Arctic (Melling et al., 2008). The inflow, to compensate for the estuarine outflow, consists of warm and saline Atlantic water concentrated to the eastern part of the North Atlantic and which cannot be distinguished from the direct wind- and thermohaline-driven surface inflow to the Nordic Seas (Hansen and Østerhus, 2000). This inflow enters the Arctic in three distinct branches: a branch to the west of Iceland, a branch between Iceland and the Faroe Islands, and a branch between the Faroe Islands and Shetland (Østerhus et al., 2005; Hansen et al., 2008).

Through heat loss to the atmosphere, the inflow is cooled as it flows northward, and this increases its density. Brine rejection from freezing of seawater also has this effect whereas freshwater input decreases density. The overall effect is to separate the inflow into an upper low-density water mass and a sub-surface high-

density water mass that is then returned to the North Atlantic.

The low-density water mass is returned as surface outflow in the East Greenland Current and through the Canadian Arctic Archipelago, which also carries most of the Bering Strait inflow of Pacific water to the Arctic Ocean (Rudels, 1989). The high-density water mass returns to the North Atlantic through several current branches, termed 'overflows' (Saunders, 2001), and passes through a number of deep channels across the Greenland-Scotland Ridge.

In terms of volume transport, the overflows export roughly twice as much water to the North Atlantic as the surface outflows of the East Greenland Current and the Canadian Arctic Archipelago. The formation of overflow water by thermohaline ventilation involves two steps: density increase and sinking. For the density to increase substantially the water has to be in contact with the atmosphere and this occurs by heat loss in the western part of the Nordic Seas and in the Barents Sea (Simonsen and Haugan, 1996). The sinking may occur through open ocean convection (Marshall and Schott, 1999) in the Greenland Sea and Iceland Sea, but this is only one of the mechanisms (Hansen and Østerhus, 2000) and convection in the Greenland Sea is not likely to be the main source of overflow water (Eldevik et al., 2009).

The overflow water is cold, much of it at temperatures below 0 °C, and is relatively saline compared to other cold water masses in the ocean. Thus, it is one of the densest water masses found anywhere in the World Ocean but, immediately after crossing the Greenland-Scotland Ridge, it mixes strongly with, and so entrains, ambient upper ocean water masses (Dickson and Brown, 1994; Østerhus et al., 2008). The resulting water mass is warmer and less dense, but also has a larger volume transport. Most of this water flows into the Labrador Sea where it contributes to the North Atlantic Deep Water together with the Labrador Sea Water formed locally by convection (Lazier, 1980; Yashayaev et al., 2008). North Atlantic Deep Water is thus from three sources: overflow water, Atlantic water entrained into the overflows, and Labrador Sea Water. Formation rates for some of these are debated (Haine et al., 2008), but there seems to be general agreement that they are of similar magnitudes.

North Atlantic Deep Water is carried south by the deep branch of the AMOC and is the main deep water source of the World Ocean, complemented by Antarctic Bottom Water formed by dense outflows from the Antarctic shelf regions. The AMOC is associated with heat transport on inter-hemispheric scales toward the ventilation areas where heat is released to the atmosphere. This transport of heat by the ocean is modest compared to the atmospheric heat transport (Trenberth and Caron, 2001) and its importance for climate has been challenged (Seager et al., 2002).

However, convincing evidence suggests that, through climate interactions, this ocean transport indeed plays a disproportionate role for fundamentals of the global climate (Rhines and Häkkinen, 2003; Rhines et al., 2008). This also finds support from tailored 'what if' climate model experiments,
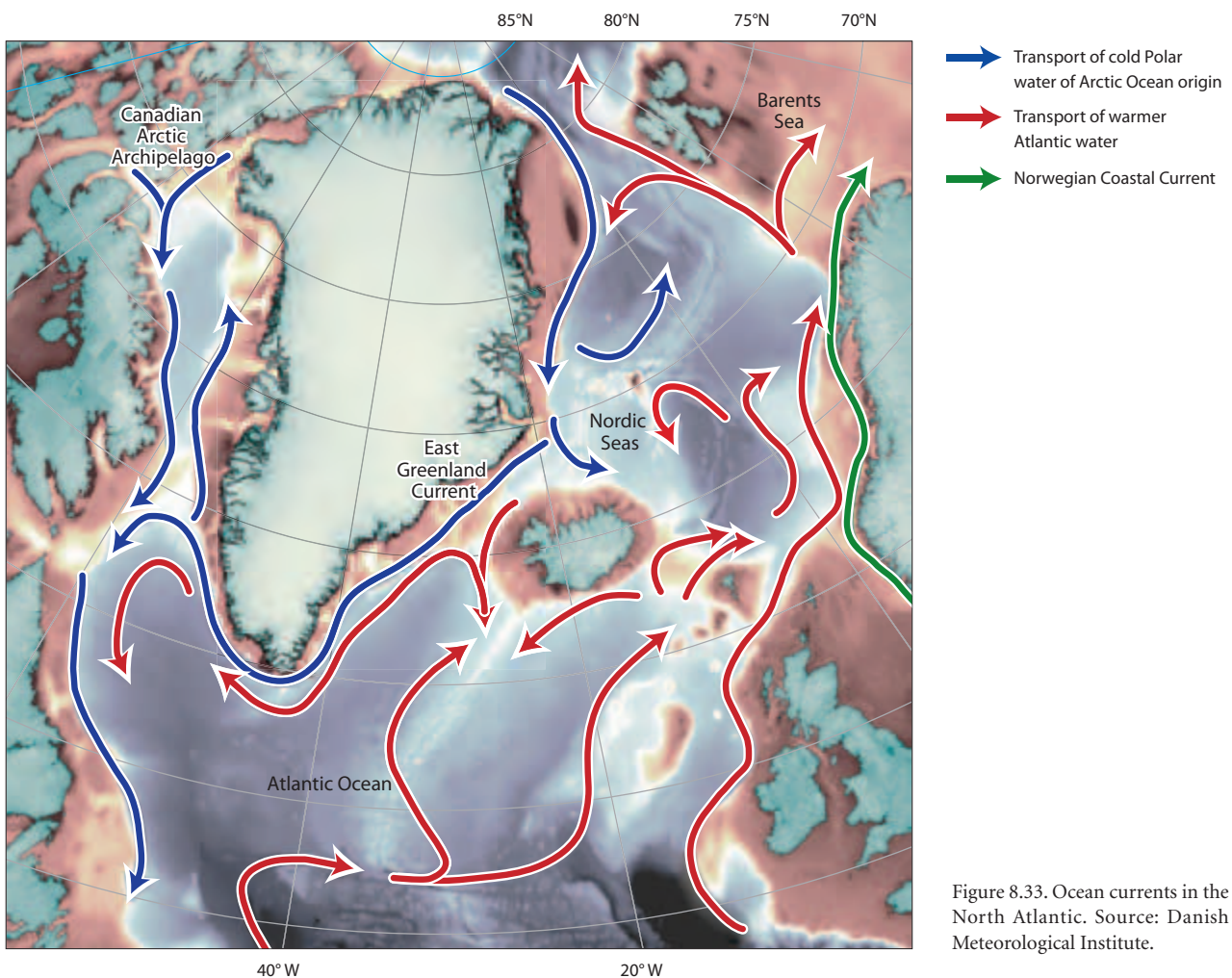

Figure 8.33. Ocean currents in the North Atlantic. Source: Danish Meteorological Institute.

where artificial shutdown of the North Atlantic Deep Water production has been shown to strongly cool the regional climate of the Nordic Seas and Barents Sea but with cooling evident over the entire Northern Hemisphere (e.g., Vellinga and Wood, 2002; Sutton and Hodson, 2005). Apparently, this has occurred in other climatic periods, for example, during the deglaciation period where reduced thermohaline ventilation was likely to have been triggered by increased extreme freshwater supply from the melting glaciers to the ventilation areas (Tarasov and Peltier, 2005). Changes in the AMOC also have indirect impacts on the global climate. The formation of deep water and its subsequent spreading is crucial for maintaining not only the density but also the chemical structure and the biological conditions characterizing the World Ocean (Schmittner, 2005) with feedbacks also on atmospheric chemistry and climate.

A number of confounding factors, however, make quantifying the impacts of increasing runoff from the Greenland Ice Sheet difficult. First, runoff is not the only freshwater source in the Arctic region. In the present-day climate, total Greenland Ice Sheet runoff (0.018 Sverdrup[1]; Dickson et al., 2007) is almost an order of magnitude smaller than the estimated freshwater (including sea ice) transport southward through Fram Strait (0.15 Sv; Carmack et al., 2008), which flows through the areas into which the Greenland Ice Sheet runoff is released. It is also much smaller than the freshwater component of the Canadian Arctic Archipelago outflow (Carmack et al., 2008).

Various changes in the freshwater cycle of high latitude lands and oceans have been reported for the period since the 1950s (Peterson et al., 2006). Changes have been widespread and perhaps dramatic in the case of the Great Salinity Anomaly of the 1970s but, by attribution, anomalous glacial melt has played a minor role, contributing only about 5% to the observed ocean freshwater storage in the North Atlantic and Arctic region. In the future, the Greenland Ice Sheet runoff may well, in relative terms, increase more rapidly than the other sources but even in the most extreme case, the projected peak rate of melting estimated from a coupled climate-ice sheet model (0.06 Sv; Ridley et al., 2005) remains considerably smaller than the contribution from other sources today (Dickson et al., 2007). In idealized anthropogenic climate change simulations, absolute changes in atmospheric moisture transport are also found to exceed changes in runoff (Mikolajewicz et al., 2007a; Vizcaíno et al., 2008), which tends to downplay the isolated role of the Greenland Ice Sheet runoff on ocean circulation.

Second, the effects of increased freshwater supply depend on other factors. The intensity of the estuarine circulation is determined mainly by the rate of mixing between freshwater and seawater thus depending primarily on stratification and winds (Stigebrandt, 2000). For the thermohaline ventilation, the amount of freshwater that can reach an area of deep ventilation is a crucial factor. Most of the Greenland Ice Sheet runoff should be able to reach the ventilation areas in the Labrador Sea, but only runoff from the east coast of Greenland, north of Denmark Strait, can directly affect the offshore regions of the Nordic Seas where overflow waters are produced.

In the present-day climate, even this runoff seems to affect these areas only weakly because most of it is carried south through Denmark Strait by the East Greenland Current

(Dickson et al., 2007; Holfort et al., 2008), depending on the wind conditions over the East Greenland Current (Stigebrandt, 2000), which may well change. Although part of the runoff from the ice sheet may initially bypass areas of ventilation, it is recirculated in the sub-polar North Atlantic and will, on decadal scales, dilute the entire sub-polar North Atlantic and spread to the Nordic Seas (Curry et al., 2003; Curry and Mauritzen, 2005; Peterson et al., 2006). The recent freshening has indeed been traced into the deep overflows from the Nordic Seas (Dickson et al., 2002).

In anthropogenic climate change projections, the effect on the AMOC of increased freshwater supply to the ventilation areas in response to an intensified atmospheric hydrological cycle (Cubasch et al., 2001) may also be partly or completely compensated for by increased salinity of the Atlantic inflow. With enhanced evaporation at low latitudes, the source waters in the sub-tropical North Atlantic are becoming more saline (Curry et al., 2003) and this effect has been shown to stabilize the AMOC (Latif et al., 2000). The pathways of subtropical waters toward the ventilation areas, however, are dependent on the circulation of the subpolar gyre (Hátún et al., 2005), which again depends on Labrador Sea convection (Häkkinen and Rhines, 2004). No such direct compensation, however, will be in effect considering an increased Greenland Ice Sheet runoff resulting from a negative mass balance of the ice sheet, or anomalous freshwater outflow from the Arctic Ocean due to retreating Arctic sea-ice cover.

Although these freshwater fluxes will certainly add to AMOC weakening under anthropogenic climate change, this effect, including changes due to intensification of the hydrological cycle, will possibly be masked by the direct effects of atmospheric warming on the AMOC. In a comparison of 11 different climate models, Gregory et al. (2005) found weakening of the AMOC in idealized global warming experiments to be induced mainly by changes in surface heat flux and to a lesser extent by changes in freshwater fluxes.

In the IPCC AR4, Working Group 1 reviewed the status of climate model projections for the AMOC and found it 'very likely' that the strength of the AMOC will decrease in the 21st century (Meehl et al., 2007). The working group did not expect the AMOC weakening to lead to cooling over Europe and considered an abrupt transition during the course of the 21st century to be 'very unlikely'. However, none of the model studies above take the effects of ice sheets into account.

Very few studies exist where GCMs have been coupled to dynamic ice sheet models (Fichefet et al., 2003; Ridley et al., 2005; Winguth et al., 2005; Mikolajewicz et al., 2007a,b; Vizcaíno et al., 2008). It may also be argued that only in the recent study by Mikolajewicz et al. (2007a) was the coupling complete. This model alone features a realistic climate, allowing reasonable simulation of ice sheets. Nevertheless, the models yield similar meltwater input rates and most, although not all, reveal only a moderate effect of enhanced Greenland Ice Sheet runoff on the AMOC. Thus, it would be premature to conclude on the effects of the Greenland Ice Sheet runoff on the thermohaline circulation on the basis of these few studies.

Whereas fully coupled models will be required to convincingly project future melt rates of the Greenland Ice Sheet and project future global sea level rise, the freshwater impact on the AMOC by other and probably more dominating sources

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[1] A Sverdrop = 1 Sv = 10^6 m³/s.

may be studied independently. This understanding is crucial for assessing the stability of the AMOC under anthropogenic climate change where a large spread in projected weakening for the 21st century among available models (Gregory et al., 2005) indicates a significant model uncertainty. In particular, differences in the projected changes in the hydrological cycle explain differences in the projections of the AMOC (Cubasch et al., 2001; Vellinga et al., 2008). This is one of the key questions to be addressed in the FP7-funded research project THOR (ThermoHaline Overturning - at Risk?) initiated in 2008.

It has been debated whether the projected AMOC decrease has already been initiated. Bryden et al. (2005) suggested that the meridional overturning circulation decreased by 30% between 1950 and 2000 at 24 °N in the Atlantic Ocean with most of the decrease being due to declining overflows. This supports the suggestion of a similar, but smaller, decrease in the overflow transport through the Faroe Bank Channel (Hansen et al., 2001).

Direct measurements have demonstrated a high stability of this flow between 1995 and 2005 (Hansen and Østerhus, 2007) and Olsen et al. (2008) found no weakening of either this overflow branch or the total overflow transport from 1948 to 2005 in a study combining measurements with an ocean model. Concerns have also been raised about the methodology of the Bryden et al. (2005) study (Baehr et al., 2007; Cunningham et al., 2007; Wunsch, 2008).

Conversely, there is evidence in the Labrador Sea of a weakened convection since the mid-1990s, but the formation rate of Labrador Sea Water exhibits large variability (Haine et al., 2008) and attribution of this weakening to anthropogenic causes would be premature and, on the present observational basis (Peterson et al., 2006), it seems unlikely that enhanced Greenland Ice Sheet runoff has played a prominent role.

8.4.1.3. Sea ice

Sea-ice formation depends on the stability of the surface layer of the ocean. Changes to the surface layer incurred by increased meltwater runoff from the Greenland Ice Sheet in a warmer climate could therefore potentially influence sea-ice production.

8.4.1.3.1. Freshwater flux

An increased freshwater flux may lead to increased stability of the upper ocean, which could precondition greater sea-ice production. On the other hand, ice cover of the marginal seas around Greenland has reduced during the present global warming period (satellite data from the period 1978 to 2008). The total effect of higher annual mean air temperatures and increased freshwater flux is not well predicted.

In the Greenland Sea, the role of local sea-ice production and advection was shown to potentially precondition the ocean for deep convection (Visbeck et al., 1995; Toudal and Coon, 2001). Brine rejection during sea-ice formation destabilizes the ocean and advection of the ice out of the area removes excess freshwater. After a period of ice formation and advection, the density gradient will be eroded and further cooling can generate deeper convection. With sufficient cooling, this mechanism can remove excess freshwater from the surface of the Greenland Sea and counteract an increased freshwater flux from Greenland in relation to deep water formation. However, as the mechanism

is reliant on subsequent cooling during winter, it is not clear if it will prevail in a warmer climate.

Few examples of coupled global climate model predictions of the response of sea ice to an increased freshwater flux from the Greenland Ice Sheet have been published. Fichefet et al. (2003) showed indications of a possible feedback mechanism where a breakdown of the thermohaline circulation leads to a significant cooling over eastern Greenland and northern North Atlantic waters, amplified by the sea-ice albedo feedback related to an expansion of the sea-ice cover in the Greenland-Iceland-Norwegian seas. The cooling subsequently reduces the freshwater flux from Greenland.

Feedback of less ice on the state of ice shelves

The reduced summer ice cover at the Arctic Ocean margins (e.g., Greenland Sea and Barents Sea) exposes the coastline of these regions to the influence of ocean swells. This may cause increased coastal erosion but may also directly influence the stability of floating glaciers, such as Storstrømmen and 79-fjorden, in northeastern Greenland.

Reeh et al. (2001) studied the interaction between sea ice and glacier, in the case of the floating tongue of the 79-fjorden glacier, northeastern Greenland (79°30' N, 22° W). The authors used information from glaciological and geological studies, expedition reports, aerial photographs and satellite imagery to document the position of the glacier front and fast-ice conditions on millennial to decadal time scales. The study indicates that stability of the floating glacier margin is dependent on the presence of a protecting fast-ice cover in front of the glacier. In periods with a permanent fast-ice cover, no iceberg calving occurs, but after fast-ice break-up the glacier responds with a large ice discharge activity, whereby several years worth of accumulated glacier ice flux suddenly breaks away.

Climate-induced changes in sea-ice conditions in the Arctic Ocean with seasonal break-up of the nearshore fast ice could lead to the disintegration of floating glaciers. The present dominant mass loss by bottom melting would then be largely overtaken by grounding line calving of icebergs and the local influx of freshwater from north Greenland glaciers to the sea would be reduced and local iceberg production would increase.

8.4.1.4. Greenland coastal currents, including fjords, circulation and stratification

The ocean currents around Greenland are part of the cyclonic sub-polar gyre circulation of the North Atlantic and the Arctic region (Figure 8.33). The surface waters in southeastern and western Greenland are dominated by two very different water masses: warm and saline Irminger Water, a side branch of the North Atlantic Current, and cold and low-saline Polar Water originating from the Arctic Ocean.

These water masses meet in the northern Irminger Basin and in the Denmark Strait. The main branch of the Irminger Current turns west toward East Greenland where it meets the Polar Water, which is transported southward along East Greenland within the East Greenland Current. The strength of these two currents determines the hydrographic conditions around southeastern and West Greenland. The two water masses flow side by side forming large meanders at the front where intensive mixing takes place. Polar Water is found at the surface on the continental

shelf whereas Irminger Water is found over the continental slope and partly below the Polar Water on the deeper parts of the continental shelf. As they round Cape Farewell the Irminger Water subducts under the Polar Water.

Adjacent to the coast, the surface water is modified by runoff from Greenland whereby the stratification in the water column increases. During summer, this forms a freshwater baroclinic jet, named the East Greenland Coastal Current, which is only a few tens of kilometres wide but with maximum velocities exceeding 1 m/s (Bacon et al., 2002; Wilkinson and Bacon, 2005). The runoff originates from melting of the Greenland Ice Sheet as well as from precipitation. Kiilsholm et al. (2003) showed that by far the greatest value (i.e., precipitation minus evaporation) over Greenland is found in southeastern Greenland, which could be the main driver behind the East Greenland Coastal Current. If so, the presence of significant, baroclinic coastal currents is limited to southwestern Greenland waters, but this is still an open question, especially due to the lack of observations in northeastern Greenland waters.

The classic time series of temperature and salinity on top of Fylla Bank describes the observed hydrographic conditions off West Greenland. For the temperature curve, the most striking feature is the three cold periods centered around 1970, the early 1980s and the 1990s; all three related to atmospheric forcing (Buch et al., 2004). The cooling in the late 1960s was associated with a large pulse of freshwater that left the Arctic Ocean through Fram Strait and moved rapidly southward in the East Greenland Current and later northward in the West Greenland Current, and it was easily recognized in the Fylla Bank salinity. This event has been labeled the 'Great Salinity Anomaly' (GSA) (Dickson et al., 1988; Belkin et al., 1998), and Curry and Mauritzen (2005) estimated the freshwater flux to be about 0.07 Sv. This value is in agreement with model simulations preformed by Haak et al. (2003) and Olsen and Schmidt (2007).

Interestingly, the freshwater anomaly associated with the GSA is similar to the expected freshwater anomaly due to the forecast maximum melting of the Greenland Ice Sheet. In other words, the natural variability observed today in freshwater fluxes is similar to the expected additional contribution from the Greenland Ice Sheet. However, events similar to the GSA are expected in the future and more importantly, as noted by Vizcaíno et al. (2008), the atmospheric moisture transport is expected to increase by one order of magnitude more than the freshwater increase due to melting of the Greenland Ice Sheet.

In a future climate with increased runoff due to increased precipitation and melting of the Greenland Ice Sheet, the East Greenland Coastal Current would be expected to increase. In a study of climate changes in Greenland waters and surrounding seas, high-resolution regional models were applied to calculate a scenario of the climate for the period 1950 to 2050 (Stendel et al., 2007; Kliem et al., 2007). The regional models were used to downscale a global AOGCM.

The ocean model was forced by the GCM simulation at the lateral boundaries and by a regional climate atmospheric simulation at the surface. Runoff was not included due to melting of the ice sheet directly in the simulations. However, an indirect effect was included by restoring the surface salinity toward the surface salinity of the GCM. Still, Vizcaíno et al. (2008) found that the increase in freshwater flux from the

Greenland Ice Sheet is one order of magnitude smaller than the increase in the atmospheric moisture transport under anthropogenic greenhouse gas forcing.

The slight increase in temperature is due to a general atmospheric warming and not caused by the retreat of sea ice, as it is rarely present at Fylla Bank. Moreover, the simulations show a clear freshening at Fylla Bank that is most likely to be a result of the increased hydrological cycle. Note that the major salinity fluctuations are due to major GSA-like events. These natural variations are an order of magnitude higher than the general freshening.

8.4.1.4.1. Fjords

Deep fjords are characteristic elements of the Greenland coastline. They constitute a key element of the land-ocean interface as they connect the ice sheet to shelf waters around Greenland. In addition to tide, runoff and local winds are the main driving forces acting on the upper water masses in a fjord system (Svendsen et al., 2002). Frequently, there is a large climate gradient from the inner parts of the fjords to the outer parts. Also, meltwater from the Greenland Ice Sheet or rivers that drain into the inner parts of the fjords also causes a strong salinity gradient along the fjord axis. Typically, meltwater discharge takes place over a few months when temperatures exceed 0 °C and the discharge often occurs in pulses (Hasholt et al., 2008).

Where a glacier meets ocean water, mass is not just lost by the visually spectacular calving mechanisms. At ice shelves and floating tongues it is known from mass conservation considerations that sub-ice melting can occur at rates exceeding 10 m/y (Rignot and Jacobs, 2002; Rignot and Steffen, 2008). Also, Motyka et al. (2003) proposed a model of freshwater-driven fjord convection that brings in warm saline bottom water and transports cold fresh surface water out of the fjord. Recent observations of high water temperatures (3 to 10 °C) in the immediate vicinity of tidewater glaciers (Motyka and Truffer, 2007; Ritchie et al., 2008; Rysgaard et al., 2008) indicate that the ocean water around Greenland has a large heat content and a corresponding melting potential. Changes in the ocean have been observed to coincide with glacier changes in this system and elsewhere in Greenland (e.g., Howat et al., 2008; Holland et al., 2008).

Numerical simulations of the hydrographic conditions under a warmer climate scenario exist from a northeastern Greenland fjord (Young Sund, 74° N) that is connected to the Greenland Ice Sheet via river runoff. However, no large floating glaciers are present in the fjord system. In a warmer climate, the runoff from land would increase and this would locally change the circulation and influence the biological production (Rysgaard et al., 2003; Bendtsen et al., 2007). The sensitivity of Young Sund to changing runoff shows that in the case of no runoff, the mixed layer thickness is about 9 to 11 m during the summer period due to a weak pycnocline in the upper part of the water column (Figure 8.34).

At a runoff corresponding to only 50% of the normal runoff, the halocline is quickly established, and the mixed layer is about 0.7 to 1.6 m deeper in July and August than under present-day conditions where the mixed layer depth is about 4.0 to 5.2 m in July and August. At a runoff two times greater than in the
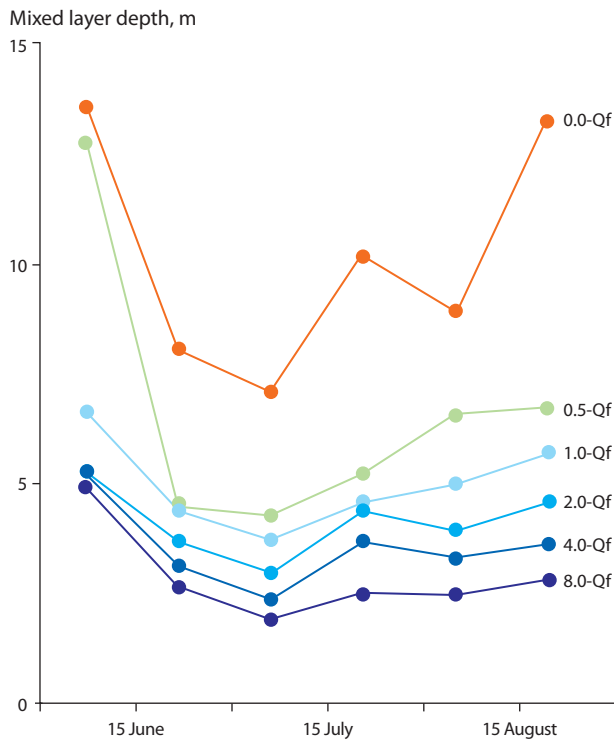
Mixed layer depth, m

Figure 8.34. Model solutions from a sensitivity study of the relation between the mixed layer depth in the central part of Young Sound and the freshwater discharge (Qf) where the present-day discharge (corresponding to 1.0) is scaled between 0 and 8. Source: adapted from Bendtsen et al. (2007).

reference case, the mixed layer becomes 0.3 to 1.1 m shallower during the summer season than in the reference case, and in the extreme case when runoff is 8 times greater, mixed layer depth decreases to about 2.2 to 3.7 m during the summer season.

These sensitivity studies show the importance of the runoff for controlling the depth of the mixed layer. Without a runoff, the mixed layer is controlled solely by wind-induced mixing and buoyancy fluxes at the surface. Even at a moderate runoff of only 50% of the present level, a freshwater controlled mixed layer is established quite early by the end of June, and is only slightly deeper than in the reference case. At the other extreme, when runoff is large, the mixed layer depth is controlled by the strength of the surface forcing on the system, that is, wind and air temperature. Thus, the case with no runoff puts an upper limit on the mixed layer depth of about 11 m in the fjord during the summer season, and, correspondingly when runoff is 8 times the current level this puts a lower limit on the mixed layer depth of about 2 m.

In a future warmer climate scenario, the runoff period might increase significantly, and this would prolong the period in which surface conditions are controlled by freshwater discharge. However, this would not be expected to decrease the mixed layer depth significantly because the balance between runoff and atmospheric forcing is established within a few weeks.

Whereas mixed layer depth would change little due to increased freshwater runoff, it may alter significantly the transport of saltwater into the fjord from the Greenland Sea due to increased estuarine circulation. Hence, in a future climate scenario primary production is expected to increase due to a combination of increased nutrient import to the fjord and improved light availability as a result of reduced sea ice (Rysgaard et al., 1999, 2003; Bendtsen et al., 2007).

8.4.1.5. Summary

Melting of the Greenland Ice Sheet will affect the physical marine environment on different scales, ranging from local scales to global scales. The most visible effect, which is also the one that attracts most attention in public debate, is the rising water level of the world ocean. The latest assessment from the IPCC (2007) predicts the water level to rise between 0.18 and 0.59 m by 2100. This forecast has been questioned by several scientists, however, because recent data show that global warming seems to accelerate the melting of the Greenland Ice Sheet, which will lead to an increased rise in sea level. Estimates are still rather uncertain but recent studies project a sea level rise of 0.5 to 1.6 m by 2100.

More locally, the most visible effect of the melting ice sheet on the marine environment may be the changes in salinity conditions especially in the Greenland fjords and coastal waters. These may result in stronger gradients between the relatively low-saline surface water and the more saline water below; stronger gradients will have a great impact on primary production.

Although the increased runoff of meltwater from Greenland will reach the areas of deep water formation in the North Atlantic (i.e., the Greenland Sea and Labrador Sea), the freshwater volume from the Greenland Ice Sheet is around one order of magnitude less than that from other freshwater sources (i.e., the Arctic Ocean), and so it seems to be of minor importance to changes in the global thermohaline circulation although this has not yet been fully proven.

8.4.2. The marine environment and marine ecosystems

Almost no research has focused on the effects of increased melting of the Greenland Ice Sheet on marine ecology. Present knowledge on the impact on marine ecology due to climate change in the Arctic region is mainly related to the melting or reduction of sea ice although increased meltwater production will also influence the marine food web especially in the fjords.

8.4.2.1. Primary and secondary production

Overall, predictions of how the marine food web will respond to climate change are challenging due to the non-linear nature of ecosystems. Changes in the phytoplankton influence the copepods that, in turn, influence the phytoplankton through changes in grazing pressure. The same is true for the relationship between copepods and their predators, as well as for relationships further up the food chain. Thus, more research is needed before the essential details of the effects of climate change on marine organisms at the base of the food web can be understood and reliable predictions of ecosystem changes can be made.

In summary, climate-mediated changes of the Greenland Ice Sheet and subsequent changes in physical properties of the marine environment may have large consequences for the succession, composition and production at the base of Arctic marine food web. These changes will propagate up the food web and ultimately affect populations of fish, birds and marine mammals with major consequences for the structure, productivity and exploitation potential of the Greenland marine ecosystem.

8.4.2.2. Important economic species

Greenland waters are important fishing grounds, especially for the Northern shrimp (*Pandalus borealis*) and Greenland halibut (*Reinhardtius hippoglossoides*) fisheries that are essential to the economy of Greenland (Buch et al., 2004). Northern shrimp, currently the most important marine resource in Greenland, accounts for more than 70% of the total fisheries revenue. Catches have gradually increased to around 150 000 t/y (Kingsley, 2007), with 90–95% harvested off West Greenland. Northern shrimp in West and East Greenland waters are considered separate stocks.

Greenland fish stock composition and stock size is believed to change as an effect of climate change but there are, however, pending research projects to further study and quantify such changes. The effect of increased melting of the Greenland Ice Sheet, however, is believed to be of minor importance compared to the effects of increasing temperature especially on the offshore fishing grounds.

8.4.2.3. Marine apex predators in Greenland

Currently, there are only a few predictions of the direct effects of the melting of the Greenland Ice Sheet on marine apex predators. However, this large-scale event will have cascading effects on the primary productivity and associated species in the marine ecosystem; thus, many indirect effects can be expected.

8.4.2.4. Summary

Generally, climate change has immediate effects on the physical environment of the ocean and in Greenland waters these effects are increased by a strong inflow of freshwater from the melting Greenland Ice Sheet. These changes may have large consequences for succession, composition and production at the base of the Arctic marine food web. The changes will propagate to levels higher up the food web and will ultimately affect populations of fish, birds and marine mammals with major consequences for the structure, productivity and exploitation potential of the Greenland marine ecosystem.

To date, there has been almost no research on the effect of the melting Greenland Ice Sheet on marine ecology; therefore much of the present knowledge on the effects of climate change is related to the melting and reduction of sea ice. Research is therefore needed to quantify how the higher trophic levels of the food chain will be affected by changes in the physical marine environment around Greenland as a result of the increased melting of the Greenland Ice Sheet. To be able to make proper and scientifically based plans and decisions for the future, such research is extremely important to the Greenland society since its economy is highly dependent on the surrounding ocean.

8.5. Socio-economic and cultural aspects of changes in the Greenland Ice Sheet

- Impacts of climate change cannot be studied in isolation from other environmental and social and cultural changes.
- Until now, the observed effects of changes in the Greenland Ice Sheet due to global warming have caused only minor impacts on society. The most marked change observed by the Greenland population has been the retreat of Jakobshavn Isbræ. This has caused an increase in tourism and has increased international attention on Greenland as an icon of global climate change.
- Planned hydroelectric power plants rely solely on runoff from the Greenland Ice Sheet and retreat of the ice sheet. To be economically viable they need to be based on reliable projections of the future evolution of the ice sheet.
- Retreat of the Greenland Ice Sheet will open areas for mining but will also result in longer walking distance in relation to hunting.

This chapter evaluates the possibility for projecting socio-economic and cultural impacts on Greenland society caused directly or indirectly by changes in the Greenland Ice Sheet. The chapter describes the modern Greenland society from a historical and contemporary perspective and introduces a number of specific cases that illustrate the propensity for change in a society that is derived from the Inuit culture; a culture that has survived over a millennium of climate change and external cultural influences.

It is essential to emphasize that impacts of climate change cannot be studied in isolation from other environmental and social and cultural changes. History shows that Arctic societies, like Greenland, have always changed and have always demonstrated a great ability to adapt to change. The implications of climate change have been the strengthening of ongoing socio-economic changes, and have contributed to outlining possible development paths. On the other hand, socio-economic and cultural changes have been reflecting a more or less pragmatic approach both to changes in the environment (including climate change) and the political reality.

8.5.1. History of change and contemporary socio-economic and cultural trends

The shift from an economy based on marine mammals to one based on fisheries has had profound consequences on society, first of all through the sedentarization process. The shift from a cod- to a shrimp-based economy furthermore consolidated an already ongoing process of concentration and centralization, but did not totally restructure the organization of the communities. The present-day diversification both of renewable resource exploitation and the economy in general may contribute to a partial reversal of the former concentration policy by making some of the settlements located near new resources and new economic opportunities (e.g., tourism) a potentially more active part of the economy (Rasmussen and Hamilton, 2001; Hamilton et al., 2003).

Although Greenland has considerable raw material deposits, commercial exploitation has been limited by adverse natural conditions and difficult logistics. Cryolite, coal, copper, marble, zinc, lead and silver as well as gold, olivine, and rubies have all been mined – and the latter three recently extracted. International interest in exploration for minerals and hydrocarbons is increasing and some deposits may prove of economic interest in the future not least because of increasing accessibility due to changes in the Greenland Ice Sheet.

Hydroelectric power is a new potential resource, not for

energy export but for domestic consumption and for energy intensive activities such as aluminum production. It is important to note that hydroelectric power is based entirely on water basins supplied by meltwater from the Greenland Ice Sheet.

8.5.2. Implications of changes

The new expectations for large-scale mineral and energy resource production as well as the dependence on fisheries and hunting activities are influenced by changes in climate and associated changes in the Greenland Ice Sheet and sea-ice conditions. Renewable resources are affected through their interactions with the environment, while non-renewable resources are affected by changes in accessibility and transportation.

The smaller settlements will experience the immediate effects of these changes because renewable resources are still of significance to them and because access to resources and to settlements will be altered by changes in accessibility, abundance and geographic distribution of these resources, as well as by changes in their economic importance (Poppel, 2007). Tourism has become an alternative source of income for many local communities, enabling a positive interaction between new economic opportunities and traditional activities that are less affected by change than exploitation.

At a longer perspective, Greenland's economy is shifting toward increased dependency on non-renewable resources and in so doing will be highly affected by the ongoing changes, including those affecting access to resources and transportation. The decisive force will be the population's ability to adapt to the changes and to continue to improve the level of education (Hamilton and Seyfrit, 1993; Paldam, 1994; Rasmussen, 1997, 1998, 2002; Winther, 1999, 2001; Zelarney and Ciarlo, 2000; Duhaime et al., 2001; Hansen, 2003; Nellemann and Vistnes, 2003).

8.5.3. Influences

8.5.3.1. Accessibility and impact on communities

Changing ice conditions, both on land and at sea, are important factors influencing everyday life in most communities in Greenland, mostly by determining the accessibility of the water bodies. Access to the sea, and sea transport to and within Greenland, are determinant factors for hunting, fishing, exchange of goods and tourism, which is becoming increasingly important for many communities.

Hunting terrestrial mammals is contributing both to subsistence and informal economic activities of many Greenlanders (Dahl, 2000; Poppel and Kruse, 2009). Muskox (*Ovibos moschatus*) and caribou / wild reindeer (*Rangifer tarandus*) are hunted by licensed commercial hunters and by sports hunters in the autumn, when as much as a fifth of the population goes hunting. Changes in the position of the Greenland Ice Sheet margin may affect hunting activities by changing the routes of rivers draining meltwater from the ice, and in conjunction with rising temperatures may affect the migration patterns of the game. In cold weather, the animals move toward the coast, where they become more accessible to hunters, while in warmer weather, the animals move toward cooler conditions closer to the ice margin in order to reduce insect harassment. As a result, the walking distance of the hunters to the animals becomes increasingly difficult to cope with, first while searching for the animals, and then especially when the meat is carried on the shoulders back to the boats anchored along the coast.

8.5.3.2. Possible consequences for new activities

Reduction of glaciers, ice sheet and sea ice in and around Greenland will increase accessibility to new areas with the potential for commercial exploitation of minerals and energy resources. A significant side effect to the retreat of the ice sheet margin will be the increased volume of meltwater that can be harnessed for producing hydroelectricity. This potential is already included in the new economic strategies for Greenland. Over the past 50 years, the hydroelectric power perspectives in Greenland have been mapped and evaluated as a potential contribution to the development of the national economy. The hydroelectric power exploitation has already emerged in the form of three hydropower plants: A 30 MW plant near Nuuk, a 1.2 MW plant in Tasiilaq, East Greenland, and a 7.2 MW plant near Qaqortoq, South Greenland. A 15 MW hydropower plant near Sisimiut, West Greenland, is under construction and will become operational in 2010.

Several other hydropower plants are in the planning process, now aiming at providing power to new economic activities connected to energy intensive large-scale raw material processing. The first project is the supply of energy to an aluminum production plant currently planned at Maniitsoq, West Greenland. Even though Greenland does not have the ore resources needed for aluminum production, import of ore (i.e., bauxite) from Australia or South America to Greenland could be a viable economic activity given the access to large and continuing amounts of meltwater from the Greenland Ice Sheet and the derived hydroelectricity.

A key question, however, is the stability of the water supply to the new power plants as the ice sheet margin retreats due to melting. Increased amounts of silt and sand in the meltwater may fill dammed lakes and affect the viability of the dams, eventually requiring additional reconstruction in order to maintain the water flow needed. Or worse, the bedrock topography exposed by the retreating ice may cause new meltwater drainage patterns, leading to a partial or complete drying out of the water supply to the hydropower plants (Anon, 1994; Udvalget om socioøkonomiske virkninger af olie- og gasudvinding samt mineralindustri, 1997; Rasmussen, 1997, 2005; Keddeman, 1998; Storey and Hamilton, 2003).

8.5.4. The Ilulissat case

The Jakobshavn Isbræ (i.e., the Ilulissat Glacier) at the head of the Ilulissat ice fjord is one of the world's fastest moving glaciers producing more than 10% of the total discharge of icebergs from the ice sheet (Figure 8.35). During the culmination of the last Ice Age 21 000 years ago, an ice sheet that reached out onto the margin of the continental shelf covered the present-day ice-free land bordering the Greenland Ice Sheet.

After the end of the Ice Age, the melt-back of the ice started about 11 700 years ago and, during a warm climatic period 8000 years ago, the ice rapidly regressed causing the glacier front to retreat inland to a position more than 50 km east of the present grounding line. With the onset of the Little Ice Age, the glacier front again advanced, and in 1851, a maximum extension was recorded in the middle of the ice fjord close to the former Inuit settlement Qajaa. Following this advance, the ice front stabilized

and a slow regression prevailed until the front suddenly collapsed in 2002/2003 causing the floating ice tongue to withdraw more than 15 km eastward in just a few years. This rapid withdrawal ceased in 2007, when the ice front reached the grounding line.

8.5.4.1. Inuit settlements, cultural changes and role of the Ilulissat ice fjord

The glacier and the ice fjord have had an immense impact on the population of the region during the history of human occupation. The glacier produces a large amount of nutrient-rich meltwater that facilitates the primary production in the fjord. The constant input of nutrients thus forms the basis for a rich fauna in the fjord, which has provided excellent hunting and fishing conditions for people inhabiting Disko Bay through the ages. Prehistoric hunting areas and settlements were established and abandoned as the location of the glacier front changed. In this way, the Ilulissat area illustrates the interplay between an ever-changing glacier environment and the development of human occupation.

Many archaeological findings testify to the long history of settlement in the Ilulissat ice fjord region, where hunting and fishing have always been the basis for life. The Saqqaq people reached West Greenland from Alaska 4500 years ago and settled in the Ilulissat area where they lived for 1500 years before they disappeared. West Greenland was then uninhabited until a new wave of paleo-eskimo settlers, the Dorset people, arrived 2800 years ago. They occupied the region for 800 years before they disappeared leaving the Ilulissat region and Greenland uninhabited for the next 1000 years. By 1200 AD, people of the Thule culture settled in the region and established two flourishing communities, Sermermiut and Qajaa, along the ice fjord. These settlements are among the largest prehistoric Inuit

settlements discovered in Greenland and, together with the dense pattern of smaller settlements in the area, indicate that the living resources around the ice fjord were rich, stable and plentiful over long periods of time (Grønnow and Meldgaard, 1988).

8.5.4.2. The Ilulissat ice fjord and present-day socio-economic effects

In 1741, the Danish colony of Jakobshavn (today Ilulissat) was established and this led to the depopulation of the rich Sermermiut settlement around 1850. Today, Ilulissat is the third-largest town in Greenland with a population of more than 4000. Half of Greenland's coastal fisheries are now centered here and the town is also home to one of the country's largest fish-processing factories. Greenland halibut are plentiful throughout the fjord system and are subject to fishing based on a quota system. Long-line fishing for Greenland halibut from the sea ice in the Ice Fjord has been an important part of the fishing tradition through centuries and it is still taking place.

The abundance of resources and the adaptability of the traditional culture to the changing conditions of the glacier and winter sea ice have been crucial to the prosperity of the social and economic life of the region. Although the remarkable present changes in the ice will constitute new challenges to economic and social activities in the region, the conditions may now be different, less predictable and more challenging than before: Icebergs are smaller and the immense discharge by the glacier in combination with a rise in sea temperature, has caused a considerable reduction in sea-ice cover during winter.

This may imply profound changes for the production of nutrients – the basis for biological life in the region – and may lead to smaller stocks of Greenland halibut and northern shrimp. These species are fundamental, not only to human harvesting,
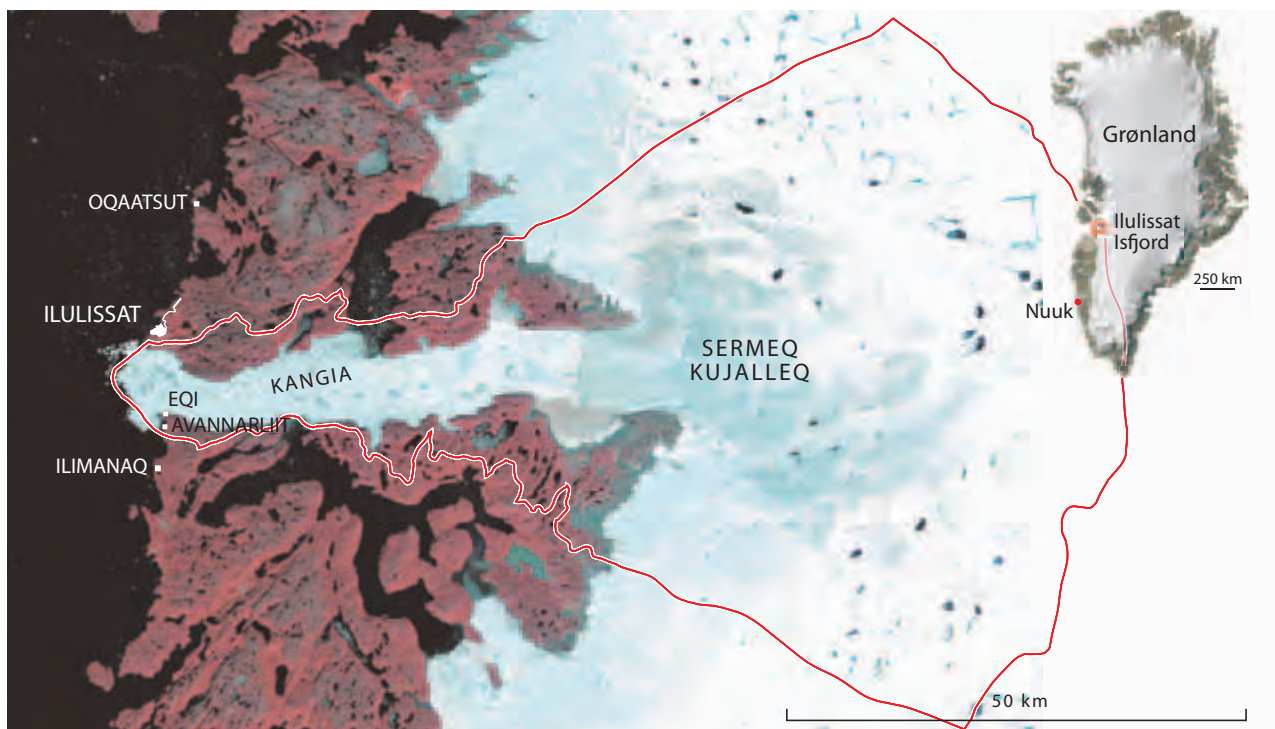

Figure 8.35. Location of Jakobshavn Isbræ and adjacent region. Place names: Ilulissat (town of Ilulissat; formerly named Jakobshavn); Kangia (the ice fjord proper, white because of tightly packed floating glacial ice and stranded icebergs); Sermeq Kujalleq (Jakobshavn Isbræ; also known as Ilulissat Glacier). Copyright by Geological Survey of Denmark and Greenland (GEUS).

but also to the seals and whales that form a significant part of the daily food supply for humans and dogs. Also, the distribution of seabird species may be affected. Other species may enter the region offering new and potentially prosperous options for economic activities. Such challenges will be met as before.

The adaptations of today are, however, far more complicated. Modern and economically viable harvest of renewable resources is highly industrialized, and to establish new production will be costly, requiring national initiatives as well as international involvement and regulations applied to managing, accessibility and rights. The local capacity for adaptation will be met by implications of today's global interaction. Changes in sea-ice cover, especially a reduction in winter sea-ice extent and thickness will inevitably alter marine transport conditions. Reduced sea ice may offer new transportation routes and prolonged navigation seasons, but will also increase marine access to natural resources and, accordingly, the potential for disturbance and for more efficient fisheries. Although the changes in sea ice may lead to significant changes in the socio-economic life of the region, it is still premature to state the nature of these socio-economic changes. The unpredictability of climate change and its local implications make planning, decision-making and management even more challenging. In spite of its richness, Arctic nature is vulnerable and poor management may lead to serious and irretrievable damage.

8.5.4.3. The Ilulissat ice fjord and its impact on science, politics and tourism

The Ilulissat ice fjord and its glacier is one of the world's best-studied glaciers with the longest observational record. The ice fjord has attracted many scientists through the centuries, the first being H.J. Rink who studied and described the glacier and the Greenland Ice Sheet in the mid-19th century. The studies were instrumental in formulating the important theory of the Quaternary glaciations. Today, the ice fjord is still the subject of detailed international research because of its exceptional manifestation of changes.

The Ilulissat ice fjord was nominated a UNESCO World Heritage site in 2004. The World Heritage List uses the name 'Ilulissat Icefjord' for the entire world heritage site, which includes land areas, the glacier front and parts of the Greenland Ice Sheet adjacent to the ice fjord, recognizing the uniqueness of the area and the extensive interplay between nature and culture.

By entering this prestigious list, tourism has virtually exploded in the ice fjord region. This has made Ilulissat Greenland's leading tourist town and the area now hosts more than 50% of the total number of tourists visiting Greenland. Thus, the ice fjord is by far the main attraction and cruise ships arrive in increasing numbers during the summer to let their passengers visit the fjord.

It is not just tourists that are congregating in the area. The Ilulissat ice fjord has also attracted a large number of politicians from around the world who wish to experience firsthand the global climate change as witnessed here. The ice fjord has now entered political discussions in a way that has never been seen for any part of Greenland before (Mikkelsen and Ingerslev, 2002).

8.5.5. Conclusions

Both local and global effects of the ongoing changes in Earth's climate will inevitably affect Greenland and its population. This chapter has attempted to assess the extent to which changes in the Greenland Ice Sheet will be a major factor in shaping the future society of Greenland. The response of Greenland society to past changes in climate has also been reported and has compared these changes to the effects of societal dynamics. The history of the Thule culture, which gradually transformed into the present-day society in Greenland, has been one of constant adaptation to changes in climate and increased interaction with European culture.

Until now, the observed effects of changes in the Greenland Ice Sheet due to global warming have caused only minor impacts on society. The most marked change observed by the Greenland population has been the retreat of Jakobshavn Isbræ. This has caused an increase in tourism and has increased international attention on Greenland as an icon of global climate change.

In the coming years, changes in drainage patterns around the Greenland Ice Sheet may affect coastal waters and thereby cause changes in the distribution of marine fish stocks and game species. Future changes in drainage patterns may also be a major concern for large infrastructure investments with decadal to centennial life spans, such as hydroelectric power plants and transmission lines, roads, airports and seaports. This chapter has not considered effects from changes in permafrost, although these may have considerable economic implications; the issue of permafrost is considered as separate from that of the ice sheet. Based on the historic observations and the strong drive for change in Greenland society today, it is likely that the consequences of changes in the Greenland Ice Sheet will be minor compared to socio-economic, cultural and demographic influences generated within the Greenlandic and international societies.

8.6. Major knowledge gaps and recommendations

8.6.1. Introduction

New research results from the Greenland Ice Sheet often lead to the identification of new gaps in knowledge, although they may become more specific over time (see Table 8.6). For example, research projects initiated under the International Polar Year 2007/2008 should eventually lead to the identification of new gaps in knowledge or to changes in the existing gaps in knowledge.

8.6.1.1. Assessment of the impacts

Earlier assessments identified some of the major gaps that are currently being addressed. The Arctic Climate Impact Assessment (ACIA, 2005) focused on improving projections of future changes and identified several major gaps in knowledge that are also valid for the Greenland Ice Sheet:

- a need to improve understanding of albedo changes and feedback mechanisms
- the need for studies of outlet glacier dynamics with an emphasis on their potential for triggering persistent and rapid thinning
- a need for improving ice dynamic models for determining

- the long-term response of the ice sheet to past climate change
- a need for improving parameterization and verification of internal accumulation models
- a need to improve understanding of the relationships between climate change, meltwater penetration to the bed, and changes in iceberg production.

Uncertainty in scenarios concerning how glaciers, ice caps and ice sheets contribute to sea level rise over decadal to centennial time scales needs to be reduced. This requires an assessment of the Greenland Ice Sheet and its stability and vulnerability to climate change, including sudden and potentially irreversible climate change. The Climate and Cryosphere Project (CliC), established in March 2000 to stimulate, support, and coordinate research into the processes by which the cryosphere interacts with the rest of the climate system, recommends that an insight is needed into the highly non-linear links between the various components of the cryosphere and their likelihood for producing significant global change. CliC recommends that the long-term ice sheet monitoring system is improved in order to make realistic representations (modeled as well as observational) of spatial and temporal variability in the surface mass budget in areas sensitive

to sea level change (Climate and Cryosphere Project, 2007).

The IPCC's Fourth Assessment Report (IPCC, 2007) stated that realistic scenarios of the Greenland Ice Sheet SMB require a resolution exceeding that of the current AOGCMs used for long-term climate experiments. Present climate models do not generally include a representation of the refreezing of surface meltwater. All models predict an increase in snow accumulation, but there is much uncertainty in its size. In projections for Greenland, an increase in ablation is important but uncertain, being particularly sensitive to temperature change around the margins. The main uncertainty is the degree to which outlet glaciers having a marine terminus respond to climate change. Further accelerations in ice flow, of the kind recently observed in some Greenland outlet glaciers (and in West Antarctic ice streams), could increase the ice sheet loss of mass substantially, but quantitative projections cannot be made with confidence.

8.6.1.2. **Summary of earlier assessments**

The research needs presented in the ACIA, CliC and IPCC assessments of the Greenland Ice Sheet vary in their degree of detail, but all focus on the importance of making more precise scenarios and filling gaps in knowledge. There is a common

Table 8.6. Major gaps in knowledge and recommendations.

Major gaps in knowledge	Recommendations
Large variability between presently used SMB models	Targeted in situ SMB data with which to calibrate/validate models
Incomplete understanding of SMB processes	Systematic studies focused on improvement in process understanding
Large spatial variability in SMB at the ice sheet margin	Better sub-grid scale quantification of the key processes; refreezing, albedo and blowing snow
Large error in determination of surface annual accumulation	Improved determination of snow densification
Discrepancy in estimating the mass change from satellite gravity measurements	Establish a community wide recognition on which processing method provides the most reliable method for tracing mass changes
Incomplete estimate of the mass loss from marine terminating outlets	More systematic mapping of ice thicknesses from outlet glaciers
Incomplete determination of the annual cycle in SMB from satellites	Sampling sufficiently frequent to capture the annual cycle in accumulation, melt and iceflow
Lack of ice discharge determination from ice streams and narrow outlets	Development of high resolution sampling methods from satellites to capture basin scale changes
Large seasonal as well as interannual variability in marginal ice fluxes	Improved theoretical understanding of the basal hydrology and its impact on ice flow
Basic understanding of coupling between ocean and marine based fronts	Theoretical understanding of the calving and basal melting at marine margins where changes occur
Lack of thickness determination around highly dynamic outlets	Develop methods to retrieve ice thicknesses at heavily crevassed margins
Incomplete understanding of flow properties deepest in the ice sheet	Determination and validation of the thickness of the basal temperate layer as well as deep ice temperatures
Incomplete cryospheric representation and determination in climate models	Validate climate models more systematic with observations of surface climate and mass balance of the ice sheet
Lack of full coupling between surface dynamics and the atmosphere	More accurate incorporation of surface albedo and snow microphysics in climate models
Major dynamical features like ice streams lack representation in ice sheet scale models	Better model resolution as well as process understanding leading to well parameterized models
Several-fold changes in velocity variations at marine based outlet glaciers	More complete determination of process leading to improved parameterizations of basal and marginal (ocean) linked process

agreement, that improved understanding of processes is needed in order to determine meltwater runoff more precisely. Also, the rapid increase in ice discharge from the outlet glaciers observed over the past decade calls for a better understanding of processes as well as a better formulation of theory before models quantifying the observed increase in ice flux can be developed. The three assessments all call for better monitoring capabilities and a more precise determination of present state of balance in the Greenland Ice Sheet before more reliable scenarios can be made.

8.6.2. Gaps in knowledge of surface mass balance

Significant differences exist between different estimates of the components of the SMB derived from numerical modeling and downscaling of climate re-analysis data. For example, there are important differences in distribution and total accumulation in the 'wet' southeastern Greenland. The differences between model estimates can be as large as the net SMB for a given year and the standard deviation of the differences is of the order of 100 Gt. Although values for total runoff may agree between models, the individual terms included in the calculations of total runoff vary much more widely, particularly when comparing melt and refreezing. Total differences are of the same order of magnitude as the inferred increase in mass loss due to changes in ice dynamics over the past decade (Rignot and Kanagaratnam, 2006). The differences suggest that mass budget calculations are seriously hindered by uncertainties in SMB.

Direct observations of increasing surface elevation in the regions above 2000 m are not confirmed by direct observed accumulation changes due to very high variability in the records. RCM reconstructions of precipitation do not predict increased values.

Uncertainties in SMB originate from a range of sources related both to lack of *in situ* data and incomplete process understanding. To reduce uncertainties to a level useful for mass budget calculations and reliable predictions will require concerted effort. Especially, (i) to collect more targeted *in situ* data to calibrate and validate models, (ii) to improve process understanding, and (iii) to model key processes such as refreezing, surface albedo, blowing snow and sub-grid scale effects. Sub-grid scale effects are particularly significant in coastal regions due to higher surface relief, which makes ablation observation uncertainties difficult to incorporate into SMB model validation. The most important parameters for controlling runoff in SMB modeling, however, are surface albedo and meltwater refreezing.

8.6.3. Gaps in knowledge of net balance or components thereof

At present, the best estimate of the mass of the Greenland Ice Sheet is probably provided by GRACE (see Section 8.2.4.2.3). However, different processing methods result in different results. One method of averaging the total mass of the Greenland Ice Sheet over 30-day periods and calculating seasonal loss (Velicogna and Wahr, 2006) provides a mass loss of 211 Gt/y, while another uses different processing algorithms for analyzing the distance between the two satellites (Luthcke et al., 2006); this second approach provides a divergent mass loss of 154 Gt for 2007 (Witze, 2008). Such discrepancies reflect

ongoing discussion on which method is most valid for detecting mass changes in the Greenland Ice Sheet using GRACE and the newer publications seem to have a better agreement of results.

Measuring volume change by altimetric methods such as laser (IceSat) or by radar (CryoSat) provides another tool for detecting changes in mass balance. However, connecting volume change to mass changes requires insight into densification processes of the surface snow and firn, which also has a strong seasonal dependence when the surface temperature reaches the melting point. Moreover, it remains difficult to produce a measuring accuracy any better than a few tens of centimetres, and this also renders it impossible to capture the seasonality of the parameters.

Satellite radar techniques measuring ice displacements and ice fluxes near calving glacier fronts are unique in providing a means for detecting iceberg production when the ice thickness is known. However, the lack of measured ice thicknesses from all outlet glaciers and the difficulty of measuring ice movement in the interior of the Greenland Ice Sheet constrain this technique.

The various techniques used to estimate the state of balance of the Greenland Ice Sheet differ in their methodology for detecting changes in specific components of the system; which means that they do not return results that are mutually compatible. Closing the gaps between the results for different techniques will make an important contribution toward a more consistent understanding of change in the ice sheet. Filling these gaps in knowledge will require common or compatible temporal and spatial sampling as well as an understanding of the annual cycles in mass balance, including its most important parameters: Accumulation, melt, runoff, and ice flow. Temporal sampling will need to be frequent enough to capture the annual cycle, or designed to mitigate the effect of interannual variations in this cycle on annual estimates.

Spatial sampling capturing variation at a drainage-basin scale would allow for a more direct comparison of changes in discharge and measures of ice sheet mass or volume. Improved sampling of surface height with new satellite missions may provide altimetric determinations that more faithfully reflect the impact of annual cycles on surface height. However, in order to secure significant progress in understanding densification processes, observation and modeling of the surface snow are also required.

8.6.4. Gaps in knowledge of ice dynamics

A better understanding of basal boundary conditions (i.e., sliding laws) is needed. This involves a more precise understanding of the mechanisms of basal motion in order to derive good parameterization, which accounts for seasonal and longer-term variability. An improved understanding of basal boundary conditions is inherently linked to an understanding of subglacial hydrology, which is assumed to drive the present speed-up of calving outlets. For instance, subglacial hydrology and the complex connection between surface hydrology (supraglacial lake drainage) and the bed are integral to basal flow, which can dominate ice flux near the margin. Modeling the ice sheet hydrology, however, is still in an early stage of development and has not yet been coupled with ice dynamics.

Concerning the ice-ocean interface, gaps in knowledge are mainly related to the assessment of melting rates under floating glacier tongues, where the ocean and heat circulation in combination with freshwater from melting ice drive the circulation and impact on the calving front (Motyka et al., 2003).

Generally, iceberg calving and basal melting at marine margins are difficult to measure and to simulate, and these processes are absent or overly simplified in current ice sheet models.

Observational gaps mainly relate to determining ice thickness at heavily crevassed outlet glaciers, due to difficulties retrieving radar signals in these regions. Also, ice temperatures and, in particular, thickness of the basal temperate layer are not well known, which in combination with ice fabric (Lüthi et al., 2002) show a strong dependence on ice deformation rates and hence dynamics of the outlets.

8.6.5. Gaps in knowledge of ice sheet and climate modeling

Models of meteorological processes, ice sheet mass balance, and ice sheet dynamics are constantly improving, but there are still many uncertainties associated with simulating the Greenland Ice Sheet and its likely response to climate change.

8.6.5.1. Climate models

Many climate models do not resolve the steep topography of the Greenland Ice Sheet margins very well, limiting model skill in simulating orographic precipitation and ice sheet ablation in this zone. High-resolution meteorological models such as PMM5 (e.g., Box et al., 2006) overcome this problem, but there are still outstanding issues with regard to temperature biases, simulation of mass balance processes such as sublimation and boundary layer winds, and the large-scale boundary conditions for regional climate modeling in future climate-change scenarios.

The dominant precipitation and ablation processes can also be event-driven in Greenland (i.e., resulting from a small number of extreme synoptic weather events), and these processes are well captured in simulations that are forced by historical (i.e., re-analyzed) climatology but do not necessarily arise in climate model simulations strong in predictions of the mean state; and they do not always represent the full range of weather variability.

Probably the most important boundary conditions for climate model simulations of the Greenland Ice Sheet are the surface properties (i.e., the physical representation of snow in the models). The inherent instability of snow crystals on the ground, the ever changing meteorological conditions, and the feedback between snow and atmosphere via snow albedo makes the snow-atmosphere coupling an important and highly dynamic one. New progress in snow physics theory and parameterizations needs to be incorporated into climate models.

8.6.5.2. Glaciological models

Ice sheet models now operate at resolutions of 5 to 20 km for the Greenland Ice Sheet, representing the overall ice sheet geometry quite well but still not capturing details of ice margin positions or some of the major fjords and outlet glaciers. This means that it is still difficult to compare model simulations with geological reconstructions of recent (i.e., 20th century) ice sheet changes. In addition, it is difficult to compare modeled ice discharge with observations, with the more distributed (i.e., less channelized) ice flux in the model. Major ice dynamical features such as the glaciers at Ilulissat and Storstrømmen (also known

as the Northeast Greenland Ice Stream) are known to account for a large percentage of ice discharge from the Greenland Ice Sheet, but these features are generally absent in ice sheet models. This is partly an issue of model resolution and partly an issue of complex dynamical processes that are presently too poorly understood to be well parameterized in models.

Ice sheet models are presently insensitive to ocean forcing, in contrast with recent observations of significant (several-fold) velocity variations in marine-based outlet glaciers, most likely to have been triggered by ice-ocean interactions and thinning at the grounding line. Models still cannot simulate this interannual flow variability, so it is unclear whether they are able to provide a credible estimate of the response time of ice sheet change to atmospheric or oceanic forcing.

8.6.6. Perspectives

During the 20th century, the mass discharge from the Greenland Ice Sheet was almost equally divided between meltwater runoff and iceberg discharge (Reeh, 1994) – making the two terms equally important. The magnitude of the present-day balance is uncertain due to lack of coherence between the different estimates relying on different methods. Closing the gaps between the results is a prerequisite for reliable estimates of the present balance. Furthermore, a firm knowledge about the present balance is needed in order to generate reliable future scenarios.

Making specific recommendations on research needs to close the gaps in knowledge requires detailed insight into the problems. However, detailed insight is still lacking for many of the major knowledge gaps, such as the processes underlying increased ice discharge and meltwater refreezing. Instead of reducing uncertainty about changes in the mass balance of the Greenland Ice Sheet, the new observational methods from space have revealed to the research community the incomplete understanding of the fundamental dynamics of calving glaciers. This lack of understanding may explain why gaps in earlier assessments are still unresolved in the present-day assessment of the Greenland Ice Sheet. Closing such gaps in knowledge, where there is a lack of fundamental understanding of the processes, demands the development of theory and methodology to validate the theory against field measurements.

Most changes appear in the marginal regions of the Greenland Ice Sheet where data coverage is most scarce due to logistical access constraints. Moreover, thin ice in these regions, combined with a complex terrain, results in highly variable ice dynamics, melt rates, and accumulation patterns. This calls for high spatial resolution modeling in combination with enough *in situ* measurements to capture adequate detail and to gain sufficient understanding of the present thinning process.

New satellites have triggered a quantum leap in surface observations of the Greenland Ice Sheet. However, *in situ* measurements are still progressing at a slow pace due to resource-demanding fieldwork. The development of new, and less labor-intensive, measuring techniques could potentially advance the process understanding over larger areas. A combined effort in advancing process theory, field methods and coverage as well as modeling efforts could facilitate major advances in understanding the present balance of the Greenland Ice Sheet and, hence, help create less uncertain future scenarios.

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9. Sea Ice

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Contents

Key Findings

- The decline in Arctic summer sea-ice extent has accelerated over the past ten years. The retreat is faster than previously predicted by IPCC models and summer ice-free conditions are likely over most of the Arctic basin by mid-century or earlier.

- Newly developed estimates of sea-ice age, basin-scale sea-ice thickness, and sea-ice volume indicate thinning of the ice cover and loss of old ice types.

- Recent research has better quantified some interactions between the sea ice, ocean, and atmosphere, particularly during the extreme summer ice loss of 2007.

- Predictions of seasonal sea-ice extent are highly variable on a regional basis. Environmental conditions and bathymetry are also highly variable on a regional basis. In combination, these factors will strongly affect the impacts of climate warming on Arctic biological systems.

- The extinction of some Arctic endemic species is likely if current sea-ice trends continue. These losses will represent biodiversity losses of global significance, and reverse millions of years of evolutionary change (since the beginning of the Pleistocene). Northward range expansions for some temperate species and potentially greater overall productivity will enhance foraging opportunities for some animals within High Arctic areas, at a cost to Arctic endemic populations.

- The responses of Arctic endemic cetaceans to changing sea-ice conditions will be influenced by food availability, competition from non-endemic migrant cetaceans and, in some regions, the potential for increased predation from killer whales.

- Indigenous and some local peoples have been challenged by changes in sea ice and those challenges will continue into the future.

- Reduced seasonal sea-ice presence is already leading to greater shoreline exposure to open water and storm waves. These changes impose greater wave and erosion hazards for shoreline, infrastructure, waterfront structures, and cultural heritage sites in some places.

- Loss of sea ice is increasing interest in access and resource rights, such as shipping, resource extraction, military operations, and tourism, which will have impacts across a wide-range of issues.

- Hazards related to sea-ice movement (ride-up and pile-up) onshore are known to have caused fatalities and significant infrastructure damage in the past.

- There are clear limits to any Arctic governance regime (regulatory or soft-law). The drivers of sea-ice change and its consequences occur largely outside the region and fall beyond the jurisdiction of any one Arctic state, and responding to sea-ice change creates profound governance challenges.

Summary

Arctic sea ice has thinned and summer extent has been at or near record low levels almost every year since 2001. Model projections made during the Fourth Assessment of the Intergovernmental Panel on Climate Change (IPCC), indicating near-complete loss of summer sea ice by the end of this century are too conservative (Stroeve et al., 2007) and summers with little or no sea ice in the Arctic are now considered likely by mid-century, possibly sooner (Wang and Overland, 2009). Such a fundamental change in the Arctic environment will have substantial impacts on the Earth's radiation balance and atmospheric and oceanic circulation patterns. Feedbacks in the sea-ice climate system are mostly positive, meaning that the decline in sea ice is having an amplifying effect on the global climate system. The effects of the changing sea-ice environment are already being felt by the plants, animals, and human societies living or operating in the Arctic. The flora and fauna of the Arctic region will need to adapt to changing sea-ice conditions or face extinction. There will also be impacts on human activities in the region, ranging from those undertaken by indigenous Arctic peoples to commercial activities (shipping, resource extraction), tourism, and sovereignty.

Climate change is occurring faster in the Arctic than anywhere else in the world. The rate of increase in Arctic surface air temperature has been about twice the global average over the past few decades (Solomon et al., 2007; Chylek et al., 2009) and this is often referred to as the 'Arctic amplification' (Serreze et al., 2009). The Arctic is a bellwether of change in global climate primarily because of the sensitivity of the thin sea-ice cover – which insulates the comparatively warm ocean from the much colder atmosphere – to changes in air temperature and winds. This sensitivity is enhanced by climate feedbacks associated with a changing cryosphere. Dramatic changes are occurring across the Arctic; one of the most visible indicators of this change is the transformation in the sea-ice cover and the associated impacts on climate, ecosystems, and human societies.

The Arctic sea-ice cover is in decline. There has been a well documented recent retreat of the summer ice cover (Serreze et al., 2007; Comiso et al., 2008), a general thinning of the ice (Giles et al., 2008; Haas et al., 2008; Rothrock et al., 2008; Kwok et al., 2009), and a transition to a younger and more vulnerable ice pack (Maslanik et al., 2007; Barber et al., 2009; Nghiem et al., 2007b). September sea-ice extent reached a record minimum in 2007 (Stroeve et al., 2007) and has since been followed by the second (2011; Figure 9.1), third (2008), fourth (2010) and fifth (2009) smallest observed ice extents since the start of the satellite record in 1979 (Fetterer et al., 2011). In fact, the seven summers since the publication of the Arctic Climate Impact Assessment Report (2005 to 2011) have experienced the six lowest sea-ice extents in the satellite record (since 1979), accelerating an already significant multi-decadal decline. While not every year has experienced a record low minimum extent, the loss of the oldest (5+ years) ice types has continued, particularly in Canada Basin (based on updated data from Maslanik et al., 2007). Because of the less extreme summers of 2008 and 2009 and low transport of ice out of the central Arctic through the 2009/10 and 2010/11 winters, there was a replenishment of younger (2- to 3-year old) multi-

year ice (ice that has survived at least one melt season). This highlights the large year-to-year variability in the Arctic sea-ice cover that is superimposed on the long-term declining trends (Haas et al., 2010).

Several factors may have contributed to the decline in sea-ice cover: a general warming (Richter-Menge et al., 2006), changes in cloudiness (Francis et al., 2005; Kay et al., 2008), changes in atmospheric circulation patterns (Rigor and Wallace, 2004), increased ice drift speed (Rampal et al., 2009) and ice export from the Arctic (Kwok et al., 2010) due to changes in ice dynamics, ocean heat advection from lower latitudes (Shimada et al., 2006; Polyakov et al., 2007; Woodgate et al., 2010), and enhanced solar heating of the upper ocean (Perovich et al., 2007, 2008; Steele et al., 2008). Key feedback processes affecting sea-ice cover include the ice-albedo feedback (e.g., Perovich et al., 2008) and the cloud-radiation feedback (e.g., Hanesiak et al., 2001; Kay et al., 2008; Liu et al., 2009), which may further amplify the projected rate of change.

Sea ice plays a key role for Arctic ecosystems by providing habitat for marine mammals and for a diverse range of productive communities that live within or in association with sea ice. The Arctic also has the most extensive coastal shelves of all the world's oceans; together these cover about 50% of the total area of the Arctic Ocean. A tight coupling of sea-ice ecosystems with pelagic and benthic ecosystems is important on these shallow shelves, some of which are among the most highly productive on Earth, despite the high latitudes. Polynyas (areas of open water surrounded by ice) and marginal ice zones (MIZ) are highly productive areas in the Arctic (Stirling, 1997; Smith and Barber, 2007). Arctic ecosystems have adapted to the presence and timing of sea ice due to the control of sea ice on light, heat, and momentum (Barber et al., 2005). Changes in sea ice affect polynya and MIZ systems and, in particular, the physical-biological coupling within these systems (Barber and Massom, 2007). Changes in Arctic ecosystems have already been observed, including a regime shift in the Bering Sea (Grebmeier

et al., 2006a), increased pelagic production in response to more open water and a longer open-water season (Arrigo et al., 2008), and a decline in the physical condition and reproductive success of polar bears (*Ursus maritimus*) in western Hudson Bay due to earlier ice break-up (Stirling et al., 1999, 2008b).

Sea ice is deeply integrated into the lives of Arctic peoples, but also plays an important role in commercial activities reaching beyond the Arctic. Ice roads are critical for wintertime transportation; sea ice serves as a platform for subsistence hunting; and the high productivity of fisheries is linked to the ice. Changes in sea ice are resulting in coastal erosion in some areas and are affecting the distribution of pollutants. On the other hand, decreased ice cover will make the Arctic more accessible, allowing for increased shipping and access to renewable and non-renewable resources as well as a rise in tourism. Increased activity will in turn have feedback effects not only on the climate but also on the economy and social fabric of the region. At the same time, it will pose challenges related to governance and safety. Despite uncertainties related to future changes in sea ice, the lives of Arctic people are expected to change significantly.

Projections from climate models show a continued, and possibly accelerated, decline in Arctic sea-ice extent, very likely leading to an ice-free summer within this century (Boé et al., 2009), and likely within the next 30 to 40 years (Wang and Overland, 2009; Zhang et al., 2010). However, there is still much uncertainty in the future predictions; this uncertainty arises from natural variations in the climate system; the range of plausible trajectories of greenhouse gas concentrations, aerosols and other climate drivers over the next century; and systematic errors arising from model formulations, particularly the parameterizations of unresolved processes.

This chapter presents current knowledge of sea-ice extent, sea-ice thickness distribution, associated snow cover, key feedback mechanisms, and impacts on Arctic marine ecosystems and Arctic communities. The Arctic Climate Impact Assessment (ACIA, 2005) did not address sea ice in detail, as observations other than for ice extent were scarce. Other climate assessments have paid little attention to sea-ice thickness, concentrating more on sea-ice extent (Solomon et al., 2007; UNEP, 2007; WWF, 2008). This report reviews the significant new information on sea ice, including ice thickness, which has become available since the ACIA report and also includes pre-ACIA findings that were not included in the ACIA report.

In addition to new sea-ice information, there have also been major changes in understanding of feedbacks within the sea ice-climate system and of impacts on biology and human society. The aim of this report is to provide a synthesis of current knowledge of observed and projected changes in Arctic sea ice and the likely impacts on ecosystems and society.
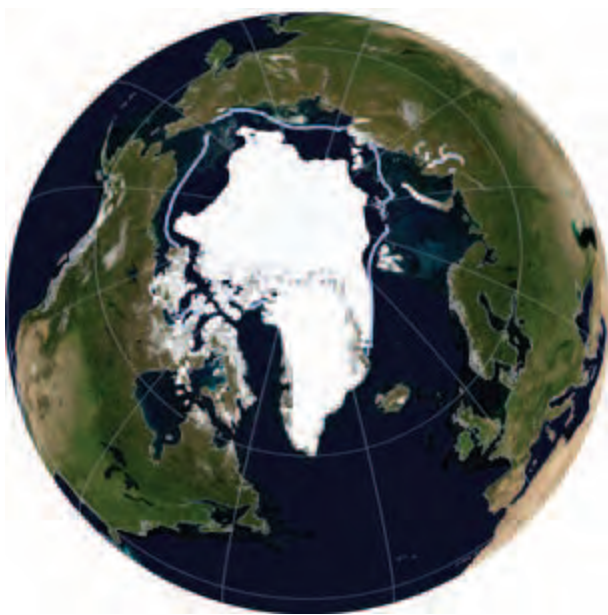

Figure 9.1. September 2011 monthly sea-ice extent (white) and 1979-2000 September average ice-edge location (blue line). Source: Fetterer et al. (2011). Base imagery from NASA Blue Marble.

9.1. Changes in the physical state of sea ice

9.1.1. Introduction

- The decline in Arctic summer sea-ice extent has accelerated over the past ten years. The retreat is faster than previously predicted by IPCC models and summer ice-free conditions are likely over most of the Arctic basin by mid-century or earlier.

  - Interannual variability in basin-wide and regional sea-ice extent is initially likely to increase over the coming decades. As sea-ice extent continues to decrease, variability will also decrease as ice-free summers are approached.

- Newly developed estimates of sea-ice age, basin-scale sea-ice thickness, and sea-ice volume indicate thinning of ice cover and loss of old ice types.

  - Most of the oldest ice types (more than five years old) have been lost.

  - It is likely that the Arctic will become increasingly dominated by first-year ice.

The large-scale physical state of sea ice is defined by the extent (areal coverage), concentration (fractional coverage), and thickness of the ice and overlying snow. Small-scale properties, such as snow density, ice porosity, brine pockets, and temperature, are important for local processes, including biological activity, and also affect the accuracy of some methods to retrieve ice extent/concentration and ice/snow thickness. However, this section focuses on the larger-scale properties of the sea ice and the changes that have been observed.

Sea-ice extent and thickness evolve in concert, both seasonally and interannually, due to the interplay of thermodynamic and dynamic forcing (Box 9.1). Here each are treated separately because the measurement techniques are largely independent, although extent and thickness do change in interrelated ways that determine the overall sea-ice volume or mass balance.

9.1.2. Arctic-wide changes in sea-ice extent, age, and thickness

Sea-ice extent has been most consistently and completely tracked over the past 33 years by satellite-borne passive microwave sensors. Several algorithms have been developed to retrieve sea-ice extent estimates and time series are produced by several organizations. While there are some important differences in the products (as described by Andersen et al. 2007), there is overall consistency in estimates of long-term trends. Time series that are available electronically are noted in Table 9.1. The sea-ice extent/concentration imagery and total extent statistics quoted in this report are derived from passive microwave imagery using the NASA Team sea-ice concentration algorithm (Cavalieri et al., 1984, 1996; Maslanik and Stroeve, 1999; Meier et al., 2006), distributed by the National Snow and Ice Data Center's Sea Ice Index (http://nsidc.org/seaice_index/; Fetterer et al., 2011). It should be noted that these observations are hemispheric and total extent estimates include regions outside of the defined Arctic region of this report (e.g., Sea of

Box 9.1. Large-scale sea-ice thickness distribution and circulation within the Arctic

Sea-ice thickness changes as a result of thermodynamic (winter freezing and summer ablation) and dynamic (thickening by deformation) processes (Box 9.2; Thorndike et al., 1975). On a basin-wide scale, sea ice in the Arctic Ocean moves with two predominant drift systems: the Transpolar Drift and the Beaufort Gyre. The Transpolar Drift advects ice from the Siberian coast region across the Arctic Ocean toward and eventually through Fram Strait. The Beaufort Gyre is a clockwise circulation that circulates ice within the western Arctic in response to average high air pressure in the region. These drift systems, as well as smaller-scale motion patterns in the marginal seas, help control the extent and thickness of the sea-ice cover. Wherever ice drift is directed southward, ice extent protrudes south with it, as in the East Greenland Sea. Wherever ice drift is directed toward obstacles like coasts, ice is pushed against these obstacles and thickens by deformation (i.e., breaking, rafting, and formation of pressure ridges). As a result, on a basin-scale, the thickest sea ice (mean thicknesses 4 to 6 m) is located north of the coasts of Canada and Greenland, while the thinnest sea ice (mean thicknesses 0 to 2 m) extends along the coasts of Siberia, from where it is advected away soon after formation. Changes in the relative strength of the Beaufort Gyre and the Transpolar Drift are important controls on the extent and distribution of multi-year ice (MYI). When the Beaufort Gyre is strong, MYI tends to be retained within the Arctic Ocean; when the Transpolar Drift is strong, more ice is advected out of the Arctic Ocean and MYI area decreases. The relative strengths of these two circulations have been linked to the phase of the Arctic Oscillation (Rigor and Wallace, 2004).
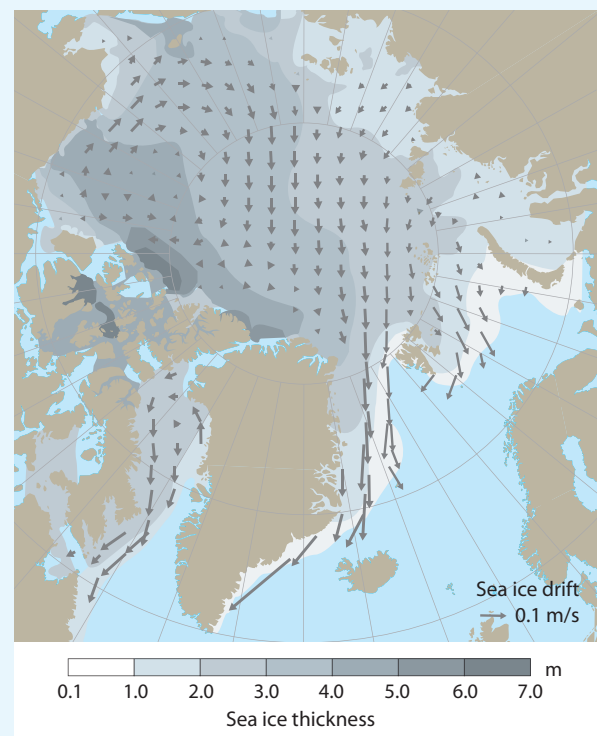
Figure 9.2. Typical winter sea-ice thickness distribution within the Arctic and large-scale circulation patterns.

Table 9.1. Selected publicly accessible sources of Arctic sea-ice extent estimates based on passive microwave satellite data.

Group	Website
Arctic Regional Ocean Observing System (ROOS), Nansen Environmental and Remote Sensing Center, Bergen, Norway	http://www.Arctic-roos.org/
The Cryosphere Today, University of Illinois, Urbana-Champaign, Illinois, USA	http://Arctic.atmos.uiuc.edu/cryosphere/
NASA Goddard Sea Ice Remote Sensing	http://polynya.gsfc.nasa.gov/seaice_datasets.html
University of Bremen	http://www.iup.uni-bremen.de/seaice/amsr/
Japanese Space Exploration Agency (JAXA)	http://www.ijis.iarc.uaf.edu/en/home/seaice_extent.htm
Danish Meteorological Institute Center for Ocean and Ice	http://ocean.dmi.dk/arctic/icecover.uk.php
Sea Ice Index, National Snow and Ice Data Center, University of Colorado, Boulder, USA (used in the present assessment)	http://nsidc.org/data/seaice_index/

Box 9.2. Local ice thickness distribution

Sea-ice thickness and sea-ice concentration vary on small spatial and temporal scales due to local dynamic and thermodynamic forcings. Divergent ice motion causes the opening of cracks, leads, and polynyas. During winter, where water is exposed to the cold air, formation of new, thin ice occurs; near the ice edge this can increase ice extent. In summer, these open water areas are characterized by increased absorption of solar energy. Convergent motion results in a retreat of the ice edge (lowering the extent), or the formation of pressure ridges (piled ice blocks above and below the water level form when ice floes collide and break) within the ice pack. Ridges can be up to tens of meters thick. As a result, the frequency distribution (histogram) of sea-ice thicknesses in a given region is typically characterized by a strong mode representing the most frequent, level ice formed by freezing and melting, and a long tail toward thicker ice representing the amount and thickness of ridges. Mean sea-ice thickness is strongly affected by the shape of the sea-ice thickness distribution, and does not normally agree with modal ice thickness, but is a result of the mixture of various amounts of thermodynamically and dynamically formed ice.
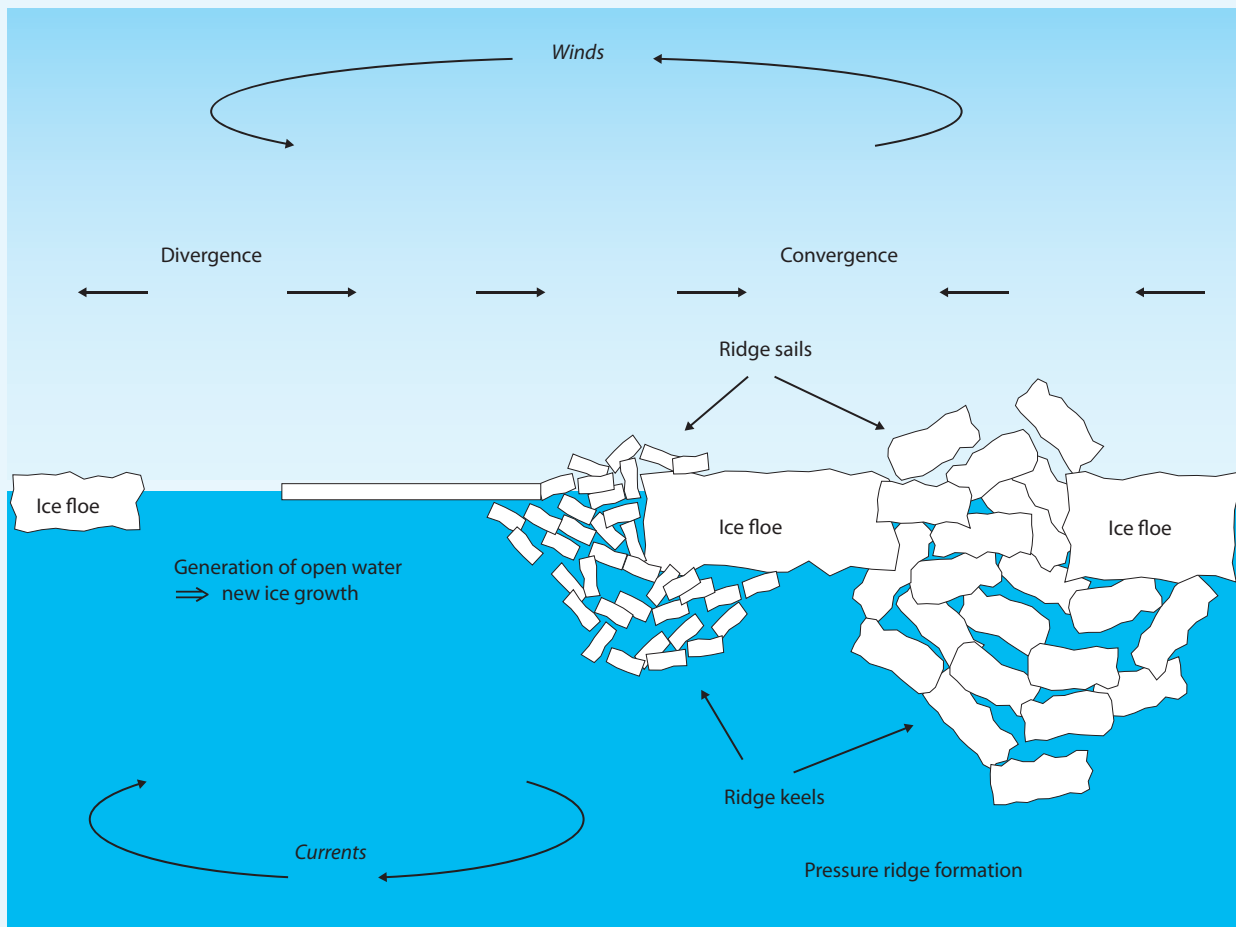
Figure 9.3. Processes that dynamically modify ice-thickness distribution. Source: Haas and Druckenmiller (2009).

Box 9.3. Uncertainties in ice-extent estimates

Passive microwave remote sensing instruments provide sea-ice extent information day or night, clear or cloudy, and have a near-complete record since late 1978. However, there are many limitations to passive microwave imagery. First, the instantaneous field of view (sensor footprint) is large at ~25 to 50 km, limiting the precision of the ice edge location and concentration at the ice edge because of inherent subpixel ambiguity (Eppler et al., 1992; Hwang and Barber, 2006). In addition, passive microwave instruments are sensitive to the effects of snow cover, liquid water (Hwang et al., 2007a), brine volume change (Hwang et al., 2007b), melt ponds (Yackel et al., 2000), and ice temperature (Hwang et al., 2008). These effects can result in either overestimation or underestimation of sea-ice amounts depending on seasonal and local conditions. Atmospheric emissions can also affect the microwave signal (Oelke, 1997; Andersen et al., 2006). Finally, thin ice also tends to be underestimated by algorithms that convert brightness temperature estimates into ice concentration (Cavalieri, 1994; Hwang et al., 2007b, 2008). In general, biases in sea-ice concentration are ~5% during winter (Steffen et al., 1992; Kwok, 2002), with higher errors in summer due to surface melt (Steffen et al., 1992). However, significantly higher errors have been found, depending on surface and atmospheric conditions (Comiso et al., 1997; Meier, 2005; Andersen et al., 2007). Different algorithms have different error magnitudes depending on conditions and no single algorithm has superior performance in all cases.

For tracking long-term variability and trends, seasonal or annual biases are not as important as the consistency of the time series, particularly when sea-ice concentration anomalies are used (Barber and Hanesiak, 2004). The greatest factor for ice concentration/extent consistency is intercalibration of the products through transitions of different generations of satellite-borne sensors. Careful intercalibration of the time series has minimized such errors (Cavalieri et al., 1999; Comiso and Nishio, 2008) and there is high confidence in the accuracy of the long-term trend estimates.

Ice charts are compiled from manual interpretation by ice analysts of the best available data with the focus on producing the most accurate operational analysis at any given time. As the quantity and quality of input data have progressed, the accuracy of the charts has varied over time and thus the charts are not necessarily consistent for long-term tracking of trends and variability (Partington et al., 2003). From the 1950s through the early 1990s, charts were produced primarily from aircraft reconnaissance imagery, ship observations, and starting in the late 1960s, visible and infrared satellite imagery. From the late 1970s, passive microwave satellite data have filled in regions where visible/infrared imagery was unavailable. Beginning in the early 1990s, synthetic aperture radar data became widely available. The higher spatial resolution and all-sky capabilities of these data significantly increased the amount and quality of data available to operational ice analyses, yielding more detailed and more accurate charts.

Okhotsk). Uncertainties in sea-ice extent estimates are reviewed in Box 9.3.

Long-term sea-ice thickness observations have been performed by drilling at various fast-ice locations along the Arctic coast. Since 1958, ice draft measurements have been performed by sporadic submarine Upward-Looking Sonar (ULS) missions crossing the Arctic Ocean along certain transects, and by moored ULS in confined regions since the early 1990s. Only since 1993 have freeboard measurements been performed by satellite radar altimeters, and since 2003 data have become available from satellite laser altimetry. Ground-based and airborne surveys using the same or other methods have been performed since the 1990s, but are regionally limited and have only rarely resulted in long-term data sets. Uncertainties in sea-ice thickness estimates are reviewed in Box 9.4.

9.1.2.1. Recent changes – since 2000

9.1.2.1.1. Extent

Arctic sea ice has experienced several extreme low summer sea-ice extents since 2000. A new record low since the beginning of the passive microwave satellite record in 1979 has occurred three times in the past decade: in 2002, 2005, and 2007. Most other years in the past decade, although not record-setting, were also extremely low. Each year in the period 2002 to 2011 was at least 12.6% below the 1979–2000 average (Figure 9.5). The seven years (2005 to 2011) since the Arctic Climate Impact Assessment (ACIA, 2005) have experienced the six lowest minimum sea-ice extents in the satellite record. Overall, the 2001–2011 average September monthly sea-ice extent is 5.49 million km^2; 22.0% below the 1979–2000 average of 7.04 million km^2. The trend over the eleven-year period is -191 000 km^2 per year (-27.1% per decade relative to the 1979–2000 average).

Winter sea-ice extents were also low through much of the decade, although the percentage decreases were not as extreme as for summer sea-ice extent. The 2001–2011 March average (the month with the highest sea-ice extent) is 15.04 million km^2; 4.5% below the 1979–2000 March average of 15.75 million km^2.

The minimum and maximum extremes are indicative of declining trends in sea-ice extent throughout the annual cycle, with larger declines during the summer months and less extreme declines in winter. The overall annual average extent for the period 2001 to 2011 is 11.18 million km^2; 7.8% below the 1979–2000 annual average of 12.13 million km^2.

9.1.2.1.2. Thickness and reduction in multi-year ice extent

Basin-scale ice thickness information has primarily been retrieved from satellite altimeter freeboard measurements (Box 9.4). The European Space Agency's (ESA) ERS and Envisat satellite radar altimeters have performed circum-Arctic observations south of 81.5° N since 1993. Envisat observations between 2002 and 2008 suggest little change until early 2007, but a large decrease (25 cm) following the September 2007 sea-ice extent minimum (Figure 9.6; Giles et al., 2008). However, this drop was regionally confined to parts of the Beaufort and Chukchi Seas, and no significant changes were found in the eastern Arctic Ocean. Observations of this thinning between 2007 and 2008 are confirmed by results

Box 9.4. Uncertainties in sea-ice thickness estimates

Sea ice floats on the water generally in isostatic balance, such that measurements of draft, freeboard, or surface elevation can be converted into total thickness. However, accurate sea-ice thickness measurements are challenging because there is high spatial and temporal variability and the amount of snow on the ice and the snow and ice density structure are usually not well known.

In situ measurements are the most direct and most accurate estimates for the point location or transect that they are measuring. Substantial uncertainties in the mean thickness estimate arise because these local measurements are not necessarily representative of surrounding conditions, since ice thickness can vary greatly over short distances. Biases may also occur because very thin ice and thick ridged ice are underrepresented due to access and sampling difficulties.

Remote observations provide considerably more coverage and potential representativeness, but there are substantial difficulties in converting the measured properties to thickness values. Electromagnetic induction (EM) instruments generally penetrate to the ice-water interface, but snow cover and internal ice properties lead to ambiguities. The spatial and temporal coverage of EM instruments, although more widespread than *in situ* drill holes, can still only cover relatively small regions.

Upward-looking sonar (ULS) measurements from submarines have more spatial coverage, but measurements have been sporadic in time. In addition, sonars measure draft and uncertainty in ice density and snow cover limits the accuracy of the measurements.

Satellite altimetry has shown the potential to provide large-scale coverage over much of the Arctic at regular intervals. However, similar to sonar, altimeters do not directly measure thickness, but rather freeboard. Because freeboard is only 10% to 15% of the total thickness, any errors in freeboard height are amplified. The problem is particularly acute for laser altimeters, such as ICESat, which measure the total snow + ice freeboard. Thus estimates of snow depth and density are necessary for accurate conversion to thickness. With radar altimeters, such as Cryosat-2, it is generally assumed that the radar signals penetrate through to the snow-ice interface to measure ice-only freeboard. Although less than is the case with laser altimetry, snow still represents a significant source of uncertainty for radar altimetry. Nonetheless, altimeter data have already provided key data on ice thickness and volume and future measurements will provide essential information on the future evolution of the ice cover.

Ground-penetrating radar (GPR) has the potential to obtain both snow and ice thicknesses, but has not yet been extensively used over sea ice.
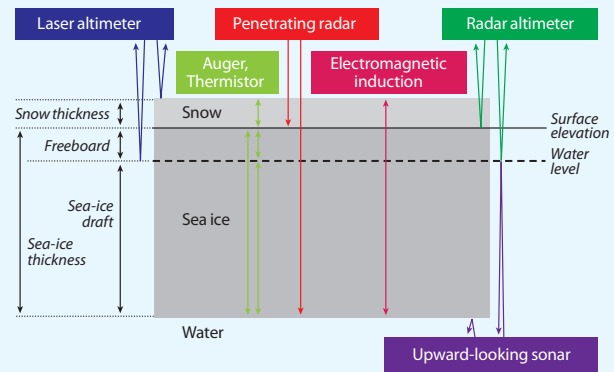
Figure 9.4. Ice floe cross-section showing definitions of ice and snow thickness, total thickness, draft, freeboard, and surface elevation. The primary measurements of individual ice-thickness observing methods are also indicated.
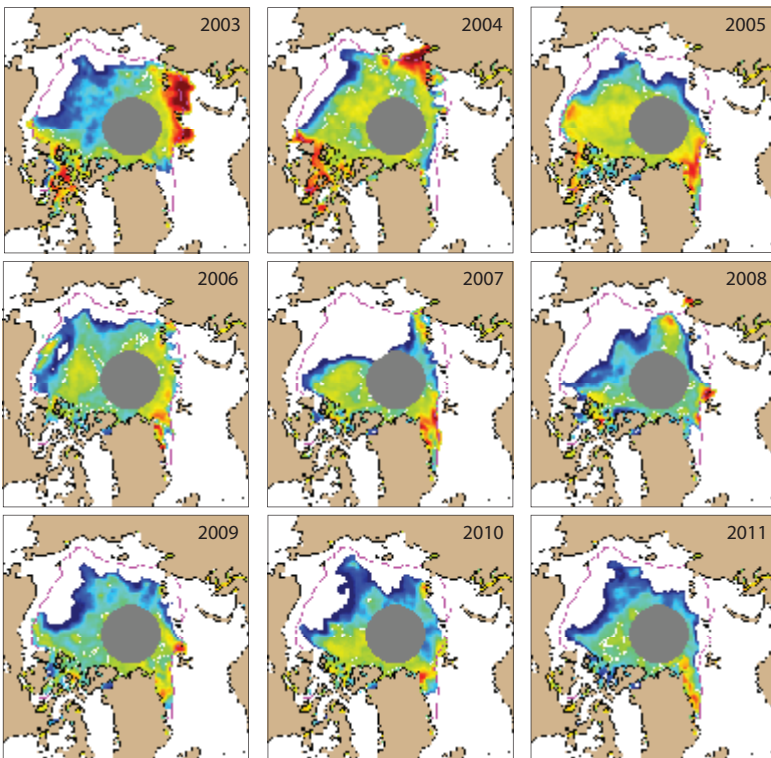
Figure 9.5. September monthly sea-ice extent and concentration (i.e., fractional area coverage) anomaly, 2003–2011. The grey circle represents areas not covered by the satellite sensors during at least part of the time series; this area has traditionally always been covered by a high concentration of sea ice. Source: based on the NSIDC Sea Ice Index (Fetterer et al., 2011).

Ice thickness anomaly, m
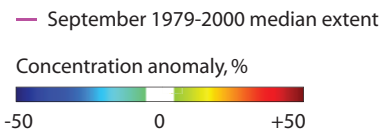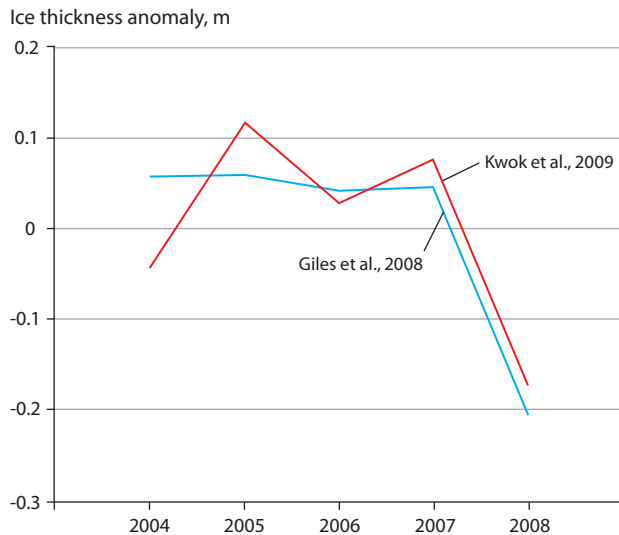
Figure 9.6. Comparison of five years of winter sea-ice thickness anomalies in the western Arctic derived from laser (ICESat; Kwok et al., 2009) and radar altimetry (Envisat; Giles et al., 2008). Anomalies are calculated for regions of the Arctic Ocean south of 81.5° N which are covered by both satellites.

Ice thickness, m

Ice volume, km³
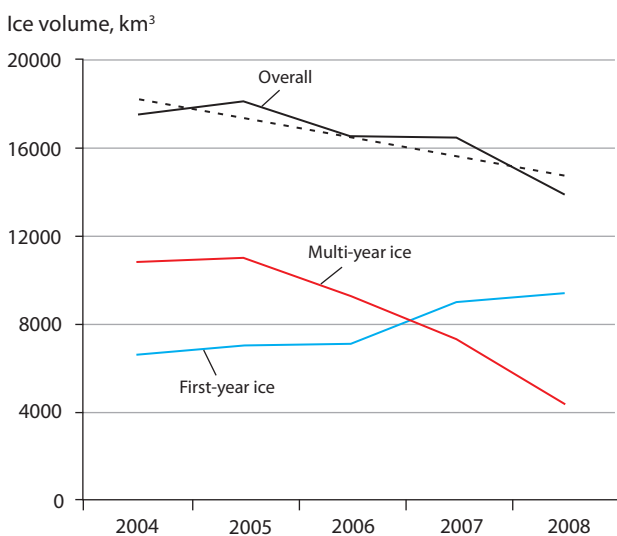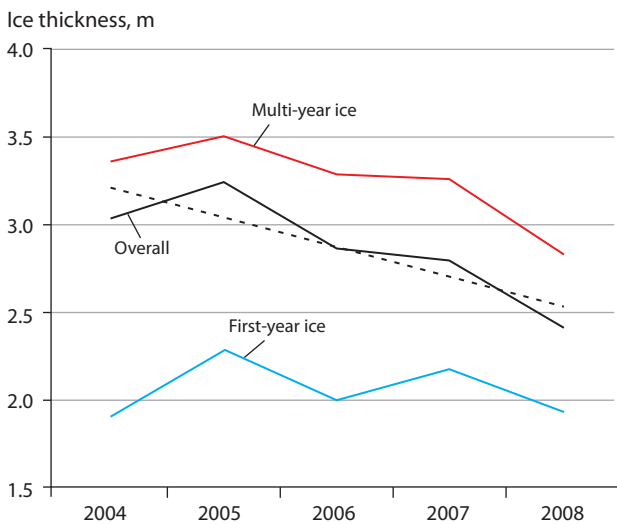
Figure 9.7. Decline in winter Arctic first-year (FYI) and multi-year (MYI) sea-ice thickness and volume between 2004 and 2008 derived from ICESat observations (Kwok et al., 2009). Overall thickness is the mean MYI and FYI thickness weighted by their areal coverage; overall volume is the sum of MYI and FYI volume.

from ICESat laser altimetry, although these data also reveal stronger interannual variability (Figure 9.6; Kwok et al., 2009).

ICESat data provide more insight as they cover regions up to 86° N and yield observations of seasonal variations between spring and autumn of each year (Kwok et al., 2004, 2007; Kwok and Cunningham, 2008). Data from ten ICESat campaigns between 2004 and 2008 (Figure 9.7) show rapid thinning and volume loss of the Arctic Ocean sea-ice cover (Kwok et al., 2009). Changes were strongest over multi-year ice regions, which thinned by ~0.6 m over four years. In contrast, the average thickness of the seasonal ice in mid-winter (~2 m), which covered more than two-thirds of the Arctic Ocean in 2007, exhibited a negligible trend.

Changes in multi-year ice (MYI) and first-year ice (FYI) thickness are accompanied by changes in the areal coverage of these ice types, which affect overall ice volume. One of the most significant changes in the Arctic Ocean is a reduction in the coverage of MYI, which has decreased by more than 42% since 2005 (Kwok et al., 2009). As a result, MYI volume in winter experienced a net loss of 6300 km³ (>40%) between 2005 and 2008, while FYI cover gained volume due to increased overall areal coverage compensating the MYI area loss (Figure 9.7). The overall decline in volume and thickness is explained almost entirely by changes in the MYI cover. Combined with a large decline in MYI cover over this short record, there is a reversal in the volumetric and areal contributions of the two ice types to the total volume and area of the Arctic Ocean ice cover. Seasonal ice, having surpassed that of MYI in winter area coverage and volume, became the dominant ice type. The average winter sea-ice volume between 2004 and 2008, including the loss of ~3000 km³ between 2007 and 2008, was ~16 420 km³.

It is presently unclear whether the reduction of old ice is a result of changed and possibly accelerated ice drift and export (Rampal et al., 2009; see also Box 9.1), or is due to reductions in summer sea-ice extent, which reduces the amount of FYI that survives melt and gets promoted to second year ice at the beginning of each winter freeze-up. The annual cycle of MYI coverage reconstructed since 1999 from satellite radar observations (Figure 9.8a) may provide further insights: MYI coverage decreases throughout the winter due to sea-ice export and divergence, and is reset at the end of every summer by the remaining FYI that has survived the summer. Balance between export, melt, and replenishment is necessary to maintain a stable MYI area. Up through the summer of 2005, the annual replenishment was sufficient to maintain a relatively stable MYI coverage of ~4 million km² (Figure 9.8a) although the variability was remarkably high. The 2007 sea-ice replenishment was the lowest after the record minimum in ice extent that summer and helped set the record for lowest MYI coverage. Although more recent ICESat data were not available for this assessment, ice age fields (Maslanik et al., 2007) indicate larger replenishment after the 2008 to 2011 melt seasons.

Figure 9.8b shows the net sea-ice export record through Fram Strait since 2000 (Kwok et al., 2009). Over the nine years, the average net area flux through Fram Strait of ~730 000 km² was close to the 29-year record average of ~700 000 ± 130 000 km². In the absence of replenishment of the ~4 million km² of MYI in the Arctic prior to 2005, it would take only a few years – at an average export rate – to export all of the MYI through Fram Strait. Therefore Figure 9.8 implies that the areal melt of FYI during summer, rather than export, is the main cause for the decreasing MYI coverage.

MY sea ice area, million km²
a

Fram Strait export, thousand km³
b
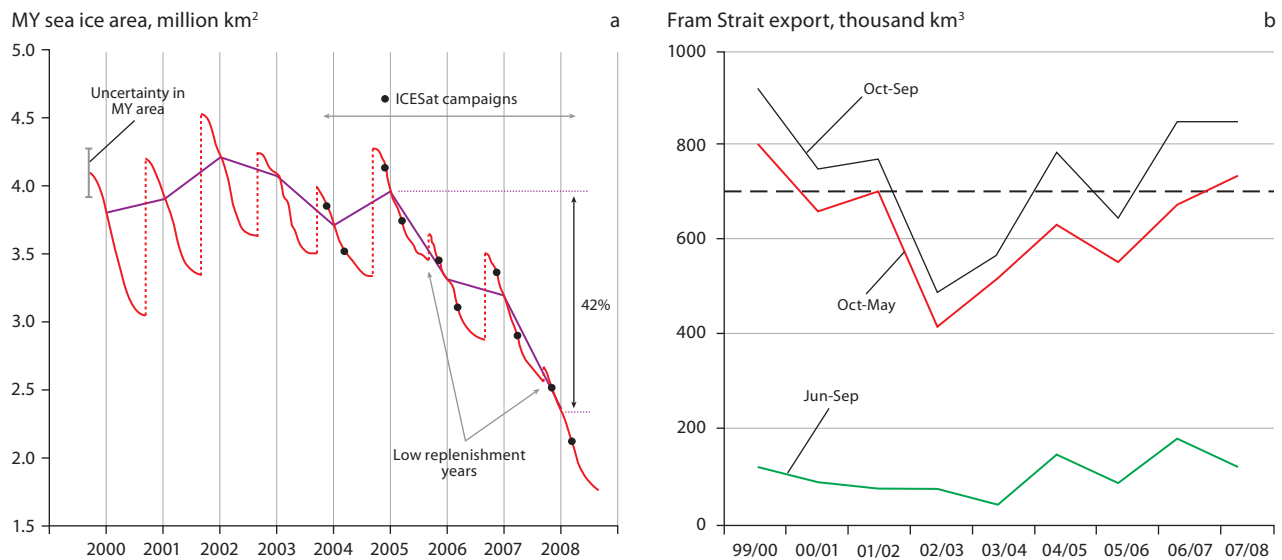
Figure 9.8. (a) Nine annual cycles of replenishment / export of Arctic Ocean multi-year ice (MYI) area constructed using QuikSCAT data and Fram Strait ice export. Black dots show the approximate times of the ten ICESat campaigns used here. The dashed vertical lines show the replenishment of the MYI reservoir by first-year ice at the end of each summer. The two near-zero replenishment years of 2005 and 2007 are indicated. (b) The annual and seasonal Fram Strait ice export over the same period. Source: Kwok et al. (2009).

9.1.2.2. Changes since the start of satellite measurements

9.1.2.2.1. Extent

Several studies (e.g., Meier et al., 2007; Comiso and Nishio, 2008; Parkinson and Cavalieri, 2008) have examined trends and variability in sea-ice extent for the passive microwave sea-ice record (Box 9.3). The over 33-year record shows a statistically significant decline in extent in all months. The 1979–2011 monthly March extent trend is -2.7% per decade relative to the 1979–2000 average, while the September trend is -12.0% per decade (Figure 9.9). The annual decline trend for 1979–2011 is -4.3% per decade relative to the 1979–2000 average. The past ten years have much larger downward trends and while there is less confidence in such short trends, they are likely to represent an acceleration of the downward trend in sea-ice extent since the beginning of the passive microwave satellite record in 1979 (Comiso, 2006).

9.1.2.2.2. Thickness

There are few satellite-derived sea-ice thickness measurements outside of the past ten years. Results from the ESA ERS satellite radar altimeters obtained for regions south of 81.5° N over the period 1993 to 2001, showed a pan-Arctic ice thickness field which was highly variable from year to year, without revealing significant longer-term changes (Laxon et al., 2003).

9.1.2.2.3. Reduction in sea-ice age

While direct retrievals of thickness from satellite are new to the past 10 to 15 years, there is a longer record of MYI and FYI extents and ice age distribution, calculated by Lagrangian tracking of ice parcels (Belchansky et al., 2004; Rigor and Wallace, 2004; Maslanik et al., 2007) that extends nearly as far back as the satellite-derived extent time series.

Sea ice extent, million km²
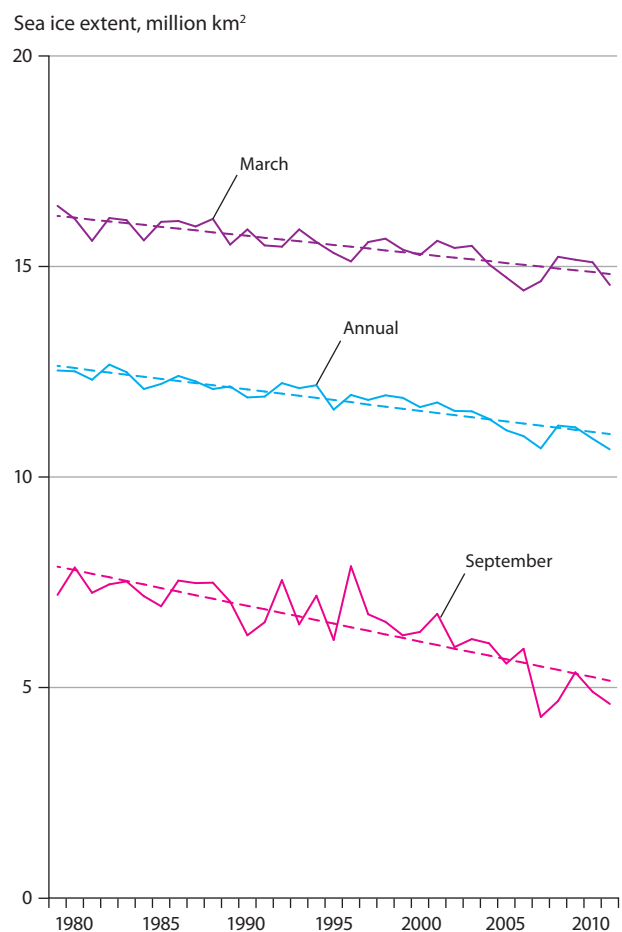
Figure 9.9. Average sea-ice extent in the Arctic in March, September, and annually for 1979 to 2011, including trend lines, from passive microwave satellite data. Source: updated from Fetterer et al. (2011).

Analysis of ice age through March 2011 (Figure 9.10) shows that the loss of older ice types within the Arctic Basin has continued and accelerated since 2007. Multi-year ice now makes up less than 30% of the spring ice cover north of 70° N. Some young (2- to 4-year old) MYI was regained in the period 2008 to 2010 due to higher survivability of FYI through summer. However, loss of the oldest ice types continued through 2011, with under 2% of the ice cover comprising ice greater than six years old compared to 20% in 1988. Such differences, in addition to ice thickness, have the potential to affect the survivability of the ice pack.

The reduction of the age and amount of old sea ice may be one of the most fundamental changes in the Arctic Ocean (e.g., Johannessen et al., 1999; Comiso, 2002; Belchansky et al., 2004; Rigor and Wallace, 2004; Kwok, 2007; Nghiem et al., 2007). The remaining ice pack now consists of more FYI and less of relatively thicker MYI, with consequences for the thickness, surface roughness, snow accumulation and albedo of the ice overall (Iacozza and Barber, 1999). This thinning of the ice pack possibly helped set the stage for the extensive retreat of summer ice extent in recent years by pre-conditioning the ice pack to be less resistant to melt and export (Stroeve et al., 2007; Lindsay et al., 2009).
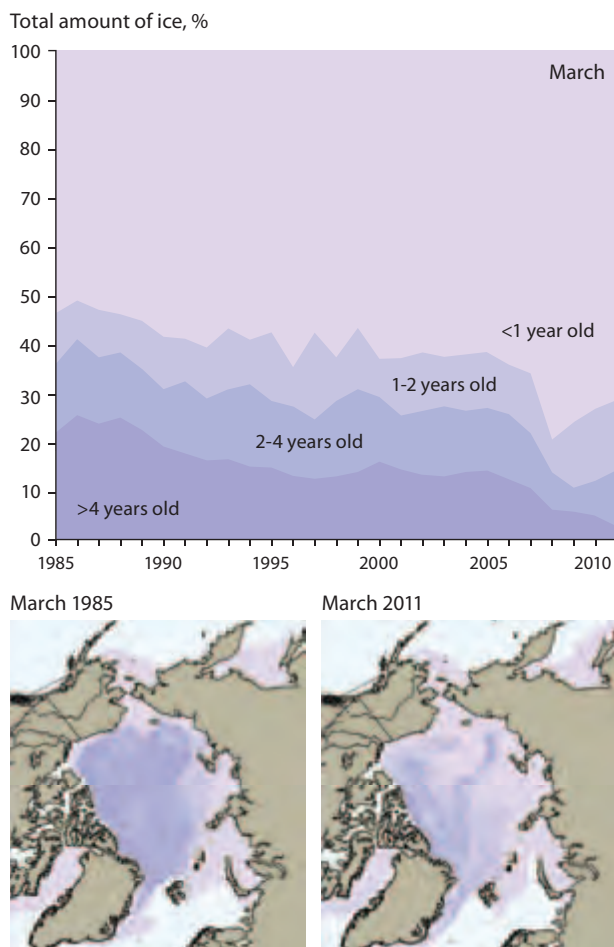
Figure 9.10. Change in fractional coverage of ice of different age categories through the end of March 2011, for the area north of 70° N with at least 40% total ice concentration (upper graphic) and the spatial distribution of ice by age category in March 1985 and March 2011. '1- to 2-year old ice' refers to ice that has survived at least one melt season. Source: updated from Maslanik et al. (2007).

9.1.2.3. Pre-satellite records of extent and thickness

9.1.2.3.1. Extent

Records of sea-ice extent before the passive microwave record become increasingly sparse going back in time. The best sources come from operational ice charting groups that used visible/infrared satellite data starting in the 1960s, aircraft reconnaissance imagery, and ship observations. These data are not as consistent as the passive microwave record because their completeness and quality vary. Extent estimates from operational ice charts and other sources have been compiled by Walsh and Chapman (2001) and were combined into a monthly extent time series since 1870 (Rayner et al., 2003; Meier et al., 2007). However, before 1953 there is little variability because the estimates were often based on climatology. Brönnimann et al. (2008) used modeling and statistical analysis to reconstruct sea-ice extent for the 1900 to 1953 period to obtain more realistic variability; the results indicate that the 1979–2011 decline, particularly the accelerated decline for 2001–2011, are unique in at least the past 100 years.

Longer observational records are only available regionally, with the most complete records from the Russian Arctic. Mahoney et al. (2008) examined ice charts for the Russian Arctic since 1933. The data indicate that during the 1930s warm period, there were some reduced sea-ice extents, but these were found primarily in summer, with little or no decline in other seasons. The current decline is year-round. Also, the 1930s decline was potentially more regional in nature, whereas the current decline is Arctic-wide.

More recent analysis of post-1900 data from the Russian Arctic by Frolov et al. (2009) showed the same general features as found by Mahoney et al. (2008). However, Frolov et al., (2009) also found cyclical patterns at several temporal scales, from 2 to 3 years up to multi-decadal frequencies of 20 years, and 50 to 60 years. These cycles are determined from only the Russian and Atlantic sectors and the long-term cycles are most prominent in the Greenland, Barents, and Kara Sea region. The analysis was based on data through 2003 and it is unclear how the recent extreme summer minimums may affect the significance of the results.

9.1.2.3.2. Thickness

Long-term, basin-wide sea-ice thickness data are only available from submarine ULS ice draft measurements, which have been collected during sporadic military US and UK missions since 1958 (Box 9.4). These data have been declassified over a data release area covering around 38% of the Arctic Ocean. By the 1980s, enough data had accumulated to allow estimates of the large-scale spatial distribution of draft in the Arctic Ocean (Bourke and Garrett, 1987). Concerns were raised as to whether the submarine dataset was sufficiently robust to distinguish any signal of long-term change from 'natural variability', which amounted to more than 1.5 m between some years (McLaren et al., 1990; Wadhams, 1990); thickness changes were found to be primarily related to variations in the origin and age of the surveyed ice and associated changes in drift patterns (McLaren et al., 1992; Shy and Walsh, 1996). Rothrock et al.

(1999) compared measurements performed between 1958 and 1976 with surveys carried out in coincident regions between 1993 and 1997. They found significant thickness reductions of 1.3 m on average, decreasing from an average of 3.1 m in the earlier period to 1.8 m in the latter (Figure 9.11). Similar reductions were observed by Wadhams and Davis (2000). The observed thinning was partially explained by changes in sea-ice drift patterns changing from a predominantly anticyclonic mode to a more cyclonic mode (Tucker et al., 2001; Winsor, 2001; Holloway and Sou, 2002; see also Box 9.1), but thermal forcings (i.e., higher temperatures) were also found to be a major factor (Yu et al., 2004).

Since 1975, data coverage has improved and it is now possible to separate the annual cycle, spatial field, and interannual signal. Over the period 1975 to 2000 the data show a decline of 1.25 m in mean sea-ice thickness (Rothrock et al., 2008). These data, together with the ICESat data, provide the best available and most extended ice thickness time series within the central Arctic Ocean (Figure 9.12; Kwok and Rothrock, 2009). The overall mean winter thickness changed from 3.64 m in 1980 to a mean of 1.89 m during the last winter of ICESat measurements in 2008, a decrease of 1.75 m. Overall, the rate of decrease accelerated to up to -0.2 m during the six-year ICESat record (Kwok and Rothrock, 2009).

9.1.2.4. Indications of pre-1900 sea-ice conditions from paleographic proxy data

Indications of pre-20th century sea-ice conditions can be inferred from proxy data, such as tree rings, ice cores, and sediment cores. A comprehensive review of proxy records in the Arctic indicate that recent sea-ice extents are the lowest in at least the past few thousand years and the recent reduction cannot be explained by variations in any known natural mechanisms (Polyak et al., 2010). Müller et al. (2009) examined sediment cores from the Fram Strait region and found evidence for seasonally ice-free conditions about 14 000 years ago. Changes in Arctic ice shelves (Copland et al., 2007; England et al., 2008) and proxy temperature reconstructions
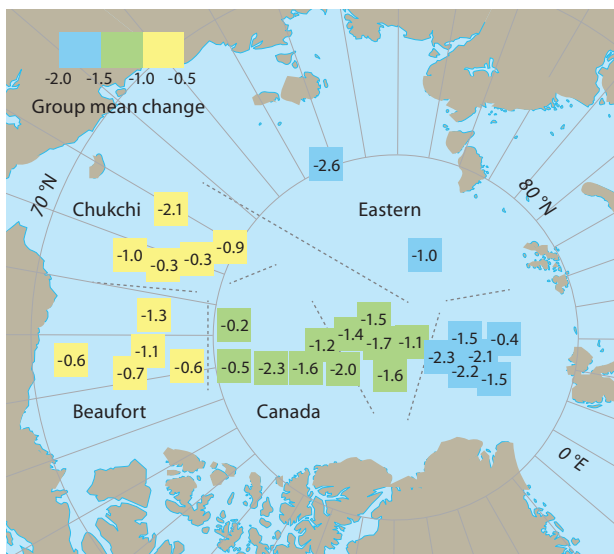
Figure 9.11. Regions of coincident submarine tracks in 1958–1976 and 1993–1997. Numbers indicate ice thickness difference between the two periods. Source: Rothrock et al. (1999).
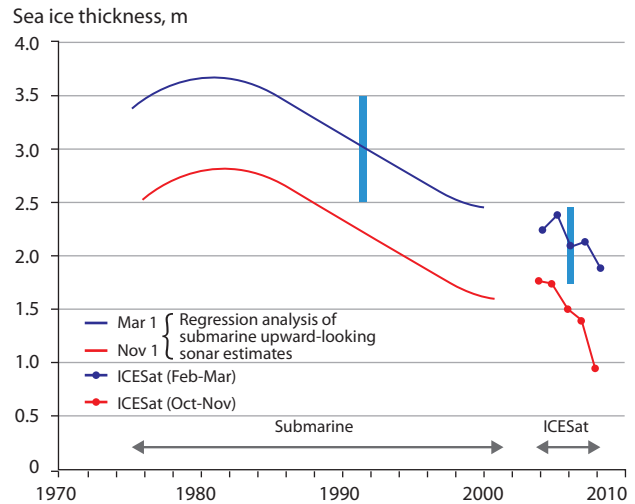
Figure 9.12. Interannual changes in winter and summer sea-ice thickness from Rothrock et al. (2008) and Kwok et al. (2009) centered on the ICESat campaigns. Blue error bars show residuals in the regression and quality of ICESat data (Kwok and Rothrock, 2009).

(Kaufmann et al., 2009) provide indirect supporting evidence that current ice conditions are low compared to historical levels over the past several thousand years. These proxy records are not complete enough to provide solid estimates of extent or to determine with high certainty when in the past the Arctic Ocean may have been ice-free during summer. More proxy data and new techniques could be valuable in more precisely determining sea-ice conditions in the distant past and allowing the current decline to be put into a long-term context.

9.1.3. Seasonal changes in sea ice

There is a strong seasonal cycle in Arctic sea ice that affects both extent and thickness. Because the majority of FYI melts completely during summer, there is much more variability in extent than thickness, with summer ice extents less than half the winter values (Figure 9.9).

The seasonal cycle of ice thickness varies considerably depending on ice type, location, and local atmospheric and oceanic conditions. Although both the ocean and the atmosphere can force ablation of sea ice throughout the annual cycle (Carmack et al., 2006; Barber and Massom, 2007), the mass balance is dominated by snow and ice ablation in summer and bottom accretion in winter (Box 9.2). This strong seasonal cycle must be taken into account when comparing thickness measurements from different seasons. The magnitude of the seasonal cycle varies strongly with latitude and regional meteorological and oceanic conditions. Surface and bottom melt of individual locations on drifting ice floes can be measured with Ice Mass Balance Buoys (Richter-Menge et al., 2006). Between 1994 and 2007 the annual average surface melt was 0.64 m in the Beaufort Sea and 0.26 m at the North Pole. Bottom melt ranged between 0.1 and 0.4 m at the North Pole, but can be highly variable and up to 2.1 m in the Beaufort Sea (e.g., Perovich et al., 2008), depending on ice concentration and the amount of shortwave radiation absorbed by the ocean surface mixed layer. Basin-scale seasonal variations (including deformed ice) retrieved from multiple submarine cruises show a similar pattern, with a peak-to-trough amplitude of 1.06 m between 1975 and 2000 (Figure 9.13; Rothrock et al., 2008).
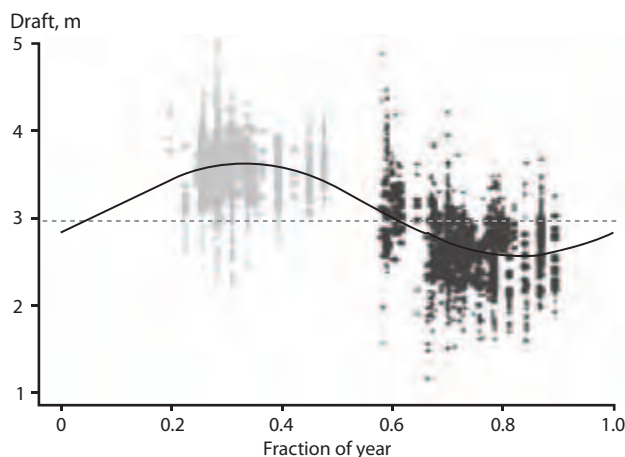
Figure 9.13. The annual cycle of sea-ice draft averaged for the period 1975 to 2000 over the Central Arctic Basin. Source: Rothrock et al. (2008). Dots show the variability of individual profile sections used for the reconstruction, black for summer/autumn, grey for winter/spring. Dots within one standard deviation from the mean are plotted heavier.

9.1.4. Regional changes in sea ice

9.1.4.1. Trends in extent

Over the satellite record since 1979, sea-ice extent has declined both during summer and winter in virtually all regions of the Arctic (Meier et al., 2007; Comiso and Nishio, 2008; Parkinson and Cavalieri, 2008). The most substantial declines have occurred during summer. Many regions in winter show

no trend because they are completely ice covered through the winter months. However in regions with winter variability, most regions show small declining trends, with the exception of increases in the Bering Sea during winter related to changes in atmospheric circulation (Table 9.2). There are few observations for many regions from the pre-satellite records, except for the Russian Arctic (see Section 9.1.2.3).

9.1.4.2. Trends in thickness

Thickness data obtained by methods other than satellite altimetry or submarine ULS are limited to smaller geographical regions due to logistical challenges. Significant data have been obtained in some regions of the Arctic Ocean using ships as operating platforms for drilling, visual/video observations, or electromagnetic induction (EM) surveys, or aircraft for laser- and radar altimetry and EM sounding (Box 9.4). However, the longest time series have been obtained from landfast ice, which is attached to coasts and therefore most accessible. As landfast ice is typically a seasonal phenomenon in the Arctic, it undergoes annual or shorter-term cycles of formation and break-up which sometimes hamper interpretation of thickness data. Many landfast ice records have not been continued in the past ten years due to closure of monitoring stations. Results from fast-ice studies may differ from results of drift-ice studies because fast-ice thicknesses primarily represent thermodynamic growth in shallow water, whereas drift ice comprises both thermodynamically grown and deformed ice, and its thickness is an integrated result of conditions experienced during its drift.

Table 9.2. Trends in sea-ice extent in percentage per decade for the subregions of the Arctic relative to 1979–2009. Statistical significance is indicated by ***bold/italic for 99.9%***, **bold for 99%**, and *italic for 95%*. Regions are as defined by Meier et al. (2007); the Arctic Ocean region includes the Laptev, Siberian, Chukchi, Beaufort, and Central Arctic regions. Regions/months without value have too little ice to yield reliable statistics. Updated through 2009 based on NSIDC data (Cavalieri et al., 1996, updated 2008; Meier et al., 2006, updated quarterly; Maslanik and Stroeve, 1999, updated daily).

Region	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Pan-Arctic Total	*-3.3*	*-3.0*	*-2.9*	*-3.1*	*-2.7*	*-3.5*	*-6.4*	*-9.1*	*-11.7*	*-5.7*	*-4.6*	*-3.3*
Bering Sea	*6.3*	2.0	-1.1	1.6	-1.7	-13.0					-22.2	-1.2
Hudson Bay	0.0	0.0	0.0	0.0	*-0.2*	-5.6					-26.7	*-1.6*
Baffin Bay	*-7.4*	*-5.8*	*-5.9*	*-6.2*	*-6.6*	*-9.9*	*-18.9*	*-28.5*	*-10.1*	*-24.6*	*-11.0*	*-11.3*
Greenland Sea	*-11.0*	*-11.6*	*-11.0*	*-9.2*	*-6.4*	*-5.6*	*-9.5*	*-14.5*	*-13.0*	*-6.5*	*-7.8*	*-10.6*
Barents Sea	*-16.8*	*-13.0*	*-10.2*	*-12.7*	*-18.0*	*-23.3*	*-30.0*	*-37.4*	-28.8	*-18.4*	*-15.9*	*-20.2*
Kara Sea	0.0	0.0	0.0	-0.1	0.0	*-1.5*	*-14.4*	*-21.8*	*-20.9*	*-7.7*	*-3.3*	*-1.0*
Laptev Sea	0.0	0.0	0.0	0.0	**0.0**	-1.2	*-4.1*	*-12.6*	*-15.9*	-0.2	0.0	0.0
Siberian Sea	0.0	0.0	0.0	0.0	0.0	0.2	-1.4	*-16.2*	*-25.6*	*-7.1*	*-0.2*	0.0
Chukchi Sea	0.0	0.0	0.0	0.0	*-0.2*	*-3.5*	*-6.7*	*-18.7*	*-31.6*	*-20.9*	*-9.2*	0.0
Beaufort Sea	0.0	0.0	0.0	0.0	0.0	-1.2	-3.1	*-6.7*	*-12.9*	*-3.1*	0.0	0.0
Canadian Archipelago	0.0	0.0	0.0	0.0	-0.1	*-1.3*	*-2.3*	*-7.5*	*-12.3*	*-4.3*	0.0	0.0
Central Arctic	0.0	0.0	0.0	0.0	*-0.1*	0.0	*-0.2*	*-0.6*	*-1.4*	-0.2	*-0.1*	-0.1
Arctic Ocean	0.0	0.0	0.0	0.0	0.0	*-0.7*	*-2.0*	*-6.9*	*-10.7*	*-3.7*	*-1.1*	0.0

9.1.4.2.1. **Results from drift ice regions**

During occasional icebreaker cruises into the central Arctic since 1991, visual and drill-hole thickness measurements have been collected, later supplemented by ground and airborne EM surveys. Haas et al. (2008) found significant thinning in the region of the Transpolar Drift. Between 1991 and 2004, modal thickness decreased from 2.5 to 2.2 m in 2004, and then dropped to 0.9 m in 2007 (Figure 9.14). Mean thickness decreased similarly strongly. The thinning was accompanied by reductions in the age of the ice, and replacement of second-year ice by first-year ice in 2007. These changes hamper the separation of thermodynamic and dynamic causes for the observed thinning. Ice thicknesses and their changes are in general agreement with other surface observations (Perovich et al., 2009) and visual and video observations from Russian icebreakers (Frolov et al., 2009), but quantitative comparisons are difficult due to different ship routes, mixtures of ice types, and effects due to proximity to the ice edge.

Most sea ice exits the Arctic Basin through Fram Strait. Moored ULS have been operated across the Strait to monitor ice export since 1990. Until 1996, the data show the export of ice mostly between 1.5 and 3 m (sometimes up to 4 m) thick on average, but with large spatial and temporal variability (Vinje et al., 1998). The latter showed some correlation with phases of the North Atlantic Oscillation. Spreen et al. (2009) combined ICESat thickness retrievals and satellite observed ice drift to update estimates of ice flux through Fram Strait, but observed no significant changes. Published data to date do not allow derivation of any thickness trends in Fram Strait, although ice thickness and volume are clearly affected by the drift and origin of the ice and by variations in the proximity of the ice edge, particularly in summer.

The Barents Sea ice is dominated by FYI and occasional older ice advected from the Arctic Ocean by the East Spitsbergen Current. Along the coast of Svalbard and in Storfjorden, ice can become very thick by deformation. Recent airborne EM surveys over Storfjorden have shown that modal thickness of coastal drift ice can be as high as 2.4 m, significantly thicker than fast ice due to heavy deformation (Hendricks et al., 2011). Deformation and thickness are strongly dependent on prevailing onshore or offshore winds.

Some of the thickest ice of the Arctic Ocean is located north of the coasts of Ellesmere Island and in the Lincoln Sea, with modal thicknesses over 4 m and mean thicknesses ranging from 4 to 6 m (Haas et al., 2006; see also Box 9.2). Due to the high degree of deformation in the region modal thicknesses may not represent thermodynamic equilibrium thicknesses. Between 2004 and 2009, there were no significant changes, but some interannual variability is obvious (Haas et al., 2010), which may be related to the variable occurrence of ice arches in the Lincoln Sea and Nares Strait (Kwok et al., 2010).

Moored ULS have been operated for 12 years in the Eastern Beaufort Sea (Melling et al., 2005). Between 1991 and 2003, ice draft has shown a small thinning trend of -0.07 m/decade, but this has low statistical significance as the net change is comparable to the uncertainty in the measurements. Mean late-winter draft was about 2 m. Airborne ice thickness surveys have also been performed in the region in different seasons (Peterson et al., 2008; Barber et al., 2009), but the data are currently too sparse and ice types too varied to be comparable.
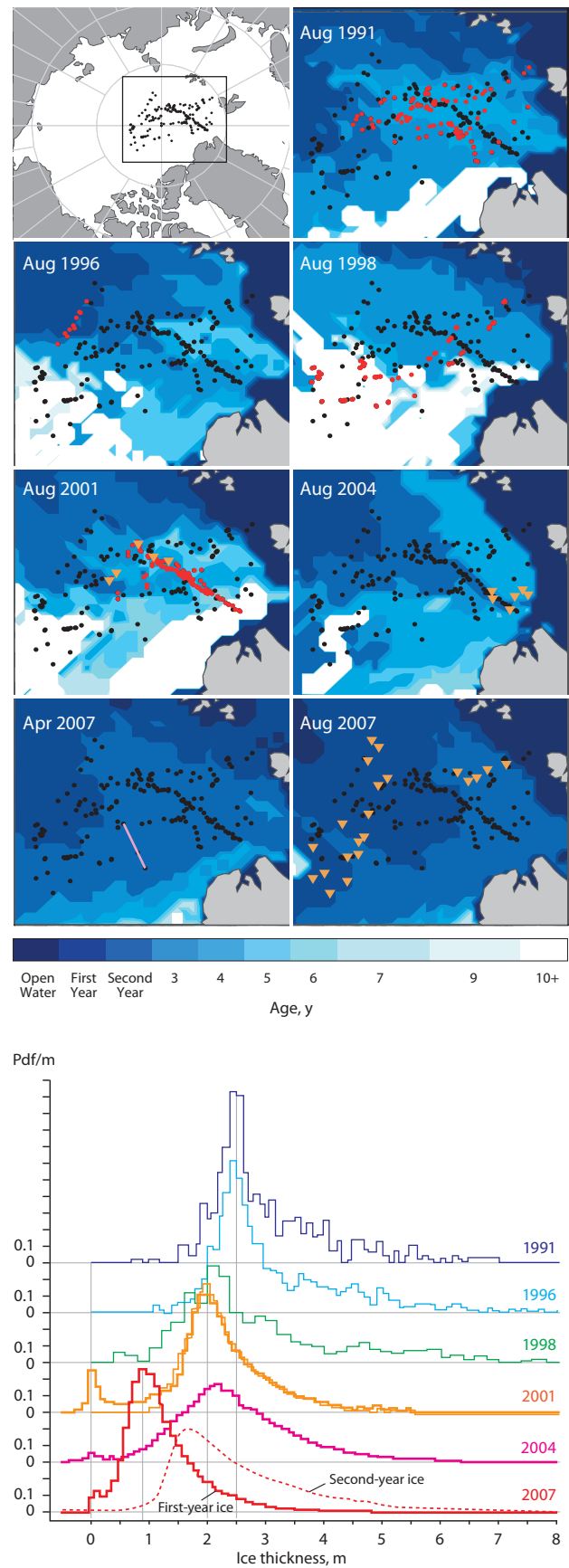
Figure 9.14. Late summer sea-ice age (above) and thickness (below) in the region of the North Pole between 1991 and 2007. Thickness was obtained by means of ground-based (thin lines, •) and airborne (thick lines, ▼) electromagnetic induction sounding. In 2007, first-year ice prevailed in August, but second-year ice was still present in April 2007. The latter distribution was seasonally adjusted to be comparable with summer conditions. Source: Haas et al. (2008).

The thickness distribution of drifting sea ice in the Alaskan Arctic is dominated by a range of processes that result in significant spatial and interannual variability (Boxes 9.1 and 9.2): production of thin, young ice in coastal polynyas (Martin et al., 2004; Melling et al., 2005); advection of MYI from the High Arctic in the Beaufort Gyre (Tucker et al., 2001; Shirasawa et al., 2009); growth of FYI with the commencement of freeze-up and surface and bottom summer melt (Perovich et al., 2003). As a result, drifting sea ice in the Alaskan Arctic is sensitive to changes in forcing, and hence significant trends are more challenging to detect. Generally, the substantial reduction in overall mean (drifting) ice thickness is due to the loss of MYI and greatly enhanced (bottom) melt (Perovich et al., 2008) resulting from a combination of changes in ice circulation taking ice out of the Beaufort Gyre and ice-albedo feedback enhancing solar heating of the region. According to ice-age data (updated from Maslanik et al., 2007), in recent years considerable amounts of old ice advected into the Beaufort and Chukchi Seas have melted completely during summer. However, the thickness of FYI close to the end of winter did not change significantly between 2007 and 2009, and modal thicknesses ranged from 1.6 to 1.8 m (Haas et al., 2010).

9.1.4.2.2. **Results from landfast ice regions**

Landfast (or simply 'fast') ice is non-drifting ice typically locked in place by islands, icebergs, or grounded ridges. The appearance and extent of landfast sea ice vary markedly around the Arctic, typically extending to the shelf break. However, fast-ice extent depends not only on direct thermodynamic boundary conditions, but also on effects of the timing and amount of snow deposition (e.g., Gerland et al., 1999; Hanesiak et al., 1999; Nicolaus et al. 2003) and ocean mixed layer properties (Carmack et al., 2006; Barber et al., 2008b). There is general evidence that the annual cycle of landfast ice is shortening as the Arctic sea-ice retreats (Reeh et al., 2001; Serreze et al., 2007) and air temperatures rise (Comiso, 2006).

In the Russian Arctic, interannual variability of landfast sea-ice extent has been linked to variability in the discharge of the large rivers that drain into the Russian Arctic (Dmitrenko et al., 1999). Landfast ice extent and thickness in the Laptev, East Siberian, and Chukchi Seas exhibited strong multidecadal variability during the 20th century. This variability correlates with both dynamic and thermodynamic atmospheric forcing, but makes it difficult to discern any clear long-term trends (Polyakov et al., 2003). Long time series of fast-ice thickness anomalies have been obtained at five sites from 1936 to 2000, and show an increase in ice thickness from the 1930s to about 1960, and then a decrease from 1960 to 2000 (Figure 9.15; Polyakov et al., 2003). The five individual stations show that trends are a few centimeters per decade and can be both positive and negative for different sites, but that the changes are not very significant.

Along the northern Alaskan coast, Mahoney et al. (2007) found no evidence of long-term change in landfast ice extent between the 1970s and the period 1997 to 2004. However, there were significant differences in the timing of the annual landfast ice cycle in this region. In comparison with observations from the 1970s (Barry et al., 1979), landfast ice formed about one month later along the Alaskan Chukchi coast and broke out on average 6 days and 19 days earlier along the Chukchi and Beaufort Sea coasts, respectively. The later formation of landfast

Ice thickness anomaly, cm
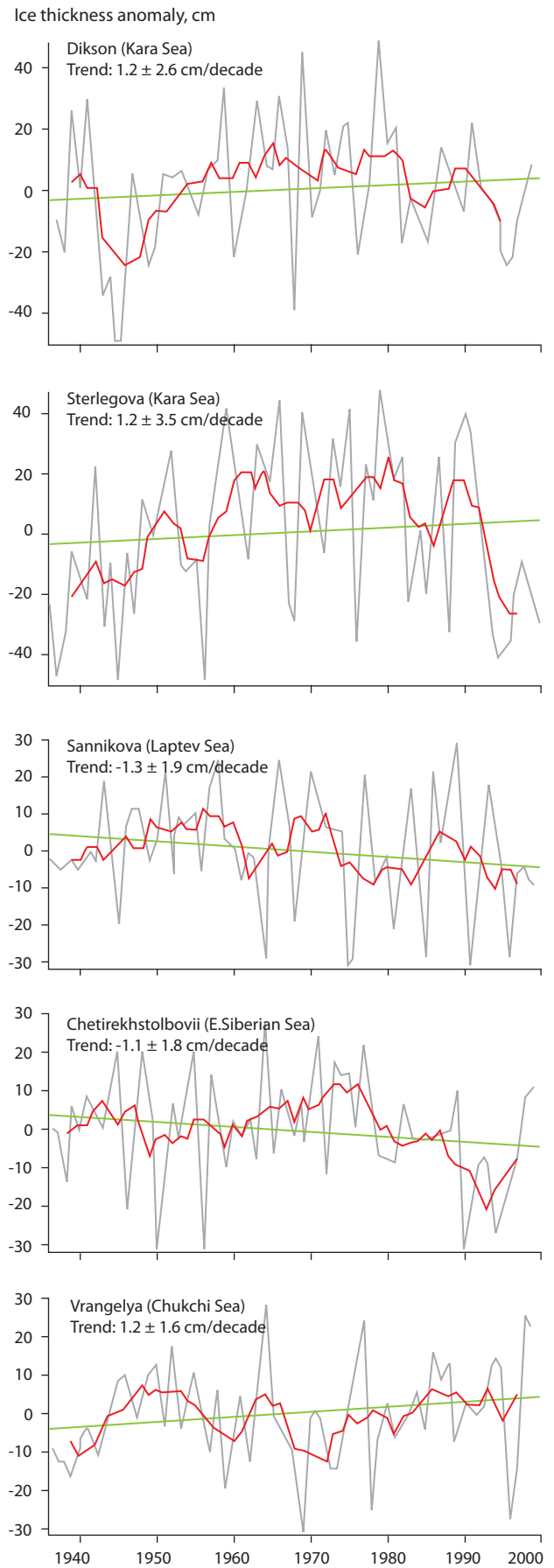
Figure 9.15. Time series of May fast-ice thickness anomalies at five locations along the Siberian coast. Grey lines show yearly May values and red lines the 6-yr running mean, green lines show linear trends (quoted limits represent 95% confidence levels). Source: Polyakov et al. (2003).

ice in recent years is likely to be related to the later incursion of pack ice into the coastal zone, whereas early breakouts of landfast ice are most closely correlated with earlier onset of thawing air temperatures in summer.

This broadly agrees with measurements collected from Barrow, Alaska over a period of about ten years (Druckenmiller et al., 2009), which indicate thinner ice compared to data published for the 1970s (Weeks and Gow, 1978). These studies also indicate that ice thickening through ridging has changed due to the absence of MYI in recent years, reducing the number of deep-draft, grounded ridges that help stabilize the ice cover (Shapiro and Barnes, 1991).

Ice of the Canadian Arctic Archipelago is a mix of MYI and FYI types and becomes landfast during winter. Extensive drill-hole data were obtained in the 1970s through oil exploration activities. The average ice thickness in late winter was 3.4 m but local means reached 5.5 m (Melling, 2002). First-year ice growth amounted to about 2.3 m at the time. Earlier data from the period 1959 to 1989 showed no clear thickness trends (Brown and Cote, 1992) with thicknesses varying interannually due to both dynamic and thermodynamic atmospheric forcing.

Seasonal fast-ice and snow-cover thickness in Svalbard fjords and coastal ice near Svalbard have been extensively studied since 1997, showing generally decreasing thicknesses (Gerland et al., 1999; Gerland and Hall, 2006; Cottier et al., 2007; Gerland and Renner, 2007). However, interpretation of these changes is difficult because fast-ice evolution in Svalbard fjords varies strongly regionally, with the west coast influenced by the relatively warm Atlantic water of the West Spitsbergen Current, and with gradients in snow deposition. Recent airborne EM surveys over Storfjorden have shown that modal thicknesses of coastal drift ice can be as high as 2.4 m, significantly thicker than fast ice, due to heavy deformation (Hendricks et al., 2011).

A long time series of drill-hole data has been obtained on fast ice off the coast of the island of Hopen (Gerland et al., 2008a). Between 1966 and 2007, seasonal maximum thicknesses decreased by 0.1 m per decade, although the interannual variability is high (Figure 9.16). The latter is also related to the high variability of Barents Sea ice extent, which can contribute to early fast-ice removal on the fairly exposed coast of Hopen.
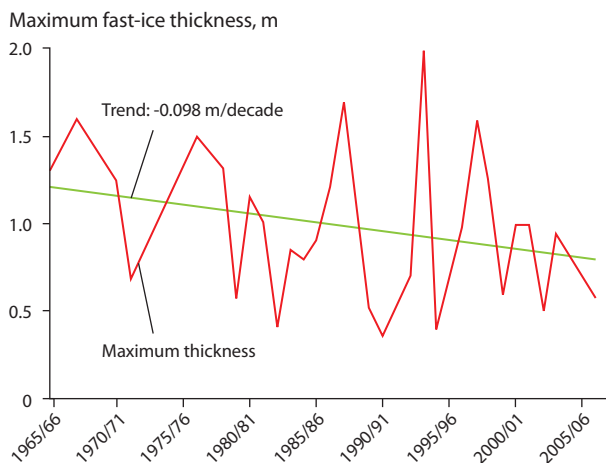
Figure 9.16. Maximum seasonal sea-ice thickness for the island of Hopen in the Barents Sea. The linear trend is significant at the 0.05 level using a 'significant zero crossings of derivatives test'. Source: Gerland et al. (2008a).

9.1.5. Observations and importance of snow thickness

The snow cover is an essential part of the sea-ice thickness profile (see Box 9.4) and plays an important role in the sea-ice mass balance. Snow strongly modifies the surface albedo of the sea-ice cover and the heat flux through the ice and the amount and distribution of snow on sea ice also affects summer melt pond evolution (e.g., Barber and Yackel, 1999; Yackel et al., 2000; Hanesiak et al., 2001). It also contributes to ice thickness through snow-to-ice transformation (superimposed ice and snow-ice formation). The various controls of the snow cover on the sea-ice mass balance and feedbacks are described in Section 9.2. In addition, snow thickness and density variations affect the accuracy of many sea-ice thickness measurement methods (see Box 9.4).

Despite its importance, there is a lack of snow thickness data over Arctic sea ice due to methodological and logistical limitations in measuring this thin but highly stratified and laterally variable layer. Most datasets are very local and provide only one snapshot in time, both seasonally and interannually. Radionov et al. (1996) and Warren et al. (1999) reviewed data from Russian Arctic drifting stations obtained between 1954 and 1991 by manual snow stake measurements. Snow accumulates rapidly in September and October, moderately in November, very slowly in December and January, then moderately again from February to May. Snow melt on sea ice in the Arctic Ocean typically commences in June, and during August the sea-ice cover is mostly snow free but covered with melt ponds. During the 37-year study period, average snow depth on MYI reached a maximum of 34 cm in May. The deepest snow was observed just north of Greenland and Ellesmere Island, peaking in early June at more than 40 cm, when the snow is already melting north of Siberia and Alaska. Similar results of the annual cycle and maximum end-of-winter snow thicknesses between 30 and 40 cm were obtained by Brown and Cote (1992), Flato and Brown (1996), and Melling (2002) in the Canadian Arctic and Archipelago, and by Sturm et al. (2002) in the Beaufort Sea. Gerland and Haas (2011) showed snow thickness data obtained during expeditions across the Arctic Ocean. In April and May 2007, snow thickness varied from 10 to 15 cm on the Siberian side to around 40 cm off the northern coast of Greenland.

Kwok and Cunningham (2008) have developed a method to construct daily Arctic-wide snow thickness fields from precipitation fields calculated by the European Centre for Medium-Range Weather Forecasts (ECMWF), and by tracking ice floes to account for different accumulation rates along the drift tracks of the ice. Results between 2003 and 2008 are in general agreement with earlier data presented by Warren et al. (1999), and show small interannual variability of only a few centimeters (Kwok et al., 2009).

9.1.6. Changes in factors controlling sea-ice extent and thickness

Sea-ice extent and thickness evolve in response to large-scale and local atmospheric and oceanic forcing on both the dynamics and thermodynamics of the ice (Boxes 9.1 and 9.2). The most direct link between climate forcing and sea ice is through surface air temperature. Increasing air temperatures have been observed in recent decades (Rigor et al., 2000; Johannessen et al., 2004;

Overland et al., 2004, 2008b). Changes in atmospheric circulation have also played an essential role, particularly in the reduction of MYI and amount of deformation. Rigor and Wallace (2004) found that a strongly positive Arctic Oscillation (AO) enhanced the export of MYI between around 1989 and 1996. However, export through Fram Strait seems to be more coherent with the smaller-scale pattern of the North Atlantic Oscillation and the position of its centers of interannual variability. During the 1980s and 1990s, these shifted eastward compared to 1958–1977, resulting in anomalous meridional winds (Hilmer and Jung, 2000). Model results by Koenigk et al. (2006) and Hilmer et al. (1998) showed that the variability of the Fram Strait sea-ice export is mainly governed by the local sea level pressure (SLP) gradient across Fram Strait. The high AO-phase in the late 1980s and 1990s led to an increased SLP-gradient due to an anomalous extension of the Icelandic Low into the Arctic. However, results from models and reanalysis data show there is no significant correlation between AO/NAO and Fram Strait sea-ice export before 1978 (Jung and Hilmer, 2001).

Drifting buoy data suggest a relationship between sea-ice drift and change of atmospheric sea level pressure pattern regressed on the linear decreasing trends of the Arctic sea-ice extent in September (Inoue and Kikuchi, 2007). In recent years the Arctic Dipole Anomaly, characterized by two SLP anomaly centers that result in increased advection of ice across the Arctic, has been strong (Wang et al., 2009). Cloud cover and its influence on the amount of incident solar and longwave radiation has also had an influence (Schweiger et al., 2008a; Kay and Gettelman, 2009).

A key factor in the seasonal evolution of the surface albedo is the date of the onset of melt, when the early spring albedo values drop from ~0.85 to 0.71 or less (Perovich et al., 2002a). In the 2000s, melt onset has begun about two weeks earlier than in the 1980s (Stroeve et al., 2006) according to measurements from satellite passive microwave radiometers (Anderson and Drobot, 2001). Trends toward earlier melt onset dates in most regions of the Arctic have been observed, from -1.0 d/decade in the Bering Sea to -7.3 d/decade in the East Greenland Sea (Markus et al., 2009). Earlier melt onset may lead to an earlier disappearance of snow cover and earlier appearance of melt ponds and maximum melt pond coverage. This significantly changes the total absorbed solar flux during the summer season, which is an important factor controlling sea-ice extent and thickness through the summer melt season. Perovich et al. (2007) found that the total absorbed flux varies linearly with the date of melt onset, increasing by 25% if the onset moves from about 15 June to 23 May. The dependence on the date of freeze-up or the total duration of the melt season is much weaker as this occurs when solar elevation is already low. Albedo is also affected by snow properties and surface contaminants such as black carbon (see Section 9.2).

In addition to the effect on melt and albedo, the atmosphere affects sea ice through wind forcing and its influence on ice dynamics (Boxes 9.1 and 9.2). Ice drift speed within the Arctic Ocean has increased substantially in recent years. Based on drifting buoy data, Rampal et al. (2009) found sea-ice mean speed increased 17% per decade for winter and 8.5% for summer over the period 1979 to 2007. Concomitant with this increase in mean speed, the strain rate of the ice also increased significantly over this period (51% per decade in winter and

52% in summer). Rampal et al. (2009) showed that changes in wind forcing were not important in determining the change in ice speed. Instead, they suggested that sea-ice kinematics play a fundamental role in the albedo feedback loop and sea-ice decline: increasing summer deformation means stronger fracturing and hence more lead opening and reductions in albedo. This accelerates sea-ice thinning in summer and delays refreezing in early winter, decreasing the mechanical strength of the ice cover and allowing even larger drift speeds and deformation, and possibly faster export of sea ice through Fram Strait. Kwok and Rothrock (1999), Kwok et al. (2004), and Kwok (2008) examined the sea-ice area flux through Fram Strait from passive microwave ice motion data, and found a small positive trend over the period 1978 to 2008. A stabilizing process is found in the increased ridging and ice production that would also accompany increased winter ice deformation rates.

Decreased ice concentration during summer melt results in significant *in situ* heating of the ocean surface, which can then further melt ice (Steele at al., 2008). Significant bottom melt of sea ice due to ocean heat has been observed in recent years (Perovich et al., 2009). The role of the oceans in forcing the observed reduction in ice thickness and extent is still a matter of debate. Most or all of the heat of the Atlantic water away from the continental shelves is well sequestered below a strong halocline created by the fresher Pacific water, river input, and a net accumulation of freshwater from precipitation minus evaporation. On the other hand, the Pacific water entering the Arctic Ocean from the Bering Strait is much closer to the surface and may play a role in the reduction of the ice extent in localized regions in the Chukchi Sea and perhaps beyond (Zhang et al., 2009). The precise role of ocean heat transport in the sea-ice decline is more ambiguous. Observations by Shimada et al. (2006) and Woodgate et al. (2010) suggest that increased inflow of warm Pacific waters strongly influenced the dramatic reductions in summer sea ice in the western Arctic. Within the Arctic basin the solar heat absorbed by the ocean may be stored so that ice that advects over the region is rapidly melted (Perovich et al., 2009) or the formation of new ice is delayed. Ocean heat has always played a much larger role in the peripheral seas, such as the Bering and Barents Seas.

The condition of the ice influences its future state. As sea ice thins, it becomes more susceptible to melting completely over the course of a summer, which leads to associated melt, albedo, and ocean forcings (Holland et al., 2006; Lindsay et al., 2009). This leads to the concept of 'pre-conditioning', where a thinner ice cover going into the summer melt season is a strong indicator of low total extent at the end of summer. Model simulations demonstrate this effect and show that the initial condition of the ice thickness distribution is an important factor in accurate predictions of summer minimum extent (Lindsay et al., 2008; Zhang et al., 2008a,b; Kauker et al., 2009).

There is still much to understand about the interplay of these factors and their influence on sea ice. It is clear that these factors work together in complex ways and that there are many feedbacks between the sea ice, atmosphere, and ocean (discussed in detail in Section 9.2). The climate forcings and feedbacks with sea ice will ultimately determine the future evolution of the sea-ice cover. Model forecasts of this future evolution and comparisons with observations are addressed in Section 9.1.7.

9.1.7. Modeling: Comparing model output with observational data

There is considerable scatter among simulated rates of Arctic sea-ice loss by most atmosphere-ocean-ice general circulation models (AOGCMs), but in general the observed summer sea-ice loss is faster than anticipated (Stroeve et al., 2007). By 2008, the Arctic had already reached a milestone of two sequential years of low summer sea-ice extent, a situation that the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4, IPCC, 2007) projected would not occur until later in the century (Stroeve et al., 2007). However, the reasons for this are complex, and are likely to be related to insufficient parameterizations of the ice-albedo feedback, ice dynamics and deformation, and heat storage in the ocean surface mixed layer. Differences between models result from their different formulations, the impact of random phasing of natural low frequency variability, and different future greenhouse gas emissions scenarios, as shown by Zhang and Walsh (2006).

Ridley et al. (2007) investigated the sensitivity of the Hadley Center Model (HadCM3) to perturbed model parameters and found a large range in the sensitivity of Arctic sea-ice retreat to global temperature change, from 11% to 18 % per °C. From simulations of the Community Climate System Model (CCSM3), Holland et al. (2006) showed that AOGCMs simulate instances of rapid September ice loss much like that which occurred in 2007. However, from the seven ensemble members considered, there is a wide range in the timing of such events with the earliest event occurring in 2013 and the latest not initiated until 2045. These ensemble members use exactly the same model but the initial conditions are slightly different. Given the chaotic nature of natural variability in the climate system, different phasing of the decadal maxima and minima of the sea-ice loss due to natural variability can be expected, in addition to the long-term trend due to external (anthropogenic) forcing. A critical result is that the apparent sea-ice response to the sum of natural and forced atmospheric and oceanic changes is not linear. Once sea ice is removed there is no return to previous conditions as long as the concentration

of greenhouse gases remains high. This is what is probably occurring in the real world after the summer sea-ice losses of 2007 and 2008. However, there is still high natural variability in the system and increases the following year from record or near-record lows, as was seen in 2009, are not unexpected in model projections (Holland et al., 2006; Goosse et al., 2009). It is notable that over 50% of the IPCC AR4 models simulate instances of rapid September ice loss similar to the 2007 observed transition (Holland et al., 2006). This suggests that while models generally underestimate the historical ice loss, they do reasonably simulate the superposition of natural variability and anthropogenically forced change that allows for instances of rapid ice retreat.

Such results are summarized in Figure 9.17 which shows projected sea-ice extent for a subset of the 23 IPCC AR4 models. The six models shown all satisfy an observational constraint, i.e., that they show a similar amount of initial (i.e., pre-industrial) September sea ice and the same seasonal range as the Hadley sea-ice dataset (Rayner et al., 2003). It is clear from the figure that the selected models project a faster rate of sea-ice loss than the average of all models. In addition, some of the time series are near the recent sea-ice minima. If a model was initialized with the observed 2007/2008 minimum extents, a future loss of sea ice to 1 million km^2 in summer would be expected to occur around 2040, as the future rate of sea-ice loss is dependent on the magnitude of initial conditions (Zhang and Walsh, 2006). However, models respond differently to sudden change and may return to the long-term trend behavior, which corresponds to the prevailing forcing, rather than continue on the trend implied by the recent observations. The ensemble average, which removes the sensitivity due to natural variability and model differences, should be close to the sea-ice response to greenhouse forcing alone. The conclusion that sea ice is retreating faster than expected is based on the comparison of the apparent fast track that the real world is on and in a few ensemble members, representing the contribution from natural variability, compared to the ensemble mean behavior.

The reasons for the large variation in simulated ice-extent loss among the different climate models are difficult to determine. Simulated ice-extent loss is related to the initial

September sea-ice extent, million km^2
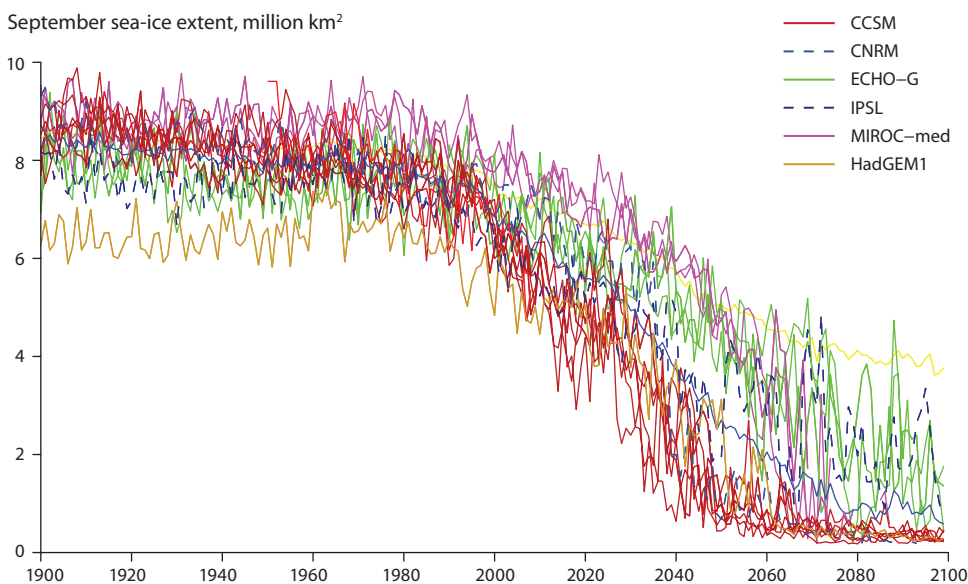
Figure 9.17. Northern hemisphere September sea-ice extent as simulated by six IPCC AR4 under the IPCC A1B emissions scenario. The models were selected for simulating mean and seasonality of sea-ice extent within 20% of the observed values. The thick blue line is the ensemble mean for the six models and the thick yellow line is the ensemble mean for all IPCC AR4 models. The thick red line shows the observed values based on Hadley Centre sea-ice analysis (HadISST). Source: after Wang and Overland (2009).

(mid-20th century) sea-ice state, the changing ice-mass budgets, and how ice-volume loss translates into ice-extent change. Holland et al. (2008a) assessed Arctic sea-ice mass budget changes over the 20th to 21st century from 14 IPCC AR4 models and related these to the changing September ice extent. As shown in Figure 9.18, the relationship between ice-volume loss and ice-extent loss is nonlinear. Models with excessively thick ice generally simulate the largest rates of ice-volume loss. However, this translates into very little ice-extent change (Figure 9.18). In contrast, when the Arctic sea ice thins below about 2 m, which is the typical maximum thickness obtainable by FYI during one winter, substantial rates of ice-extent loss occur as large regions of the ice pack can completely melt out during summer. Models do vary considerably in the September ice-extent change realized by a similar ice-thickness loss even in this thinner ice regime. As shown by Holland et al. (2008a) one contributing factor appears to be the spatial distribution of ice within the Arctic; models with a smaller spread in the March Arctic ice-thickness distribution tend to simulate larger ice-extent loss for a similar mean ice-thickness change. Thus, to simulate a reasonable Arctic ice-extent response, it is necessary to realistically simulate both the mean and spatial distribution of ice within the Arctic.

Bitz (2008) also investigated some of the reasons for the uncertainty in simulated Arctic sea-ice response to external forcing. This author determined that the role of uncertainties in the strength of the surface albedo feedback for sea-ice thickness loss were small. However, the influence of the mean state on uncertainty in sea-ice thickness response (and by extension on simulated sea-ice extent loss) was considerable. This is likely to be due to the importance of the sea-ice thickness for the strength of the stabilizing sea-ice thickness – sea-ice growth rate feedback. This agrees with the results of Holland et al. (2008b) that the initial sea-ice state is critically important for the simulation of the sea-ice thickness and sea-ice extent response to external forcing.

Regional sea-ice models have been used to investigate the potential causes of the 2007 summer sea-ice loss. Kauker et al. (2009) and Wang et al. (2009) looked at anomalous wind patterns, sea-ice thickness and ocean transport. Lindsay et al. (2009) suggested a long-term trend in sea-ice volume. Gorodetskaya et al. (2008) noted a range of simulated cloud properties in different IPCC models which have an impact on the surface radiation balance for sea ice. The Arctic Ocean Model Intercomparison Project (AOMIP) compared sea-ice concentrations and sea-ice drift in 20th century hindcasts (Gerdes and Koeberle, 2007; Johnson et al., 2007; Martin and Gerdes, 2007). Compared to observations, differences were a result of changes in both the atmospheric forcing used and model characteristics. Model to model uncertainties become larger when sea ice is investigated on a regional basis (Parkinson et al., 2006; Overland and Wang, 2007).

Looking forward, it is clear that models indicate large year-to-year variability in the total summer sea-ice extent, making short-term predictions difficult. As summer sea-ice extent continues to decline, model simulations indicate that this variability will initially increase due to the larger fraction of thinner, FYI that is more susceptible to variable seasonal atmospheric and oceanic forcing (Goosse et al., 2009). As the decline continues further, the variability will decrease due to less summer sea ice overall.

The amount of year-to-year climate variability may be key to the timing of a transition to a seasonally sea ice-free environment. A modeling study investigating future sea-ice response to warming found that, as expected, a stronger average temperature trend makes an earlier loss of summer sea ice more likely (Zhang et al., 2010). However, the year-to-year variability in temperature was also found to play a crucial role. Higher variability, similar to that of the past two decades, increases the likelihood of summer ice-free conditions by 2050, while lower variability (similar to conditions between the late 1940s and the late 1980s) results in a reduced chance of ice-free conditions by 2050.

Understanding the processes influencing sea ice and feedbacks with the climate system, discussed in Section 9.2, are important to developing better models and improved predictions. Accurate predictions are crucial for anticipating and responding to impacts on Arctic biological systems (see Section 9.3) and human society (see Section 9.4).
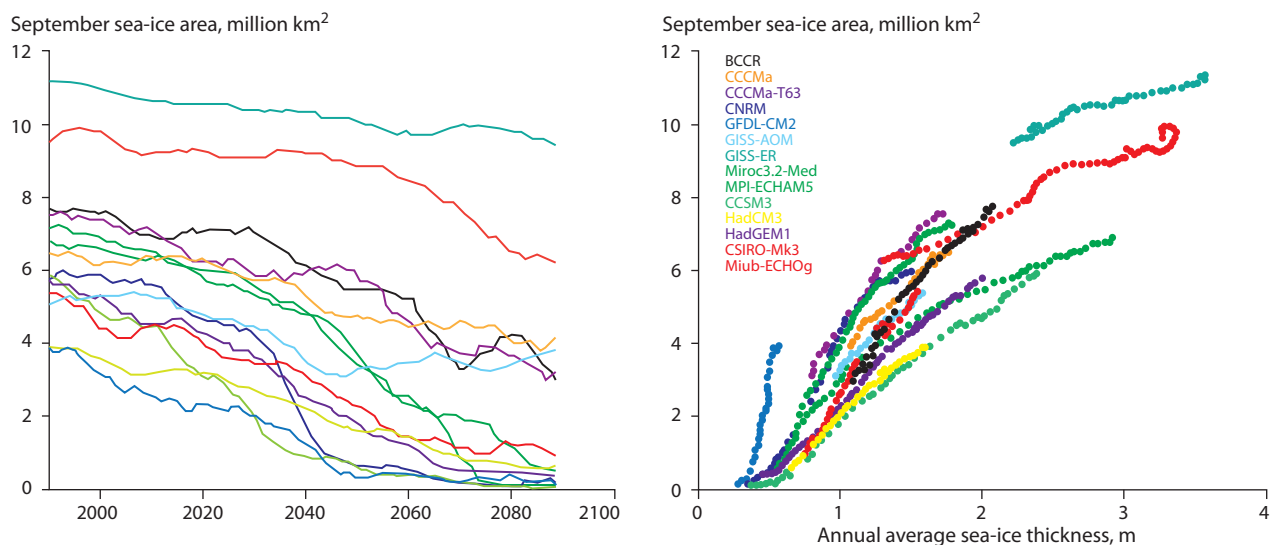
Figure 9.18. Projected change in September sea-ice area within the Arctic (left) and the 21st century September sea-ice area *vs* annual average sea-ice thickness (right) for 14 different CMIP3 models. Source: Holland, et al. (2008a).

9.2. **Physical feedbacks and forcings**

- Recent research has better quantified some interactions between the sea ice, ocean, and atmosphere, particularly during the extreme summer ice loss of 2007.

  - Trends in cloud cover have suppressed some surface warming over the Arctic Ocean.

  - Loss of summer sea ice and increased upper ocean heat storage is modifying regional wind patterns, with the potential for impacting the mid-latitudes over the coming decades as sea ice retreats further.

  - Reduced sea-ice cover has increased solar heat input to the ocean and the amount of melting: an ice-albedo feedback.

  - Heat fluxes from Bering Strait and warmer Atlantic waters are playing a larger role in the decline of sea-ice cover.

9.2.1. **Introduction**

The observed decline in sea-ice cover is directly linked to dynamic and thermodynamic forcing and is impacted both by positive and negative feedback processes. Polar amplification of climate warming is attributed predominantly to positive feedbacks, including the ice-albedo feedback, and cloud-radiation feedback. Although there is much evidence for significant warming in the Arctic, the attribution and interdependence of the snow and ice albedo and cloud feedbacks on polar amplification is still an area of active research.

This section discusses the feedbacks and forcings that are contributing to the observed decline in Arctic sea-ice cover, as well as those which may be slowing this decline. Specifically, the ice-albedo feedback and the cloud-radiation feedback are examined. The consequences of black carbon on sea-ice albedo are considered and there is a discussion of the impact of surface morphological features on sea-ice conditions and feedback processes. Projections of future ice decay due to feedbacks are presented. Finally, the concept of sea-ice tipping points is introduced and discussed.

9.2.2. **The ice-albedo feedback**

As ice, or the snow on ice, melts its albedo decreases (see Box 9.5). This results in additional absorption of solar radiation by the ice and more melting, further decreasing the albedo. This positive loop is the ice-albedo feedback. Under warming climate conditions, this feedback can act on longer time scales to enhance the climatic changes. An initial warming of the climate will cause the onset of summer melt to begin earlier and the autumn freeze-up to start later. This leads to a larger area of the Arctic Ocean being covered by open water and darker sea-ice surfaces for a longer period of time. This longer melt season and reduced ice area leads to an increase in the absorption of solar energy, which goes to melting more ice or to heating the ocean. Heat in the ocean will delay the onset of autumn freeze-up, possibly resulting in thinner ice. This process acts over years to decades as a positive feedback on a larger, regional scale. A cooling climate would reverse the process, also leading to a positive feedback, with more ice driving more cooling.

Box 9.5. Sea-ice albedo

Albedo is the fraction of incident sunlight reflected by a surface. A surface that is a perfect absorber of sunlight has an albedo of zero, while a perfect reflector has an albedo of one. Albedo is a key parameter for the surface heat budget, as it strongly influences the radiation budget. Snow-covered sea ice has a much larger albedo than an ice-free ocean surface, meaning that it reflects much more of the sun's energy, limiting the ability of an ice-covered region of ocean to heat up, relative to ice-free ocean areas. Figure 9.19 shows a simulated look at the Arctic from space in summer after melting has begun. The snow-covered sea ice is bright, reflecting roughly 85% of the sunlight incident on the surface (Perovich et al., 2002a), while the open ocean is dark, reflecting only 7% of the incident sunlight (Pegau and Paulson, 2001). As the ice edge retreats, the white, snow-covered sea ice is replaced by black, open water; greatly increasing the input of solar heat to the ice-ocean system. The summer increase in the area of open water, together with the decrease in ice albedo caused by melting, combine to form the ice-albedo feedback (Perovich et al., 2002a).
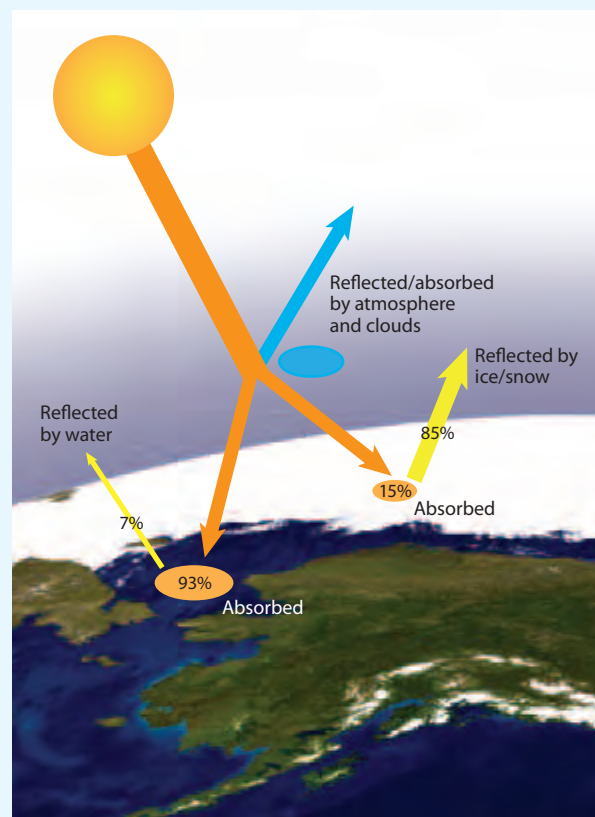
Figure 9.19. A simulated view from space of Arctic sea ice in August. The view shows the north coast of Alaska, an area of open water where the ice has melted, and the main ice pack.

Figure 9.20 shows representative albedos for different surfaces in the ice pack (Grenfell and Perovich, 1984, 2008; Pegau and Paulson, 2001; Perovich et al., 2002) as well as the resulting amounts of solar energy absorbed by the surfaces on a typical cloudy day in early July (Intrieri et al., 2002) and the estimated thickness of sea ice that the solar energy absorbed in one day could melt.

Daily-mean solar flux absorbed
by surface, W/m²

Ice melted by total solar energy
absorbed in a day, cm
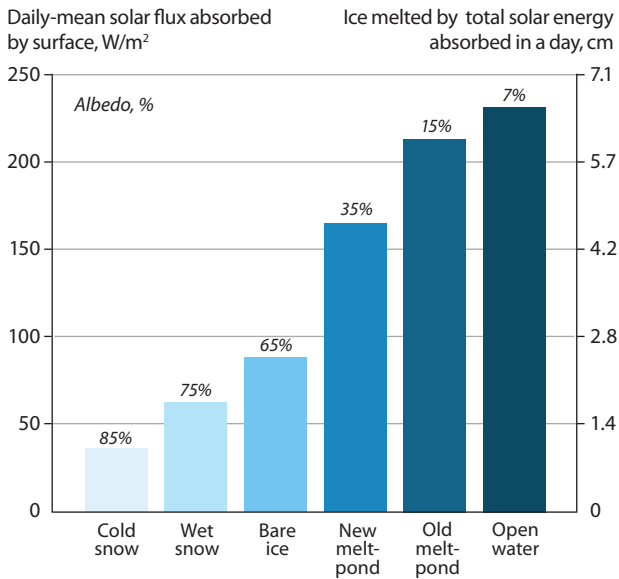
Figure 9.20. The albedo of different types of sea-ice surface (numbers above bars) and the resulting daily absorbed shortwave flux for a typical cloudy day in early July. The right-hand axis converts the total solar energy absorbed in one day to the thickness of sea ice that this energy could melt. Source: based on Grenfell and Perovich (1984), Pegau and Paulson (2001) and Perovich et al. (2002a).

The large differences among albedos of different sea-ice surfaces also result in an unequal distribution of the melt progression. Once formed, a melt pond will absorb much more energy than the bare ice around it, leading to enhanced melt in and around the pond, growth of the pond, and a further reduction in albedo (Perovich et al., 2002a; Eicken et al., 2004). This aspect of the ice-albedo feedback as well as the lateral melt of ice floes increases its effectiveness by concentrating melt in areas where it will cause the greatest change in albedo.

The albedo feedback due to ice and snow cover has been examined in climate models at least since the work of Budyko (1969) and Sellers (1969). Coupled GCMs have been used to show that the albedo feedback is a major contributor to the projected amplification of future warming in the Arctic, and that it also contributes to warming globally (e.g., Hall, 2004). Quantifying the effect accurately with GCMs is difficult because of significant uncertainties in sea-ice parameterizations in the models (Curry et al., 2001) and because cloud cover must also be accurately modeled. As discussed in Section 9.2.8, there are countervailing factors, such as cloud cover (Covey et al., 1991), which can temper the influence of the ice-albedo feedback.

A pair of recent studies tried to quantify the strength of the ice-albedo feedback by fixing the seasonal cycle of the surface albedo in the climate models and comparing the results to model runs with variable albedo under a doubling of atmospheric carbon dioxide (CO_2) levels. Bitz (2008) found that the impact of the ice-albedo feedback factor on ice thickness was small in the NCAR Community Climate Model (CCSM3) because of the countervailing thin ice growth feedback. This author estimated the feedback factor f by holding the seasonal cycle of the surface albedo of sea ice and ocean fixed while doubling CO_2. Ice-albedo feedback causes sea ice to thin about 26% more compared to a model run without ice-albedo feedback. A reduction of 26% corresponds to a feedback factor of only $f = 0.21 \pm 0.02$, where the error here is an estimate of uncertainty in the model. As might be expected, the feedback was strongest

in the seasonal ice zones. Also using CCSM3 but fixing the albedo of all surfaces, not just the ocean, Graverson and Wang (2009) found that the surface albedo feedback amplifies the surface-temperature response in the Arctic area by about 33%, whereas the corresponding value for the global-mean surface temperature is about 15%. Even though the surface-albedo feedback is an important process underlying excessive warming at high latitudes, the Arctic amplification is only 15% greater in the variable than in the locked-albedo experiments. These authors found that an increase in water vapor and total cloud cover lead to a greenhouse effect that is greater in the Arctic than at lower latitudes and may explain a part of the Arctic surface air-temperature amplification.

Perovich et al. (2007) investigated how changes in the amount of open water in the ice cover affect the total solar heat input to the ocean. They took a synthetic approach, combining field observations of ocean albedo, satellite-derived ice concentrations, and incident irradiances determined from meteorological models to calculate the amount of solar heat input directly to the upper ocean for every day from 1979 through 2005.

Trends in annual solar heat input directly to the ocean for each grid cell from 1979 to 2005 are mapped in Figure 9.21. Positive trends are pervasive over much of the Arctic with values as large as 4% per year. A smaller region with a negative trend in solar heat input is evident along the northern edge of the Canadian Archipelago. In this region, ice motion from the eastern Arctic has resulted in increases in ice concentration. Overall, 90% of the area shows a positive trend of increasing solar heating while 10% shows a negative trend. Trends were typically modest in magnitude, with the median and mean trend equal to an increase of 0.6 and 0.8 % per year respectively. The mean and median trends appear small; a few centimeters per year of ice thinning and < 1 W/m² of additional heat flux per year. The median solar heat input increased by 17% from 1979 to 2005, while the region of greatest increase showed a more than doubling of the solar heat input.

The enhanced solar heating has had an observable impact on the ice cover. There was a record minimum sea-ice extent in September 2007, with a large ice retreat in the Beaufort Sea. *In situ* ice mass balance measurements in this region showed anomalously large amounts of bottom melting (Perovich et al., 2008). Ice mass balance measurements made from an autonomous buoy showed 2.10 m of bottom melt in 2007 in the Beaufort Sea. This was more than six times the annual average of 0.34 m for the 1990s and two and a half times the observed bottom melt in 2006. This observation indicates that bottom melting was a major contributor to the 2007 ice loss in the Beaufort Sea.

The extreme amount of bottom melting observed in the Beaufort Sea in 2007 required considerable heat from the upper ocean. Earlier work has established the importance of solar heating of open water on bottom melting of the ice (Maykut and McPhee, 1995; Perovich, 2005). The solar radiation deposited in areas of open water was a major contributor to the large amount of ocean heat in 2007. Perovich et al. (2008) showed an ocean solar heat anomaly in the Beaufort Sea for 2007 that was 400–500% higher than the average for 1979–2005 and that this was a direct consequence of a large increase in the area of open water. The solar heat input through open water was sufficient in magnitude and timing to generate the observed bottom melting, illustrating the impact of an ice-albedo feedback on bottom melt.
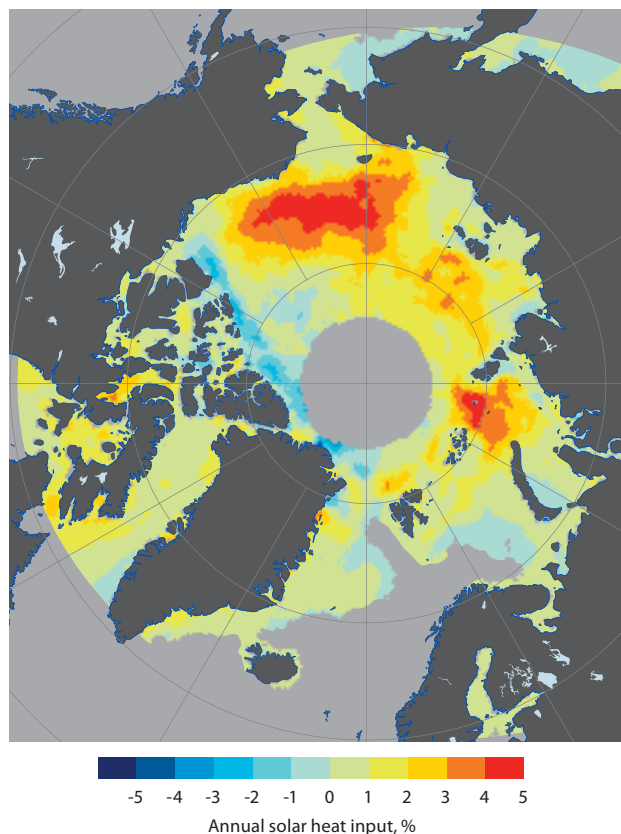
Figure 9.21. Linear trend of modeled annual total solar heat input directly to the ocean from 1979 to 2005. Source: Perovich et al. (2007).

9.2.3. Impacts of black carbon on the sea-ice albedo

There has been growing concern in recent years over the impact of black carbon on the albedo of sea ice, particularly when the ice is snow covered. Black carbon (BC or soot) is composed of small particles emitted by fossil fuel and biomass combustion. The sources are both natural (forest fires) and from human activities (industry, transport, household). The particles, which are black, absorb sunlight both in the atmosphere and where deposited on the surface, directly affecting the radiative energy balance. In addition, the absorbing BC particles may affect clouds and act on the climate through semi-direct and indirect effects (Hansen et al., 1997; Ackerman et al., 2000). Most of the particles stay close to the sources, mainly at lower latitudes, but some are transported to the Arctic. In the Arctic, some particles are deposited on the snow and ice, where even very low concentrations of BC will reduce the surface albedo.

9.2.3.1. Processes

The transport and deposition of polluted air masses from the lower latitudes to the Arctic varies seasonally and interannually (Shindell et al., 2008). The transport processes are not fully understood and different models give somewhat different results (Koch and Hansen, 2005; Stohl, 2006; Shindell et al., 2008). The phenomenon known as 'Arctic haze' is believed to be the main transport mechanism for air pollutants to the Arctic. It occurs in late winter and springtime when the meteorological conditions allow intrusions of pollution plumes from lower latitudes. Near the source, BC tends to be hydrophobic, but with time the particles are transformed and become more

hydrophilic. This makes them prone to take up water and be removed by clouds (Ogren et al., 1983, Ramanathan et al., 2007), be incorporated into snowflakes, or be scavenged as snow falls through the air. Transformation processes and removal mechanisms influence the mixing state of the particles, which in turn influences their optical properties.

Snow- and ice-covered surfaces have a high albedo (Wiscombe and Warren, 1980; Grenfell et al., 1994) and the deposition of BC and other dark particles will reduce this. However, BC is estimated to be about 50 times as effective as soil dust, and 200 times as effective as volcanic ash in its ability to reduce albedo (Warren and Wiscombe, 1980; Warren, 1982, 1884). Model results indicate that nanogram per gram amounts of BC can significantly reduce snow albedo (Warren and Wiscombe, 1980). Concentrations of BC measured in Arctic snow in 1983/84 suggest a possible broadband albedo reduction of 0–4% depending on snow grain size (Warren and Wiscombe, 1985). Radiative forcing by BC in snow can have a significant climatic effect, because of its high 'efficacy' resulting from several positive feedbacks (Hansen and Nazarenko, 2004; Jacobson, 2004). Flanner et al. (2007) computed a large radiative forcing over snow-covered regions, particularly in spring.

9.2.3.2. State and trends

The main global sources of BC emissions are China, Pacific Asia, North America and Europe (Bond et al., 2007). Globally, BC emissions are decreasing and projections show a drop from 8.0 Tg in 1996 to 5.3–7.3 Tg by 2030, under different IPCC emissions scenarios (IPCC, 2007), and with different spatial distributions (Streets et al., 2004). Levels of light-absorbing BC have been measured in Arctic air (Sharma et al., 2006) at concentrations of 25 to 75 ng/m³ (Point Barrow, Alaska, 1989–2003), 20 to 40 ng/m³ (Alert, Nord-Canada, 1989–2003), and 15 to 90 ng/m³ (Ny-Ålesund, Svalbard, 2002–2008). Concentrations are highest in winter and lowest in summer. Measurements of the all-year mean BC levels from 1989 to 2003 indicated significant reductions at Alert and Barrow, while more recent data show indications of an increasing trend over the past few years (Sharma et al., 2006). Ny-Ålesund shows a similar recent trend reversal where a decadal reduction until 2006 shifted to a significant increase over following three years.

Observations show relatively low levels of elemental carbon (EC; a proxy for BC) in snow both at Svalbard and Barrow in 2007. Levels in snow depend on both the amount of precipitation and the BC levels in air, as precipitation efficiently cleans the air of hydrophilic carbon. The total deposition for the entire year is important for the albedo, particularly in spring when the snow is melting, as there are indications that the particles concentrate on the surface when the snow is melting. Current EC levels measured in Arctic snow are generally lower than published data from 1983/84 (Clarke and Noone, 1985). The lower values are not entirely due to reduced emissions as different analytical methods were used for the recent and earlier measurements (EC_a and BC_e, respectively; Bond and Bergstrom, 2006). Studies have shown that apparent elemental carbon (EC_a), although noted to be a proxy for BC, is of the order of a factor of two lower than equivalent (BC_a) (Watson et al., 2005). Results from Svalbard 2007 (Forsström et al., 2009) used in the Warren and Wiscombe (1985) model showed that 11% of new snow samples and 63% of old snow samples had

enough BC to reduce the spectral albedo by 1% in new snow at the most sensitive wavelength.

9.2.3.3. Future projections

Changes in BC deposition may exacerbate or moderate future changes in Arctic sea-ice cover. Although the global trend indicates BC reductions, current understanding of deposition processes is not sufficient to reproduce observations using numerical transport models (International Workshop on Black Carbon in Snow, Tromsø 13-14 August 2009).

Transport of BC to the Arctic, rather than emission rates, is currently the factor limiting BC levels in the Arctic. As the temperature difference between the Arctic and lower latitudes weakens due to Arctic amplification of warming, changes in the typical transport patterns can be expected and thus possible shifts in the relative contribution of BC to the Arctic. In addition, the possible increase in Arctic shipping associated with a reduced sea-ice cover would result in more emissions released directly within the Arctic. This has critical implications for climate forcing (Law and Stohl, 2007) as studies show that shipping is already affecting the Arctic atmosphere in summer (Xie et al., 2006; Lack et al., 2008).

Most of the ongoing and predicted rapid changes in the Arctic are a direct consequence of increased levels of greenhouse gases with long atmospheric lifetimes. The only viable way to reduce their impact is to reduce their atmospheric concentration. This is a long-term commitment. Reducing short-lived pollutants, such as BC, holds promise for achieving an impact on a shorter time scale, thereby buffering some of the detrimental effects of long-lived greenhouse gases. In particular, the reductions can slow warming of the atmosphere and melting of the snow and ice (Law and Stohl, 2007).

9.2.4. Importance and role of local-scale sea-ice features

The sea-ice cover is not a flat slab of uniform thickness, but a complex mosaic of ice of various thicknesses, ages, and characteristics. In winter, the sea-ice floes are frozen together to form a flawed, but continuous ice cover. During summer, the ice cover breaks up to form a granular ensemble of individual floes. Meltwater collects on the surface forming ponds. Spatial heterogeneity is found at all length scales in sea ice, but is particularly apparent on lateral length scales from tens of meters upward.

9.2.4.1. Thermodynamic processes

Differential rates of freezing and melting result in the creation of horizontal heterogeneity on length scales of the order of meters or more. One of the most dramatic processes, and that which most clearly impacts sea-ice thickness and extent, is the formation of melt ponds in summer. Melt ponds form as meltwater accumulates on the surface of sea ice from snow and sea-ice melt. The area covered by ponds depends on surface topography, melt rates, and ice permeability. Melt ponds affect the heat and mass balance of the ice cover in various ways (Taylor and Feltham, 2004), for example, the latent heat released in freezing of melt ponds keeps the ice relatively warm and reduces winter ice growth. The most important effect that melt ponds have on the heat and mass balance of the ice cover is through their impact on albedo. Pond-covered ice has a lower albedo, absorbs more solar radiation, and therefore melts more quickly, than bare sea ice. Melt rates for ponds are estimated to be up to two to three times greater than for bare ice (Fetterer and Untersteiner, 1998; Perovich et al., 2002a). As shown in Figure 9.20, the albedo of pond-covered ice is much less than bare ice or snow-covered ice (e.g., Grenfell and Maykut, 1977; Perovich et al., 2002a; Eicken et al., 2004), and is principally determined by the optical properties and thickness of the ice beneath the pond.

The fractional area of the sea-ice surface covered in ponds is needed to determine the area-averaged albedo of the ice cover, an important quantity in the sea-ice components of GCMs. The pond fractional area has been most extensively studied from visual photography using aircraft and balloons (Derksen et al., 1997; Perovich and Tucker, 1997; Tschudi et al., 2001, 2008; Perovich et al., 2002b; Eicken et al., 2004), but also from satellite imagery (Fetterer and Untersteiner, 1998; Barber and Yackel, 1999). These studies show that the fractional pond coverage has been highly variable, with values ranging from 5% to 80% depending on such factors as ice type (FYI or MYI), time elapsed since the beginning of the melt season, surface roughness, and snow cover.

Current understanding of melt pond evolution comes mostly from field experiments, with the Surface Heat Budget of the Arctic (SHEBA) year-long camp from autumn 1997 to autumn 1998 (e.g., Perovich et al., 1999; Eicken et al., 2002,2004) providing some of the most detailed information to date. Various attempts have also been made to model melt ponds. Taylor and Feltham (2004) developed a 1-D model of melt ponds that consisted of heat and salt balance equations, and a two-stream radiation model. A subsequent cellular automata model was developed to study the area evolution of melt ponds on sea ice (Lüthje et al., 2006). This model divides a sea-ice floe into square cells, inside which the pond height and ice depth are recorded and allowed to evolve according to the state of the surrounding cells. While the two models have led to improvements in understanding of the processes governing melt pond evolution, their detailed complexity makes them unsuitable for use in GCMs. Flocco and Feltham (2007) and Pedersen et al. (2009) have both developed melt pond models for use in the sea-ice component of a GCM by building on the existing theory of the evolution of sea-ice thickness distribution, taking into account vertical drainage of meltwater, and other factors.

Field observations and modeling studies both indicate that the spatial distribution, and extent, of melt ponds is strongly influenced by ice topography, and thus the age of the ice. Since FYI is relatively flat, the ponds tend to be of larger area and shallower, compared to ponds on MYI (e.g., Eicken et al., 2004). In recent years, the areal fraction of FYI has been increasing relative to MYI (Cavalieri et al., 2003). For a given volume of melt water, shallow ponds covering a large area result in greater melting than deeper ponds covering less area (e.g., Eicken et al., 2004; Lüthje et al., 2006). Since ponds on FYI cover a larger area than those on MYI, the increase in the relative abundance of FYI indicates that the significance of melt ponds to the heat and mass budget is likely to increase.

9.2.4.2. **Dynamic processes**

The forces acting on the ice cover, predominantly due to winds and ocean currents, can cause ice convergence with thinner ice buckling and over-riding to form pressure ridges or rafted ice. Pressure ridges can extend horizontally over several kilometers, and in cross-section consist of a triangular sail and keel, with the keel typically five to six times deeper than the sail is tall, with total thickness typically between 5 and 20 m. Since thin ice will raft and ridge more easily than thicker ice, ridging can act as a kind of negative feedback process in regions of convergence whereby thinner ice is transformed into thicker, ridged ice and more open water is exposed in adjacent regions of divergence. This open water can be a region of intense ice growth in winter or increased solar absorption in summer.

Sea ice is relatively weak in tension and divergence of the ice cover can result in the formation of cracks, which can join up to form leads. Winter leads are typically long, linear features covering only 1–10% of the area. They are of significant importance since they expose the ocean directly to the atmosphere and are regions of rapid ice growth, large heat loss to the atmosphere, and considerable salt rejection to the ocean.

Sea-ice extent is affected by processes occurring in the marginal ice zone. The seasonal retreat of sea ice at the ice edge is caused by a combination of advection, mechanical floe break-up, and melting at the floe edges and upper and lower interfaces (Squire, 2007). As floes break up, the total perimeter, or edge length, of a given area of sea ice increases. The reduction in floe size affects the response of the ice cover to the incoming wave field, promoting further jostling between the floes, and more break-up. This in turn promotes further lateral melting creating a positive feedback between dynamic and thermodynamic processes.

9.2.5. **The cloud-radiation feedback and its impact on the surface-heat budget**

Clouds play a dominant role in determining shortwave and longwave radiative transfer in the atmosphere. Cloud area, height, thickness and water content, among other properties, all influence radiative fluxes (Curry and Ebert, 1990). For sea-ice modeling, having an adequate description of cloud cover is especially important for estimating radiative fluxes at the surface because surface melting and ice growth are sensitive to shortwave and longwave radiation (Maykut and Untersteiner, 1971; Curry and Ebert, 1990). As a result, there are strong couplings between the sea ice, surface albedo, and clouds that have been termed the ice-albedo (Section 9.2.2) and cloud-radiation feedbacks (Barry, 1984; Makshtas, 1991; Curry et al., 1995; Makshtas et al., 2003, 2007).

Over the course of a year, the net effect of clouds over the central Arctic Ocean is to warm the surface (Curry and Ebert, 1992; Schweiger and Key, 1992; Intrieri et al., 2002). Only in the middle of summer does the cloud cooling effect overwhelm the warming effect for a period of time that is determined largely by cloud type, sun angle, and surface albedo. Observations over the past decade have revealed that liquid water clouds impart the greatest radiative influence on the Arctic surface radiation budget. In general, low-level stratiform liquid and mixed-phase clouds exert the most significant cloud effect on the surface radiation budget (Sedlar and Tjernstrom, 2009).

Several studies have attempted to evaluate the influence of ice crystal precipitation, also called 'diamond dust' (Overland and Guest, 1991; Curry and Ebert, 1992), Arctic haze (Blanchet, 1989), and aerosols (Leck et al., 2002; Prenni et al., 2006; Leck and Bigg, 2008) on the radiation and thermal state of the Arctic atmosphere and surface. For example, Curry and Ebert (1992), building on the cloud statistics of Huschke (1969), found surface longwave fluxes to be very sensitive to the presence of lower-tropospheric ice crystal precipitation. Overland and Guest (1991) compared model results with experimental data and found a discrepancy. They suggested that the missing modeled downward longwave radiation might be explained by ice crystal precipitation. They did not, however, rule out other explanations such as blowing snow or optically-thin clouds.

Despite the intensive development of methods for remotely sensing the atmosphere from space, uncertainties can be large for certain conditions, so *in situ* visual observations continue to play a critical role in the assessment of Arctic cloudiness. Many studies have used the Russian 'North Pole' (NP) drifting station cloudiness data, with about sixty years of observations, to verify their findings (Schweiger and Key, 1992; Lindsay; 1998; Makshtas et al., 1999; Lindsay and Makshtas, 2003). One of the most interesting findings from analysis of NP cloudiness data is the distribution of total cloud amount in winter and summer (Makshtas et al., 1999) (Figure 9.22), having a well-pronounced 'U' shape with fewer clouds in winter and a very high frequency of overcast conditions in summer. Both circumstances illustrate
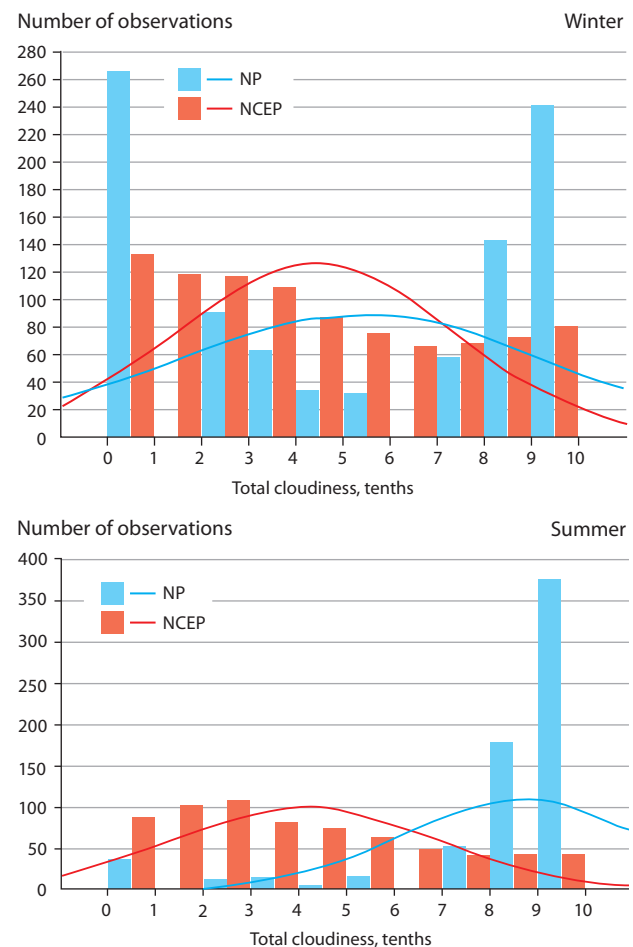
Figure 9.22. Frequency of total cloud amount in winter and summer observed at the Russian drifting station 'North Pole' for 1954 to 2005 in comparison to NCEP-NCAR Reanalysis data. Source: Makshtas et al. (1999).
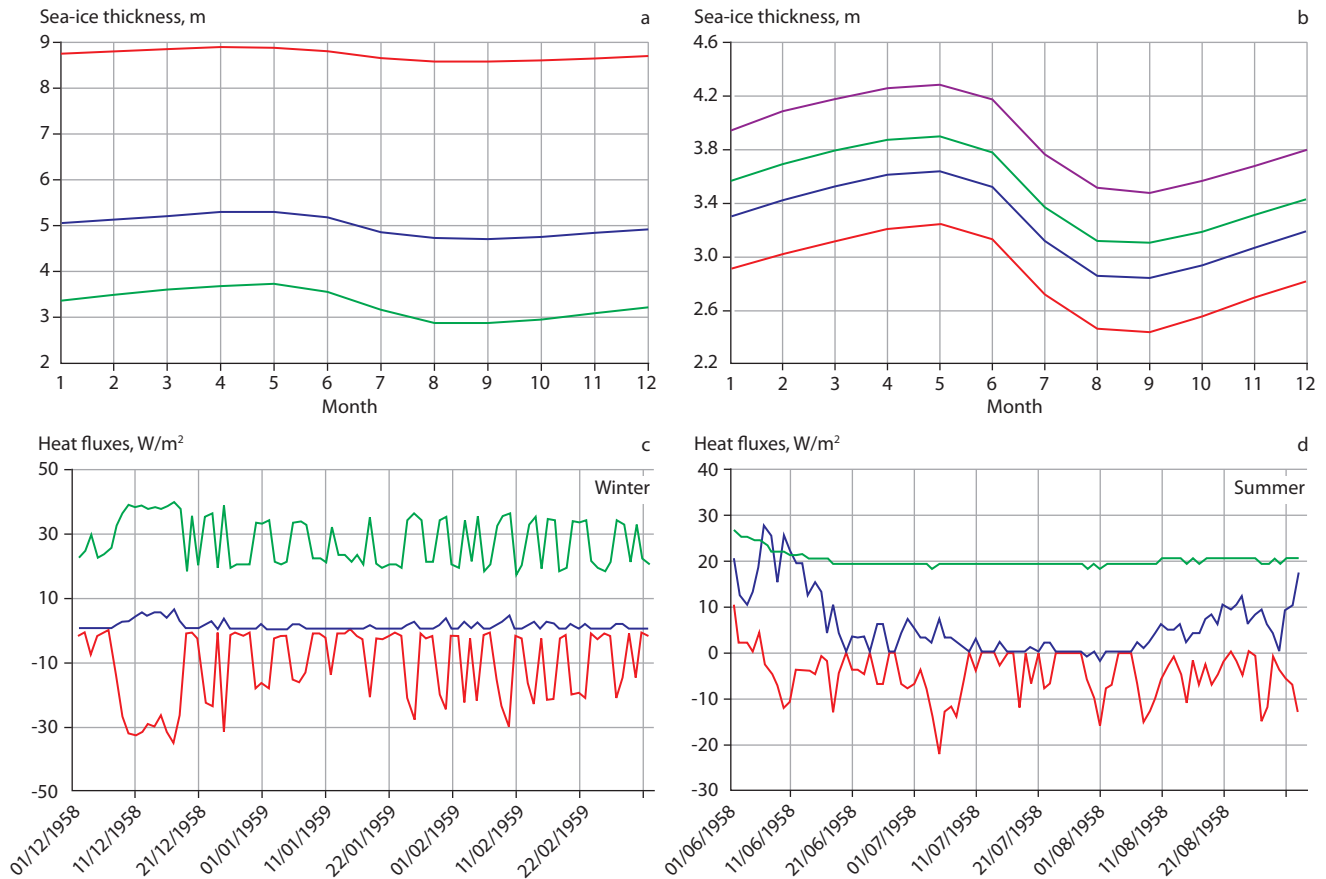
Figure 9.23. (a) Seasonal variability in equilibrium sea-ice thickness near the North Pole, calculated with the thermodynamic sea-ice model for surface-level air temperature Ta (green) and also when air temperature is restricted to Ta = 0 °C if Ta > 0 °C (blue) and Ta = -0.5 °C if Ta > -0.5 °C (red). (b) The same as for 'a' except that during winter the surface-level air temperature differs from the data by 10 °C (red), 2.5 °C (blue), -2.5 °C (green) and -10 °C (purple). (c) Values of the turbulent sensible (red) and latent (blue) heat fluxes and the longwave radiation balance (green) during winter 1958/59 near the North Pole calculated with the thermodynamic model. (d) The same as for 'c' except for the summer of 1958.

the variability of cloudiness with predominately clear (0 to 2 tenths) and overcast (8 to 10 tenths) skies. Analyses by Ikeda et al. (2003) and Makshtas et al. (1999) showed a strong negative trend of clear sky amount in winter and a modest negative trend of overcast conditions in summer. These trends would imply a reduction in sea-ice thickness due to a decrease in longwave radiative cooling in winter and an increase in incoming shortwave radiation in summer. Some statistical support of this supposition was given by Ikeda et al. (2003).

Makshtas et al. (2003, 2007) applied a zero-dimensional thermodynamic sea-ice model, similar to Semtner's (1976), to investigate the mechanisms responsible for the sensitivity of sea ice to external parameters. In the model, the energy fluxes between the atmospheric surface layer and the sea-ice surface had been calculated following Jordan et al. (1999) and Radionov et al. (1996). The model was integrated for 50 years at the North Pole with the same atmospheric forcing and an oceanic heat flux of 1 W/m². The results of the standard run (Figure 9.23) show good agreement with the results of Maykut and Untersteiner (1971) for equilibrium sea-ice thickness and Lindsay (1998) for the mean values of energy fluxes during winter and summer. Numerical experiments show that the equilibrium sea-ice thickness is much more sensitive to changes in summer air temperature than winter air temperature. Figure 9.23a demonstrates that restricting the summer air temperature to 0 °C or less and 0.5 °C or less leads to 1.5-fold and 3-fold increases in the equilibrium sea-ice thickness, respectively. In

contrast, changing the surface air temperature in winter by +10 °C or 10 °C changes the equilibrium sea-ice thickness by only 30% (Figure 9.23b). Figures 9.23c and 9.23d partly explain these results. In winter there is a strong negative feedback between the turbulent sensible heat flux (H) and the longwave radiation balance (R) (e.g., Makshtas et al., 1999). As a result, the conductive heat flux in the sea ice is not very sensitive to changes in air temperature; and, in turn, neither is the growth of ice on the bottom of the ice cover.

In winter, the negative feedback between the sensible heat flux and the net radiation together with insulating properties of snow stipulate the weak dependence of the conductive heat flux through sea ice on air temperature changes and, in turn, ice growth. In summer, on the other hand, the surface temperature of the ice is at or near the freezing point, and the negative feedback between the turbulent and longwave radiative heat fluxes cannot operate. As a result, any increase in the downward sensible heat flux caused by a small increase in the surface-layer air temperature and an increase in absorbed shortwave radiation due to a decrease in mean surface albedo (positive feedback) leads directly to increased melting of the ice cover.

Satellite studies show that winter cloud amount in the central Arctic appears to have been generally decreasing since the early 1980s, corresponding to a decrease in surface temperature (Wang and Key, 2003, 2005; Liu et al., 2008). The same studies show that spring cloud amount in the Arctic has been increasing. While this is somewhat in contrast to the

studies described by Ikeda et al. (2003) and Makshtas et al. (1999) for winter trends based on NP drifting station data, the difference is likely to be one of the spatial domain and period of study. A benefit of using satellite data to diagnose feedback mechanisms is the uncertainty in model simulations. For example, Gorodetskaya et al. (2008) used a variety of GCMs to demonstrate that the discrepancies in the modeled surface net shortwave flux resulting from the large variability in model representations of cloud properties partly explain the range in the model's sea-ice area changes.

The relationship between winds, radiative forcing, heat advection and the dramatic decrease in sea-ice cover over the past two and a half decades has been examined with satellite data. Francis et al. (2005) found that there are distinct regional differences in the relative roles of these parameters in explaining the variability of the ice edge position. The downwelling longwave flux anomalies explained about 40% of the variability, with northward wind anomalies important north of Siberia. Anomalies in solar insolation were negatively correlated with ice retreat in all regions, so the solar flux anomalies are overwhelmed by the longwave influence. Figure 9.24 illustrates the importance of the various parameters for different regions of the Arctic.

A number of recent studies have addressed the relationship between changes in cloud cover and sea ice. They are particularly relevant in light of the record minimum sea-ice extent in 2007. On the time scale of a single season, changes in cloud amount may have minimal influence on summer sea-ice melt. Schweiger et al. (2008a) used an ice-ocean model to examine the relationship between unusually clear skies during summer 2007 and the record sea-ice extent minimum. They found that the negative cloud anomaly and increased downwelling shortwave flux from June through August did not contribute substantially to the record sea-ice extent minimum in September. In contrast, Kay et al. (2008) calculated that their satellite-derived 16% decrease in summer cloud cover in 2007 led to an increase in the downwelling shortwave flux sufficient to melt 0.3 m of ice or warm the surface ocean by 2.4 K.

The influence of trends in sea-ice concentration and cloud cover on surface temperature over the Arctic Ocean from 1982 to 2004 was investigated analytically, and evaluated empirically with satellite products by Liu et al. (2009). They found that changes in sea-ice concentration and cloud cover played major roles in the magnitude of recent Arctic surface temperature trends. Significant surface warming associated with sea-ice loss, over 0.9 K/decade, was found over the Chukchi and
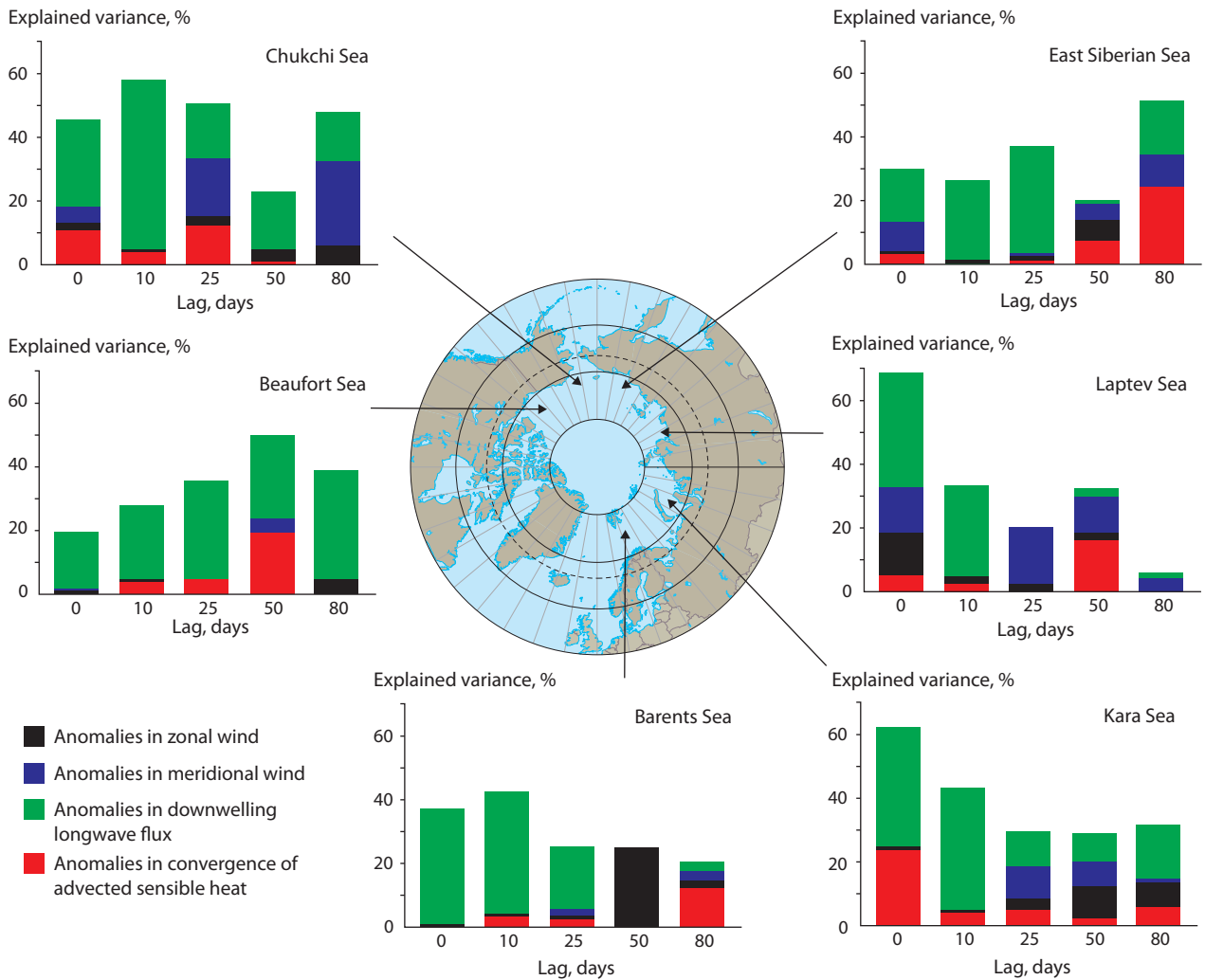
Figure 9.24. Percentages of variance in anomalies of sea-ice maximum retreat explained by anomalies in zonal wind, meridional wind, downwelling longwave flux, and the convergence of advected sensible heat in each peripheral sea of the Arctic Ocean. The bars represent explained variance at lags of 0, 10, 25, 50, and 80 days, where the ice edge anomaly lags the forcing anomaly. Source: Francis et al. (2005).

Beaufort Seas in autumn, accounting for most of the observed 1.1 K/decade warming trend. In winter, surface temperature trends associated with changes in cloud cover were found to be negative over most of the Arctic Ocean, with cloud cover trends explaining -0.91 out of -1.2 K/decade of the surface temperature cooling. In spring, 0.55 K/decade of the total 1.0 K/decade warming can be attributed to the trend associated with cloud cover changes. Figure 9.25 shows the relationship between trends in surface temperature, sea-ice concentration, and cloud cover.

While the ice-cloud relationship of primary interest in ice studies is the radiative impact of cloud cover on the surface energy budget, the influence of changes in ice cover on cloud properties is an important part of the feedback process (see also Box 9.6). Satellite atmospheric sounder data and the ECMWF Re-Analysis (ERA-40) have both shown that sea-ice retreat is linked to a decrease in low-level cloud amount and an increase in mid-level clouds (Schweiger et al., 2008b). The changes in cloud cover are a result of changes in the atmospheric temperature structure and stability. However, the radiative effect of these changes is relatively small, as the cloud radiative forcing changes are compensated for by changes in the lower tropospheric temperature and humidity structure. Kay and Gettleman (2009) reached a similar conclusion about the relationship of cloud cover changes to modifications in near-surface static stability as a result of decreasing ice extent,
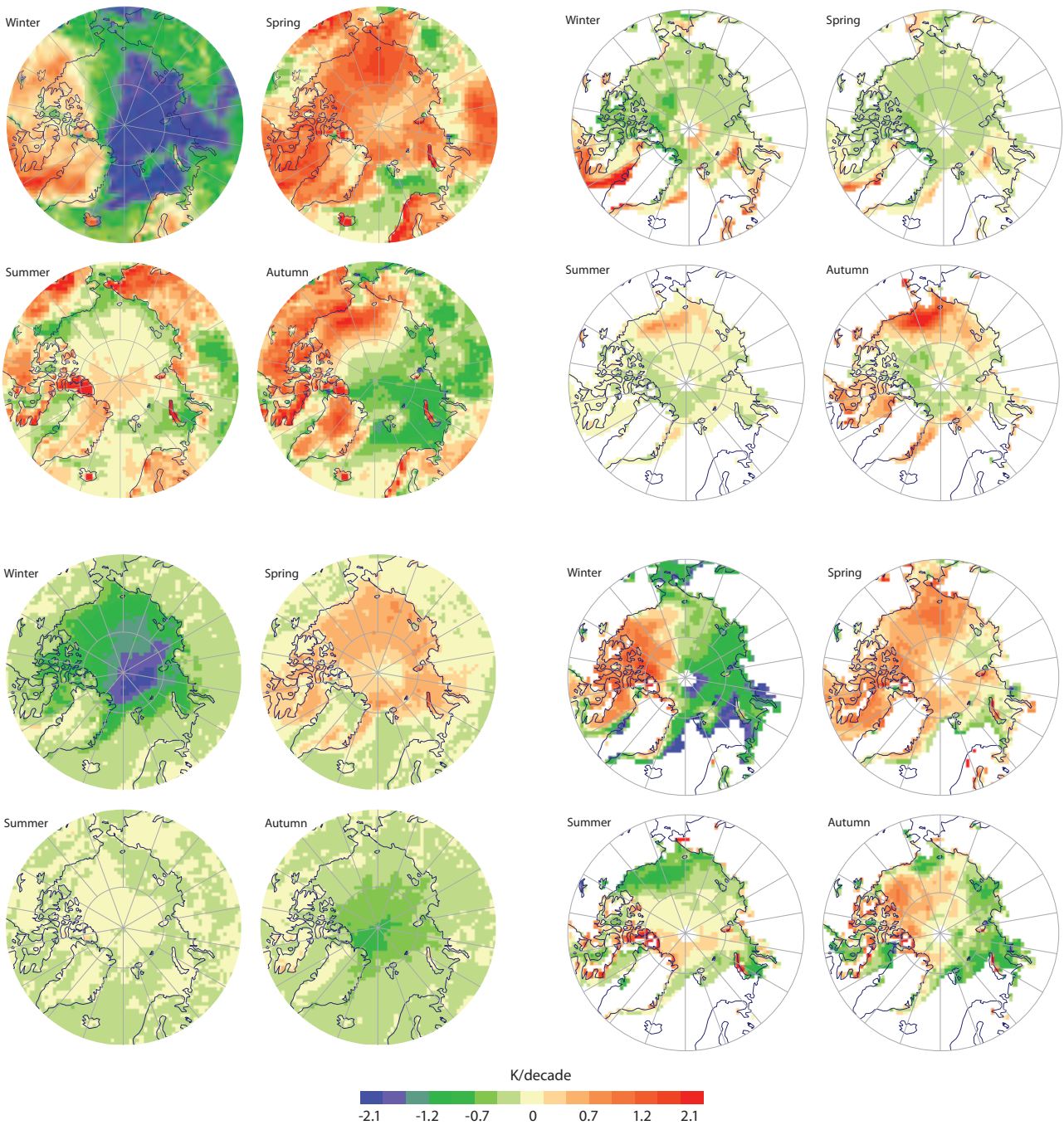
Figure 9.25. The seasonal all-sky surface temperature trend (top left), the temperature trend caused by sea ice concentration changes (top right), the temperature trend caused by cloud cover changes (bottom left), and the residual trends (bottom right) from 1982 to 2004. The residual trend is the difference between the all-sky surface temperature trend (1982–2004) and the sum of the two other trends. Source: Liu et al. (2009).

Box 9.6. Scientific challenges for the cloud-radiation feedback

Observations from satellites and surface. Continuing observations of clouds from the surface and improvement in satellite cloud-property retrievals, such as cloud amount, particle size, and water content, are crucial to improving understanding of cloud statistics and surface radiative impact.

Improvements in cloud modeling. The cloud-radiation feedback remains highly uncertain due to the spread in climate model simulations at high latitudes. Accurate treatment of clouds in models and improvements in cloud model parameterizations are needed to reduce uncertainties in future Arctic and global climate change projections.

Feedbacks and processes taking place on the local scale. Understanding the sensitivity of cloud processes to regional changes at high latitudes (e.g., aerosol, sea-ice extent, snow cover) is required to accurately project climate changes in the Arctic region and their global impacts.

particularly during autumn. It was found that owing to their seasonal timing, the response of cloud cover to sea-ice loss plays a minor role in regulating the summer ice-albedo feedback, but a larger role in the cloud-ice feedback in autumn.

9.2.6. Feedbacks and forcings related to changes in atmospheric transport

9.2.6.1. Wind forcing and ice loss

Atmosphere pressure fields and winds drive ice motion and so can have a large impact on the overall state of the ice cover. To illustrate this, consider the example of 2007 when the Arctic sea-ice extent reached a record minimum (Box 9.7). While this record low of total ice extent was set in September 2007 (Maslanik et al., 2007; Comiso et al., 2008; Drobot et al., 2008; Perovich et al., 2008), the distribution and extent of different sea-ice type and thickness classes in spring (March to May) before summer melt, hold critical information for understanding the many varied and complex causes of the record reduction (Lindsay et al., 2009). Arctic sea ice consists of two major classes: perennial or MYI defined as sea ice that survives at least one summer (Armstrong et al., 1973), and seasonal or FYI that grows in the current freezing season and is typically thinner than perennial ice (Haas et al., 2008). The old and thick MYI is important to the stability of the Arctic sea-ice cover, as it is less affected by interannual variations in ice growth and melt. The SeaWinds scatterometer aboard the QuikSCAT satellite has been proven particularly useful in delineating MYI and FYI (Nghiem, 2003; Nghiem et al., 2005, 2006, 2007; Nghiem and Neumann, 2007).

Atmospheric forcing continued to push the ice cover from the Russian Arctic toward Greenland and to enhance the ice transport out of the Arctic via both Fram Strait and Nares Strait in winter and spring 2008. Ice export was also facilitated by the greater sensitivity of thin ice to wind forcing (Gascard et al., 2008; Rampal et al., 2009). By 1 March 2008, the extent of MYI was reduced by one million km^2 compared to that

Box 9.7. The Polar Express

Wind forcing plays a significant role in changes in the sea-ice cover. In summer 2007, wind forcing was driven by a dipole anomaly pattern of the surface level pressure with a coexistence of a pronounced atmospheric low pressure over the Barents Sea together with a strong high pressure over Canada Basin. Clockwise winds around the high pressure merged together with counterclockwise winds around the low pressure to set up the persistent wind anomaly along the Transpolar Drift Stream (TDS) that enhanced the ice transport out of the Arctic Basin via Fram Strait (Nghiem et al., 2006; Gascard et al., 2008). This phenomenon is referred to as the 'Polar Express' (Figure 9.26) (Nghiem et al., 2007). A strong Polar Express forces ice loss by four processes: (i) compressing sea ice from both sides of the TDS, (ii) pushing sea ice from the Russian Arctic toward Greenland while accelerating the TDS, (iii) transport of sea ice by the TDS out of the Arctic, and (iv) melting of the exported sea ice in the Greenland Sea by warm Atlantic waters originating from the south (Nghiem et al., 2006, 2007). The Polar Express was strong in 2007 and together with other factors contributed to the record low 2007 summer ice extent.
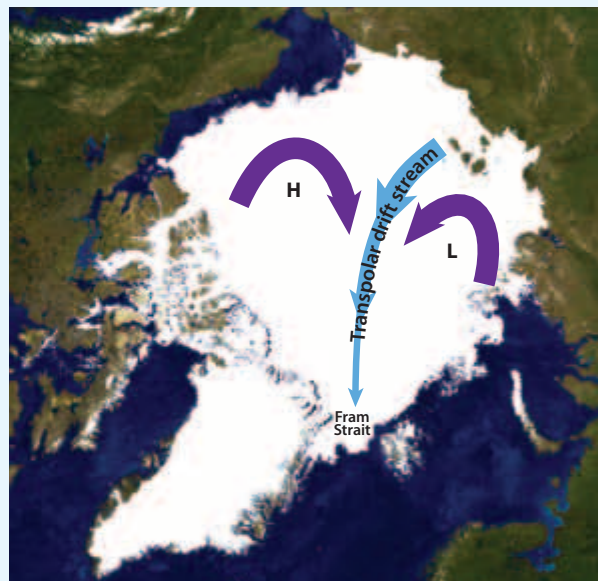
Figure 9.26. Schematic illustration of the Polar Express phenomenon.

at the same time in 2007 (Figure 9.27). QuikSCAT results, together with the buoy-based Drift-Age Model estimates in half a century (1957–2008), showed that MYI extent decreased in the 2000s at the rapid rate of 1.5 million km^2 per decade, which tripled the decreasing rate in the previous three decades (1970 to 2000) (Nghiem et al., 2008). Observations in 2009 suggested a split between two main perennial ice packs (Figure 9.27): the Transpolar Drift perennial ice pack and the north Greenland perennial ice pack, extending from the North Pole to Fram Strait.

A strong Polar Express event requires: (i) the coexistence of low and high surface level pressure (SLP) anomalies; (ii) the low and high pressure anomaly centers to locate at appropriate places such as the low above the Barents Sea and the high over the Canadian Basin; and (iii) the low and high not to be very far apart. When the locations of the SLP anomaly centers switch to
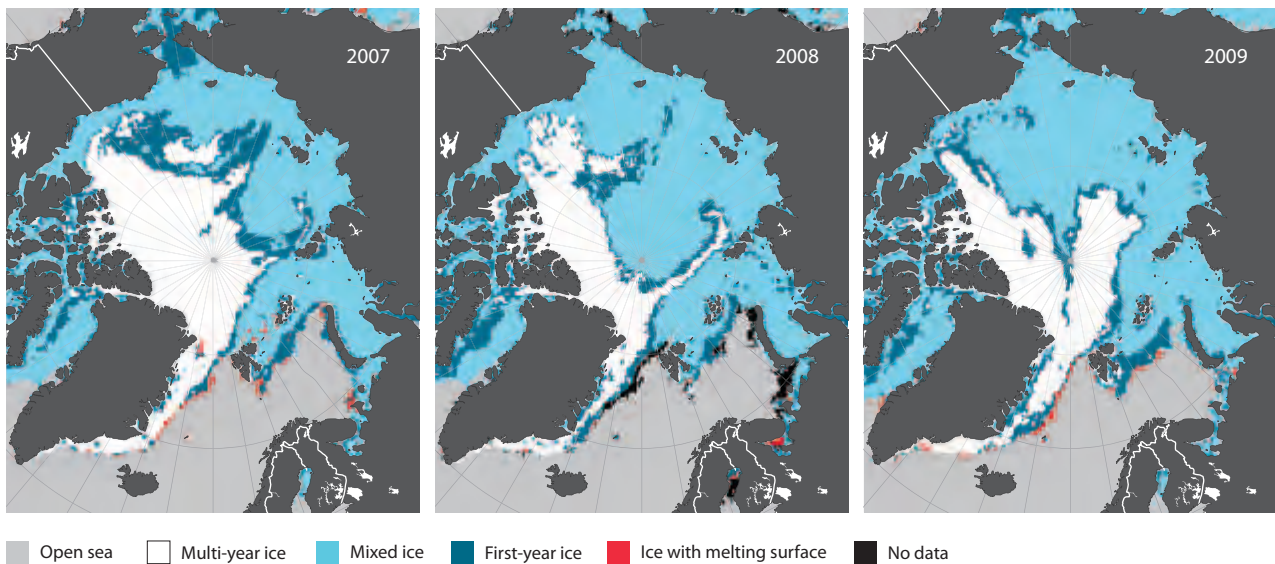
Figure 9.27. Comparison of Arctic sea-ice age distribution on 1 March 2007, 2008, and 2009.

different places, such as the low over the Russian Arctic Ocean and the high above the north of the Greenland Sea, the dipole pattern may reduce ice transport out of the Arctic, slowing the sea-ice decline.

Previous studies have illustrated connections between sea ice, SLP, and stratospheric forcing on weekly timescales in the Beaufort Sea region from 1979 to 2000 (Lukovich and Barber, 2006). In particular, it was shown that reversals in the Beaufort Gyre occur throughout the annual cycle, with variability in strength and timing of cyclonic circulation of sea ice between years having implications for ice convergence and divergence in the region. Reversals in the Beaufort Gyre were shown to coincide with reversals in SLP for the region of interest. In addition, maximum correspondence was observed between sea ice and stratospheric relative vorticity with a lag of the order of two to six weeks, while negative correlations were observed between surface relative vorticity and sea-ice motion, indicating that sea-ice motion is governed by surface and stratospheric atmospheric phenomena. Noteworthy is the anticyclonic circulation sea-ice regime in summer 2007, characteristic of the SLP high that dominated Canada Basin prior to the record minimum in ice extent in September 2007. Comparatively weak circulation is observed during late summer in 2008.

Recent studies also highlight a correspondence between stratospheric sudden warming (SSW) events and surface synoptic types in the Beaufort Sea region and demonstrate an increase in anticyclonic types during SSWs in winter that persists for up to 20 days (Lukovich et al., 2009). Furthermore, investigation of connections between surface synoptic types and ice motion in the Beaufort Sea region illustrates correspondence between surface cyclonic types and summer reversals in the Beaufort gyre.

Many models have underestimated Arctic sea-ice loss (Stroeve et al., 2007). Recently, Zhang et al. (2008b) and Lindsay et al. (2009) conducted a model study to show that preconditioning, anomalous wind forcing, and ice-albedo feedback were the main contributors responsible for sea-ice loss in 2007, and that Arctic sea ice has become vulnerable to the anomalous atmospheric forcing. Kauker et al. (2009) showed the importance of preconditioning using an adjoint analysis of a coupled sea ice-ocean model. Wang et al. (2009) examined the SLP fields and used a coupled ice-ocean model to investigate the small extents of sea ice including the record minima observed in 2005 and 2007. What causes the change in the SLP field; how such change can be related to global temperature change; and whether the impacts can be amplified or suppressed by positive or negative feedbacks remain to be investigated.

9.2.6.2. Sea-ice impacts on atmospheric circulation

While it can be argued that decadal shifts in atmospheric wind patterns over the past decade contributed to recent reductions in summer Arctic sea-ice extent (Overland et al., 2008a; Wang et al., 2009), the increase in late summer open-water area is, in turn, directly contributing to a modification of large-scale atmospheric circulation patterns. Once there is a reduction in sea-ice cover in late summer, additional heat is stored in the ocean which is released to the atmosphere during autumn. Beginning in 2002, extensive regions in the Arctic during late autumn have surface air temperature anomalies of more than 4 °C and temperature anomalies above 850 hPa of 1 °C (Serreze et al., 2009). The transfer of heat from the surface to the troposphere is more effective as the strength of the normal Arctic stable boundary layer is eroded from below (Schweiger et al., 2008b).

These higher temperatures reduce the density of the air and contribute to an increase in the 500 to 1000 hPa atmospheric thickness field in every recent year with reduced sea-ice cover. Horizontal gradients in the thickness field impact pressure fields and winds over the Beaufort Sea. Compositing wind fields for late autumn 2002 to 2008 helps to highlight this contribution; for the years with diminished sea-ice cover there were extensive anomalous tropospheric easterly winds of ~1.2 m/s, relative to climatological easterly winds near the surface and upper tropospheric westerlies of ~3 m/s. (Climate Data Center, www.cdc.noaa.gov/cgi-bin/data/composites/printpage.pl). Changes in the Arctic atmosphere also impact farther afield in northern mid-latitudes (Francis et al., 2009). Responses can be complex,

as loss of sea ice north of Eurasia may result in a cooling effect over eastern Asia (Honda et al., 2009). With continuing loss of summer sea ice to less than 20% of its climatological mean over the next decades (Wang and Overland, 2009), increased modification of atmospheric circulation patterns and impacts on the global heat budget are anticipated (Rinke et al., 2006).

9.2.6.3. Sea-ice impacts on carbon dioxide concentration in atmosphere and ocean

Many studies have investigated the variability in atmospheric concentrations of (CO_2) over high-latitude Arctic Ocean areas and the role of sea ice in CO_2 variability. Anderson et al. (2004) found that dissolved inorganic carbon (fCO_2) increases with depth below the surface mixed layer. They suggested that this excess is a result of enhanced air-sea exchange of CO_2 caused by sea-ice formation and the rapid transport of high salinity brine out of the surface layer. Semiletov et al. (2004) showed that CO_2 concentration increases from the coastal region toward the ice-covered sea. Alekseev and Nagurny (2005) related winter variation in CO_2 concentration over the Arctic Ocean to ice dynamics. Alekseev and Nagurny (2007) showed that polar amplification of the annual cycle of CO_2 concentration over the Arctic Ocean is connected with active influence of sea ice on formation of a seasonal CO_2 cycle. The formation and growth of ice during winter is accompanied by allocation of CO_2 into the sub-ice water layer and directly to the atmosphere. They speculated that the Arctic Basin is more likely to be a source than a sink of CO_2 over an annual cycle.

9.2.7. Projection: Impact of feedbacks on rate of sea-ice decay

It is difficult to quantify the impact of climate feedbacks on rate of sea-ice decay because of their complexity. Surface-albedo feedback, for instance, is closely linked to atmospheric and oceanic processes. The impact of surface-albedo feedback on changes in sea ice depends on the dynamic and thermodynamic air–ice–sea interactions and is strongly influenced by clouds. It is also related to the preconditioning of sea ice (Zhang et al., 2000; Lindsay and Zhang, 2005). Under certain atmosphere, ocean, and sea-ice conditions, surface-albedo feedback may substantially enhance sea-ice decay.

The impact of surface-albedo feedback may best be illustrated by the record retreat of Arctic sea ice in summer 2007. The dramatic sea-ice retreat occurred after years of shrinking and thinning (e.g., Rothrock et al., 1999, 2008; Meier et al., 2007), linked to an increase in surface air temperature and changes in wind circulation (ACIA, 2004; Serreze et al., 2007). Changes in wind circulation are often associated with the Arctic Oscillation. The AO was strongly positive from 1989 to 1995, which advected a considerable amount of MYI out of the Arctic into the North Atlantic (e.g., Hilmer and Jung, 2000; Zhang et al., 2000; Lindsay and Zhang, 2005; Tsukernik et al., 2010). Since then, a more meridional atmospheric circulation pattern (southerly wind anomalies from the Pacific sector of the Arctic Ocean) has been observed (Overland et al., 2008a).

As a result of increasing surface air temperature and changing atmospheric circulation, MYI, particularly the oldest and thickest ice within the multiyear ice pack, has been rapidly replaced in recent years by thinner FYI (Kwok, 2007; Maslanik et al., 2007; Nghiem et al., 2007) that is more sensitive to changes in atmospheric and oceanic forcing. The reduction in areal extent, thinning of sea ice, and the increased presence of young ice have significantly increased the surface absorption of solar radiation because of the positive ice-albedo feedback (Perovich et al., 2007, 2008).

During summer 2007, the meridional atmospheric circulation pattern was characterized by a strong southerly wind anomaly, which drove ice out of the Pacific sector of the Arctic Ocean and created a large area of thin ice and open water where surface albedo is lowered considerably. This surface-albedo feedback leads to enhanced surface solar heating, which results in increased melting at the ice surface and increased heat deposit in the ocean mixed layer (Steele et al., 2008; Zhang et al., 2008b). A warmer ocean mixed layer causes significant lateral and bottom melting (Perovich et al., 2008). About 70% of the Arctic sea-ice loss in summer 2007 is found to be due directly to the enhanced surface, lateral, and bottom melting, with 30% due to the wind-driven ice advection (Zhang et al., 2008b). The amplified ice melting suggests that under certain atmospheric and oceanic forcing the positive surface-albedo feedback may play a dominant role in sea-ice retreat in the Arctic.

Regional climate models (RCMs) also provide useful information on feedback processes in the Arctic. By prescribing forcing at the model boundaries, these studies can explicitly separate the influence of internal Arctic feedback processes and the forcing from lower latitudes. For example, Döscher et al. (2009) assessed the role of regional *vs* large-scale atmospheric circulation for interannual Arctic climate variability and found that recent sea-ice trends are largely associated with large-scale forcing outside the Arctic region. However, internally generated variability is not negligible and can be more important for some variables and locations. Regional climate modeling studies have also addressed biases in simulated fields through model intercomparison of specific Arctic climate features, such as the Arctic boundary layer structure (Tjernström et al., 2005). These types of study have provided insight into the major uncertainties associated with simulating future Arctic climate change.

9.2.8. Negative feedbacks

Changes in sea-ice cover result from the direct influence of forcing anomalies combined with feedbacks that act to amplify or damp the initial perturbation. As previously discussed, the Arctic is home to strong positive feedbacks making it susceptible to change. However, negative feedbacks must also be present given the relative stability that has presided in the Arctic system over long periods of the geological past (e.g., Polyak et al., 2010). Understanding the role of these negative feedbacks and their interaction with other aspects of system change is critical for developing accurate projections of the future state of the Arctic.

As has been recognized for some time, the conduction of heat through sea ice is dependent on ice thickness, causing thin ice to grow more rapidly than thick ice subject to the same atmospheric and ocean forcing. This relationship is nonlinear, having an inverse dependence on ice thickness. This basic fact of sea-ice thermodynamics gives rise to a fundamental stabilizing

feedback on ice thickness (L'Heveder and Houssais, 2001; Bitz and Roe, 2004). This reduces the net ice mass budget change in simulated future climate projections. Indeed a number of climate models simulate increased ice growth in future climate projections that partially compensates increases in ice melt (Holland et al., 2008b).

There are also indications that sea-ice dynamic processes may stabilize the Arctic sea-ice cover. In uncoupled sea-ice modeling studies, ice dynamics generally reduces the sensitivity of the ice cover to forcing perturbations (e.g., Hibler, 1984; Arbetter et al., 1999). This is related to the ice strength-ice thickness relationship, which causes thinner ice to more easily converge resulting in more ridging and mechanical thickening of the ice pack. This can also modify the ice velocity field and observational evidence does indicate an increase in ice speed associated with the thinning ice pack (Rampal et al., 2009). Coupled model simulations (Hewitt et al., 2001; Vavrus and Harrison, 2003) have found that ice dynamics reduces the climate sensitivity to increased CO_2 concentrations. However, the mechanisms responsible for this vary in different models.

Sea-ice growth, melt, and transport redistribute water and salt in the Arctic with potential climate feedbacks. Increased ice growth and brine rejection can enhance convection, resulting in deeper mixed layers and modified heat entrainment. This can influence the ice-ocean heat flux and provide a feedback to the sea-ice cover. Modeling studies have shown this to be an important mechanism in polynya formation in the Southern Ocean (e.g., Martinson et al., 1981; Goosse and Fichefet, 2001). Changes in sea-ice growth, melt and transport also have potential implications for the large-scale Atlantic meridional overturning circulation (e.g., Holland et al., 2001; Komuro and Hasumi, 2003) and the extension of this circulation into the Arctic (Bitz et al., 2006). Feedbacks associated with sea-ice insulating properties also modify the ocean circulation response to external forcing perturbations (Levermann et al., 2007). Changes in ocean circulation in turn influence the poleward ocean heat transport with consequential effects on sea-ice mass budgets. However, considerable uncertainty remains in the magnitude, importance, and sometimes even the sign of these sea ice-ocean circulation/mixing feedbacks.

9.2.9. Thresholds, tipping points and rates of change

With the recent rapid decrease in summer sea-ice extent in the Arctic and with the stark contrast seen between white, snow-covered ice and black open ocean, the question naturally arises as to whether the dramatic change in the surface means the entire system is poised to change rapidly and irreversibly. *Has the Arctic sea-ice system passed a tipping point?* was a question raised but not answered by Lindsay and Zhang (2005). This question has renewed scientific, political and economic interest after the surprisingly large decrease in Arctic sea-ice extent that occurred in summer 2007. It has evolved into the broader question of how Arctic sea-ice extent will evolve in the future. This section briefly describes current understanding of this issue and also touches upon the question of a possible tipping point for the loss of Arctic sea ice.

Considerable discussion has ensued about how vulnerable the Arctic sea-ice system is to passing a tipping point and indeed

about what constitutes a tipping point in this context. The notion is adopted here that a tipping point exists for sea ice if the ice does not recover from a certain ice loss caused by climatic warming even if the climate forcing were to return to the colder conditions that existed before the onset of that specific ice loss.

A dynamic system may be said to be in a stable state, where changes in the system remain small when it is perturbed; an unstable state, where the system changes rapidly, irreversibly, and of its own accord; or a metastable state, where the system is stable as long as it does not change, but becomes unstable if the state of the system moves a small amount in the right direction. A tipping point may be regarded as a metastable state.

The major forcing parameters that determine the state of the sea ice are the thermodynamic forcing (air temperature, humidity, cloud cover, ocean heat) and the dynamic forcing (winds, currents). The possibility of a tipping point depends on the nature of the feedbacks that result from changes to these forcings. A feedback parameter 'a' may be defined in terms of the changes seen in a system with and without the feedback operating:

$$\Delta X_\alpha = \Delta X_0 + a\,\Delta(X)_\alpha$$

where ΔX_α is, for example, the change in the September ice extent including the albedo feedback and ΔX_0 is the change expected in the extent due to global warming in the absence of feedbacks (Curry et al., 1995). The change measured might also be the mean surface temperature or the mean ice thickness. The feedback parameter can only be determined from modeling studies where the feedbacks can be turned off. If all of the feedbacks are linear with the forcing, then they only help determine the rate at which the ice changes as a given forcing parameter changes and do not cause a rapid and irreversible change. However if the feedbacks are sufficiently strong and sufficiently nonlinear, it is conceivable that they could contribute to forming a tipping point when the forcing conditions change.

Curry et al. (1995) conducted a study of the ice-albedo feedback in several 1-D models and emphasized the importance of evaluating the sensitivity of a model to perturbations of the forcing. Further, they emphasized that the ice albedo feedback is dependent on a range of physical parameters characterizing the ice cover, including ice extent, ice thickness distribution, snow depth, and melt pond coverage. They determined the feedback parameter 'a' for ice thickness and for surface temperature and found a wide range of values, from 0 to 10, depending on the model used and the parameter 'X' measured.

If the trajectory of the system does not retrace its original path when forcing conditions revert to normal, then the change is said to be irreversible. Eisenmann and Wettlaufer (2009) showed how the sea-ice system in a simple model can exhibit hysteresis when the thermal forcing first warms and then cools the polar regions. They showed that the trajectory reaches a bifurcation point (tipping point) under extreme warming where all of the winter ice is lost. In this case, owing to the albedo feedback, winter ice does not form again until an extreme amount of cooling is imposed.

Flato and Brown (1996) constructed a 1-D thermodynamic model that had two stable states: thin seasonal ice and thick perennial ice. A relatively modest change in climate was

sufficient to cause the model to jump from the thick perennial state to the thin seasonal state, thus exhibiting a tipping point.

Stern et al. (2008) examined the trajectory of a sea-ice system as described by the fractional monthly coverage of thin and thick ice. They examined this trajectory in the CCSM model and in a retrospective model simulation. They also constructed a simple empirical linear model with exchange parameters for each month of the year that converges to a stable annual cycle from any initial state. When a quadratic nonlinear term is included to simulate a crude ice-albedo feedback, the model exhibits two stable states, one with perennial ice and one with ice-free summers, resembling the projections of some climate models for the late 21st century. The existence of two possible stable cycles implies the existence of a tipping point between them.

A number of other studies have not found evidence of a tipping point as the amount of summer open water increases. Holland et al. (2006) found that several GCMs exhibit periods of abrupt ice loss in which the late summer ice extents decline rapidly for a period of five years or more, but the rapid decline does not continue until all of the summer ice is gone and is often followed by partial recoveries, not what would be expected for a tipping point.

A tipping point according to this definition has been suggested to exist for the loss of sea ice owing to various destabilizing feedbacks. Such feedbacks can lead to additional sea-ice loss once the ice cover has decreased by a certain amount. Among these destabilizing feedbacks is the ice-albedo feedback, which is probably the most often cited climatic feedback loop. As discussed in Section 9.2.2, less ice will lead to more absorption of solar radiation, and hence to a warmer Arctic with less ice. According to this simple picture, the ongoing loss of sea ice in the Arctic can only be stopped if the climate were to become significantly cooler. This automatically implies the existence of a tipping point for the loss of Arctic summer sea ice.

However, recent research consistently suggests that despite the ice-albedo feedback, no tipping point exists for the loss of Arctic summer sea ice (Winton, 2006a; Eisenman and Wettlaufer, 2009). There are two important reasons for the possible non-existence of a tipping point for the loss of Arctic summer sea ice. First, recent research suggests that the ice-albedo feedback is potentially less efficient than previously thought (Winton, 2006b; Bitz, 2008; Winton, 2008; Graverson and Wang, 2009). For example, the annual sea-ice minimum is reached in September at a time when incoming solar radiation is already rather weak.

The second reason for the possible non-existence of a tipping point for the loss of Arctic summer sea ice lies in the very high growth rate of thin ice, which provides an efficient resistance of the sea-ice cover to persistent ice loss. With a decreasing summer sea-ice cover, larger and larger areas of the Arctic Ocean become ice-free during summer. As long as winter remains relatively cold, sea ice will start to form in these areas at some point during the growth season. The forming thin ice grows much more rapidly than thick ice, which leads to a recovery of the sea-ice cover during winter. An additional contributor to this recovery is that there is often little snow on the newly forming sea ice. Since snow is an efficient insulator, ice with little snow on its surface can grow much faster than ice with a thick snow cover. This allows for the newly forming

sea ice to reach a thickness that might even be larger than that of ice that survived the summer in the same area (Figure 9.28).

The results imply that if the climate were to become colder, the loss of sea ice that occurred during the past decades can probably be reversed. If, however, the climate continues to warm, the loss of sea ice is likely to decline further in the future. This is caused by the increasingly thinner sea-ice cover becoming increasingly vulnerable to ice loss during summer: With a thinner sea-ice cover, a given amount of energy that is available for melting during summer will cause a larger fraction of the ocean to become completely ice-free (Holland et al., 2006).

The winter ice extent may not be so robust in a warming world. Eisenman and Wettlaufer (2009) used a simple seasonal model of the Arctic sea-ice system to show that the ice-albedo feedback does not cause a bifurcation to all ice-free conditions (a tipping point) as the summer ice extent declines to all open-water conditions, however if the winter maximum thickness declines too far in a seasonally ice-covered regime, the system becomes unstable and all of the ice disappears, even in winter.

Winton (2006a) examined the output of two GCMs that had been run to a quadrupling of the CO_2 where all sea ice disappeared, even in winter. The author found no evidence of acceleration of the ice decline or of polar warming when the summer ice disappeared. However, one of the models showed some indication of an acceleration as all of the winter ice disappeared.

Evidence suggests that summer sea ice is vulnerable to abrupt changes not because the feedbacks are so powerful but because when much of the pack is very young and very thin, small perturbations in the thermal or dynamic forcing can
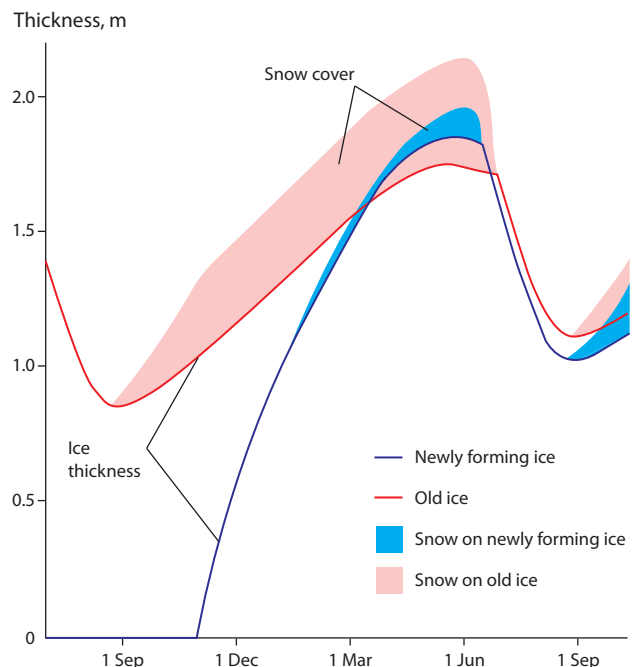
Figure 9.28. Schematic illustration comparing the ice thickness evolution of newly forming ice and of ice that survived the summer. The snow on top of the ice is represented by the shaded blue and red areas. Ice formation of the newly forming sea ice was prescribed to start on 1 November, two months after the end of the melting season of the ice that survived the summer. Note that nevertheless the newly forming ice reaches a larger maximum thickness than the old ice because of the thinner isolating snow-cover on top of the former. Source: Notz (2010).

cause very large changes in ice extent. This was the case in 2007, when the cumulative effects of many years of gradual warming and episodes of large anomalous ice export from the Arctic Ocean in the early 1990s created large amounts of thin ice in the spring that was vulnerable to melting the following summer. This condition plus unusual winds that blew the remaining ice to one side of the basin caused the abrupt decline of the September ice extent (Lindsay et al., 2009). As the late summer ice extent continues to decline, large variability could be expected in ice extent, but not a tipping point.

This vulnerability of a thinner ice cover to widespread ice loss during summer was one of the reasons for the surprising record low in sea-ice extent in summer 2007. It is likely that similar surprises will occur in future, with the sea-ice cover in the Arctic apparently being in a state of transition to mostly ice-free conditions in late summer. Current understanding of the physics of sea ice and of its interaction with the ocean and the atmosphere is still too limited to reliably forecast the future trajectory of this transition. For example, recent studies suggest possible changes in the atmospheric circulation in a warmer climate (Zhang et al., 2008c). Such changes might cause stable high-pressure systems to form more often in the Arctic during summer, which in turn would accelerate the sea ice to retreat in the coming decades. In addition, the oceanic heat transport into the Arctic might increase in a warmer climate, which could also cause sea ice to disappear much more rapidly than currently projected.

9.3. Biological impacts of changes in sea ice in the Arctic

- Predictions of seasonal sea-ice extent are highly variable on a regional basis. Environmental conditions and bathymetry are also highly variable on a regional basis. In combination, these factors will strongly affect the impacts of climate warming on Arctic biological systems.

  - The central Arctic basins are likely to become more productive under conditions of decreased MYI and increased open-water extent and duration. It is not possible to predict the magnitude of this increase in production or its persistence as a shift is anticipated from light limitation to nutrient limitation.

  - Changing sea-ice and snow patterns will shift the balance between primary production by ice algae and phytoplankton, with unknown consequences for energy flow in Arctic food webs.

  - The timing of sea-ice retreat will affect the availability of ice-associated zooplankton and lipid-rich forage fish, which are critical resources for seabirds, marine mammals and some commercially important fish species (e.g., Atlantic cod).

  - Sub-Arctic species are expanding northward and are likely to compete with Arctic species. Ice-associated bird species are in some cases already declining (e.g. ivory gulls); others are likely to do so in the future.

- The extinction of some Arctic endemic species is likely if current sea-ice trends continue. These losses will represent biodiversity losses of global significance, and reverse millions of years of evolutionary change (since the beginning of the Pleistocene). Northward range expansions for some temperate species and potentially greater overall productivity will enhance foraging opportunities for some animals within High Arctic areas, at a cost to Arctic endemic populations.

  - It is highly likely that distribution and abundance reductions will take place among ice-associated pinniped populations (ringed seals, bearded seals, walrus) due to reductions in the extent and thickness of sea ice. Pack ice-breeding seals (hooded, harp, ribbon, spotted) will experience reproductive failure more frequently as their late winter / early spring breeding habitat becomes less stable. This is already evident in the Atlantic sector of the Arctic (West Ice/Fram Strait and White Sea breeding areas, impacting harp and hooded seals).

  - Model results suggest that polar bears are highly likely to be extirpated from seasonally ice-covered regions (where two-thirds of the world's polar bears currently live) within 50 to 70 years.

- The responses of Arctic endemic cetaceans to changing sea-ice conditions will be influenced by food availability, competition from non-endemic migrant cetaceans and, in some regions, the potential for increased predation from killer whales.

9.3.1. Sea ice and Arctic marine ecosystems

Arctic marine ecosystems have several unique physical characteristics that set them apart from marine systems at lower latitudes. Some of the most important include: strong seasonality in light levels, from complete winter darkness to continuous daylight in summer; low overall temperatures with winter extremes; and the presence of extensive shelf seas around a deep central ocean basin. However, the most defining characteristic is perhaps that for the last five million years or more there has been a 'permanent cap' of multi-year ice that extends and retracts seasonally over vast ocean areas with the formation and melt of first-year ice (see Polyak et al., 2010 for historical sea-ice patterns). Together, the MYI and the FYI comprise a unique habitat type that has become home to a wide range of Arctic species (e.g. polar bears *Ursus maritimus*) that have evolved within the sea-ice environment or have joined it over the millions of years of its existence (Figure 9.29).

In addition to its direct role as habitat for Arctic biota, sea ice also plays a major role in terms of structuring Arctic ecosystems (e.g., Alexander, 1980; ACIA, 2004). Sea ice influences all processes operating across the ocean–sea–atmosphere interface (see Sections 9.1 and 9.2). From a biological perspective, one of the key issues is its influence on radiative exchange, which is critical for photosynthesis and surface heat fluxes. Sea ice also influences freshwater fluxes, and momentum exchange between the atmosphere and the oceans, which in turn influences surface stratification and mixing. Ice margins, as well as polynyas (openings within ice-covered areas) are thus typically areas of high productivity where large numbers of marine birds and mammals congregate to feed.

The ongoing changes in Arctic sea ice (e.g., Overland, 2009) are expected to have major consequences for marine ecosystems. The objective of this section is to provide a brief summary of current knowledge of sea ice-associated ecosystems in the Arctic, to report on documented changes that have already been observed in relation to changes in sea-ice conditions, and to predict likely future changes in Arctic ecosystems in relation to current scenarios for projected changes in Arctic sea ice over the 21st century.

9.3.1.1. **Biota in sea ice**

The sea-ice interior is habitat for an abundant and diverse microscopic community that forms the base of sea-ice associated food webs. The sea-ice biota consists of a complete food web with primary producers, a microbial food chain and three to four trophic levels (e.g., Horner et al., 1992; Deming, 2010).

Microorganisms, nutrients and other inorganic and organic constituents are incorporated into sea ice as the ice is formed. Larger cells are selectively scavenged from the water column into the sea ice at the time of its formation (Gradinger and Ikävalko, 1998; Riedel et al., 2007; Różańska et al., 2008). The organisms incorporated into sea ice are assumed to be the founding members for the development of the ice-algal bloom that takes place in spring with the seasonal increase in solar radiation. Light availability determines the onset of ice-algal production in spring (e.g., Lavoie et al., 2005). Since snow strongly attenuates light penetration throughout the snow-ice matrix, snow-cover thickness and patchiness influence the temporal and spatial variability in ice-algal distribution, affecting the timing of ice-algal growth (e.g., Riedel et al., 2006) and horizontal distribution of ice algae (e.g., Rysgaard et al., 2001; Mundy et al., 2007). While light determines the onset of the bloom, nutrients are essential to sustain sea-ice production, and can become limiting toward the end of ice-algal blooms. Self-shading in the dense ice algal layer can also, at times, limit ice-algal production (e.g., Smith et al., 1988). Recent results from the Chukchi and Beaufort Sea (Gradinger, 2009; Różańska et al., 2009) suggest that nutrient supply from the water column determines the upper limit of ice-algal production (also see Michel et al., 2006).

Organisms are more abundant in Arctic FYI than MYI. This is due to the greater presence of brine pores and channels in FYI that offer more colonization or habitable space for a diversity of microorganisms. In the Arctic, sea-ice algae flourish primarily at the ice–water interface (bottom-ice communities), and algae located here represent up to 95% of the total chlorophyll *a* biomass in FYI in spring, but they can also be found higher up toward the surface of the ice (e.g., Mock and Gradinger, 1999). Ice algae can represent between 4% and > 50% of total primary production in the Arctic on a regional basis (Legendre et al., 1992; Gosselin et al., 1997) and can play a key role in Arctic food webs as an early and concentrated (in space and time) food source for zooplankton grazers at a time when little or no other food is available (Bradstreet and Cross, 1982; Michel et al., 1996; Gradinger, 2009).

Sea-ice communities include hundreds of protist species (Von Quillfeldt et al., 2003; Różańska et al., 2009). A viable
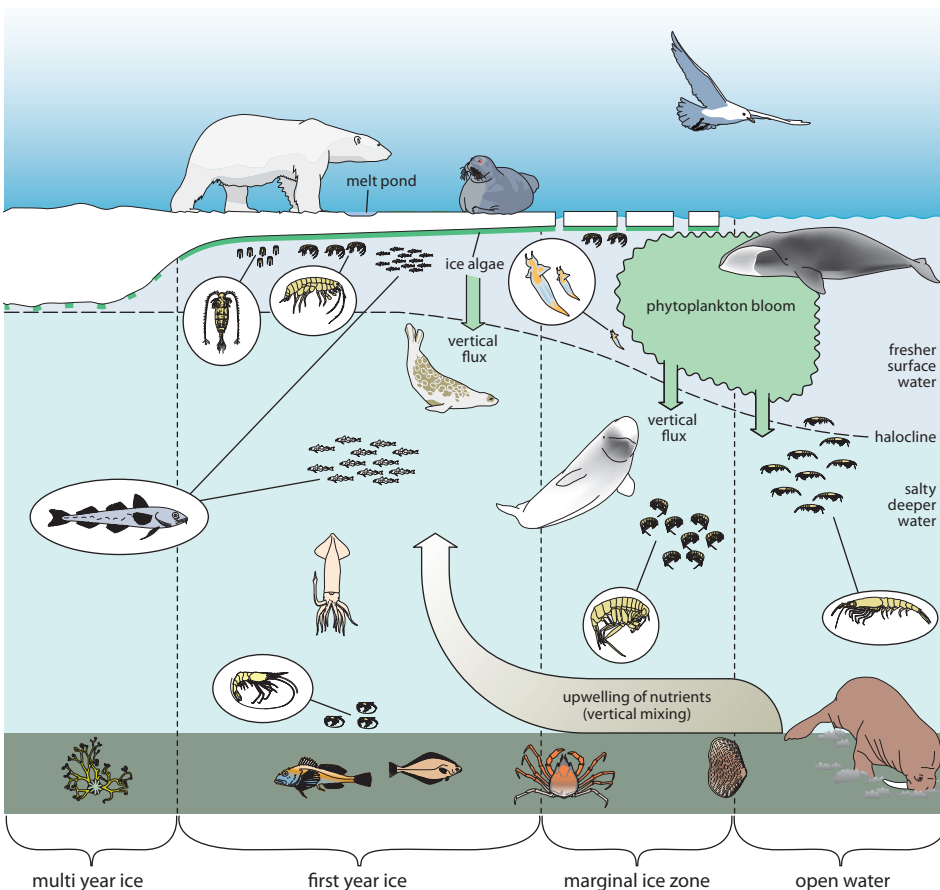

Figure 9.29. A schematic view of the major biological components and physical features of the Arctic sea-ice community during the spring ice retreat/melt period. Source: drawn by Audun Igesund, Norwegian Polar Institute.

sea-ice protist community may be present throughout the sea-ice season, with a characteristic species succession from flagellate dominance (dinoflagellates and cryptophytes) in winter followed by pennate diatoms during the bloom period (Mikkelsen et al., 2008). The colonial diatom *Nitzchia frigida* is a pivotal species of landfast ice across the Arctic as it is the most frequent, most abundant and most consistently observed species in FYI (Różańska et al., 2009). Bacterial abundances in the ice can be high and their activity is often linked to dense algal aggregations in bottom communities (Meiners et al., 2008). Heterotrophic protists are considered important bacterivores within the ice. There is also a diverse community of multi-cellular organisms living within the ice (often referred to as ice meiofauna), including several species of Crustacea (mainly Harpacticoidea), Nematoda, Acoela, Rotifera and Cnidaria (Gradinger, 2002; Bluhm et al., 2009a). In addition to these 'permanent' members of the ice communities, meroplanktonic larvae and juvenile stages of benthic Polychaeta and Gastropoda inhabit the ice for periods of a few weeks to months (Gradinger et al., 2009a). Despite many of the metazoan species found in sea ice being herbivores that feed on ice-algal biomass, typical daily ingestion rates are below 10% of the algal standing stock (e.g., Gradinger, 1999; Michel et al., 2002).

Overall, grazing by in-ice meiofauna and macrofauna does not limit the seasonal build-up of ice-algal biomass in the bottom layers of Arctic sea ice, so the bulk of the ice-algal biomass is available for under-ice grazers or is released into the water column at the time of ice melt (Michel et al., 2002). This coupling between sea ice and pelagic (water column) and benthic (seafloor) ecosystems is important in structuring Arctic marine food chains (Bluhm and Gradinger, 2008).

9.3.1.2. Biota under sea ice

The inherent difficulties in measuring ice-algal production (see Glud et al., 2007), and its patchy distribution, has led to considerable uncertainty as to its role within Arctic ecosystems. Clearly, ice-associated communities are capable of exploiting the newly produced organic matter, but it is only fairly recently that the direct and indirect roles of ice algae for pelagic and benthic food webs have been shown to be significant (Michel et al., 1996, 2002; Renaud et al., 2006; Tamelander et al., 2008; Gradinger et al., 2009b).

Arctic under-ice fauna is dominated by crustaceans, mainly gammarid amphipods, copepods and fish (*Boreogadus saida* and *Arctogadus glacialis)*. In some areas, pelagic copepods, most notably *Calanus* and *Pseudocalanus* species, migrate toward the under-ice surface to feed (Runge et al., 1991; Fortier et al., 2001). Recently, swarms of krill have been observed under Bering Sea ice (Gradinger, Bluhm and Iken, University of Alaska, pers. comm.). Pressure ridges often harbor higher concentrations of amphipods and meiofauna compared to level sea ice (Hop et al., 2000; Hop and Pavlova, 2008; Gradinger et al., 2009a). This may be because ice floes that make up the ridging reach below the low-salinity water layer observed during ice melt in summer. Under-ice copepods and amphipods play an important role in linking sea ice-algal production to higher trophic levels. The abundance and feeding ecology of amphipods vary widely. However, in areas of high abundance, amphipods can consume 30% to 60% of the ice-algal biomass (Siferd et al., 1997; Werner, 1997).

Several studies have reported copepods feeding on ice algae, providing some insight into the potential importance of this early food source for zooplankton (e.g., Conover et al., 1986; Fortier et al., 2001). The strongest evidence for just how important this food source may be for the pelagic ecosystem, however, comes from recent studies using stable-isotope and fatty-acid biomarkers. Ice algae contain polyunsaturated fatty-acids (PUFAs), many of which are essential for consumers and must be derived from their diets. Søreide et al. (2006) and Tamelander et al. (2008) showed that ice algae make up significant fractions of the diets of many zooplankton taxa during winter and spring. PUFA biomarkers also directly link ice algae to the diets of key members of Arctic pelagic food-webs (Auel et al., 2002; Søreide et al., 2008).

Søreide et al. (2010) recently showed that ice algae and the timing of its production are not just important for additional food intake, but also play critical roles in the development and survival of *Calanus glacialis*, a key Arctic zooplankton species. *C. glacialis* is among the most important large copepod species across the Arctic, and its predators (fish, seabirds, seals, bowhead whale *Balaena mysticetus*) rely on its high lipid levels to build their own lipid stores (Karnovsky et al., 2003). Its use of ice algae early in the season, therefore, is significant for the functioning of large parts of the Arctic ecosystem.

With body sizes of 1 to 6 cm, sea-ice associated amphipods are a major dietary item for the only ice-associated fish taxa in Arctic waters, the Arctic cod (*Boreogadus saida*) and polar cod (*Arctogadus glacialis*) (Bradstreet and Cross, 1982). In addition to under-ice amphipods, other common prey items for fishes under the ice include the copepods *Calanus glacialis* and *C. finmarchicus*, the hyperiid amphipods *Parathemisto libellula* and *Themisto abyssorum*, and the mysid *Mysis oculata* (Bradstreet et al., 1986; Lønne and Gulliksen, 1989; Süfke et al., 1998). Polar cod also feeds on juvenile Arctic cod (Joensen, 2008).

Arctic cod occurs across the Arctic in variable-sized schools ranging from small groups to several million fishes (Andriyashev, 1954; Craig et al., 1984), and is often associated with sea ice (Bradstreet et al., 1986; Lønne and Gulliksen, 1989). Under-ice aggregations of Arctic cod have recently been reported in the Beaufort Sea (Benoit et al., 2008). Its apparently eurythermal and euryhaline physiology (Craig et al., 1984) allows Arctic cod to inhabit and take advantage of a wide range of habitats. Polar cod occurs in the Arctic and North Atlantic at depths from 0 to 1000 m (www.fishbase.com, retrieved Oct 9, 2009) and is also often ice-associated (Dorrien von et al., 1991). Where the two species co-occur, dietary segregation may occur (Joensen, 2008). Both species are important food items for top trophic animals. Arctic cod is a major food source for ringed seals (*Pusa hispida*), spotted seals (*Phoca largha*) and ribbon seals (*Histriophoca fasciata*), narwhal (*Monodon monoceros*) and beluga (*Delphinapterus leucas*), and provides important food during seasonal migrations of seals feeding in northern waters, while polar cod is common in diets of beluga, narwhal and ringed seals (Bluhm and Gradinger, 2008). Arctic seabirds (murres, guillemots, kittiwakes) prey on Arctic cod throughout the Arctic (Lønne and Gabrielsen, 1992; Hobson, 1993; Weslawski et al., 1997), and this single species can comprise 85% or more of the diet of ringed seals in Svalbard (Labansen et al., 2007).

Ice algae can provide an important food supply to the benthos as a large part of the biomass accumulated during the bloom can sink to the seafloor (e.g., Fortier et al., 2001). Sediment trap studies have identified pulses of algal export from sea ice (e.g., Michel et al., 2006; Forest et al., 2007; Tamelander et al., 2009), which may occur more than one month before the typical phytoplankton pulse (Ambrose and Renaud, 1997; Juul-Pedersen et al., 2008). Benthic communities are fully capable of consuming and assimilating ice algae. Results from a winter-to-summer time series in the Beaufort Sea showed a rapid response by the benthos to ice-algal deposition (Renaud et al., 2008). Experimental studies in the Arctic indicate that seafloor communities respond to a pulse in ice algae within 12 hours (McMahon et al., 2006), and that ice-algal-derived PUFAs are assimilated preferentially (Sun et al., 2007). Rapid, intense response and assimilation suggest that ice–benthic coupling plays a significant role in benthic food webs on Arctic shelves. Increasingly, it is thought that ice algae are important for fueling growth and reproduction of pelagic and benthic taxa, and for ecosystem functioning in general.

9.3.1.3. **Biota on the sea ice and at its edges**

Ice-edge environments are regions of intense seasonal production and a focal point for trophic transfer to the upper trophic levels during bursts of primary production because seabirds and marine mammals, which must surface to breathe, can easily access concentrated marine production along ice edges. Intermittent wind-driven upwelling at ice edges serves to fuel primary or new production along the ice edges by replenishing nutrients to the surface layer (e.g., Alexander and Niebauer, 1981). As the ice retracts northward during spring, exposing new areas with a build-up of yet unused nutrients (or in some polynyas, cracks and leads) the result is seasonal bursts of productivity that occur in a concentrated fashion along the edges of the ice within the spatially transient marginal ice zone (MIZ) (see Box 9.8). The high productivity of these regions plays a key role in Arctic food chains, particularly because it represents production early in the season when other food for secondary consumers is limited (e.g., Hunt et al., 1996; Karnovsky and Hunt, 2002). The presence of top predators in the MIZ, including seals and whales is largely mediated by the abundance of lower trophic level prey organisms (zooplankton and fish) that are concentrated along the ice edge (e.g., Kaschner et al., 2006). There is little competition from other species for marine mammals that are able to access ice-associated prey beneath the ice (Burns, 2009a). The surface of the sea ice is also important in that it provides habitat for birthing, molting and resting, as well as a mating platform or staging area for some ice-associated seals; a hunting and transportation platform for polar bears; and the ice provides protection from predators and a less turbulent environment that is shielded from strong winds both for seals and whales (Tynan and DeMaster, 1997; Burns, 2009a). Sea ice is a unique habitat for mammals in a global context; it has few surface predators and is virtually free of disease vectors (Burns, 2009a).

Polar bears have a circumpolar Arctic distribution. Nineteen populations of polar bears are currently recognized, varying in size from a few hundred to a few thousand animals within a global population estimated to be between 20 000 to 25 000 animals (Aars et al., 2006). Polar bears are heavily dependent on sea ice for foraging and on ice corridors for transportation to and from terrestrial denning areas, which often occur along slopes near shore-lines where snowdrifts accumulate sufficient snow (Stirling, 2009). Female polar bears emerge from their dens in spring, following a fasting period of four or five months. They depend on immediate access to sea ice and a plentiful supply of ringed seal, particularly pups, at this time of year. Polar bears are scarce in the permanent MYI of the central Arctic Ocean as they prefer FYI that forms over shelf areas where seals, their primary prey, are most abundant. Ringed seals are the dominant prey of polar bears, but bearded seals (*Erignathus barbatus*) and other ice-associated seals such as harp seals (*Pagophilus groenlandicus*) and hooded seals (*Cystophora cristata*) are important regionally. Even walruses (*Odobenus rosmarus*) are consumed by polar bears in some areas (see Thiemann et al., 2008).

The largest of the Arctic seals, the walrus, ranges across the circumpolar Arctic, but the species' distribution is disjunct and two subspecies are recognized, one in the Pacific (*O. r. divergens*) and the other in the Atlantic (*O. r. rosmarus*). The actual distribution of walruses is restricted by their narrow ecological niche. They depend on shallow water (<100 m) with suitable bottom substrate that can support high bivalve abundances, reliable open water over the rich feeding areas, and haul-out areas near feeding areas (Kovacs and Lydersen, 2006). For much of the year, the haul-out platform is sea ice (e.g., Freitas et al., 2009). Atlantic walruses of all ages and sexes use terrestrial haul-outs in summer and autumn, as do Pacific walrus males. But, Pacific walrus females and calves remain in the MIZ during summer. There are probably in excess of 200 000 walruses worldwide, but many population assessments are out of date and in some areas are non-existent (e.g., Franz Josef Land, Pechora Sea, Laptev Sea). A key sea ice-related sensitivity with walrus is that, at least seasonally, all populations use the MIZ as a platform to move over foraging areas that are too far from land-based haul-out sites to be energetically feasible sites for feeding. The use of the two habitats broadens the feeding distribution markedly, which permits greater overall walrus abundances.

The smallest of the Arctic seals, the ringed seal is also a circumpolar species. The world-population of ringed seals probably numbers in the millions, but few areas have been systematically surveyed (Hammill, 2009). This species is the only northern seal that can maintain breathing holes in thick sea ice and thus has the broadest distribution among the Arctic pinnipeds, ranging north to the Pole. This species is extremely dependent on sea ice, which is its exclusive breeding and haul-out platform (Smith, 1987; Burns, 2009a). Snow on the surface of the sea ice is essential for the construction of lairs (small caves above a breathing hole), in which the seals routinely rest during the winter months and also give birth in the early spring (e.g., Furgal et al., 1996). Ringed seal pups are only about 4 kg when born, so are thermally dependent on having a lair environment. The lair also provides some protection against polar bears, Arctic foxes and other potential predators. Lack of snow build-up on the ice or high spring temperatures and spring rains can result in reproductive failure in this species (e.g., Smith and Harwood, 2001; Stirling and Smith, 2004). Ringed seals

Box 9.8. The role of polynyas and flaw leads in Arctic sea-ice communities

Open-water areas within ice-covered seas are regions of high biological productivity compared to the surrounding ice-covered waters; in addition, the growth of phytoplankton in these areas starts early in the season making them important biological 'hot spots' (e.g., Bursa, 1963; Hirche et al., 1991; Stirling, 1997). In contrast, where Arctic sea ice remains mobile throughout the winter (e.g., Hudson Bay and Strait, Barents Sea) small open-water leads occur unpredictably with tidal and wind-driven ice movements, making the small open-water areas that result too ephemeral to support most birds or marine mammals that feed within the water column (e.g., Hunt et al., 1996). An exception is the black guillemot (*Cepphus grylle*), which overwinters in the mobile pack, apparently shifting from lead to lead as conditions change (Gaston and McLaren, 1990).

In some areas, offshore winds and the movement of pack ice may create more extensive shore leads. These leads attract a diverse array of birds which remain in such areas over winter, especially common eiders (*Somateria mollissima*), and to a lesser extent long-tailed ducks (*Clangula hyemalis*) and king eiders (*S. spectabilis*) (Gilchrist and Robertson, 2000), as well as ice-associated seals and whales (Kingsley et al., 1985; Heide-Jørgensen and Laidre, 2004; Moore and Laidre, 2006). The overwintering of ice-associated whales within mobile pack ice and temporary polynyas occasionally causes entrapments (Nerini et al., 1984; Finley, 2001), sometimes involving hundreds of whales (Harwood and Smith, 2002). Later freeze-up as a result of climate change may alter the frequency of formation of these types of open-water areas and reduce the predictability of conditions and the reliability of prior experience (Laidre and Heide-Jørgensen, 2005b).

In regions of strong tidal currents, such as those produced in narrow inlets, small tidal polynyas remain open throughout the winter, providing refuges. Such predictable openings although typically small, serve as concentration areas for a wide variety of marine birds, and even whales and seals, if they are large enough (e.g., Kingsley et al., 1985; Richard et al., 1998). These aggregations of potential prey attract ice-based predators (polar bears and Arctic foxes *Vulpes lagopus*) and raptors such as gyrfalcons (*Falco rusticolus*) and snowy owls (*Bubo scandiaca*) (Brown and Nettleship, 1981; Stirling, 1997). If loss of sea ice results in the marine birds and other animals remaining dispersed, the predators that rely on these concentrations may have to seek other food sources.

Numerous small polynyas of this type occur near the Belcher Islands (Sanikiluaq) where they are essential to the year-round existence of a large population of common eiders (Freeman, 1970). The eiders spend the night on the tidal polynyas, flying out at dawn to the open water of shore leads for feeding. When shore leads close, the whole population may be forced to feed in the tidal polynyas and then benthic prey (e.g., bivalves) becomes rapidly depleted (Gilchrist et al., 2006). Similarly, the entire world population of spectacled eiders (*S. fischeri*) overwinters in leads and polynyas south of St. Lawrence Island, in the eastern Bering Sea (Petersen et al., 1999), an area of strong benthic–pelagic coupling that supports an important bivalve community with a high density of prey for the eiders (Lovvorn et al., 2003; Grebmeier et al., 2006b). The ability to rest on the ice at night, when they are not feeding, is essential to the energy balance of the eiders (see Figure 9.30; Lovvorn et al., 2009).

Polynyas of all kinds are particularly important because they provide early season productivity. In summer, the region of the North Water Polynya, in Baffin Bay, supports some of the largest concentrations of seabirds anywhere in the Arctic (Stirling, 1980, 1997), dominated by little auks (*Alle alle*) which breed in the millions in northwestern Greenland (Kampp et al., 2000). In the North Water Polynya, the distribution and diets of the little auks reflect physical processes responsible for the timing and spacing of polynya opening in spring and subsequent food-web dynamics (Karnovsky and Hunt, 2002). The eastern side of the polynya is a sensible heat polynya (driven by the upwelling of warmer water) and primary production there precedes production elsewhere (Tremblay et al., 2006a). Little auks arrive first in the open waters of the eastern side, and only later in summer occupy the western side, which is predominately a latent heat (wind-induced) polynya. Little auks time their arrival to coincide with the availability of copepods (*Calanus* spp.), their primary prey in May to July. Thus, the timing and location of little auks in the region is determined by the timing of primary production and the upward migration of copepods to crop that production, rather than by the availability of open water itself, although of course the birds must have enough open water to access their prey (Karnovsky and Hunt, 2002).

Changes in the location or predictability of polynyas due to changing sea-ice conditions could have significant impacts on seabird and marine mammal populations. Large polynyas, particularly where they occur over continental shelves, are currently important to all trophic levels of ice-associated ecosystems.
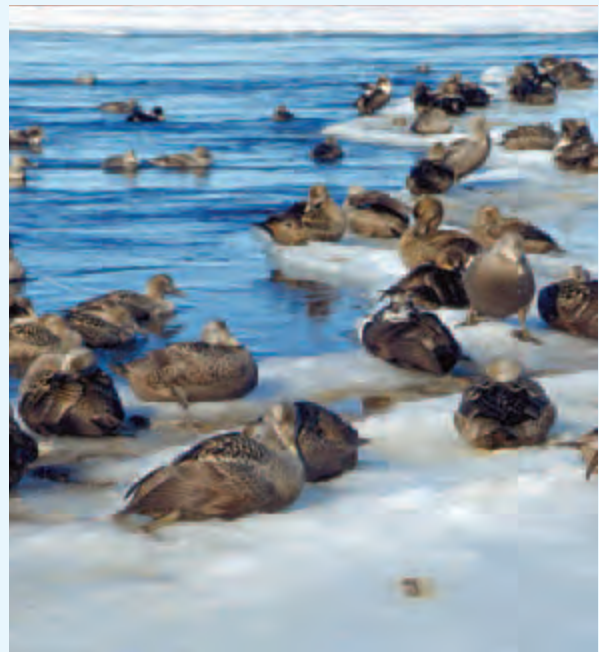

Figure 9.30. Common eiders resting on the ice edge in a polynya in early spring. Source: Grant Gilchrist, Environment Canada.

have the longest lactation period of any of the northern true seals, and need stable ice throughout the period of neonatal dependence (Lydersen and Kovacs, 1999). Ringed seals appear to prefer landfast FYI for breeding, but can be found on all ice types including free-floating pack ice and MYI (Burns, 2009a). Ringed seals remain associated with sea ice throughout the year and feed predominantly on ice-associated prey, which are particularly important to young ringed seals (Freitas et al., 2008a). Polar cod, Arctic cod and large zooplankton species such as *Parathemisto libellula* make up most of the ringed seal diet (e.g., Wathne et al., 2000; Labansen et al., 2007). A key sea ice-related sensitivity with ringed seals is that this species relies on early formation of landfast sea ice in the autumn and on the formation of ridges and rubble zones in FYI, which can capture enough snow for birth lair formation.

Bearded seals have a patchy distribution throughout the Arctic, occurring at low densities across their range. The global population of bearded seals has not been assessed, but probably numbers in the hundreds of thousands (Kovacs and Lydersen, 2006). This species is largely solitary, although small groups can be seen on the ice during late spring and early summer, when they are breeding and then molting. Similar to walruses, bearded seals forage mainly on benthic organisms (Kovacs, 2009a). Therefore, they prefer to reside in drifting pack-ice over shallow water. They are largely coastal animals but can also be found in drifting pack-ice far from shore. Bearded seals avoid densely packed ice unless open-water leads are available. During winter, they concentrate near polynyas or in areas where leads are frequent, or they stay near ice edges. Bearded seals prefer to occupy areas with drifting ice in shallow coastal environments on a year-round basis that have a rich bottom fauna to support their energy requirements. But, in areas where ice disappears entirely, they will haul out on shore for resting or molting (Burns, 2009a).

Four other ice-associated seals live within Arctic and sub-Arctic pack-ice areas. These include harp seals and hooded seals in the North Atlantic region and spotted seals and ribbon seals in the Bering Sea, Chukchi Sea and Sea of Okhotsk. These species all give birth to their young on the pack ice and spend most of their lives associated with sea ice (Burns, 2009a,b; Kovacs, 2009b; Lavigne, 2009; Lowry and Boveng, 2009). With the exception of the spotted seal, these seals only haul-out on ice. Spotted seals are relatively recently evolved from harbor seals and in summer can be seen hauled-out on shore in harbor seal groups. All four of the Arctic pack-ice seals display broad diets, consuming a wide variety of small fish and invertebrate species. They favor lipid-rich schooling fishes such as polar cod, Arctic cod and capelin (*Mallotus villosus*). Hooded seals also eat larger prey such as Greenland halibut (*Reinhardtius hippoglossoides*) and deep-dwelling red fish (*Sebastes* spp.), and they share a diet of cephalopods with ribbon seals (Haug et al., 2007). They are all sensitive to the availability of vast fields of pack ice in late winter / early spring.

Three species of cetacean are endemic to the Arctic: the narwhal, the beluga (or white whale) and the bowhead whale. Beluga and bowhead whales a have a circumpolar distribution, while the narwhal occurs primarily in the Atlantic sector (Heide-Jørgensen, 2009; O'Corry-Crowe, 2009; Rugh and Shelden, 2009).

The narwhal is the most specialized of the three endemic species (Laidre et al., 2008). They are typically found in waters north of 60° N in the eastern Canadian High Arctic, offshore eastern and western Greenland, Svalbard and Franz Joseph Land. Narwhal overwinter in deep, ice-covered habitat along the continental slope and spend roughly two months in summer in ice-free shallow bays and fjords. These disjunct seasonal distributions are connected by annual migrations of roughly 1000 km that last about two months. Narwhal appear to feed little during summer, but some polar and Arctic cod can be found in their stomachs at this time as well as *Gonatus* squid, which seem to be eaten in all seasons in all localities (Heide-Jørgensen, 2009). Peak feeding for narwhal takes place from November to March in dense pack-ice habitat, which is focused on Greenland halibut, and which may be key to their survival (Laidre and Heide-Jørgensen, 2005a). In contrast, beluga occupy estuaries, continental shelf and slope waters as well as deep ocean basins in conditions that range from open water to dense FYI. Some beluga undertake long migrations between summer and wintering sites, while others remain in the same region year-round (e.g., Lydersen et al., 2001; Richard et al., 2001). Migratory beluga occur along the western and northern coasts of Alaska, in the Canadian High Arctic and western Hudson Bay. Non-migratory beluga often make seasonal shifts in distribution and in winter can be found near the MIZ. They are generalist feeders that eat a wide variety of prey including invertebrate and vertebrate species found in benthic, pelagic and ice-associated environments (Seaman et al., 1982; Dahl et al., 2000; Loseto et al. 2009; O'Corry-Crowe, 2009).

Bowhead whales are well adapted to ice-covered waters, spending summers in the High Arctic and winters in sub-Arctic seas. Bowheads occur in polynyas in winter, but can move easily through extensive areas of nearly solid sea ice and break though ice up to 45cm thick. Although strongly ice adapted, bowheads often feed in open water areas, especially in late summer and autumn, when sea ice can be hundreds of kilometres offshore (Moore et al., 2010). Bowheads eat invertebrate prey including copepods, euphausiids, mysids and gammarid amphipods (Lowry, 1993). A variety of other baleen whales and a few toothed whales also migrate into Arctic waters during summer, but sea ice represents a distributional barrier to these animals rather than useable habitat. The key ice related sensitivity for the ice-adapted whales is how the sea ice structures the ecosystem and influences prey availability. But, sea ice also provides a sheltered environment where whales can seek refuge from turbulent seas and escape predation by killer whales (*Orcinus orca*) (George et al., 1994; Burns, 2009a; see Section 9.3.3.4).

Seabirds are also a major biological feature of Arctic ice edges. Being highly mobile, birds are in a good position to exploit the seasonally available prey of the MIZ. In addition, a few species remain throughout the winter at high latitudes, taking advantage of polynyas or recurrent flaw-leads. Some seabirds use sea ice for resting both during migrations and foraging excursions and several species of seabird can be considered more-or-less obligate sea-ice associates: the thick-billed murre (*Uria lomvia*), a deep-diving species of auk (to 150 m in the Atlantic Arctic; Elliott and Gaston, 2009); the ivory gull (*Pagophila eburnea*), a small, tern-like, surface-feeding gull and; in the non-breeding season, Ross's gull (*Rhodostethia rosea*), another small surface feeder. In addition, in the north Pacific, Kittlitz's murrelet (*Brachyramphus brevirostris*), is considered

a specialist on tidewater inlets where glaciers reach the sea. Although its range extends southward well beyond the limits of the Arctic, the black guillemot is common in the High Arctic where it is found especially in narrow cracks and leads in sea ice and remains at high latitudes throughout the winter (Renaud and Bradstreet, 1980). Likewise, several species of eider (large benthic-feeding diving ducks) remain in ice-covered areas throughout the winter.

In summer, many other marine birds make temporary use of open water areas created by the retreat of winter ice cover. The most common are several species of gull (glaucous *Larus hyperboreus*, Iceland *L. glaucoides*, Thayer's *L. thayeri*, Sabine's *Xema sabini*), the black-legged kittiwake (*Rissa tridactyla*), three species of jaegers (parasitic *Stercorarius parasiticus,* pomarine *S. pomarinus,* long-tailed *S. longicaudus*), the little auk, the Arctic tern (*Sterna paradisea*), two puffins (Atlantic *Fratercula arctica*, horned *F. corniculata*) and several species of auklet (parrakeet *Cyclorhynchus psittacula,* crested *Aethia cristata,* least *A. pusilla*).

The position of birds within Arctic marine food webs of the seasonal ice zone has been investigated using nitrogen isotope ratios (Hobson, 1993; Hobson et al., 2002a,b), and fatty acid composition of tissues, as well as more traditional methods involving stomach or throat pouch content analyses (Bradstreet, 1980; Roby et al., 1981; Gaston and Bradstreet, 1993; Lønne and Gabrielsen, 1992; Karnovsky et al., 2008). The large gulls, glaucous gull and Thayer's gull, occupy the highest trophic levels, commensurate with their status as predators of smaller birds and scavengers on the kills of polar bears. The ivory gull also occupies a high trophic level (Karnovsky et al., 2009), presumably because it frequently scavenges marine mammal remains from polar bear kills and sites where people hunt marine mammals. At the opposite end of the spectrum, the smallest auks, little auk and least auklet have a diet composed primarily of copepods. Other seabirds exhibit diets comprising a mixture of large crustaceans (e.g., amphipods, euphausiids) and small fish, especially polar cod. The diet of thick-billed murre has been investigated extensively throughout the circumpolar Arctic. Arctic cod appears to be an important constituent of adult diets as well as making up the bulk of food fed to nestlings (Gaston and Hipfner, 2000). In early summer, as sea ice breaks-up and the murres shift into newly opened areas, birds concentrate at the margins of consolidated pack ice. The pattern of distribution agrees with evidence that these birds feed under, as well as beside the fast ice, making use of under-ice fauna (see Section 9.3.1.2). In the Bering Sea, thick billed-murres feed their chicks a broader diet including juvenile walleye pollock (*Theragra chalcogramma*), squid, and even euphausiids when fish are not available (Sinclair et al., 2008).

Timing of breeding of thick-billed murres at the large breeding colony on Prince Leopold Island, in Parry Channel (Northwest Passage) has been shown to be strongly correlated with the timing of ice break-up, with breeding generally more successful in years of early break-up (Gaston et al., 2005a). The same is true for black-legged kittiwakes, northern fulmars (*Fulmarus glacialis*) and glaucous gulls breeding at the same site (Gaston et al., 2005b).

9.3.2. Ice as a structuring element of polar marine ecosystems

Sea-ice cover greatly affects element cycling and regulates primary production in polar environments. In ice-covered areas, the propagation of light to the underside of the ice is impeded by strong backscattering and attenuation, primarily in the snow cover but also within the sea-ice matrix. The attenuation of light by snow, and to a lesser extent by ice, and by ice algae and other particulate and dissolved material in the ice, influences the amount and spectral composition of light reaching the ice under-surface (e.g., Perovich et al., 1998). This results in very low under-ice irradiances and under-ice phytoplankton biomasses typically lower than 0.5 mg chl$a$/m^3 (e.g., Conover et al., 1999), and has major ecological significance, reflected by a positive correlation between annual primary production and the length of the open-water period in polar and sub-polar regions (Rysgaard et al., 1999; Rysgaard and Glud, 2007).

9.3.2.1. Ice-associated *vs* open-water food webs

Ice-associated food webs in the Arctic are based on intense and localized primary production at the ice–water interface or at receding ice edges (see Section 9.3.1), which starts when solar radiation is sufficient for photosynthesis. This production is concentrated in space and time, although there is a large interannual and regional variability in the timing of ice-algal blooms (e.g., Fortier et al., 2002) and the timing and location of ice-edge blooms. Primary production in the water column (excluding receding ice edges) typically takes place after ice break-up and depends on a balance between stratification and mixing, the former keeping phytoplankton cells within the surface layer where enough irradiance is available for photosynthesis, and the latter fuelling nutrient supply that supports the production (Sverdrup, 1953). While light is a key factor determining the levels of primary production in the ice and open waters, it is ultimately the nutrient supply delivered to the surface waters through tidal-mixing (e.g., Cota et al., 1987), wind-mixing (Tremblay et al., 2006a) or upwelling (Mundy et al., 2009) that determines the maximum biomass produced by the algae and other photosynthesizing organisms. Therefore, early in the season, removal of light limitation by decreasing the snow and ice cover, or via the release of particulate and dissolved substances from the sea ice, allows the onset of a phytoplankton bloom and potential carbon flow to higher trophic levels. This has been shown experimentally by Glud et al. (2007) (Figure 9.31), and has also been observed in polynyas (Tremblay et al., 2002, 2006a) and ice-covered (Fortier et al., 2002) or open-water (e.g., the Barents Sea, Wassmann et al., 2006) regions in the Arctic.

When light is not limiting, nitrogen-based nutrient supply, mainly through vertical mixing, becomes a key factor for sustaining primary production (Tremblay and Gagnon, 2009). For this reason, not all open-water areas sustain high primary production, as stratification and mixing determine nutrient supply in these regions (e.g., the North Water Polynya, see Tremblay et al., 2002). As an example, the former Northeast Water Polynya (no longer considered a polynya due to changed ice conditions) had much lower productivity than the North

Water Polynya due to little vertical replenishment of nutrients in the former area (Kattner and Budéus, 1997; Tremblay et al., 2002b).

It has been suggested that Arctic systems export a greater fraction of primary production to the deeper layers (and ultimately the sea floor) than temperate and tropical locations (Petersen and Curtis, 1980). Seasonal ice melt releases ice algae and mucopolysaccharides that can sink rapidly to the benthos, and a temporal mis-match between phytoplankton production at ice edges and zooplankton grazer populations may also result in the sinking of large quantities of phytoplankton-derived detritus to the sea floor (Tamelander et al., 2008). It is important to note, however, that there is a lot of variability in the quantity and composition of ice-associated material reaching the sea floor, as it depends on a combination of factors including the abundance of zooplankton grazers, the rapidity of melt, and water depth (Fortier et al., 2002; Michel et al., 2006). Benthic communities in seasonally ice-covered areas are typically well-developed as they are tightly coupled to the productivity regime. This is reflected in stocks, diversity and food-web structure (e.g., Grebmeier, 1993; Hobson et al., 1995; Piepenburg, 2005; Carroll et al., 2008; Conlan et al., 2008; Ambrose et al., 2009; Cochrane et al., 2009). Current estimates suggest that one-third to one-half of new production is cycled through the benthos annually on Arctic shelves (Renaud et al., 2008; Blicher et al., 2009). The vertical flux of organic material in open-water areas can also be high, but is often more refractory (see Wassmann et al., 2006, Wassmann, 2008). Consequently, the pelagic-dominated food web of open waters often exhibits higher retention in surface waters and *in situ* degradation by zooplankton and the microbial food web compared to ice-covered regions.

9.3.2.2. **First-year *vs* multi-year ice**

The relative contribution of MYI and FYI to total sea-ice extent has changed dramatically over the past decade (see Sections 9.1 and 9.2). From an ecological perspective, a major difference between FYI and MYI is their transient *vs* permanent nature, respectively. While FYI melts to leave ice-free waters during summer, MYI has traditionally been a 'permanent fixture' within the Arctic that has provided a refuge/habitat for birds, marine mammals, and under-ice fauna (see Section 9.3.1). Differences in the structure and surface characteristics of FYI and MYI (e.g., crystallographic structure, melt ponds, snow accumulation etc., see Section 9.2) are important influences on their role in Arctic food webs. Generally speaking, FYI is much more productive than MYI, although the species diversity of ice algae has been found to be higher in MYI than FYI in the Canada Basin (Melnikov et al., 2002). In addition, melt ponds, which can cover extensive areas of MYI in spring (see Section 9.2), are habitat for a variety of freshwater and marine algal species (Melnikov et al., 2002). Phytoplankton production under MYI is very low, about an order of magnitude lower than in neighboring open-water areas, mainly due to low light intensity underneath the ice (Lee and Whitledge, 2005). By combining production estimates under MYI and in open-water areas, Lee and Whitledge (2005) obtained a revised estimate of annual phytoplankton production of 5 g/m^2, on average, for this region. This estimate does not include the long strands (up to 4–6 m) of *Melosira arctica* that attach to the under-surface of MYI (Melnikov and Bondarchuk, 1987; Booth and Horner, 1997), nor production associated with the presence of melt ponds.

While much of the ecological work in the Arctic has been carried out in areas of FYI, MYI, primarily in the central Arctic
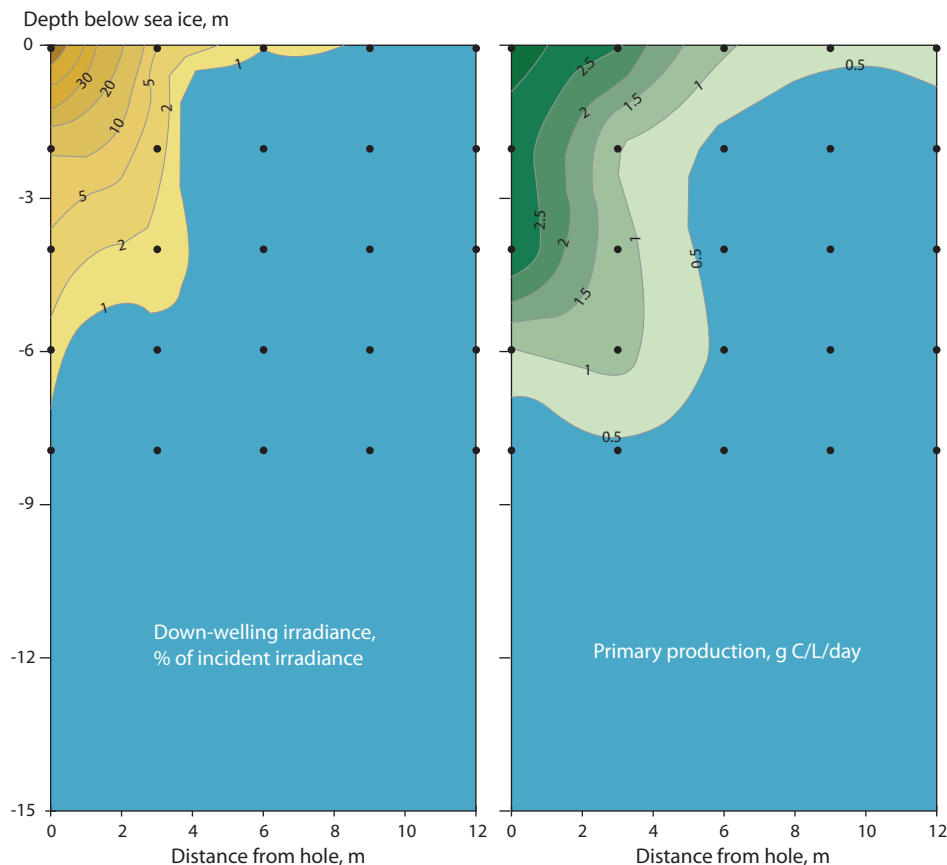
Figure 9.31. Effect of increased irradiance at an ice edge for under-ice production. (a) Scalar irradiance relative to downwelling irradiance at the sea-ice surface as measured around a diver hole (hole size about 1 m^2) and (b) primary production measured at the same positions. Zero '0' indicates the position of the ice edge and dots the points of actual measurements. Source: Glud et al. (2007).

Basin, has been shown to be important for various elements of the ecosystem. These include: the diversity of algal species in melt ponds and in MYI, the under-ice accumulations of mucilagenous diatoms such as *Melosira arctica* (Melnikov et al., 2002), the higher contribution of ice algae to the total annual primary production in areas of MYI (Gosselin et al., 1997; Bluhm and Gradinger, 2008) and the role of MYI as critical habitat for long-lived ice-associated species (e.g., *Gammarus wilkitzkii*; Hop and Pavlova, 2008). The thick, multi-year nature of this ice has, in the past, served as refugia for marine mammals, including polar bears, during summers in years with extensive melt of FYI.

9.3.3. Observations to date regarding ecosystem impacts

Sea ice in the Arctic has already changed markedly. From 1979 to 2011, there has been a significant decline in sea ice extent (-12.0% per decade for September) relative to the 1979–2011 average (see Section 9.1.2.2.1) and a significant loss of MYI has occurred. Sea-ice thickness has also decreased overall, largely due to the reduction in MYI extent and volume as the thickness of FYI appears to have remained fairly constant. Biological impacts related to the decrease in Arctic sea ice are being documented in many regions of the Arctic at a variety of trophic levels.

9.3.3.1. Primary production

A major consequence of the decrease in sea-ice extent is the increase in open-water area, which models suggest will dramatically increase the duration of the phytoplankton growing season, and hence result in increased phytoplankton

biomass and total production (Arrigo et al., 2008; Pabi et al., 2008; see Box. 9.9). Two recent remote-sensing studies indicate that reduced sea-ice cover over the past decade resulted in an increase in phytoplankton biomass and total Arctic primary production, with most change in the Eurasian sector (Arrigo et al., 2008; Pabi et al., 2008). These observations agree with the earlier prediction by Sakshaug (2004) that a positive phase of the AO/NAO would increase total primary production in the Arctic Ocean, especially on the Eurasian side. Alternatively, Grebmeier et al. (2010) reported that the effect of extreme ice retreats and seasonal asymmetry in sea-ice loss on primary production is equivocal in the Pacific Arctic sector, with no clear shift in satellite-derived chlorophyll concentrations for the period 2003 to 2008. These divergent observations can be reconciled within the context of regional differences in sea-ice changes and associated physical forcings, and in the type of ecosystems and their responses to these changes, thereby emphasizing the importance of regionality in observed and predicted changes in the Arctic.

Changes in the size-structure of phytoplankton communities have been documented in Canada Basin, with increasing abundances of small species and decreasing abundances of large species during the past decade (2004 to 2008), attributed to stronger stratification associated with surface melt (Li et al., 2009). Changes in species composition were also documented for ice-associated microalgae living within and under the ice in a study comparing observations in the late 1990s with those for 1979/80 (Melnikov et al., 2002).

The timing of ice melt and retreat is a key structuring element for Arctic marine food webs, and for the reproductive success of marine birds and mammals (see Sections 9.3.1.3 and 9.3.3.4). Fortier et al. (2002) showed that early snow melt and

Box 9.9. The low ice year of 2007

The lowest sea-ice cover on record occurred in 2007 (Overland, 2009). Based on remote-sensing data and a primary production algorithm parameterized for Arctic waters, the estimated annual primary production of 513 TgC/y in 2007 exceeded the 1998–2002 average by 23% (see Figure 9.32; Arrigo et

al., 2008). Total Arctic primary production increased by 35 TgC/y from 2006 to 2007, with the highest increase observed in the Siberian sector of the Arctic. Most (70%) of the 2006 to 2007 increase was explained by a longer phytoplankton growth season, while the remainder (30%) was explained by an increase in open-water area (Arrigo et al., 2008).

-300 -225 -150 -75 0 75 150 225 300
Difference in annual production, g C/m²/y

-100 -75 -50 -25 0 25 50 75 100
Difference in growing season, days
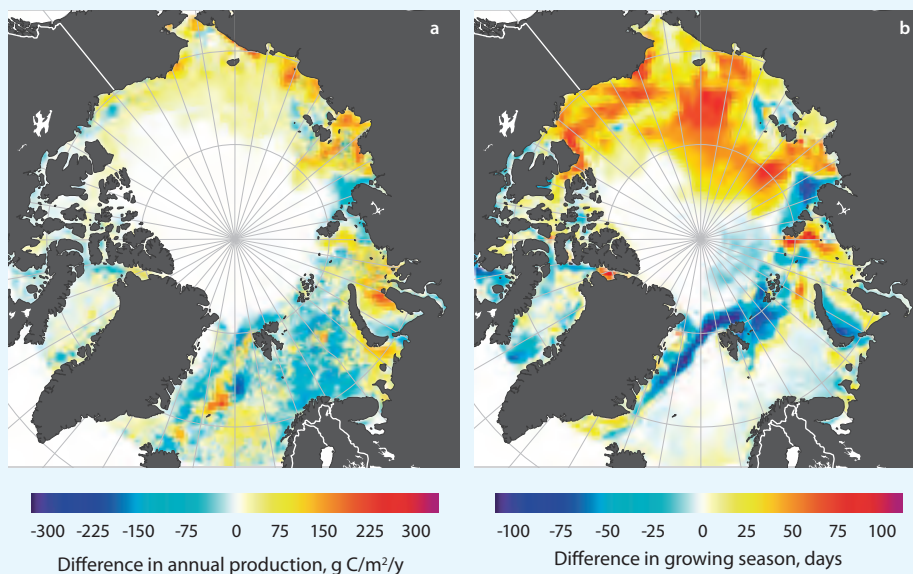
Figure 9.32. Change in (a) annual primary production and (b) length of the phytoplankton growing season in the Arctic between 2006 and 2007. Source: Arrigo et al. (2008).

release of ice algae induced an under-ice phytoplankton bloom and high sinking fluxes of organic material to the benthos, representing a net addition to the annual flux of carbon in the absence of nutrient limitation. Importantly, atmospheric forcings (e.g., rainfall, solar radiation) are instrumental in determining the rapidity of snow and ice melt (see also Section 9.2), and therefore influence the release of material from the ice into the water column, and the timing and amount of under-ice phytoplankton growth (Fortier et al., 2002; Michel et al., 2006). It is estimated that melt began 13 days earlier, on average, in the 2000s compared to the 1980s (Stroeve et al., 2006, 2007) and that, for the entire Arctic, the length of the melt season has increased by about 20 days (Markus et al., 2009).

The timing of ice formation in autumn is also believed to be an ecologically critical element of physical-biological coupling in the Arctic marine system. There is evidence for later ice formation (see Section 9.2) which is likely to influence the incorporation of material in the sea ice during autumn, as well as snow accumulation.

9.3.3.2. Zooplankton and benthic invertebrate communities

Changes in invertebrate community structure, biomass and oxygen consumption have occurred in Arctic seas, especially the sub-Arctic seas, over the past decades (e.g., see Box 9.10). It is difficult to attribute the extent to which these changes are due to warming, sea-ice changes, increased freshwater runoff or to other environmental and biological conditions. Some examples of these changes include declines in ampeliscid amphipod biomass in the Chirikov Basin (Moore et al., 2003; Coyle et al., 2007); declines in infaunal biomass south of St Lawrence Island (Dunton et al., 2005; Grebmeier et al., 2006a,b); increases in intertidal biomass in Svalbard (Weslawski et al., 2010); and increases in epifauna biomass within areas influenced by the Alaska Coastal Current in Norton Sound (Hamazaki et al., 2005) and Kotzebue Sound (Feder et al., 2005), as well as in the (heavily fished) southeastern Bering Sea (Conners et al., 2002). Recent investigations of the annual growth of the bivalve *Clinocardium ciliatum* (Sejr et al., 2009) and the sea urchin *Strongylocentrotus droebachiensis* (Blicher

Figure 9.33. Growth performance of the bivalve *Clinocardium ciliatum* estimated as maximum annual shell growth plotted against the productive open-water period. Data for the sea urchin *Strongylocentrotus droebachiensis* recalculated from Blicher et al. (2007) are included. Linear regression for the combined dataset of both species: $G_{max} = 0.24 \times [POWP] + 19.4$, $p < 0.01$, $R^2 = 0.78$. Source: Sejr et al. (2009).

et al., 2009) along a climate gradient from the sub-Arctic to high-Arctic showed a significant positive correlation between the growth of these benthic species and the average productive open-water (ice-free) period (Figure 9.33). Furthermore, local variations in growth correlated well to local variations in sea-ice conditions. These results indicate that sea-ice cover influences growth rates of benthic fauna, probably through light limitation of primary production (Sejr et al., 2009). At sites with pronounced sea-ice cover, food availability may be linked to sea-ice dynamics through the bottom-up regulation exerted by sea ice on phytoplankton production, which renders such areas especially susceptible to future climate change.

Benthic community changes, such as switches from long-lived, slow-growing Arctic to faster-growing temperate species can be expected in the southern areas of the Arctic. Northern range extensions, possibly due to climate change, have recently been documented in the Chukchi Sea for some epifaunal megafauna (Sirenko and Gagaev, 2007). Other species shifts have been noted for bivalves (Richman and Lovvorn, 2003). On a pan-Arctic scale, concentrations of the commercially valuable snow crab (*Chionoecetes opilio*) appear to have shifted northward in the Bering and Chukchi Seas (Orensanz et al., 2004; Bluhm et al., 2009b), and in the European Arctic a self-sustained population now occurs in the Barents Sea (Alvsvåg et al., 2009) where the first snow crabs were documented in 1996 (Kuzmin et al., 1998). These shifts are not only potential ecological indicators of climate warming, but also affect the current sub-Arctic crab fishery and possibly other Arctic fisheries in the future.

In the fjords of Svalbard, only minor changes have been observed in soft-sediment benthic communities over the past 10 to 20 years, despite evidence of intense warming of fjord

Box 9.10. Climate driven variability in the Bering Sea

A recent, well-documented example of climate-driven variability is the Bering Sea regime shift complex (Grebmeier et al., 2006a). The regime shift that took place in 1997 included changes in sea-surface temperature, a reduction in sea-ice extent and wind forcings of ocean currents. Since that time, several coccolithophorid blooms have occurred in the Bering Sea. These are thought to have replaced the previously occurring summer flagellate community (Schumacher et al., 2003). Zooplankton communities have changed from crustacean-dominated, to a dominance of gelatinous plankton (which subsequently decreased again). The prey of the dominant jellyfish species during the late 1990s and early 2000s, *Chrysaora melanaster*, included euphausiids, copepods and amphipods, as well as juvenile pollack and hence overlapped with some Arctic marine mammals (Bluhm and Gradinger, 2008).

waters during this period (e.g., Renaud et al., 2008; Kedra et al., 2010). The presence of a sill within some fjords has been suggested to determine the influence of the warmer shelf waters on the biota of these areas to some extent. A retrospective study of decapod crustacean communities during investigations spanning 100 years has shown an increase in generalist feeders at the expense of specialist shrimps, but without a direct causal link to climatic fluctuation (Berge et al., 2009).

9.3.3.3. **Fish species and fisheries**

Loss of Arctic sea-ice cover and regional warming will have significant impacts on fish species of ecological and economic importance. Across the Arctic, the planktivorous capelin and the omnivorous polar and Arctic cod play vital roles in the transfer of pelagic production from zooplankton to higher trophic levels such as marine mammals and seabirds. These fish derive much of their high lipid con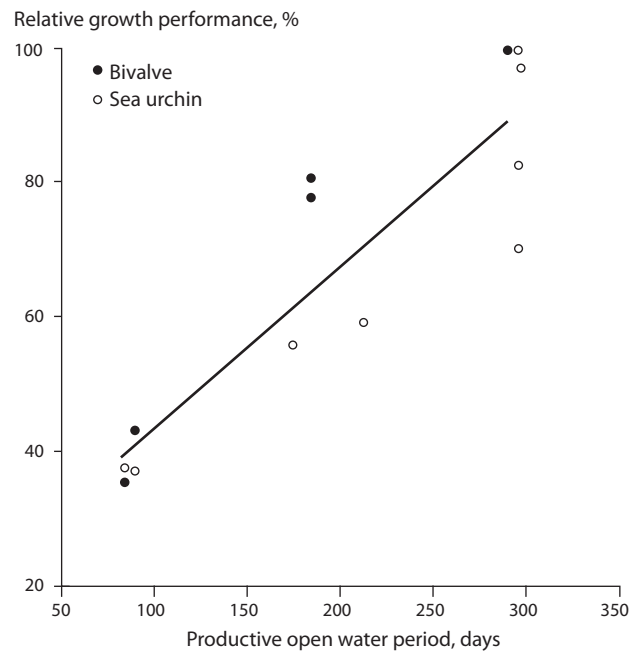tent from lipid-rich Arctic zooplankton species (e.g., *Calanus glacialis, Themisto libellula*), and it is this fat content that is critical for sustaining and fueling reproduction in their predators. In the Barents Sea, these fish species are highly dependent on the roles of sea ice and shore-fast ice as habitat and as a driver of planktonic processes, although it should be noted that capelin do live in the Gulf of Alaska and other areas where there is no FYI or MYI.

In their seasonal migrations, capelin follow the receding ice edge northward, feeding on developing zooplankton populations which themselves are taking advantage of intense but ephemeral ice-edge algal blooms. Population size and feeding success of capelin depend on many environmental and ecological factors; water temperature and seasonal ice distribution are important as they affect the degree of match or mismatch in timing of capelin arrival in an area and the availability of food (e.g., Gjøsæter, 1998; Orlova et al., 2002, 2005, 2009). Polar and Arctic cod are also dependent on zooplankton food and the influence of ice on developmental phenology of zooplankton will also influence these species (Lønne and Gulliksen, 1989; Hop et al., 1997; Fortier et al., 2006; Orlova et al., 2009). Flexible life-history strategies by polar cod, however, may help them persist with earlier melting of sea ice and increased warming of Arctic waters (Fortier et al., 2006). At least during some stages of their life, these fish may be closely associated with sea ice, using the irregular underside of the ice as refuges from predators and as occasional feeding areas. It is unclear how important this use of the sea-ice habitat is to persistence of the species. It is clear, however, that success of forage fish such as capelin and Arctic cod is critical for the commercial fish stocks that feed on them. Atlantic cod *Gadus morhua* feed heavily on planktivorous fish in the Barents Sea; and it is the availability of lipid-rich Arctic copepods that results in optimum body condition of the forage fish (e.g., Gjøsæter et al., 2002; Orlova et al., 2009).

Changes in planktonic community structure due to warming of Arctic waters are likely to negatively impact capelin and polar/Arctic cod as the fat-rich Arctic zooplankton are replaced by boreal species (Beaugrand et al., 2010; Hopcroft et al., 2010). This interaction between climate, zooplankton community composition, and condition of planktivorous fish is a complex but important ecosystem dynamic. For example, young walleye pollock in the Bering Sea grow more slowly but accumulate sufficient energy reserves to overwinter during cold years, when larger *Calanus* and euphausiids are abundant (Coyle et al., 2008; Hunt et al., 2008). During warm years, however, small neritic copepods dominate the zooplankton and, while providing resources for rapid growth of early life stages of pollock, result in age-0 pollock with energy stores too low to survive their first winter (Moss et al., 2009). Furthermore, lack of larger zooplankton in warmer years may make young walleye pollock more susceptible to predation (Farley et al., 2007; Farley and Trudel, 2009). Thus, climate impacts on lower trophic levels may propagate through the ecosystem and eventually affect seabirds, marine mammals, and commercial fish species (e.g., Karnovsky et al., 2003). Finally, northward expansion of potential competitors and predatory species could also have negative impacts both on forage-fish populations and the benthic fauna on which walrus, bearded seals and some sea ducks forage (Grebmeier et al., 2006a). A northward shift in boreal species is widely predicted (ACIA, 2005), and has already been observed in the North Pacific and in the Bering and Barents Seas (e.g., Hare and Mantua, 2000; Mueter and Litzow, 2008; Drinkwater, 2009). Whether these species successfully reproduce in Arctic waters (i.e., true expansion shifts) will affect their long-term ecological role. For example, *C. glacialis* advected from the Bering Sea into the Arctic Ocean appears to play only a transient ecological role as their viability declines as they penetrate further into Arctic waters (Nelson et al., 2009).

Target species of commercial and subsistence fisheries in the Arctic are also likely to be impacted by loss of sea ice and warming. Atlantic cod and Greenland halibut are important species in the European Arctic, including around Greenland. The former is a boreal species that has expanded into northern Greenland and Barents Sea waters during historical warm periods (see Renaud et al., 2007, and references within), and has been observed in increasing densities in west Svalbard fjords since 2004 (Berge et al., 2008). Reduction of ice cover and warming of waters in these northern regions may lead to greater populations and enhanced fisheries value of this species (Drinkwater, 2005, 2006), but it is unclear how the channeling of high levels of pelagic production to this top predator will influence food supply for seabirds and marine mammals. Greenland halibut, on the other hand, is an Arctic species and could be expected to decline substantially in response to rising temperatures, as it has during warm periods in the Bering Sea (NPFMC, 2009). A strong positive AO/NAO phase (meaning warmer conditions in the Barents/Norwegian Sea) between 1989 and 1994 corresponded with very low stock and catch values for Greenland halibut (Albert and Høines, 2003).

9.3.3.4. **Marine mammals and seabirds**

The earliest warning signs of climate change in the Arctic gave rise to immediate concern for the potential impacts on marine mammals of the region (Stirling and Derocher, 1993; MacGarvin and Simmonds, 1996; Tynan and DeMaster, 1997; Carmack and McLaughlin, 2001; Kelly, 2001). There is particular concern for 'ice-loving' Arctic marine mammals (e.g., ACIA, 2005; Johnston et al., 2005; Laidre and Heide-Jørgensen, 2005b; Moore and Laidre, 2006; Simmonds and Isaac, 2007; Kovacs and Lydersen, 2008; Laidre et al., 2008; Moore and

Huntington, 2008). Loss of sea ice represents a reduction in total habitat area available to ice-associated marine mammals that is already affecting some species, and in the longer term, it is expected that foraging success, fertility rates, and mortality rates etc. will be impacted for additional populations and species (see Section 9.3.4). Theoretically speaking, specialist feeders are likely to be more heavily impacted by changes in Arctic food webs than species that are generalist feeders, and ice-breeders that require long periods of stable ice in late spring are likely to be impacted more rapidly than late winter ice-breeders that require ice for shorter periods of time (Kovacs and Lydersen, 2008; Laidre et al., 2008).

Alterations in distribution would be expected to be among the first changes observed due to changes in ice extent and seasonal availability. Vibe's (1967) analysis of the relationship between multi-decadal environmental fluctuations and marine mammal harvest success in West Greenland was one of the first studies that explored decadal-term patterns of distribution of Arctic marine mammals in relation to ice type and distribution. Vibe (1967) clearly showed the importance of availability of specific ice types in determining the distribution of marine mammals of the region. More recent analyses of decadal patterns of sea ice and their influence on marine mammals suggest that changes in recent years are likely to impact resident marine mammal populations at regional and hemispheric scales (Barber and Iacozza, 2004). While northward expansions are possible for temperate species, for example grey whales (*Eschrichtius robustus*) in the Pacific sector (Stafford et al., 2007; Moore, 2008a,b), high-Arctic endemics have limited capacity for range shifts northward. But, changes are clearly taking place. Vast herds of harp seals have been observed off western Greenland and eastern Svalbard during the period from January to March, times of year that they are not normally in these regions (Rosing-Asvid, 2008; Norwegian Polar Institute Marine Mammal Sighting Data Base) and a new harp seal whelping ground has been discovered off southern Greenland (Rosing-Asvid, 2008). Grey seals have also been conclusively documented in southern Greenland as of summer 2009, which is a notable range expansion northward for this species (Rosing-Asvid, 2010). The degree to which animals are capable of 'relocation' of breeding areas is unknown; tradition plays a significant social role in being in the right place at the right time for many ice-breeding species that disperse broadly outside the breeding period.

In the Canadian High Arctic, killer whales appear to be extending their season of Arctic habitation and expanding their range northward, at least in periods when ice conditions permit (e.g., Ferguson, 2009; Higdon and Ferguson, 2009). Early season sightings of this species have also occurred in recent years at very high latitudes in the Northeast Atlantic (early March at above 80° N, Norwegian Polar Institute Marine Mammal Sighting Data Base), although the lack of comparative historical data limits interpretation of these sightings. Other anecdotal data suggest that some temperate cetaceans are shifting their summer distributions northward. For example, sei whales (*Balaenoptera borealis*) and harbor porpoises (*Phocoena phocoena*) have been seen in Svalbard waters in recent years for the first time, and unusually high densities of fin whales (*B. physalus*) and minke whales (*B. acutorostrata*) have occurred in July and August along the shelf break west

of Spitsbergen recently (Norwegian Polar Institute Marine Mammal Sighting Data Base). Blue whales (*B. musculus*) are also being sighted with some regularity near Svalbard, although population increases, rather than range shifts due to changing ice conditions, might be the reason for the increased incidence of this species in this area (Norwegian Polar Institute Marine Mammal Sighting Data Base).

Major declines in abundance or pup production have recently been documented for hooded seals in the Northeast Atlantic, harp seals in the White Sea and ringed seals in Hudson Bay, which have been attributed largely to climate change impacts on ice conditions (Ferguson et al., 2005; Chernook and Boltnev, 2008; ICES, 2008). Low ice years have also been correlated with low body condition indexes for female ringed seals and ovulation rates as low as 50% of the norm in extreme low ice years (Harwood et al., 2000). Ice-seals in other areas are also likely to be experiencing declines in reproductive success, and perhaps also condition, but lack of monitoring activities make it impossible to assess the current extent of change. For example, ringed seals on the west coast of Svalbard have not had sufficient ice for normal breeding levels to occur since 2005 (Kovacs and Lydersen, Norwegian Polar Institute, pers. obs.) and are likely to be in decline. In western Hudson Bay, a clear link was established between climate warming, reduced ice presence, and observed declines in polar bear abundance and reductions in their physical and reproductive condition (e.g., Stirling et al., 1999) and similar finding have now been reported from other regions (see Box 9.11). Alternatively, there has been no dramatic change to bowhead whale distribution or migratory timing in the Pacific Arctic sector. The Bering-Chukchi-Beaufort population has increased at a rate of 3% per year since the late 1970s (George et al., 2004), with indications from calf counts and harvested whale body condition that reduced sea ice in the eastern Beaufort Sea has (at least for the time being) actually been good for this population (George et al., 2006; IWC, 2010).

A growing body of data suggests that Pacific walruses are also showing negative impacts of sea-ice reductions (http://alaska.usgs.gov). Abandoned calves have been reported at sea (Cooper et al., 2006), which suggests that females with dependent young might be experiencing nutritional stress with their usual sea-ice resting platform retracting during summer north of the continental shelf, separating them from feeding areas. Mothers and calves are certainly spending more time on land (see Arnbom, 2009), where stampede incidents have recently caused significant mortality of both adults and young animals (from hundreds to thousands of individuals) (e.g., Fischbach et al., 2009). There are also suggestions that Pacific walruses have shifted their diet toward eating more seals and fewer benthic invertebrates (Rausch et al., 2007; also see Gjertz, 1990; Sheffield et al., 2001, but also see Sheffield and Grebmeier, 2009), which may lead to higher prevalence of trichinellosis (Rausch et al., 2007); this parasite has shown marked increases in the past decade in ice-associated marine mammals in Svalbard (Jensen et al., 2010). However, the magnitude of the impact of this dietary shift by walruses in the Pacific is difficult to assess. Some harvest data show that the proportion of females in the catch has increased while the relative proportions of pregnant females have declined and the age of first reproduction has shifted (Chad Jay, U.S. Geological Survey, pers. comm.). These changes are suggested to be related to harvest

Box 9.11. Sea ice decline and polar bears

Polar bears are dependent on sea ice as their primary hunting platform; their diet is predominantly made up of ice-associated seals. Most polar bears travel thousands of kilometres over sea ice searching for mates and hunting for seals each year. In one of the southernmost populations of polar bears, in western Hudson Bay, changes in the date of break-up of the sea ice were suspected in the early 1990s (Derocher et al., 2004). Early ice-break up and delayed freeze-up, has resulted in a shorter sea-ice season in the region which has caused the bears to spend more time ashore resulting in reductions in condition, reproductive rates, survival, and population size (Stirling et al., 1999, 2004; Parks et al., 2006; Stirling and Parkinson, 2006).

Litter production rate and natality declines, as well as reductions in body length, have also been detected in Svalbard polar bears, which show a relationship to large-scale climate variation (Derocher, 2005). However, in this region density-dependent factors cannot be ruled out as being part of the cause of these changes (see Aars et al., 2009). Reduced prey availability for polar bears due to changing ice conditions has been suggested to be the case for bears in eastern Greenland, where bears are now smaller than they were some decades ago; although contaminant increases may also be playing a role in the observations in this region (Pertoldi et al., 2009). Most recently, polar bears in the Beaufort Sea have exhibited reduced breeding rates, lower cub litter survival and reduced adult survivorship, that are correlated with increasing duration of the ice-free period (Regehr et al., 2006, 2007, 2010). These conclusions follow earlier reports from this area of cannibalized and starved bears and suggestions that Beaufort Sea bears were periodically nutritionally stressed during the early part of this century because of increasing duration of the open-water season and possible decreases in seal populations, or their accessibility to bears (Stirling et al., 2008a).

Polar bear distribution has shifted in the Alaskan Beaufort Sea in recent decades, with more bears being associated with land and open water, particularly in early autumn. These changes are thought to reflect a behavioral response by polar bears to changes in ice type and cover and the timing of ice formation and ablation (Gleason and Rode, 2009). Polar bear denning locations have shifted in some regions in response to changing ice conditions, with fewer dens in MYI and more ashore (e.g., Fischbach et al., 2007). The tendency for bears to spend more time ashore (e.g., Schliebe et al., 2008) has the potential for an increase in the number of human-bear interactions (e.g., Towns et al., 2009) and may give false impressions of regional densities (Stirling and Parkinson, 2006). In more southerly areas such as western and southern Hudson Bay, it also puts denning bears at risk of new threats such as forest fires (Richardson et al., 2007).

Polar bears are capable of opportunistically altering their foraging behavior to take advantage of locally abundant prey, and to some degree compensate for a decline in a dominant prey species (Thiemann et al., 2008). Dietary shifts have been documented in the Hudson Bay polar bear population, with fewer ice-seals and more open-water seals in the diet of bears over the period 1991 to 2007 (Mckinney et al., 2009) and an increased consumption of terrestrially-based food such as large quantities of goose eggs (Smith et al., 2010). Although polar bears will consume a wide variety of food types, it seems unlikely that current population sizes could be maintained on alternative dietary energy sources if the diet does not include significant quantities of marine mammals, which are normally accessed from a sea-ice platform (see Figure 9.34; also see Hobson et al., 2009; Rockwell and Gormezano, 2009; and Dyck and Kebreab, 2009, for preliminary discussions of this topic).

Failures in polar bear management systems to respond to population changes in a timely manner have resulted in severe overharvesting in Baffin Bay and Kane Basin, populations which are shared by Canada and Greenland (e.g., Taylor et al., 2008), and highlight the need for responsive management procedures for this and other species under the currently rapidly changing environmental conditions (Wiig et al., 1995; also see Clark et al., 2008; Dowsley and Wenzel, 2008; Dowsley, 2009). The observed negative consequences of sea-ice declines on polar bears to date, and future sea-ice predictions (see Section 9.3.4), led to the species being listed on the IUCN (International Union for Conservation of Nature) Red List and the Endangered Species Act of the United States, in 2008.

Figure 9.34. Polar bears stranded on land during summer often have difficulty getting enough food to restore their fat reserves for the following winter period (left). Polar bears normally use sea ice as their primary hunting platform, where they maintain good body condition on their diet of ice-associated seals (right). Photos by Kit M. Kovacs and Christian Lydersen, Norwegian Polar Institute (left) and Hans Wolkers, Norwegian Polar Institute (right).

management and changing environmental conditions (Garlich-Miller et al., 2006). Atlantic walruses might be somewhat more resilient to changes in sea ice because they utilize more nearshore areas for feeding (Born, 2005), but less extensive seasonal ice coverage is almost certain to have a long-term impact on benthic food resources that walruses of both subspecies depend upon and if walruses are spatially restricted because of distance to suitable haul-out areas, abundances are likely to decline (Kovacs and Lydersen, 2008).

Changes in the harvests of several species have been noted with respect to changing sea-ice conditions in the North Atlantic Arctic. West Greenland harvests of harp seals have increased 100-fold for adults and 10-fold for juveniles in the past 20 years; in the 1980s only some hundred adults were taken and fewer than 2000 young animals, this has increased to in excess of 20 000 of each age group in the period 1995 to 2005 (Rosing-Asvid, 2008). Given the size of the stocks, this does not represent a specific management problem in this case. However, other increased catches might be more significant from a conservation point of view. Increases in narwhal catches by hunters in Siorapaluk, Greenland, have taken place since 2002 due to changed sea-ice conditions, with lighter ice coverage allowing easier access for hunters (Nielsen, 2009). This increased harvest has unknown implications because of a lack of information about the status of this stock; both scientific surveys and ecological knowledge suggest that it is an independent stock unit, but there is insufficient information to provide management advice. In the rapidly changing conditions in the Arctic, it is essential that management systems be provided with up-to-date stock status, and that they have the capacity to respond quickly to changes (see Section 9.3.5).

Timing of ice break-up is known to have a strong effect on the timing and success of breeding for thick-billed murres in northern Hudson Bay, where early ice years tend to be sub-optimal for chick rearing (Gaston et al., 2005a; see Box 9.12). For black guillemots in the Beaufort Sea, reduction in ice

cover during summer has been linked to reproductive failure due to a reduction in the availability of Arctic cod when sea ice is far from the shore (Moline et al., 2008). A reduction in sea ice in the Bering Sea during the late 20th century affected the timing of breeding of seabirds and has been implicated in population declines of murres and kittiwakes at St Paul Island (Byrd et al., 2008a,b). In general, during warm years with little ice, planktivorous species seem to be more challenged than piscivorous species in finding food (Kitaysky and Golubova, 2000; Stempniewicz et al., 2007; Benowitz-Fredericks et al., 2008). In all the areas mentioned above, immigration has been observed by auk species that normally have more southerly distributions (horned puffin, razorbill *Alca torda*, ancient murrelet *Synthliboramphus antiquus*).

Conversely, in more High Arctic areas, such as Barrow Strait, early ice break-up has been associated with earlier breeding and enhanced reproductive success for black-legged kittiwakes, glaucous gulls and thick-billed murres (Gaston et al., 2005b). Thick-billed murre populations have shown high rates of population increase during periods of moderate climate warming in many areas (Irons et al., 2008). However, the population trend for this species throughout northern regions in Norway, including the High Arctic archipelago of Svalbard, is currently declining, despite a general rise both in air and water temperatures in the region (Lorentsen and Christensen-Dalsgaard, 2010). Populations of black-legged kittiwakes are increasing in the Canadian High Arctic (Mallory et al., 2009) possibly as a result of enhanced reproductive success due to earlier ice break-up. The cause of the rapid decline in ivory gulls in the Canadian High Arctic and possibly elsewhere, since the 1970s (Gilchrist and Mallory, 2005) is unknown. However, the core range of the remaining birds in the Canadian population has retreated northward, suggesting that, irrespective of the cause of the population trend, the remaining birds are concentrating in areas of prolonged summer ice cover (Gilchrist and Mallory, 2005).

Box 9.12. Breeding and feeding of thick-billed murres in Hudson Bay

In northern Hudson Bay, synchronization of breeding with the timing of ice break-up appears to be important in determining breeding success of thick-billed murres. Between 1988 and 2008, the timing of sea-ice break-up advanced by about 17 days. Over the same period, the timing of breeding by the murres has advanced by less than a week, apparently causing a mismatch between the time at which chicks hatch and the optimal timing with respect to food availability (Gaston et al., 2009).

The diets of thick-billed murre nestlings at Coats and Digges Island have been monitored since the 1980s. In the early years, the diet at both colonies was dominated by Arctic cod, but since the 1990s, capelin has made up a much larger share of the diet, up to 70%, at Coats Island (Figure 9.35; Gaston et al., 2003). At Digges Island, where Arctic cod comprised more than 80% of the diet in the early 1980s, capelin made up 85% of the nestling diet in 2004. Arctic cod being an ice-associated fish, these changes may be linked to progressively earlier break-up of sea ice in Hudson Bay since the early 1990s (Stirling and Parkinson, 2006).
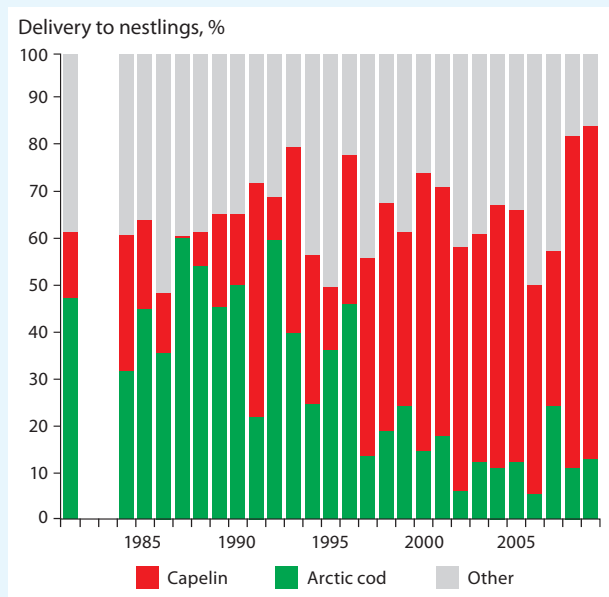
Figure 9.35. Proportion of Arctic cod and capelin fed to nestling thick-billed murres at Coats Island between 1981 and 2009 (no data for 1982 and 1983). Source: Gaston et al., (2003) and A.J. Gaston (Canadian Wildlife Service, unpubl. data).

9.3.4. **Looking into the future**

By their very nature, predictions are speculative, but given the potential for major change in the Arctic due to sea-ice decline, it is important to prepare mitigation strategies (see Section 9.4) and management plans for flora and fauna with the best available models and expert predictions (see Table 9.3 for a summary). The physical models that predict sea-ice extent still contain a lot of variability (see Section 9.1), but all of the models converge to predict a – more or less rapid – decline in September sea-ice extent during this century. Observations of sea-ice extent and thickness (see Section 9.1) indicate that conditions are moving toward a seasonally ice-free Arctic more rapidly than GCMs (Section 9.1) and high resolution coupled ocean-sea ice-atmosphere models (Section 9.2) have predicted. Current estimates suggest that the Arctic may be free of MYI somewhere during the period 2013 to 2030 (Wang and Overland, 2009). Recent results by Barber et al. (2009) and Kwok et al. (2009) support the earlier part of this temporal range. The variability in model results underscores the challenges posed by the complexity of the Arctic climate-ocean system, including variability in large-scale and regional circulation patterns and feedback processes between the atmosphere and ocean.

In Arctic ecosystems, food-web structure dictates much of the fate of organic matter, as reflected in differences in biogeochemical cycling regimes at various spatial and temporal scales. Therefore ecosystem responses to changes in physical forcings are neither direct nor linear; they are therefore difficult to predict with confidence. Additionally, the impacts of changes in Arctic sea ice, which are the focus of this review, cannot be considered in isolation from other changes taking place in the Arctic. Superimposed on sea-ice associated changes in light availability, stratification and mixing, timing of ice formation/melt, and ice-edge position, other major changes such as increased riverine input, increased sea-surface temperature, increased air temperatures, melting and receding glaciers, changes in precipitation and atmospheric forcings all contribute to modify Arctic marine ecosystems. In addition, ecosystem responses will vary regionally, depending on different regimes of oceanic and climatic forcings but also in relation to regional variation in bathymetry and variation within the biological communities themselves, including pivotal species and carbon flow pathways. Based on variability observed in the past which is high both on interannual and decadal scales in the Arctic (e.g., Barber and Iacozza, 2004), and with some background studies such as those reporting the impacts of cold and warm years in the Barents Sea (Orlova et al., 2009), and in the Bering Sea (Hunt et al., 2008), predictive scenarios for the more extreme conditions that are forecast for the coming decades can be extrapolated with modest levels of certainty (e.g., Box 9.13).

Table 9.3. Some key potential biological impacts in response to decreasing sea-ice extent and lengthening of the open-water period in the Arctic. Likelihood (very unlikely, unlikely, possible, likely, very likely) and confidence (low, medium, high) are defined according to ACIA (2005).

Predicted impact	Likelihood of prediction	Confidence in prediction
Increase in phytoplankton production	Very likely	High
Increase in total primary production	Very likely	Medium
Increase in benthic primary production on shelves	Likely	High
Increase in sedimentation of organic carbon in areas of increased production	Very likely	Medium
Advance in the timing of the phytoplankton bloom on shelves	Very likely	Medium
Development of an autumn phytoplankton bloom in areas where the open-water period is over five months long	Possible	Low
Decrease in ice-algal production on shelves	Likely	Low
Increase in ice-algal production in central basins	Very likely	Medium
Increase in bacterial respiration and remineralization of organic matter	Very likely	High
Increase in nutrient supply through wind mixing and upwelling on shelves	Very likely	High
Disruption in zooplankton life-histories	Very likely	Medium
Decline in populations of MYI faunal taxa	Very likely	Medium
Decline in feeding conditions for benthos and demersal species in nearshore areas	Likely	Low
Shift in the abundance of pivotal copepod and fish species	Very likely	High
Change in diet of ice-associated birds	Likely	High
Shift in distribution of marine birds	Likely	High
Northward shift in breeding distributions of pack-ice breeding pinnipeds	Very likely	Medium
Declining abundance of ice-associated endemic marine mammal populations	Very likely	High
Increase in abundance of temperate migrant marine mammals during summer (and longer resident times in the Arctic)	Very likely	High
Dietary shifts for migrant baleen whales in Arctic water	Likely	High
Increased mortality of Arctic endemic species due to killer whale predation in some areas	Very likely	High
Northward spread of diseases and parasites of marine mammals and other fauna	Very likely	High

Box 9.13. Case study: Effect of advanced melt in the Canadian Archipelago on primary production

A recent high resolution ice-ocean regional model for the Canadian Archipelago (Sou and Flato, 2009) predicts spring ice break-up to occur ten days earlier, on average, in the period 2041 to 2060. Regionally, ice break-up is predicted to be up to two to five weeks earlier in northern Baffin Bay, Nares Strait and the Amundsen Gulf. Bringing together published data on the typical development of the ice-algal bloom and its sinking export, and of the summer phytoplankton bloom in the Archipelago, makes it possible to draw a conceptual model of the effects of a two- to five-week earlier melt period. Advancing ice melt by a two-week period does not affect the onset of the ice-algal bloom but does shorten the duration of the bloom. The direct consequences are a decrease in the overall production and possibly biomass accumulation of ice algae, and an earlier and smaller release of ice algae biomass from the ice. Assuming that zooplankton grazers would still be synchronized with the availability of ice algae, this simple model indicates that ice–pelagic coupling (i.e., flow of carbon toward pelagic grazers) would be favored at the expense of ice–benthic coupling since little biomass would be available for sinking after zooplankton grazing.

Under the scenario of a five-week advance in ice melt, melt occurs prior to the development of the ice-algal bloom, shunting carbon flow pathways from sea ice production to the pelagos and benthos. Under such a scenario, the potential development of an ice-algal bloom would basically depend on the availability of light early in the season, determined by latitude and by conditions at the snow-ice surface. As many climate scenarios predict an increase in precipitation in the Arctic (ACIA, 2005), it is very likely that a five-week advance in ice melt north of 70° N would completely remove the potential for an ice-algal bloom to take place.

This case study illustrates one current conceptual model for how Arctic shelves may respond to decreased sea ice, that is, by shifting from the current benthic-dominated system to one where more energy is directed toward pelagic food webs (Carroll and Carroll, 2003). There is some evidence that this may be occurring in the Bering Sea (Hunt et al., 2002, 2008; Grebmeier et al., 2006a; Mueter et al., 2009), but this region is not necessarily representative of the entire Arctic. Differences in physical drivers, trophic structure, and biotic components in different regions must be considered carefully before extending predictions to other Arctic shelves (see Tremblay et al., 2008). For example, recent modeling studies using the latest climate forecasts have predicted modest increases in annual primary production in the Barents Sea, but decreased grazing pressure, perhaps leading to enhanced food for the benthos (Ellingsen et al., 2008). The apparent requirement for a specific phenology in the primary production regime for zooplankton development (Baier and Napp, 2003; Søreide et al., 2010) also suggests that a change in timing of blooms may create a mismatch with grazer populations, and perhaps more export of algal material (see Figure 9.36, see also Michel et al., 2006).
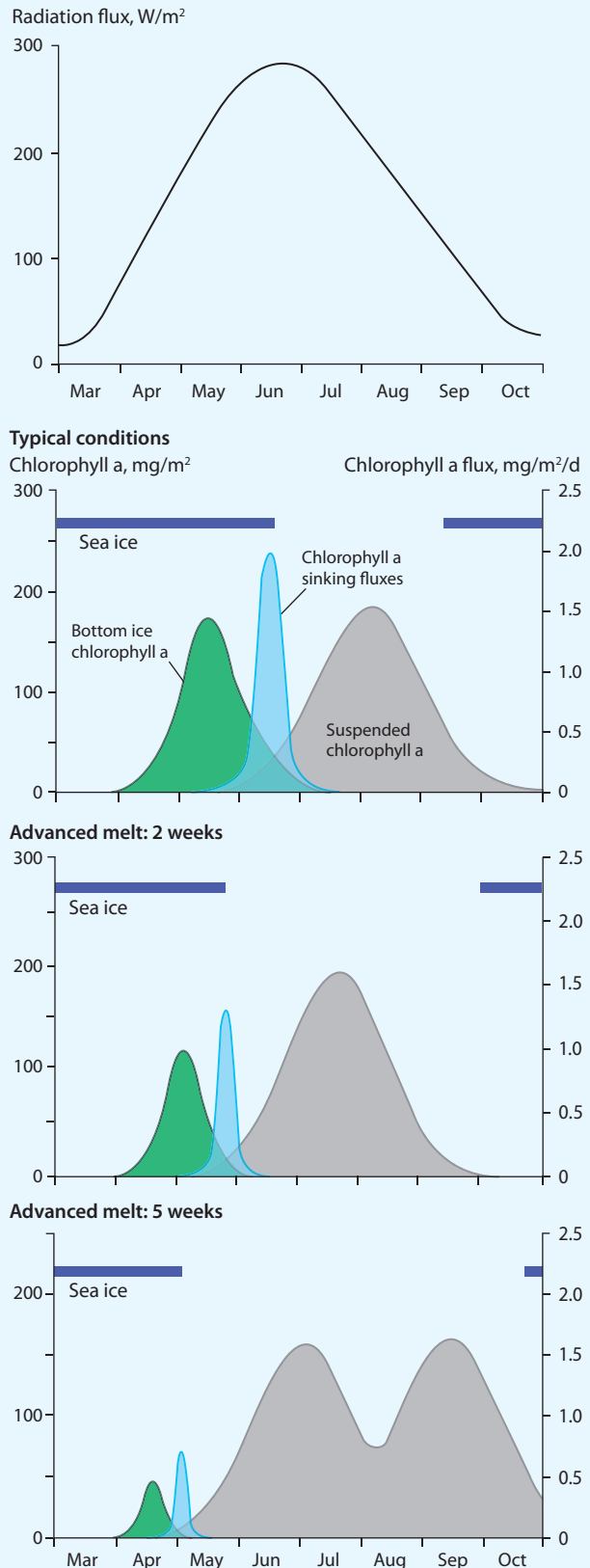
Figure 9.36. Present and predicted trends in bottom ice chlorophyll a (chl a), chl a sinking fluxes and suspended chl a concentrations in the Canadian Archipelago (Resolute Passage) for scenarios of a two-week and five-week advanced melt period (Sou and Flato, 2009). The trend in suspended chl a is from Gaussian regression adjusted to all data points in Figure 7 of Michel et al. (2006); trends for bottom ice chl a and chl a sinking fluxes are adjusted to data collected in 1992, adapted from Michel et al. (1996).

9.3.4.1. Change in seasonality and lengthening of the open-water period

Owing to the overriding influence of light on primary production (in the absence of nutrient limitation), it is predicted that an increase in the duration of the open-water period will increase phytoplankton production. It is also likely that the lengthening of the open-water period in spring, thus far seen only in the Atlantic sector, will advance the period of phytoplankton production (Wiktor, 1999; Rysgaard and Glud, 2007). In autumn, if freeze-up is delayed (predicted to occur two to four weeks later in the Canadian Archipelago; Sou and Flato, 2009), a second phytoplankton bloom may take place, if it is fuelled by nutrient replenishment through vertical mixing (Figure 9.36). This has been observed in the North Water Polynya where the open-water period extends over about six months (Tremblay et al., 2006b; Garneau et al., 2007), and is similar to boreal regions. The late formation of FYI can also lead to increased dynamic processes creating increased rubbling, ridges and other forms of surface deformation. This could increase catchment of snow on top of FYI, thereby limiting photosynthesis at the ocean-ice interface (Barber et al., 2008a).

However, since nutrient inventories and the nutrient supply to surface phytoplankton ultimately determine the maximum phytoplankton yield and the biomass available to the food web, a combination of factors, and not just the duration of the open-water period, will interact to determine the extent (and location) of increased phytoplankton production (e.g., Tremblay and Gagnon, 2009). Higher freshwater input from sea-ice melt and terrestrial runoff, together with warming of the surface layer will increase surface stratification, thereby possibly offsetting the potential for increased nutrient supply from deep waters to the surface layer due to increased storms and wind mixing in open-water areas. On the shelves, allochthonous input of nutrients and denitrification contribute to nutrient inventories and fluxes to primary producers. Additional factors such as increased transport and resuspension of dissolved and particulate material, including sediments, modify the quantity and spectral quality of light available to primary producers on shallow shelves. Such modifications in the physical environment are likely to trigger changes in the species composition of phytoplankton communities and different productivity regimes and pathways of cycling of organic matter.

9.3.4.2. Shift from multi-year to first-year ice

According to model predictions (see Section 9.1), FYI will become the dominant ice type in the Arctic over the 21st century. The Arctic has already lost a very significant proportion of its MYI (Barber et al., 2009; Kwok et al., 2009) and is poised to lose the remaining MYI in the near future if the strength and direction of current feedbacks remain unaltered (see Section 9.2).

The biological impacts associated with a shift from MYI to FYI are many and complex and will derive from the strong modifications in physical forcings associated with MYI melt (e.g., increase in freshwater content and stratification, increase in light availability) together with direct impacts associated with the loss of permanent sea ice. It has been suggested that there will be a reorganization of Arctic food chains such that they will become more similar to those in the Antarctic where FYI

makes up 80% of the sea ice (Moline et al., 2008). In the central Arctic, increased numbers and size of leads and openings, and the increase in the duration of the open-water period are likely to increase primary production. However, increased surface stratification due to ice melt will act in an opposing manner since a small change in salinity may counteract the potential for wind-induced mixing (see Sakshaug, 2004; Carmack et al., 2006). Freshening of the surface layer is expected to impact the delivery and cycling of nutrients to surface waters, thereby influencing the amount and type of primary production. As proposed by Li et al. (2009), if reported changes in phytoplankton size structure (i.e., an increase in small phytoplankton) were to persist, this would entail major cascading effects on carbon flow and trophic pathways in Arctic ecosystems.

Over the past three decades, it is estimated that the atmosphere-ocean flux of CO_2 has tripled in the Arctic Ocean (from 24 to 66 TgC/y), attributed to direct (increased atmosphere-ocean exchange) and indirect (increased primary production) effects of the loss of sea ice (Bates et al., 2006). It has also been suggested that outgassing of CO_2 will increase with increased sea-surface temperature (8 gC/m^2/y per 1 °C; Anderson and Kaltin, 2001).

The location of fast-ice and MIZ edges relative to the shelf slope break and regional bathymetry is also an important aspect of ecosystem response. Modeling work (Carmack and Chapman, 2003) has shown that an ice edge can have an 'optimal' location, which results in stimulating upwelling of nutrient-rich waters into the euphotic zone. Significant increases in production can be expected as long as the ice edge withdraws no further than a location that permits the continuation of wind-forced upwelling of nutrient-rich waters along the shelf break, as proposed by Sakshaug (2004). The position of the ice edge and its associated effects on physical and chemical characteristics will also influence Arctic shelf communities. As the ice edge retreats northward, the delineation between Arctic and boreal communities is also likely to retreat to the north resulting in range expansions of temperate species, and range contractions of Arctic species. Range expansion by boreal species was documented during previous warm periods in the North Atlantic (reviewed by Renaud et al., 2008) and the Bering Sea (Grebmeier et al., 2006b; Mueter and Litzow, 2008). But when boreal communities reach the shelf break, shelf-dependent benthos and plankton will have few available refuges and High Arctic species will face increased competition as well as changing environmental conditions that are not in their favor and are likely to decline, with largely unknown implications for ecosystem functioning (Renaud et al., 2008).

Within the sea-ice realm, pressure ridges might be important habitats, although little can be said currently because they have received limited research attention. Ridges form when ice piles up under pressure, reaching drafts of over 20 m even in summer (Eicken et al., 2005) and these ice piles accumulate snow on the surface that can be important to some animals such as ringed seals, they also create texture under the ice that can serve as shelter for small animals. Based on recent findings of higher faunal abundances related to rafted sea ice compared to level ice during summer melt (Hop and Pavlova, 2008; Gradinger et al., 2009a,b), it is possible that pressure ridges will be especially relevant for the survival of sea ice-related invertebrates over the next decades in areas of dramatic sea ice loss.

Based on current understanding of vertical flux patterns in MYI *vs* FYI, it has been predicted that as ice melts and the FYI zone expands, the vertical flux of organic matter to the sea floor may increase (Wassmann, 2008). Because this shift from MYI to FYI will occur primarily over oligotrophic deep basins, an increase in productivity and vertical carbon flux could have important implications for carbon remineralization in the water column and for deep-sea benthic communities. Kirchman et al. (2009) postulated that decreasing ice and increased temperatures in the Arctic will increase the flow of carbon toward bacteria at the expense of larger organisms.

9.3.4.3. Predictions for marine mammals and seabirds

Changes observed to date in marine mammal populations under changing ice regimes strongly suggest that impacts will continue to intensify. Range shifts are already taking place and are expected to continue. Population reductions of many ice-associated species already documented in the Atlantic Arctic sector are expected to continue, perhaps at escalated rates. Compression of the ranges of endemic ice-associated species, as well as competition from temperate species moving north, are likely to result in increased competition for food. For example, Pacific walruses are already spending more time at land-based haul-outs (see Arnbom, 2009). If this trend continues they could deplete nearshore benthic resources, which would result in a lower carrying capacity for this species, even if benthic community health is otherwise unaltered. One small area within the range of Atlantic walruses might however be enhanced as a foraging area by changing ice conditions. Thick ice in eastern Greenland currently limits access by walruses to plentiful inshore bivalve banks to the short summer period. Less ice in this region might extend the foraging season in this area and permit more Atlantic walruses to occupy the region (Born, 2005).

Species that are strongly tied to sea-ice habitats, such as the polar bear, ringed seal and ivory gull, may become limited to small sea-ice refugia (e.g., summer sea ice is predicted to persist for longer in the Canadian Arctic Archipelago than elsewhere; see Pfirman, 2009). However, such refuges will support only limited numbers of animals. In general, the decline in sea ice is going to be detrimental to ice-adapted species and advantageous to migrant marine mammal species that currently occupy the Arctic seasonally (Moore and Huntington, 2008). But, the latter group of animals may also face challenges, with future food resources being more dispersed both spatially and temporally compared to the MIZ situation.

Northward range contractions or a shift to breeding earlier in the season are likely to be initial responses to protracted periods of reproductive failure among ice-associated pinnipeds unable to find ice-habitat(s) in traditional areas at the respective breeding times of the various species (Burns, 2009a). Shifting to a terrestrial-breeding situation would require a remarkable degree of behavioral plasticity that has not been seen to date in regions/years where ice reductions have been rapid and major (Kovacs and Lydersen, 2008). And such a shift, if it were possible for some species, is likely to be accompanied by very high levels of mortality for the pups, particularly among species where pups do not swim prior to weaning. Outside the breeding season, in years when sea ice is not available, spotted seals, bearded seals and ringed seals in some few areas do haul out to rest onshore (Burns, 2009a), but this is clearly not their preferred resting habitat and long-term consequences of such a haul-out habitat shift at a greater scale are unknown in terms of predation or disease-related mortality. Energetics modeling suggests that ringed seals in Svalbard will no longer benefit from offshore summer migrations to the ice edge when the retraction of ice from the archipelago reaches distances of 600 to 700 km (Freitas et al., 2008b). Despite the ability of the pack-ice seal species to remain pelagic, without ice for extended periods, it is likely that populations of these seals will have an increased probability of decline in the relatively near future because of breeding habitat deterioration (e.g., Johnston et al., 2005) given the situation already exhibited among Northeast Atlantic hooded seals and White Sea harp seals. Although projections for ice conditions in the Bering Sea during late winter suggest that the pack ice in that region might continue to sustain ribbon and spotted seal breeding.

Large future reductions in most subpopulations of polar bears are expected, such that this species will be lost from many areas where it is common today (e.g., Wiig et al., 2008; Regehr et al., 2010). Habitat losses are expected to be greatest in the southern seas of the polar basin, especially the Chukchi and Barents seas, and least along the Arctic Ocean shores of Banks Island over to northern Greenland. Density and energetic effects are likely to become important as polar bears make increasingly long migrations from traditional winter ranges to remnant high-latitude summer sea ice. The impacts are likely to be sex- and age-group specific and may ultimately preclude bears from seasonally returning to their traditional ranges (Amstrup et al., 2008; Thiemann et al., 2008; Durner et al., 2009). Expected abundance declines range between 30% and 70% for the next half century and, in the extreme case, it is expected that the circumpolar population of polar bears will exist only as a few distinctly isolated populations within relatively few decades (e.g., IUCN, 2008). This sort of population subdivision is expected to reduce gene flow among population clusters in the future (Crompton et al., 2008). These alarming projections for the future of this iconic wildlife species have not gone unopposed (see Dyck et al., 2007, 2008; Armstrong et al., 2008; O'Neill et al., 2008), but the polar bear science community has refuted these 'procedural' issues (e.g., Stirling et al., 2008b; Amstrup et al., 2009) and stand united in their projections for polar bear declines (e.g., Amstrup et al., 2008; Durner et al., 2009).

Unlike Arctic pinnipeds that rely on sea ice as a platform, the relationship of cetaceans to sea ice rests principally on how ice structures trophic pathways (Bluhm and Gradinger, 2008). Possible changes in pelagic–benthic coupling are expected to impact marine mammals that use benthic resources. The question is, of course, how will this impact Arctic species, including cetaceans? The answer lies within the ability of each species to adapt to changing food webs. In the case of narwhal, loss of pack ice in their wintertime benthic feeding habitat may be of particular concern. This factor contributed their extreme sensitivity index to climate change when ranked among all Arctic marine mammals (Laidre et al., 2008). Conversely, the flexibility in beluga diet may provide them with some resilience to changing prey. Similarly, the generalist feeding habits of

grey whales may give them an advantage over the largely planktivorous bowhead whale in a changing Arctic, although recent observations of bowheads feeding on euphausiids alongside grey whales suggests these whales may be capable of rapid adjustments to shifts in prey availability (Moore et al., 2010). Overall, the feeding ecology of marine mammals (e.g., generalist *vs* specialist feeders) and their capacity to adapt to new or different food resources will play a key role in their success in a changing environment.

Similar to other marine mammals, projections for the future with regard to cetaceans and sea-ice decline vary by region, with the greatest changes anticipated for the 'inflow' sectors of the Pacific and Atlantic Arctic where an additional 125 ice-free days are expected (Moore and Huntington, 2008). While it is anticipated that these regions may become more productive, existing food webs may be significantly altered. Predictable, high concentrations of prey, such as those that occur following phytoplankton blooms along retracting ice edges, are important for efficient foraging for baleen whales, so the effect of sea-ice reduction on potential prey is a key link in forecasting impacts (e.g., Tynan, 2004). Further, it has been surmised that sea ice also provides protection for Arctic cetaceans from predators such as sharks and killer whales. Increased presence of the latter may have significant impacts on endemic Arctic seals and whales in an Arctic with less sea ice (Ferguson, 2009; Higdon and Ferguson, 2009). Thus, Arctic cetaceans face multiple challenges in that they will have to adapt to altered food webs, while potentially dealing with increased competition for prey with seasonally migrant species that remain in Arctic waters for longer periods, and in some areas cope with increased levels of predation.

In addition to the direct impacts of habitat loss, marine mammals in the Arctic are likely to face increased disease and parasite risks as a result of other currently temperate marine mammals moving north and the warming both of air and seawater (Harvell et al., 1999; Rausch et al., 2007; Van Bressem et al., 2009) and perhaps also increased risks from contaminants (MacDonald et al., 2003, 2005; Noyes et al., 2009). Over the coming decades it is also highly likely that Arctic marine mammals will face increased impacts from human traffic and development in previously inaccessible, ice-covered areas (Kovacs and Lydersen, 2008; Fuller et al., 2008; Arctic Council, 2009). In combination, these various changes are likely to result in substantial distributional changes and/or abundance reductions for many endemic Arctic marine mammal species. The behavioral plasticity that they will exhibit (or not) is impossible to predict. These species have never before been challenged by the extreme environmental change predicted for Arctic sea-ice communities. If species are fixed in traditional spatial and temporal cycles, and are unable to shift them within decadal time scales, some populations will be threatened with extirpation. In somewhat longer time frames, species extinctions can also be envisaged.

Because most marine birds are highly philopatric, changes in breeding range are likely to be slow for colonial species. Nevertheless, in Canada, the range of the ivory gull has contracted severely over the past 30 years, corresponding with a reduction of about 80% in the overall population (Gilchrist and Mallory, 2005). Smaller, but significant reductions have also been noted for glaucous gulls in several parts of their range (CAFF

Seabird Group, unpubl.). A pattern of population expansion for ice-associated species in the northernmost parts of their range, along with reductions and range contractions at the southern edge will almost certainly continue, being accompanied by range expansions of more southerly species (e.g., razorbill, Gaston and Woo, 2008). The extent to which competition from more southerly species will impact populations of Arctic species is unknown, but negative impacts are likely (e.g., Irons et al., 2008). Less predictable spatial and temporal primary production in the future is likely as MYI is replaced by FYI and seasonal shifts occur. This could impact seabirds which require dense concentrations of prey, particularly during the critical chick-rearing period for successful reproduction. Change from highly concentrated production associated with ice edges and MIZ to more dispersed pelagic production may result in advantages to wide-ranging, more K-selected species (e.g., northern fulmar) and reductions in those species following a strategy of higher annual reproductive output (e.g., black-legged kittiwake).

9.3.5. Knowledge gaps and critical research needs

There are currently large gaps in knowledge that limit understanding of the complex interactions between Arctic biota and their physical environment. Over the past few decades, many scientific papers have solicited increased research effort in the Arctic to fill these gaps (e.g., Hansell et al., 1998). International Polar Year (IPY) studies will provide a vast amount of new knowledge and a vital foundation for future work, but many questions remain to be answered, which will require monitoring activity over a protracted period on the scale of ocean basins. Specifically, long-term commitments to ocean observatories that explicitly examine biological responses to changes in the physical structuring of the Arctic are urgently needed (e.g., Grebmeier et al., 2010). Such long-term series, together with process studies, will improve understanding of the response of sea ice to changing physical forcing mechanisms and feedbacks within and between the physical and biological components of the system.

At present, there is poor understanding of many of the key environmental forcing factors, their temporal and spatial scales of variability, and of ecosystem structure and function within the Arctic Ocean. This alone makes it difficult to predict ecosystem change due to loss of sea ice.

Currently, the Arctic Ocean is considered a net sink for atmospheric CO_2 (Bates and Mathis, 2009). However, it is only possible to speculate with respect to the direction and magnitude of future changes in the role of the Arctic Ocean as a source or sink of CO_2. The Arctic may be in a period of transition toward a new equilibrium, which will depend on a balance of biogeochemical processes operating at multiple scales, for example, atmosphere–ocean–ice exchanges, primary production and export to depth, respiration and remineralization of organic material. It has been proposed that sea ice plays a significant role in CO_2 drawdown and redistribution through ice formation, transport and melt (Jones and Coote, 1981; Anderson et al., 2004; Rysgaard et al., 2007, 2009). Furthermore, the recent discovery of calcium carbonates in sea ice (Dieckmann et al., 2008) and the release of dissolved

inorganic (and organic) carbon during ice formation, speaks to biogeochemical processes taking place in sea ice. These processes require further attention to have a fuller comprehension of their functioning and to be able to quantify the role of sea ice in the global carbon cycle. Reductions in sea ice are expected to directly affect these processes and future CO_2 drawdown. This knowledge is essential to constrain global greenhouse gas emissions scenarios and climate model predictions.

Few long-term biological sampling series exist because of the difficulty associated with re-sampling communities throughout the year or over multiple years or even decades. The various scales of variability that characterize Arctic marine ecosystems and the complexity of physical factors which, in isolation or in combination, influence biological organisms and communities, make it difficult to establish cause-effect relationships for predictions and adaptation. Seasonal changes in algal and faunal abundance and diversity are dramatic in sea ice and the water column (Makabe et al., 2009; Story-Manes and Gradinger, 2009), with dampened variability in benthic communities because of their typically slower growth rates and longer life-spans. However, interannual and interdecadal alterations have been detected both in pelagic (Brodeur et al., 2002; Hunt et al., 2008) and benthic habitats (Grebmeier et al., 2006a,b). In ice-covered areas, the need for long-term time-series using a combination of both traditional and recently developed sampling technologies is essential for developing a better understanding of the changes occurring in the Arctic.

Changes in the productivity of a summer ice-free Arctic will largely depend on a balance between factors that influence nutrient supply (i.e., water column stratification) and light availability. Therefore, changes in sea ice cannot be assessed in isolation from other environmental changes such as increased river runoff and increased erosion. As an example, because it is not yet possible to assess or predict the effect of changes in the delivery of nutrients via storms / shelf-break upwelling, it is not yet possible to predict changes in the magnitude of primary production, or the location of biological 'hot spots', on Arctic shelves. Such hot-spots are of vital importance to top trophic predators that require considerable energy intakes over the short summer period of productivity. MIZs have traditionally been a focal point for production and energy transfer to top trophic levels. But, if the timing of ice retreat or melt comes earlier, there is uncertainty as to whether animals will be able to track these phenological changes quickly enough. Key attributes such as plasticity in the timing of reproduction and foraging patterns are difficult to predict based on current knowledge. Ecological studies of how animals are adjusting to changing conditions in the Arctic are vital for increasing predictive capacity.

In addition, since community size structure plays a key role in food-web transfers, it is essential to improve characterization of bottom-up (physical factors) and top-down (e.g., grazing) controls on community composition and carbon flow pathways. The widespread occurrence and diversity of microbes in all Arctic habitats (e.g., Lovejoy et al., 2006; Le Romancer et al., 2007; Deming, 2010) warrants a better understanding of their physiology and ecological roles in Arctic food webs and in the cycling of carbon and other materials.

At all levels within the food web, range expansion of boreal species into the Arctic will affect community structure, strength of links in the food web, and food availability for current Arctic species. It is important to determine how food webs will adjust to changes in species diversity and competitive stress, and to understand the sensitivity of the Arctic ecosystem to this aspect of changes in ice regimes.

Among ice-associated marine mammals, poor baseline data on even the most basic population parameters, such as population size, make tracking global changes in Arctic populations, and extrapolating to future scenarios of environmental change challenging. Abundance is a basic metric for measuring impacts and 'health' of mammalian populations. But, the logistical difficulties in enumerating dispersed marine mammal populations make population surveys very expensive and hence infrequent, incomplete or simply non-existent. Despite the financial and logistic challenges, comprehensive monitoring plans must be put in place immediately for key ice-associated species. Several international plans have been developed, but remain unfinanced (Simpkins et al., 2009; IWC, 2010). Lack of knowledge regarding how some of these animals currently interact with their environments also limits the capacity to make accurate predictions regarding the impacts of climate change. A first-step in predicting likely consequences of future climate change for Arctic marine mammals is developing a detailed understanding of how these species use their environment today (see O'Corry-Crowe, 2008). Finding the means to monitor key populations of Arctic marine mammals as ecosystem sentinels can provide a window into a rapidly changing ecosystem (Moore, 2008a). In some areas of the Arctic, one means of doing this is to form research partnerships with Arctic residents that rely on marine mammals for subsistence. These collaborations provide access for attachment of satellite transmitters to track animal movements and, via harvest monitoring, to investigate changes in diet, body condition and contaminant burdens concomitant with ecosystem variability (Metcalf and Robards, 2008; Moore and Huntington, 2008). Integration of findings from community-based investigations with those from ocean observations is fundamental to any effort to predict long-term ramifications of sea-ice loss for marine mammals (Grebmeier et al., 2010).

The rapid ecosystem changes currently taking place in the Arctic (see Post et al., 2009) could cause significant biodiversity losses in this unique biome (e.g., Markham, 1996; Kovacs and Lydersen, 2008). In this context, it is important that management systems have up-to-date information on all harvested stocks, so that they can be responsive to changing conditions (Prowse et al., 2009; also see Section 9.4). Precautionary harvest levels are in order for all ice-associated species given the predictions for the direction and magnitude of change expected in their sea-ice habitats.

Co-ordination of major national and international efforts is of paramount importance at this time of rapid change in the Arctic to achieve the best use of limited resources in the research community. Currently, there is considerable redundancy among activities performed by agencies exploring climate change impacts, although new data becoming available for assessments are often limited. For example, IUCN is updating its 2008 Red Lists for marine mammals because of the impacts of climate change, which were not incorporated into the recent assessment process. Concurrently, the Marine Mammal Commission of the United States is doing a global assessment of the status of

marine mammals, largely because of the risks posed by climate change. Several independent non-governmental organizations have also begun similar initiatives. The CAFF (Conservation of Arctic Flora and Fauna) CBMP (Circumpolar Biodiversity Monitoring Programme) has also focused significant attention on this faunal group. But, the reality is that for Arctic areas few new data are available for these assessments and many populations, spread over vast regions of the Arctic, have never been surveyed and are not currently being monitored.

Integration of multidisciplinary research in a broad Arctic context has made major steps forward during the IPY period, but there is still a great deal of improvement that could take place across the interdisciplinary Arctic scientific research community.

9.3.5.1. Recommendations

- The top priority for immediate action by the Arctic Council is to ensure that the governments of the Arctic countries arrange long-term commitments to ocean observatories, and process studies that explicitly examine biological responses to changes in the physical environment within the Arctic.

- If sea ice declines occur in the manner predicted in the coming decades, consideration should be given by Canada and Greenland, to creating a World Heritage Site in the Northwest Greenland / Northeast Canadian Archipelago region as a refuge for ice-associated species, as it is likely to be the only area within the Arctic to retain some MYI as well as reasonably extensive summer sea-ice cover.

- Regional scientific assessments and monitoring activities of biological community components across the Arctic, using standardized methodologies among areas, are highly recommended. The rate and extent of sea-ice change currently taking place in the Arctic has not occurred in recorded history and is of significant, broad scientific interest. Results from such monitoring and assessment activities will support decision-making for the sustainability of ecosystem services and biological conservation, and help human populations in the North make transitions to alternate resource use (Berkes and Jolly, 2001; see also Section 9.4).

- Concerted international efforts should be undertaken to preserve endemic Arctic flora and fauna.

9.4. Effects of sea-ice change on human society

- Indigenous and some local peoples have been challenged by changes in sea ice and those challenges will continue into the future.

  ◦ Predictions based on traditional and local knowledge have become less reliable and Arctic residents have to act with greater caution and develop new knowledge and strategies.

  ◦ Increased use of technology (GPS, satellite imagery) is a potential local adaptation strategy.

  ◦ Subsistence and natural resource use activities must adapt further to changing natural resource bases, including changes in timing, distribution, and abundance of animals and fish.

  ◦ Travel safety is being reduced. In addition to increased hazards, some smaller communities may become more isolated, and social interactions negatively impacted.

  ◦ Barriers to adaptation include limited access to financial resources, wildlife management regimes, reduced outdoor experience and applicable knowledge among younger generations, and institutional memory.

  ◦ Cultural resources are being compromised in some locations, a result of coastal erosion, permafrost thaw, and constrained access and availability of traditional/local foods.

- Reduced seasonal sea-ice presence is already leading to greater shoreline exposure to open water and storm waves. These changes impose greater wave and erosion hazards for shoreline, infrastructure, waterfront structures, and cultural heritage sites in some places.

  ◦ Some ice-bound coasts with negligible past wave climate will be highly susceptible to wave reworking.

  ◦ Even where seasonal open water has been extensive in the past, later freeze-up and earlier break-up increase the probability of storm wave action, even with no change in storm activity.

- Loss of sea ice is increasing interest in access and resource rights, such as shipping, resource extraction, military operations, and tourism, which will have impacts across a wide-range of issues.

  ◦ Resource management planning does not comprehensively anticipate impacts of a declining sea-ice extent.

  ◦ There is a general lack of preparedness, across local, national and international levels, to respond to increased pollution and hazards (e.g., search and rescue needs) as sea ice decreases and access to ice-free areas increases.

- Hazards related to sea-ice movement (ride-up and pile-up) onshore are known to have caused fatalities and significant infrastructure damage in the past.

  ◦ Changing climate will reduce the probability of multi-year shore ice pile-up.

  ◦ Thinner ice may be more susceptible to pile-up and a shortening of the ice season will not decrease the probability of occurrence.

- There are clear limits to any Arctic governance regime (regulatory or soft-law). The drivers of sea-ice change and its consequences occur largely outside the region and fall beyond the jurisdiction of any one Arctic state, and responding to sea-ice change creates profound governance challenges.

  ◦ Existing international and national institutions (e.g., Arctic Council, co-management agencies) can play a critical role as boundary organizations to bring various sub-governmental (e.g., villages, states, provinces) and non-governmental actors (e.g., Inuit Circumpolar Council) together to enhance the flow of information and knowledge.

9.4.1. **Introduction**

Estimates of the Arctic population vary between 4 million (AHDR, 2004) and 9.9 million (Duhaime and Caron, 2009), depending on geographic definition and seasonal variations, with the indigenous population accounting for between 400 000 and around 1.3 million. Roughly half of all Arctic residents live in the Russian Federation, where the population fluctuates seasonally through work related migration. Many Arctic inhabitants, both indigenous and non-indigenous, live near the coast where sea-ice variability is a dominant feature of the environment. This variability has consequences for subsistence livelihoods, as well as for industrial and economic activities such as commercial fisheries and hydrocarbon development.

Ecosystem services are the benefits humans derive from ecological processes and their functions. Sea ice in the Arctic is critical for ecosystem services and human activities, but may also impede both. Figure 9.37 illustrates the interaction between different types of sea ice and human activities. Eicken et al. (2009) grouped sea-ice system services (SISS) into four broad categories (see Box 9.14). These services/impediments have a spatio-temporal dimension relevant to humans with respect to seasonality and sea-ice type.

In addition, there are important linkages between the Arctic sea-ice system and the rest of the world; these include physical (Arctic climate system feedbacks globally), chemical (pollutant transport to the Arctic), biological (migratory biota and ecosystem connections), and societal (tourism, resource extraction, policy and politics) linkages. Sea ice dominates many aspects of these linkages, in some cases controlling the direction, nature and rates of change, as well as controlling processes and systems within and among Arctic areas. Thus, immediate and long-term changes in sea ice will directly and indirectly affect human activities worldwide. Reviewing the consequences of sea-ice change, it is essential to distinguish between changes to landfast ice and pack ice, as these are used differently by humans. It is also important to realize that challenges to some will represent opportunities to others.

The aim of this section is to identify and link the key effects of sea-ice changes with key aspects of human society (indigenous and non-indigenous) and activities in the Arctic. The balance among these effects will determine the net influence of sea-ice change on human society and activities, regionally and at a pan-Arctic level.

Projections of the consequences of sea-ice change for human society are limited by an accumulation of uncertainties. Uncertainties in understanding and modeling the physics of sea-ice change are amplified when projecting biological effects, and even more so when considering consequences for society. Societal, political and economic influences operating independently of climate change further increase uncertainty.
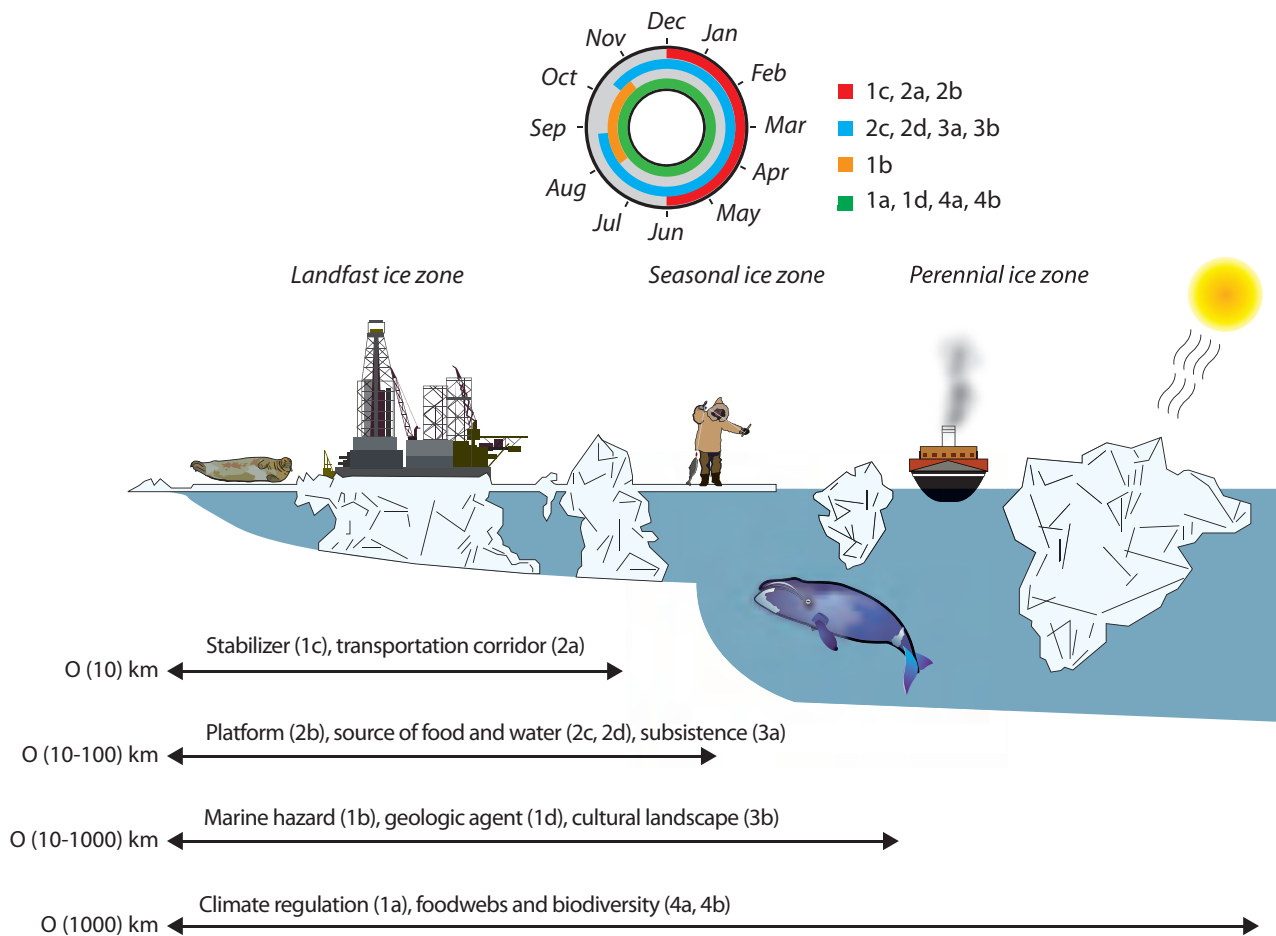
Figure 9.37. Interactions between sea ice and human society. The graphic indicates various ice types extending from the shore (left) to the deep ocean (right) and associated services to humans and ecosystems. The spatial scale of service action is indicated by the double-headed arrows with which the services are associated and cross-referenced to Table 9.4 by number/letter combinations. Source: based on Eicken et al. (2009).

Box 9.14. Sea-ice system services

Following the general structure presented in the Millennium Ecosystem Assessment (2005), Eicken et al. (2009) grouped sea-ice system services (SISS) into four broad categories: regulating, provisioning, cultural, and supporting (Table 9.4). Most services (or impediments) are captured within these categories. There is temporal variation in the services (or impediments) due to changes in sea ice through the year (presence and quality) as well as spatial variation due to the location of the sea ice (offshore ice *vs* landfast ice) and at different scales of influence.

Table 9.4. Sea-ice system services and impediments affecting human society in the Arctic. Source: modified from Eicken et al. (2009). Service category numbers and nature of service letters are cross-referenced on Figure 9.37. Decreased sea ice will result in generally positive (+) or generally negative (-) effects on humans in the particular example. Effects with uncertain direction are indicated by a question mark.

Service category	Nature of service	Comments
1. Regulating	a. Climate regulator – Arctic and global	Albedo feedback (-); carbon dioxide dissolution regulation (-); thermal feedback (-); nearshore weather regulator/moderator (e.g., fog) (-)
	b. Hazard/impediment to shipping, fishing fleets and coastal infrastructure	Likely decline in physical hazard, thus widening seasons or areas of operations (+, but may require specialized adaptations to exploit); increased ice variability may increase hazard (-)
	c. Stabilizing element for coasts, infrastructure and activities	Platform for human activity (e.g., shelf hydrocarbon development) (-)
	d. Geological role of ice	Protects coastlines from storms (shore-fast ice buttresses land-based permafrost (-); floating ice reduces ocean wind fetch(-); sediment rafting (-?); bottom disturbance (-); thinner ice in the nearshore zone will result in less extensive coastal permafrost – implications for wharves and construction on piles into permafrost (-)
	e. Contaminant regulator	Ice-stored contaminants released (-)
2. Provisioning	a. Transportation corridor	Human ice travel alongshore (-); marine mammals via flaw leads and on ice surface (+?)
	b. Platform for local subsistence and industrial activities	Subsistence hunting of surface (polar bears, seals) or floe-edge (narwhal, walrus) foods (-); stable fishing platform (e.g., long-lining for Greenland halibut) (-); ice islands for hydrocarbon development (-); ice roads (-)
	c. Multi-year ice (as freshwater source)	Decreased or more distant supply (-)
	d. Access to food sources	Travel to floe edge rougher (-), floe edge closer (+), but shorter duration of stable ice (-)
3. Cultural	a. Subsistence activities on and near ice	Activities lessened, teaching opportunities decreased (-)
	b. Cultural and spiritual landscape	Ice loss (-)
	c. Tourism, recreational fishing and hunting	Ice-associated fauna as focal point decreased (-); ship-based tourism access increased (+)
4. Supporting	a. Sea-ice based ecosystems	Habitat for ice flora and fauna on surface, interstitially and on under surface decreased (-); productive food webs decreased/changed (-/+)
	b. Biodiversity reservoir	Extremophiles (bacteria, viruses) and ice-adapted biota decreased (-)

9.4.2. **ACIA status of knowledge**

The Arctic Climate Impact Assessment (ACIA, 2005) was the first comprehensive effort to assess climate change and its effects in the Arctic region. The assessment focused on changes expected to occur within the 21st century, based on observed data and a moderate scenario of projected future warming (IPCC emissions scenario B2). Although an assessment of economic impacts and community vulnerability to climate change was limited, ACIA projected major cultural, ecological and economic impacts, including higher production costs resulting from increased costs of infrastructure maintenance, and reduced ocean transportation costs due to longer shipping seasons (see Chapter 1, Section 1.1, for a list of the ACIA key findings).

ACIA projected a summer sea-ice decline of more than 50% over the next 50 to 100 years which would extend the Northern Sea Route navigation season "by two to four months from present" (Walsh et al., 2005) and lead to increased exploration for hydrocarbons and minerals (Loeng et al., 2005). Increased

activity could enable overall economic growth, increase investment options, provide for extensive social services, help create public trust funds for future benefit, and result in local-scale economic stimulation.

Although ACIA noted that a longer ice-free open water season in summer would benefit some local communities, the overall impacts on Arctic residents, especially indigenous traditional lifestyle, health and well-being, were projected to be negative. Sea-ice loss was projected to result in marine mammals being less accessible to hunters, to changes in or losses of traditional hunting grounds, and to modifications of inter-generational transfer of traditional knowledge and culture. Decreased sea-ice coverage was also projected to increase the probability of storms in open water, which would affect near-shore productivity (Walsh et al., 2005). With access to traditional/local foods declining, social and cultural values were considered likely to change or weaken. Shifts in human diets may also have economic consequences. Nuttall et al. (2005) stated that changes in natural resource use may lead to major changes in economic conditions, demography, societal structure and cultural values. Projections of sea-ice change were corroborated by local indigenous observations and scientific monitoring, confirming that many of the changes have already taken place and are now having an impact on travel, safety and lifestyles (Huntington and Fox, 2005).

ACIA also emphasized cross-cutting issues such as effects of sea-ice change on human health (Berner et al., 2005), infrastructure and coastal erosion (Instanes et al., 2005; Walsh et al., 2005), climate change in the context of multiple stressors (McCarthy et al., 2005), and multiple interactions between effects and processes in and beyond the Arctic. The ACIA report concluded that there was a need for qualitative analyses of the economic impacts of climate change, with impact assessments taking a local focus and including all Arctic residents, thereby increasing the relevance for Arctic residents.

9.4.3. Current local and scientific observations

Documentation and use of traditional knowledge has been central to climate change research at the community level. Many Inuit elders and experienced hunters have noted changing sea-ice conditions over the past decade, including later ice freeze-up, earlier ice break-up, a longer ice-free open water period in summer, and altered ice dynamics. These changes have been attributed to rising temperatures and changing temperature variability, especially during the transition periods between seasons, and changes in wind patterns (Ogi et al., 2010). Recent years have witnessed extreme ice conditions, in many ways unprecedented in local knowledge. For example, experienced hunters have noted significant changes in Igloolik, Nunavut, the site of climate change research for over a decade (Fox, 2002; Ford et al., 2006, 2009; Laidler and Ikkumaq, 2008; Laidler et al., 2009):

> "When I was younger I remember that the ice freezes at the end of September … Now it freezes [in] late October even [the] first week of November." Harry Ittusujurat, 2005.

> "This is the year [the ocean has] been really, really late in freezing up. It has been later than I ever remember. The reason we call this month Tusaqtuut is because this is the time [when] everything is frozen and communication between families would be easier at this time … but we're still not allowed to go straight [to the mainland] because the ice is very dangerous [and] thin." Augustine Taqqaugak, December 2006.

Sea-ice data for the Igloolik region derived from the Canadian Ice Service and Environment Canada weather stations correspond with Inuit observations (Ford et al., 2009; Laidler et al., 2009), showing a trend of later ice freeze-up of ~0.7 d/y, or one week per decade, between 1969 and 2005. Detailed analysis indicates that the bulk of the change has occurred since 1990 (Laidler et al., 2009). Consistent with later freeze-up there has been significant warming in October and November, increasing by almost 0.2 °C/y between 1977 and 2002. Similar correspondence between Inuit observations and instrumental records has also been found at other times of the year (Ford et al., 2009; Laidler et al., 2009).

Changes in sea-ice extent and condition appear to be occurring much faster than projected in the ACIA report. Sections 9.1 and 9.2 report an acceleration in sea-ice decline in extent and age over the past ten years, both for landfast- and pack-ice, although regional differences exist. The physical changes in sea ice in combination with their biological impacts (Section 9.3) are of direct relevance to human society, individuals and activities, in the north. The changes will also require regulation, governance and organization of services at all scales, including at the pan-Arctic level. Although climate change will bring some opportunities, whether these are sustainable over the long-term is a key issue (e.g., Anisimov et al., 2007).

9.4.4. Current human interactions with sea ice

9.4.4.1. Indigenous peoples

For many indigenous peoples (see Figure 10.1) sea ice is fundamental to their culture and identity and is used as a platform for travelling (mainly on landfast ice) and for local subsistence activities, such as hunting and fishing. Across the Arctic, hunting and fishing practices (which mostly take place at the floe edge where landfast ice and pack ice create open-water leads) are adapted to local sea-ice conditions and the seasonal and behavioral patterns of the species harvested. Harvesting is important for nutritional health, economy, identity, and cultural, spiritual and social well-being (Huntington et al., 1998; AHDR, 2004; Hovelsrud et al., 2008). Traditional/local foods are a valuable source of energy, nutrients and minerals and make an important contribution to the diet in many communities. These foods generally have a higher nutritional value than typical store-bought foods. Although store-bought foods are common, subsistence harvesting remains important in all coastal communities, and in areas where income is scarce or where people are living below the poverty line (Nuttall et al., 2005), but can be found across Arctic households independent of income (Aslaksen et al., 2009). Polar bear and seal skins, and sport hunting of polar bears and walrus are an important source of income and employment in some communities (Mathias

et al., 2008). Major changes in, or even loss of, sea ice and the consequent changes in traditional/local food availability will have significant impacts on health, social relationships, and hunting activities (Nuttall et al., 2005; Furgal and Seguin, 2006; Ford, 2009c; Ford et al., 2010) as well as on the personal economies of coastal inhabitants.

In many communities, thinner sea ice and active cracks in the ice, combined with less predictable weather have implications for hunting, harvesting and travelling (Gearheard et al., 2006; Ford et al., 2008; Bravo, 2009; Ford and Pearce, 2010). Access to key hunting areas has been constrained particularly in spring and autumn, resulting in shortages of traditional/local foods such as seal, walrus, and caribou. In parts of the Canadian Arctic, hunters are required to travel further to access hunting areas (Ford et al., 2010; Pearce et al., 2010). Preliminary findings of an ongoing IPY project on the effects of global warming on Arctic marine mammals describe how sea-ice decline in Hudson Bay drives a shift in the ecosystem from a polar bear / seal system with Inuit hunters at the apex to one dominated by cetaceans with killer whales at the apex, thus eroding Inuit traditional subsistence culture (DFO, 2010:11). In Greenland, a number of local and regional sea-ice travel routes can no longer be used and hunters traveling by dog team are forced to travel overland, which is more time consuming and can be more dangerous than sea-ice travel (Goldhar et al., 2010). Changing sea-ice conditions, beyond expected natural variability, have increased the risk of hunting and travelling on the ice, and have led to a greater incidence of accidents and death (Ford et al., 2006, 2008; Hovelsrud et al., 2008).

> "*We have lost the skills so much. What would not have been dangerous 50 years ago is now dangerous [to young Inuit].*" James Ungalak, Igloolik, 2004.

> "*Nowadays [the younger generations] just want to go out [hunting] without realizing what the weather is going to be like … Consequently it's more dangerous for them.*" Peter Awa, Igloolik, 2004.

> "*GPS enhances your navigation, but if it runs out of batteries it doesn't enhance navigation, it disables you. It can be your blessing and your downfall at the same time.*" Theo Ikkumaq, Igloolik, 2004.

In this context, it has been argued that climate change effects experienced to-date are 'dangerous' and compromising the human rights of some indigenous peoples (Crump, 2008; Ford, 2009c).

The relationship between changing sea-ice conditions and local impacts is complex, mediated by the interaction of multiple biophysical and human stresses operating over different spatial and temporal scales. Thus, the increasing danger of using the ice needs to be linked to changes in risk-taking associated with reduced time to hunt, diminishing land skills and decreased traditional knowledge among younger generations, decreasing reliability of traditional knowledge as conditions change, and use of new hunting technology. Likewise, the implications of changing sea-ice conditions for food security are exacerbated by rising commodity prices, increasing costs of hunting (fuel prices, equipment), and weakening of sharing networks. Harvest

and access changes are compounded by community isolation, restrictions and uncertainty about travel, and increased stress in communities (e.g., Berner et al., 2005). Although the local effects appear overall to be negative, a longer ice-free open-water season in summer could also be beneficial for hunting and travelling activities, but this assumes that prey are available, skills for open-water hunting are developed, and appropriate boats and financial capacity are available in the communities (Wenzel, 2009).

While research has documented significant vulnerabilities in local communities to sea-ice change, recent work has also noted emerging adaptive capacity which is moderating some of the negative impacts. Inuvialuit hunters and fishers in the western Canadian Arctic, for example, are changing their mode of transport (e.g., using boats instead of snowmobiles) to adapt to spring break-up occurring earlier each year (Ford et al., 2010; Pearce et al., 2010). Some studies highlight how social learning and responsive local institutions have emerged as the community adapts to changing ice conditions, thus helping reduce the dangers of using the ice in a changing climate (Ford, 2009a; Ford et al., 2009).

Many individuals and communities adapt by taking advantage of new conditions and minimizing impacts by using new technology, changing resource use patterns, and/or switching to the consumption of store-bought foods. In many cases these adaptations require monetary investments. The ability of households to make these investments is constrained by limited financial capacity: for example, an average household income in Igloolik is two-thirds the Canadian average, living costs are significantly higher (store-bought food is on average twice as expensive as the same food bought outside the Arctic), and unemployment is high (officially 16%, unofficially over 30%) (Ford and Berrang-Ford, 2009).

The Sea Ice for Walrus Outlook (SIWO) is an ongoing effort that combines local knowledge and observations with scientific assessments and satellite data, to present weekly information on sea-ice condition relevant to walrus in the northern Bering Sea and southern Chukchi Sea regions of Alaska (www.arcus. org/search/siwo). In addition, the IPY 'Sea Ice Knowledge and Use' (SIKU) project (Krupnik et al., 2010) has produced the first detailed insight and comparative documentation of the patterns of indigenous knowledge and use of sea ice across from the Bering Strait to Greenland. Such information is of great value to hunters as well as others traveling on sea ice. Although there are also groups in the community that do not depend on hunting and/or have access to financial resources to adapt, the changing ice conditions threaten the cultural and spiritual identity as well as the quality of nutrition for all local peoples.

A prevalent reaction is a sense of dislocation as familiar conditions give way to those that are as yet not well understood, continue to change, and become more variable. Recovery from temporary environmental change may be possible for some time, as shown by the return of hunting practices and the continuation of associated knowledge and skills in some communities. Savoonga, Alaska, successfully began whale hunting again from the traditional location in the early 1970s, despite a break of nearly a century, in part because some whalers had continued to hunt in the neighboring village of Gambell (Noongwook et al., 2007). Owing to considerable interannual variability in the past, the indigenous peoples of the Arctic are highly adaptable and flexible in their practices and general

outlook (Huntington et al., 2007; Forbes and Stammler, 2009; Wenzel, 2009; Ford et al., 2010). However, adaptive efforts are currently strained by the extent and speed of recent changes in sea ice, because traditional knowledge may no longer be as relevant for understanding the environment and there is a high degree of uncertainty about what will come next.

9.4.4.2. **Shipping**

Current shipping in the Arctic is mostly local in nature (Chircop, 2009). Arctic marine shipping currently consists of tankers transporting natural gas from northern Norway to the United States, hundreds of cruise ships visiting Greenland and dozens visiting the Canadian Arctic each summer, while a few travel to the geographic North Pole. Adventurers are also heading north in growing numbers, some sailing small boats through the Northwest Passage in remote and incompletely charted and sometimes stormy waters (Byers, 2009). In addition to tourism and servicing offshore oil and gas installations, marine shipping also includes the larger fishing vessels, warships and other government ships operated for non-commercial purposes (Molenaar and Corell, 2009a). In the Canadian Arctic and around the east and west Greenlandic coasts most transport occurs in summer. In the Russian Arctic, year-round marine transport has been maintained since 1978/79 between the port of Dudinka on the Yenisey River and Murmansk (Molenaar and Corell, 2009a).

The Arctic sea ice has only recently retreated enough to make trans-Arctic navigation a possibility (see Section 9.4.5.1.3). Since 2000, a small but increasing number of trans-Arctic summer voyages have taken place mainly in relation to science and tourism. In September 2009 the first commercial ships navigated the Northern Sea Route. The current level of trans-Arctic shipping is moderately high through the Northern Sea Route and low through the Northwest Passage (northern Canada).

Arctic navigation is challenged by drifting icebergs, a lack of accurate charts and insufficient experience in navigating in ice-covered areas (Molenaar and Corell, 2009a). While most tankers are ice-strengthened, many tourist ships operating in the Arctic were not designed for ice-covered waters, increasing the likelihood of shipping accidents. The International Maritime Organization (IMO) has recently approved a 'Polar Code' for all ships operating in polar waters, covering ship standards, safety and reporting requirements (IMO, 2009) and this is expected to become mandatory in 2012. These guidelines form the first internationally recognized standards for the construction and operation of ships in ice-covered Arctic waters. Although international policies are being developed to support development in the Arctic (e.g., Commission of the European Communities, 2008), infrastructure (e.g., charting, communications, ports, environmental monitoring, search and rescue, salvage, aids to navigation, oil spill response, supplies and spares) is currently limited or lacking to support high levels of marine traffic.

9.4.4.3. **Local and commercial fisheries**

Fisheries in Arctic waters are of great economic importance, at both the national and international level, and contribute substantially to the economy of many Arctic countries, in particular Greenland, Iceland, the Faroe Islands and Norway (Glomsrød et al., 2009). Although the economic importance of fisheries may vary widely at the regional level, it is the backbone of local economies for many northern communities.

The marginal ice zone is a nutrient-rich zone with high levels of primary production (Section 9.3) that support a range of species higher up the food chain (fish, seabirds, marine mammals). Except for Greenland halibut, polar cod, capelin, and Greenland cod, most commercial fish species are found in waters south of the Arctic Ocean (Loeng et al., 2005). The area around the Polar Front in the Barents Sea, where warm Atlantic water meets cold polar water, is one of the most productive marine areas in the world and supports a large international fishing industry (Sherman et al., 2009). The main commercial fisheries presently occur in the Barents Sea and to the east and south of the Norwegian Sea (Commission of the European Communities, 2008), although a substantial amount of commercial fishing also occurs in the Bering Sea and in Icelandic and Greenlandic waters. The area in the Bering Sea (Alaska) formerly covered by sea ice has become a favorable habitat for many subarctic species, and as a result, there has been an increase in biomass for most fish stocks in the area (Mueter and Litzow, 2008). There has also been a continued trend toward more subarctic ecosystem conditions in the northern Bering Sea, which may have profound impacts on commercial and coastal fisheries as well as on marine mammal and diving seabird populations (Grebmeier et al., 2006a). However, a recent federal vote by the North Pacific Fishery Management Council in Alaska (Stram and Evans, 2009) placed a ban on commercial fishing in its Arctic waters mostly north of the Bering Sea, stressing the need to assess how global warming is affecting the region before allowing fishing for stocks such as Arctic cod, saffron cod and snow crab.

> "*There's concern over unregulated fishing, there's concern about warming, there's concern about how commercial fishing might affect resources in the region, local residents and subsistence fishing and the ecosystem as a whole.*" Bill Wilson, a North Pacific Fishery Management Council aide, Washington Post, February 6, 2009.

Warming ocean waters and a retreating ice edge are leading to new opportunities as new species appear and existing fish stocks are showing horizontal and vertical changes in distribution as their seasonal migration is related to the formation and melting of the sea ice (Sundby and Nakken, 2008; Sherman et al., 2009; West and Hovelsrud, 2010). This has consequences for coastal fishermen in terms the need for different fishing gear, increased risk with longer trips at sea, problems with delivering the catch to landing facilities, and fishing regulations limiting adaptation to changing fisheries. Drifting sea ice and icebergs are an obvious problem for offshore fisheries, but also pose challenges to marine-based aquaculture in fjords and bays, and the booming whale-watching industry, which is attracting an ever increasing number of visitors and is supplementing fishing as source of livelihood in some coastal communities (Einarsson, 2009).

While many commercial and locally harvested fish species are sensitive to climate change, the main impacts on fisheries are of a non-climatic nature, such as shifting fisheries management and regulations, closing of landing facilities with resulting

job loss, globalization, and market integration (Keskitalo and Kulyasova, 2009; West and Hovelsrud, 2010). Other fisheries (and shipping) concerns associated with sea ice include poor infrastructure for search and rescue, especially in more northerly areas, and the lack of charts and iceberg monitoring.

9.4.4.4. Non-renewable resources

9.4.4.4.1. Hydrocarbons

Within the Arctic Circle, one-third of the area is above sea level, one-third comprises continental shelves at depths of 500 m or less, and the remainder consists of deep ocean basins historically covered by sea ice. The onshore areas account for almost 10% of the world's known conventional petroleum resources (cumulative production and remaining proved reserves) and by 2007 more than 400 oil and gas fields, containing 40 billion barrels of oil, 1136 trillion cubic feet of natural gas, and 8 billion barrels of natural gas liquids had been developed, mostly in the West Siberian Basin of Russia and on the North Slope of Alaska (Gautier et al., 2009).

Most of the Arctic however, especially offshore, is essentially unexplored with respect to hydrocarbons and it is estimated that around 84% of the undiscovered oil and gas in the Arctic occurs offshore (USGS, 2008). While deep oceanic basins have relatively low hydrocarbon potential the Arctic continental shelves are one of the world's largest remaining prospective areas. Until now, little exploration has occurred offshore. Even where offshore wells have been drilled primarily on shelf areas, in the Mackenzie Delta, the Barents Sea, the Sverdrup Basin, and offshore Alaska, most of the discoveries remain undeveloped (Gautier et al., 2009). This is mainly due to the remoteness of the area, costs and technological challenges associated with operations in ice-infested environments, and abundant relatively low-extraction-cost petroleum.

To date, Arctic hydrocarbon exploration, development and production make use of ice-islands, Gravity Base Structure (GBS) platforms in relatively shallow water or detachable floating systems. Adaptations in the Arctic offshore oil and gas industry include a clear increase in amounts of production equipment being installed on the seabed, eliminating the need for a floating platform and thus avoiding sea-ice related hazards. Significant progress has been made in design, construction, transportation, installation, and removal of offshore structures, while remaining challenges include an increase in safety and emergency response, improved infrastructure, the need to address icing problems, improved ice and iceberg data mapping, and protection of structures from icebergs and ice pressure ridges.

Challenges for oil and gas facilities in the Arctic fall into two categories: (i) overcoming the physical environment (e.g., sea ice and wave action) and (ii) meeting regulatory requirements and increasing societal expectations related to environmental performance and social benefits (Efthymiou and Winkler, 2009).

9.4.4.4.2. Minerals

Current practices in Arctic metal mining usually proceed through a seasonal cycle of winter extraction, concentration of ore, and storage followed by a summer export season once shipping lanes are open. Thus, challenges and opportunities linking mining and sea ice are mainly related to increased opportunities for shipping as sea-ice changes occur.

9.4.4.5. Tourism

Accessibility in the Arctic has increased through improved transport technologies ranging from cruise ships and air transport to specialized vehicles to access remote areas. Polar tourism has developed substantially, both in terms of numbers and in geographical and seasonal reach (Stewart and Draper, 2006; UNEP, 2007). The sector is diverse and is aimed at the mass market. Activities include sports fishing and hunting, ecotourism, adventure tourism, and cultural and heritage tourism. Iconic Arctic species such as the polar bear draw tourists northward (e.g., Forgione and Martin, 2008) but also hold considerable 'existence' value for many people who will never visit the Arctic but enjoy knowing that Arctic species exist. Sport hunting of polar bears has become an important means of contributing to economies for certain Inuit communities, whereas in other areas polar bear hunting is allowed for subsistence use only (e.g., Hovelsrud et al., 2008). Cultural tourism is an important part of the economy in Nunavut, Greenland, and northern Scandinavia, and is currently being developed in northern Russia by the Russian Association of Indigenous People of the North, Siberia and Far East (RAIPON) (UNEP, 2007).

Tourism is an integral part of local economies, personal income, revenues and financial capital for northern infrastructure. Arctic economies and gateway cities in the Southern Hemisphere have seen tourism evolve from an incidental activity to a vital sector upon which they increasingly rely (UNEP, 2007). Challenges for tourism can be divided into three categories: technical challenges, mainly related to shipping; infrastructure, including search and rescue; and sustainability (Stewart and Draper, 2008). Ironically, the first ship to transit the Northwest Passage in 1984 was also the first cruise ship to sink in polar waters, off the coast of the Antarctic Peninsula, in 2007. The incident illustrates the importance of contingency planning for emergencies and the need for circumpolar safety and environmental regulatory mechanisms to be in place for the expected increase in tourism and shipping. In most regions of the Arctic, the number of seasonal tourists exceeds the population in rural settlements. This puts a stress on the community and its population; in many instances an annual income must be earned in only a few months. Many Arctic residents are challenged by balancing cultural and social norms with the need for a viable economy (UNEP, 2007).

9.4.5. Anticipated consequences of sea-ice change

Arctic sea ice is declining much faster than previously predicted by IPCC models and the Arctic Ocean is likely to be ice-free for short periods during summer within the next few decades (see Section 9.1; Stroeve et al., 2007). High interannual variability in sea-ice extent will continue, but will decrease as ice-free summers are approached. The Arctic basin is projected to be mostly ice-free in summer by mid-century or before. The winter pack ice is projected to be increasingly dominated by first-year ice (Koç et al., 2009) and there are estimates suggesting that

the Arctic may be free of multi-year ice over the next couple of decades.

The retreat, thinning and temporary absence of sea ice will have profound impacts on ice-dependent species and as yet unforeseen effects on the ecosystem. This is likely to affect people and sectors active in the Arctic alike, through new challenges and new opportunities, requiring further economic and societal adaptation. Most of the likely losses will be through the decline and seasonal shifts in landfast sea ice with particular consequences for indigenous lifestyles and coastlines. Most of the likely benefits will be through both landfast- and pack-ice decline, with consequences for industry, shipping and fishing (re-supply, hydrocarbons etc.). Impacts and responses for human societies are summarized in Figure 9.38 and Table 9.5.
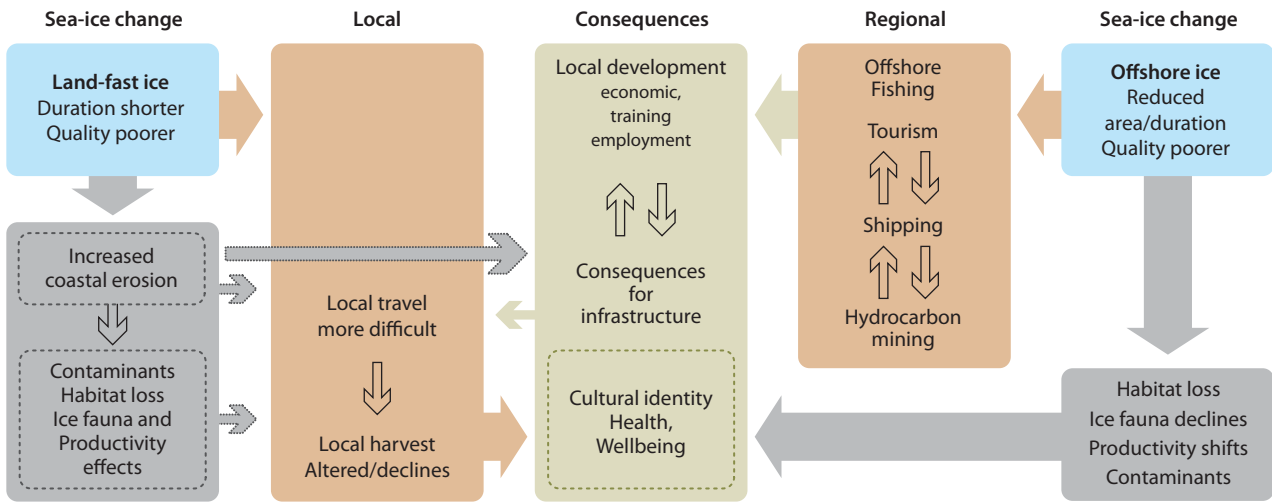
Figure 9.38. Interactive follow-on and cumulative effects of sea-ice change on humans, illustrating the direct and indirect human relationships with sea ice use (and potential changes). Large arrows of the same color as the block indicate effects from/on the whole block. Smaller arrows and dashed blocks indicate effects between components between blocks (e.g., habitat loss, ice fauna decline and contaminants have consequences for local harvest, cultural identity and health and wellbeing). This figure should be viewed in association with Table 9.5.

Table 9.5. Results, responses and rate of change of effects of sea-ice change on humans.

Type of sea-ice change	Direction of effects	Rate of change	Result	Response
Shorter duration of landfast ice	Landfast ice travel (locals)	<20 years	Personal safety and food security compromised	Changes in transport mode. New equipment needed
Poorer ice quality	Harvesting opportunities	<20 years	Personal safety and food security compromised	Changes in transport mode. New equipment needed
Melt of MYI	Release of legacy contaminants	<30 years	Personal safety and food security compromised	Change in diet
Longer/larger open-water season at the coast	Wider fetch (combined with increased storminess)	< 20 to 30 years	Coastal erosion Harbor degradation	Shoreline stabilization, resettlement Engineering options
Longer/larger areas of ice-free ocean	Increased possibilities for shipping Increased possibilities for non-renewable resource exploitation Changes in species composition	30 to 50 years	More traffic More tourism More disturbance Better re-supply More jobs (?) Ballast/Pollution More accidents Changes in fishing opportunities	Improved: mapping; infrastructure; harbors; search and rescue; emergency, prevention, preparedness and response; training Coordination of tourism International legal agreements

9.4.5.1. Challenges and opportunities

9.4.5.1.1. Travel, safety and food security

Observations indicate that sea-ice conditions are changing in such a way that the value of traditional knowledge is decreasing. The changing sea-ice conditions are increasing risk of travelling on and between sea-ice floes. For example, hunters must travel increasing distances to access the animal populations they are hunting (Ford et al., 2007; Hovelsrud et al., 2008) and a reduction in landfast ice in the western Canadian Arctic combined with more open water has resulted in unpredictable fog events, making coastal travel treacherous and communication between communities more difficult (Barber et al., 2008b).

An increase in human activities such as fisheries, tourism, and industrial development will be followed by an increased demand for seasonal icebreaking of harbors and shipping routes. Feedback effects on the environment through increased shipping and industrial activities (pollution, habitat fragmentation by icebreakers, noise) represent a major challenge for local communities (Koivurova and Hossain, 2008). Regulation will be needed to avoid disturbing hunting activities and destroying travel routes over the sea ice. At the local level, less sea ice may lead to increased transportation by boat, which may lead to higher fuel costs, straining the household economy. Changing sea-ice conditions and increasing concern about hunter safety create the need for costly new technology (GPS, satellite phone). Increased use of technology may further erode the utility and transfer of traditional knowledge, with major social and cultural implications (Ford et al., 2010; Pearce et al., 2010).

The projected changes in sea-ice distribution and extent and the formation of polynyas will affect the distribution and ranges of many species, directly through habitat change and indirectly through changes in predator-prey relationships. Such changes will have consequences for local communities dependent on a mixed cash-subsistence economy. The trend of shorter seasonal duration of landfast ice and poorer ice quality is expected to continue. While there is great uncertainty about the type and rate of changes, it is likely that hunting and travelling activities will be increasingly affected, compromising personal safety, food security, and culture.

Changes in the availability of traditional/local foods and in food security may have negative impacts on well-being, physical and mental health, family and network relations, and the social and cultural fabric and identity, and may increase lifestyle diseases such as diabetes and obesity (e.g., Ford and Berrang-Ford, 2009). Female community members have been identified as being at particularly high risk of food insecurity (Beaumier and Ford, 2010). However, changes in sea ice could also improve access to resources such as fish, and combined with more rapid technological development this may positively affect harvesting activities and income. Nevertheless, and despite the uncertainties in future projections, the changes in sea ice are likely to require adaptation strategies that go beyond those that are currently taking place. The projected increasing sea-ice variability and the decreasing predictability of conditions, combined with social, economic and political drivers of change may increase vulnerability of local communities and peoples.

The decline in sea ice may result in rapid changes in Arctic weather, and severe weather events are projected to become more prevalent in future (Kolstad and Bracegirdle, 2008). Strong wind events can have major effects on coastal sea ice and may create open-water areas that affect travel and subsistence activities on the ice. It can also lead to open water and the formation of low clouds and fog which, when formed during autumn and early winter, can result in freezing drizzle and significant icing on aircraft and vessels (Hanesiak et al., 2010). The North-East Atlantic is particularly prone to marine cold-air outbreaks (MCAOs). As sea-ice retreats, the highest frequency of MCAOs will also retreat northward, resulting in a decrease in the strength of MCAOs along the densely populated coastlines of Norway, Iceland, the British Isles and northern Europe in general. Initially, traditional sea-ice use and commercial activities in the Arctic (fisheries, oil and gas industry, shipping) will be more exposed to storms and icing, but over the long-term this type of weather will move northward following the retreat of the sea ice (Kolstad and Bracegirdle, 2008).

9.4.5.1.2. Fisheries

Although global aggregated marine primary production is not expected to change substantially over the next four or five decades, there is a stronger basis for projecting changes in primary production at the regional level. This is supported by observational evidence, particularly for the North Pacific and North Atlantic, but also for the Bering Sea; a general northward shift in species is expected with increased biomass production in Arctic waters and associated declines in fish species with more southerly distributions (Sherman et al., 2009; WMO, 2009). However, species are limited in how far north their range distributions can shift, and there is potential for the extirpation, and even extinction of some species (Loeng et al., 2005). A northward shift in lower trophic-level species will impact higher trophic-level species, and ultimately humans, while species in the marginal ice zone may benefit from earlier ice break-up (Anisimov et al., 2007). The impacts of changes in sea ice on future Arctic productivity cannot be considered in isolation from the other changes taking place in the region. Ecosystem responses will vary regionally, depending on oceanic and climatic forcing, and require consideration of many rate and yield-limiting processes.

Qualitative changes in species composition as a result of northward migration may lead to increased fish stocks regardless of changes in sea-ice related primary production. Together with wind-induced flux of warm Atlantic water into the region, changes in stratification, light penetration, salinity, pH and oxygen in the upper water masses, will act directly on the abundance, distribution and composition of key commercial fish species and plankton species in the Arctic (Stenevik and Sundby, 2007; Brander, 2010). New species moving into the Arctic could lead to competition with indigenous species, new predators, diseases and parasites (Furgal and Prowse, 2008).

The implications of change may differ for coastal and industrial fisheries. Sea-ice retreat will increase the area of open water and this may have implications for food security through a decline in ice-associated coastal fishing, such as Greenland halibut long-lining on the frozen fjords (Côté and Williams, 2008). Less sea ice may also lead to greater fetch and

storminess, resulting in greater risk for coastal fishermen at sea and economic losses for those staying on shore (West and Hovelsrud, 2010). On the other hand, a decline in sea ice may generate new possibilities for coastal fishing from small boats and a larger seasonal window for fishing, but this will come at the cost of new investments. A shift in the freshwater budget is likely to result in Greenland halibut moving from deep waters to shelf areas, requiring changes to the fishery (Loeng et al., 2005).

For industrial fisheries, a decline in sea-ice dependent fish species but an overall increase in pelagic productivity in the coastal seas and improved access to the Arctic Ocean could mean a substantial net benefit over the long-term. This is dependent on which fish species flourish under the new conditions and how management regimes and policies respond (e.g., Molenaar and Corell, 2009b; Sherman et al., 2009). Anisimov et al. (2007) suggested that increased productivity in the marginal ice zone is likely to benefit many commercially important species. Biomass is likely to increase for key commercial species in the Barents Sea (e.g., Atlantic cod and herring) and these species are likely to shift their centers of summer distribution eastward in response to sea-ice loss, thus increasing their availability to Russian fishing fleets. There is also the potential for an eastward colonization of new spawning areas, and this may forecast a wider geographical distribution for these species (Stenevik and Sundby, 2007). With increased importance of the Norwegian and Russian sectors of the Barents Sea, it is likely that fishing rights, cooperative management plans and catch limits will need to be renegotiated.

While the opening of new areas earlier in the season may offer new opportunities, changes in the sea-ice regime may also pose challenges to coastal fisheries in terms of drifting ice and icebergs. Other concerns relate to access to and sustainability of species including Greenland cod (particularly in Greenland), Pacific salmon (western North American Arctic), deep-water shrimp (Greenland), and pollock (Bering Sea). Fish stock abundance is the result of a combination of human and environmental factors. As it is difficult to predict many of the consequences of sea-ice change on commercial stocks, the implications for fisheries are extremely uncertain (Mueter and Litzow, 2008).

9.4.5.1.3. **Shipping**

The decline in sea ice will progressively open longer seasonal windows for navigation through Arctic waters, increasing the opportunity for trans-polar commercial shipping and considerably shortening the route between Europe and the Pacific (Figure 9.39). Climate models project that the loss of summer sea ice will extend the navigation season in the Northern Sea Route from three to six months and from two to four months for the Northwest Passage by the end of the 21st century according to the IPCC A1B emissions scenario. Through the 21st century, models indicate a steady increase in the Northern Sea Route navigation length of about 4.5 (±1.3) and 3.6 (±1.7) months averaged over the 'best' and all models, respectively. This suggests that transit via the Northern Sea Route from Western Europe to the Far East may be up to 15% more profitable in comparison to Suez Canal transit by the end of the 21st century (Khon et al., 2010). Such an increase in navigation season length and the increased potential for Arctic shipping will potentially save billions of dollars a year (Borgerson, 2008). Ships will reduce travel distance by 40% (compared to current shipping routes) by using the Northern Sea Route, and it is expected that the transit transport from Russia will increase. This could save energy, reduce emissions, promote trade, diminish pressure on the main trans-continental navigation channels and have great economic benefit (Commission of the European Communities, 2008).

Sea-ice decline recently made it possible for two merchant ships to pass through the Northern Sea Route, albeit with icebreaker support (Barents Observer, September 2009). The predicted volume of transport through the Northern Sea Route, primarily associated with transporting natural resources, gas and oil, may increase to 5.5 million tonnes in 2010, and to 12.8 million tonnes by 2020 (Ho, 2010).

While less ice in the Arctic Ocean will increase the potential for naval and maritime operations, several obstacles remain, including seasonal drift ice and icebergs, annual variations in ice cover in the central Arctic Ocean, and local sea-ice formation along sea routes. However, this increased potential must be seen
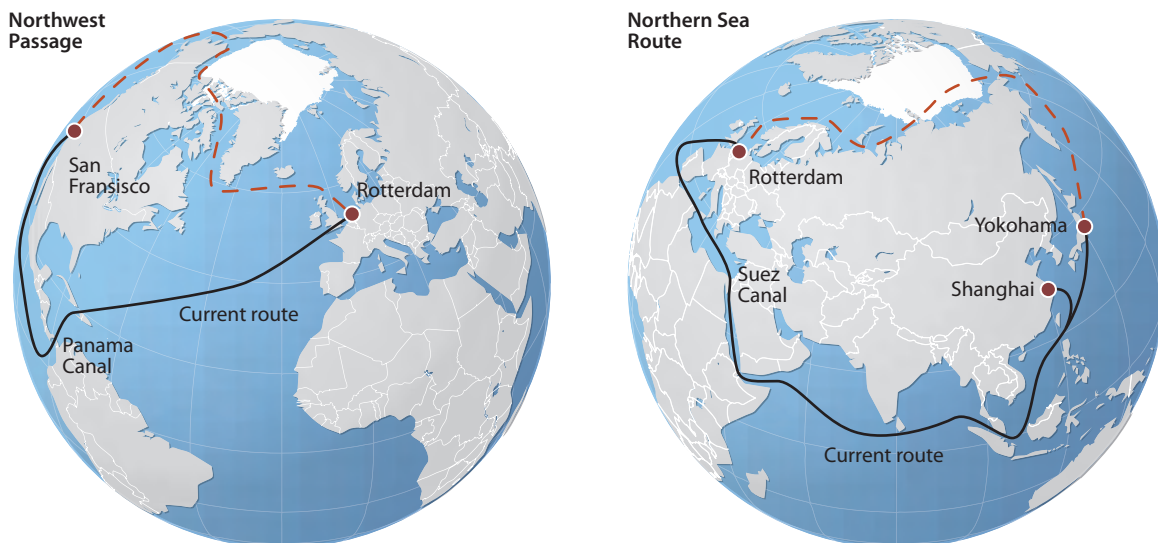

Figure 9.39. Northern Sea Route and the Northwest Passage compared with currently used shipping routes.

within the context of factors such as global demand and prices for natural resources, changes in ship design and technology, a growing world economy, and the size of vessels compared with the capacity of the Panama or the Suez canals, as well as the increasing emphasis on shipping due to diminishing numbers of ice roads (Mathias et al., 2008). Although the sea routes along the Siberian coast may open earlier, the depth of water along the coast may limit the size and freight capacity of the ships (Johannessen and Pettersson, 2008). The East Greenland Current, the main current leaving the Arctic, will continue to transport sea ice long distances southward along the coast of Greenland and through Icelandic waters (Jakobsson, 2003). In the United States, the Coast Guard polar icebreakers are aging and need replacing, and the country has limited capacity and ability to expand and operate in the Arctic environment (US National Ice Center and US Arctic Research Commission, 2007). Other challenges include a lack of infrastructure, environmental risks, and uncertainties about future trade patterns. Ships will need to deal with an increase in storm surges, wave heights and wind direction, resulting in icing on equipment, hull, propulsion systems and appendages (Molenaar and Corell, 2009a). As a result, there is a higher risk of shipping incidents. Arctic marine shipping may also cause adverse impacts on the marine environment through accidental discharges of polluting substances, operational discharges and emissions, navigation impacts on marine species, the introduction of non-indigenous organisms through ballast-water release or attached to vessel hulls, and icebreaking (e.g., physical habitat effects) (Molenaar and Corell, 2009a). Due to the remoteness of the area, the response time will be relatively long once shipping incidents do occur and may be inadequate to address impacts to the marine environment and marine biodiversity.

Risk-assessments and the marine insurance industry are likely to be crucial factors in the economic viability of Arctic marine shipping (VanDerZwaag and Chircop, 2008; Molenaar and Corell, 2009a). In particular, the following issues need to be resolved (Ho, 2010):

- Environmental monitoring and forecasting services providing meteorological, oceanographic and sea-ice information to support shipping all year round will need to be significantly enhanced.

- Search and rescue as well as icebreaker support services, with increased seasonal and regional access, will need to be provided in a comprehensive manner.

- Experienced mariners trained for Arctic operations and accurate hydrological charts are needed to operate the ships.

- New ship technology is required for independent ship operations in ice-covered waters, which will remain present for most of the year.

- When traffic gets heavier, it may be necessary to implement vessel traffic systems along narrow straits that may end up as choke points, such as the Barents Strait.

- An integrated governance and regulatory framework is needed based on the United Nations Convention on the Law of the Sea (UNCLOS).

Different solutions have been investigated for building safe and cost-efficient transport chains. One solution is to use dedicated ice-strengthened vessels for the ice-covered part of the route and to apply ship-to-ship transfer in open water (e.g., for commodities such as liquefied natural gas). Policy objectives include the development of infrastructure (on land, for air support, and for dredging of Arctic ports), use of satellite navigation systems for improved maritime surveillance, and promotion of full implementation of existing obligations concerning navigational rules, maritime safety, route networks, and environmental standards in the Arctic, in particular those drawn up by the IMO. The potential to provide specially-designed, environment-friendly ships, including icebreakers, is an important asset for the future (Commission of the European Communities, 2008).

Although the Arctic may be navigable before 2050, and routes along the coast of Siberia even earlier, routine ocean-to-ocean commercial navigation is not expected to develop significantly for quite some time (US National Ice Center and US Arctic Research Commission, 2007). More investment is required in infrastructure and the provisioning of marine services to ensure safe and secure ship transit with minimal environmental impact (Ho, 2010). For the near future, Arctic shipping is expected to take place largely between points within the Arctic rather than ocean-to-ocean transit, and commercial maritime activity in the Arctic will probably be driven more by economics, dominated by the development of natural resources and tourism, than by declining sea ice. High prices for hydrocarbons will be an important driver of shipping activities, because of the cost-benefits of shorter trans-Arctic shipping routes and the expected exploration and exploitation of hydrocarbon resources in the Arctic marine area. Increased tourism to see 'iconic species' before they become extinct may also drive an increase in Arctic traffic. As the Arctic sea ice thins, the difference between Arctic residents' view of change and the shipping industry's view of opportunity may become more pronounced. Future expansion of Arctic marine shipping is likely to attract diverse stakeholders with different interests in the Arctic, from countries outside the eight Arctic nations, and rules of engagement with northern peoples will need to be renegotiated. Trans-Arctic marine shipping is expected to be an important driver of this development (Commission of the European Communities, 2008; Molenaar and Corell, 2009a).

9.4.5.1.4. Non-renewable resources

The United States Geological Survey (USGS, 2008) concluded that about 30% of the world's undiscovered gas and 13% of the world's undiscovered oil may be found north of the Arctic Circle (1669 trillion cubic feet, 90 billion barrels of oil, 44 billion barrels of natural gas liquids), mostly offshore under less than 500 m of water. Undiscovered natural gas is three times more abundant than oil in the Arctic and is largely concentrated in Russia. The study included only those resources believed to be recoverable using existing technology, but with the important assumptions for offshore areas that the resources would be recoverable even in the presence of permanent sea ice and oceanic water-depth. More than 70% of the mean undiscovered oil resources are estimated to occur in five provinces: Alaska (the largest with 31%), Amerasia Basin, East Greenland Rift Basins, East Barents Basins, and West Greenland / East Canada. More than 70% of the undiscovered natural gas is estimated to occur in three provinces: the West

Siberian Basin, the East Barents Basins, and Alaska. However, these first estimates are, in many cases, based on very scant geological information, and understanding of Arctic resources will certainly change as more data become available (Gautier et al., 2009).

A loss of sea-ice cover will affect offshore exploration drilling in several ways. Key challenges for the future include (AMAP, 2007; Efthymiou and Winkler, 2009; Seidlitz, 2009):

- Combination of water depth beyond the feasible range of GBS platforms and moderate to heavy ice conditions

- Combination of sea ice and icebergs

- Potential for increased wave action and storm surges

- Narrow weather and seasonal window for exploration and development

- High costs (although variable between regions and onshore and offshore structures generally much higher than equivalent southern developments) due to improving design, infrastructure, logistics and transportation, waste management, labor productivity, etc.

- Hard, rock-strewn seafloors

- Emergency response (spills, safety and rescue)

- Winterization of facilities including design for fire and blast footprint reduction.

Arctic hydrocarbon development is viewed by the industry as an opportunity to develop new technology to improve efficiency and reduce costs. Advances have already been made in current focus areas of Arctic subsea processing, compression and power transmission, drilling and pumping technologies, and communication technology. Other focus areas are ice-load and ice-scour prediction, offshore Arctic drilling systems, Arctic tanker technology and emergency, evacuation and spill response systems and technology, and cost effective logistics. Although subsea production has potential, given the right reservoir conditions and sufficient open-water season for drilling, inspection, maintenance and repair operations, challenges remain. These include flow assurance, protection from deep-keeled ice pressure ridges and icebergs, and inaccessibility during the ice-bound season.

Hydrocarbon discoveries will change the economic landscape and way of life of local inhabitants. Addressing the regulatory and stakeholder issues, including the study of impacts on environment and local communities, should be a key focus area for industry, communities, and government (Seidlitz, 2009). Local public resistance to the locations of large-scale, obtrusive facilities, such as oil refineries and GTL (gas to liquids) plants, is an issue to be resolved (IEA/OECD, 2009), although pragmatism is illustrated by the quotes below:

> "*Climate change will be very beneficial to society [in Greenland]. It will improve fishing and above all make it easier to drill for oil and gas. The US Geological Survey calculates that the greatest unused oil reserves on Earth are in the Greenland waters – and they are in the east, where the ice is melting fastest.*" Jesper Madsen, DMU environmental research institute in Roskilde, Denmark, in The Times, May 7, 2008.

> "*The agenda is the agenda of the oil companies, we cannot stop the development but we want a part of the benefits and want to play a role in resources development.*" comment of Geir Tommy Pedersen, head of the Sami Council, to AFP on the sidelines of the Arctic Frontier Conference in Tromsø, Norway, January 2008.

Impacts of oil and gas development on people, communities and governments can be both positive and negative (AMAP, 2007). They could drive social and economic change, create employment opportunities and revenues, impact households as workers may be away for long periods, and many of the jobs will be filled with employees from outside the region. Impacts on the environment and traditional practices are very likely to be disruptive. In many countries, attempts to counteract harmful effects on the environment have resulted in the development of environmental regulations and restrictions for where drilling may occur. In the United States, for example, drilling has not been allowed on offshore coastal zones for many years. Environmental catastrophes in ice-infested environments (e.g., shipping spill, oil well blow-outs) are not likely to be contained within the same season in which they occur, but will require remedial action at least into the following season. Low temperatures may further exacerbate the clean-up because the material (especially heavier crude oil) congeals rather than breaks down. While the oil and gas industry in general does not consider the consequences of climate change as a major concern, the local installations are sensitive to the expected increase in storm surges, increasing wave heights, change in wind direction, and rising sea level, which will have consequences for icing on equipment and installations and for environmental safety (West and Hovelsrud, 2008).

Although oil and gas activity in the Arctic is projected to increase (AMAP, 2007), risk of oil spills, spills remediation costs, and the likelihood of further changes in environmental regulations discourage investment in Arctic oil facilities. Development will depend on market conditions, technological innovation, the sizes of undiscovered accumulations, and regulatory issues such as the US presidential decision (May 2010) to extend the moratorium on new offshore hydrocarbon developments as a result of the Gulf of Mexico incident. In combination with the Arctic weather conditions and enormous investments in infrastructure needed to develop new fields in the Arctic region, exploitation for oil or gas in these areas is unlikely to start much before 2020 (IEA/OECD, 2009). The estimated oil resources therefore, if found, would not come into production at once but would be added to reserves and produced incrementally over decades (Gautier et al., 2009). This is also reflected in recent model projections (Lindholt and Glomsrød, 2009) for various Arctic regions, which indicate that although global gas production will increase in most regions outside the Arctic up to 2030, the Arctic share of global gas production will actually decline from around 21% in 2008 to around 9% in 2030. Projections for oil likewise indicate that the Arctic share of global oil production will drop from about 10% to roughly 7% in 2030. Until 2030, the vast majority of Arctic oil and gas will be Russian, 81% and 94% respectively, while around 14% of the oil will come from Alaska.

9.4.5.1.5. Tourism

Although facilitated by improved accessibility, Arctic tourism is likely to increase independently of sea-ice change. Latent demand and the tendency for cruise visitors to make return visits form the basis of the predicted moderate increase in cruise tourism (Stewart and Draper, 2006). On the other hand, cruise ship tourism worldwide is developing rapidly, and loss of sea ice may open up waterways and possibilities for increased cruise traffic (Glomsrød et al., 2009). However, several recent accidents have shown that sea ice, as well as response time in the case of accidents, remains a hazard (Stewart and Draper, 2008). To counteract the hazards, icebreakers are used to assist cruise ships that are not ice-strengthened. The increasing volume of icebreaker and cruise traffic is of local concern. Similar to Arctic shipping, accidents and lack of search and rescue infrastructure are key issues. In addition, other types of life-saving technologies used in the North are being questioned, such as how long a basic lifeboat can permit survival in ice-filled Arctic waters.

> "And if it can't [permit survival in ice-filled Arctic waters], why are we letting all the cruise ships go up north with lifeboats on them?" Mary Williams, director general of the National Research Council of Canada's Institute for Ocean Technology, CBC news, June 5th, 2008.

Loss of Arctic sea ice will affect wildlife habitat, and the numbers and migratory behavior of wildlife species. To minimize further impact, careful consideration of extended seasons, cruise tourism transport schedules, and fisheries conditions are of critical importance to ensure the cultural and economic well-being of Arctic people (UNEP, 2007). In addition, there are environmental, economic, social and cultural impacts ranging from wildlife disturbances (e.g., noise, habitat change) to garbage, waste and pollution problems that need to be regulated. While the potential economic benefits from tourism increases for local communities, such an increase may also negatively affect those communities. For example, ice-dependent polar bear populations will, with reduced sea-ice, have increasingly smaller habitats and with more people in the area human-bear interactions will increase. Given international pressure for sustainable management of remaining populations, indigenous communities depending on hunting polar bears (for tourism or subsistence) may have to adopt alternative livelihoods (Hovelsrud et al., 2008). Further sea-ice melt and expansion of tourist seasons may result in extended use of infrastructure and longer duration of employment and income benefits. How Arctic communities allow their natural and cultural resources to be used by large numbers of visitors and whether or not this is viewed as a disruption, depends on the attitude and resources of the community. Impacts may be mitigated through collaborative management with tour operators and in some instances, tax revenues and special fees may offset local costs.

Although there are many protected areas on land in the Arctic, there are few if any marine protected areas (UNEP/GRID-Arendal, 2007). Opening up of waterways and further development of Arctic tourism should therefore take place in a sustainable manner. An EU policy brief (Commission of the European Communities, 2008) proposed the promotion of sustainable Arctic tourism through three main policy objectives: protecting and preserving the Arctic in unison with its population, promoting sustainable use of resources, and contributing to enhanced Arctic multilateral governance. Two main considerations pertain to both protection of the environment and ensuring benefits to local coastal communities. Concrete actions include support that increases cruise ship safety, improves guiding, restricts access to highly vulnerable areas, and encourages environmentally friendly tourism, involving local communities. This corresponds well with the proposed principles and a 'code of conduct' for tourists and tour operators (UNEP/UNWTO, 2005), stipulating guidelines for ensuring good management, income for communities, and protection of culture and environment, which are of key importance in an increasingly accessible Arctic.

9.4.5.2. Overarching issues

9.4.5.2.1. Coastal change and infrastructure

Sea ice can both prevent and cause erosion of coast and infrastructure, and plays an important role as a regulating element in coastal sediment dynamics. Sea ice protects the coast from the erosive action of storms (Atkinson, 2005) and preserves the thermal state of subsea permafrost (Solomon et al., 2008). ACIA (2005) reported that coastal erosion rates along the Arctic coast have increased over the past 30 years. Low-lying coastal plains, which are not tectonically active, are especially vulnerable to coastal erosion. Although rocky (lithified) coasts predominate in the European and western Russian Arctic and in parts of the Canadian Arctic Archipelago, human settlements are often associated with unlithified sectors of the coast because these provide more suitable locations for human activities. Rapid erosion is typically associated with unlithified, frozen sections of the Arctic shoreline, and projections of sea-ice decline and progressively longer open water seasons imply increasing exposure of Arctic coasts to erosion. Coastal erosion rates vary considerably between and within regions and over time, and erosion presents a significant problem for communities, infrastructure of various types, cultural heritage sites, and in some cases protective coastal landforms.

The complex interactions between declining sea ice and other consequences of climate change (rising sea level, shifting river discharges, run-off, altered sedimentation rates in coastal areas, permafrost degradation) will increasingly affect Arctic coasts, coastal infrastructure and coastal marine ecosystems, and potentially human resource use (Martin et al., 2009; see also Chapter 5). Major forcing parameters are waves, currents and water levels, and, especially for the Arctic, sea surface temperature, salinity, decreasing sea ice and increasing open-water fetch, ground temperatures, and excess ground-ice content. With sea ice forming later in the season, the coast is more exposed to a projected increased number of autumn storms, and to storms with longer fetch. The subsequent coastal erosion may threaten some Arctic communities (Forbes et al., 2008; Perovich and Richter-Menge, 2009). Even a small increase in the intensity of storms and coastal surges is likely to increase infrastructure damage and costs substantially (Stern et al., 2006). Recent time-lapse photography of crumbling Alaskan coastlines shows the combined eroding action of declining sea ice, warming ocean waters, and wave action (Livescience, 2009).

The erosion rates from these forces are greater than anything seen along the world's coastlines, resulting in the coast between Alaska's Point Barrow and Prudhoe Bay eroding by several tens of meters per year.

In addition, a decrease in landfast ice increasingly exposes coastal permafrost to wave action and increased temperatures, leading to thaw (see also Chapter 5). This can have serious consequences for existing infrastructure (e.g., structures or buildings depending on ice-bonded surface for strength could be at risk over the decadal time scale). Although new coastal permafrost may form when sea ice freezes to newly aggraded sediments, it is likely that this permafrost will be unstable because of increased temperatures of the seabed. Rising ground temperatures will increase seasonal thaw depth and enhance coastal erosion processes. Thermal niche development and ice-wedge thaw can promote rapid failure of low coastal bluffs even in the absence of significant wave action (Hoque and Pollard, 2009).

Other hazards involve ice ride-up, ice pile-up and the formation of nearshore pressure ridges, primarily as a result of wind action, which may provide a mechanism for nearshore scour and landward sediment transport (Reimnitz et al., 1990). Ice ride-up has been reported extending more than 100 m inland (Taylor, 1978; Kovacs and Sodhi, 1980); Ogorodov (2005) cited a case of at least 150 m and suggested evidence for much more. While the effects are often superficial (Kovacs and Sodhi, 1980; Kovacs, 1983), they include an example of damaging ride-up of 0.4 m of ice that knocked a lighthouse off its foundation and destroyed standing infrastructure on fishing harbor wharves (Forbes et al., 2004). Ice pile-up in extreme events has been reported as high as 36 m (Sverdrup et al., 1904) and 50 m (Zubov, 1945). Pile-up typically occurs rapidly (~30 minutes), has been known to cause fatalities, and can cause significant damage, impinging on waterfront residences, breakwaters, navigation infrastructure, coastal roads, and even overtopping dunes (Forbes et al., 2004).

Because of the highly episodic and rare occurrence of damaging ice motion, such as ice ride-up and pile-up, little effort has been devoted to mitigate and forecast such hazards (Druckenmiller et al., 2009), and many communities where damage has occurred may not have anticipated the event before it happened. Current community responses to increasing ice-free fetch, longer open-water seasons, and increasing impacts from erosion run the gamut from protection (e.g., dynamic beach nourishment) to retreat. The situation in some communities is sufficiently dire that they are considering immediate relocation (e.g., Shishmaref: www.shishmarefrelocation.com). In other cases (e.g., Tuktoyaktuk, Northwest Territories, Canada, see Figure 9.40) phased retreat to a new location is an option which is now being considered (Johnson et al., 2003). While engineering solutions are available for shore protection, these measures may address one problem but create another by altering the dynamics of erosion and deposition processes (ACIA, 2005).

Some coastal landforms (e.g., spits and barrier islands) are highly sensitive to the balance between sediment supply and loss, such that rapid change can occur once a threshold is exceeded. Some barrier islands on the Alaskan Beaufort coast have diminished in surface area since 1955 and recent (1990–2007) migration rates of 24 m/y greatly exceed earlier

Figure 9.40. Beach erosion at Ukkuqsi, Barrow Alaska (upper). Photo courtesy of Anne Jensen. Erosion at Tuktoyaktuk, Northwest Territories, Canada (middle). The shore has eroded as much as 200 m in 55 years in some locations. Concrete and rock revetments now protect parts of the outer coast (lower), but erosion continues at the unprotected locations. Erosion of the uninhabited and unprotected island at the harbor mouth will threaten the long-term viability of the harbor within the next 50 years. Middle and lower photos: Martin Fortier/ArcticNet.

(1955–1990) rates of 5 m/y (Ravens and Lee, 2009). Longer open-water duration and increased occurrence of large waves (longer fetch), perhaps in combination with negative effects from permafrost degradation, may contribute to this coastal erosion (Martin et al., 2009).

Projected changes in climate and consequent oceanographic forcing include continued sea-level rise, decreased sea-ice extent and duration, and increased air and sea surface temperatures. Reduced length of the winter ice season (in areas with only seasonal sea ice) will limit the window for ice motion onshore, while increasing shore exposure to storm waves. In High Arctic regions, the progressive loss of multi-year ice will reduce the probability of thick ice pile-up, but overall thinning of seasonal ice may increase the frequency of ice movement onshore. Increased fetch may also expose coastlines to ocean swell – a condition to which many Arctic coasts have had limited exposure and which could alter shoreline processes. Although some recent studies (e.g., Jones et al., 2008) present erosion rate

measurements based on historical and modern observations covering the past 175 years, allowing detection of loss of both cultural and historical sites and modern infrastructure, those types of detailed analyses are still lacking for most of the Arctic coast. At present there are no predictive models that can be used to quantify projections of increased rates of coastal erosion, and although there are likely to be physical limits to the rate at which erosion products can be removed from the shore face, which will put an upper limit on erosion rates, sea-ice decline in combination with climate-change driven changes is likely act to increase the risk associated with coastal erosion.

9.4.5.2.2. Contaminants and human health

The Arctic serves as a major heat sink for the Northern Hemisphere, and wind and water deposited contaminants can be stored in sediments and sea ice (both multi-year legacy contaminants and depositions in younger ice). The relationship between sea-ice change and contaminants is difficult to ascertain, but an increase of contaminants in the Arctic is possible both through direct and indirect effects of melting sea ice. Melting of sea ice will result in a release of contaminants from ice and frozen sediments and possibly lead to their incorporation into food webs. Increasing rates of coastal erosion will increase input of contaminants sequestered there (e.g., mercury – Outridge et al., 2008). Using DDT and PCBs (polychlorinated biphenyls) as an example, it appears that melting sea ice will add little to the contaminant load of the Arctic Ocean (NCP, 2003; AMAP, 2009).

Conversely, a reduction in sea ice is likely to lead to increased transport of contaminants into the Arctic following a variety of pathways. A projected increase in precipitation due to an extended ice-free area in the Arctic Ocean will probably lead to increased transport of airborne contaminants such as PCBs, toxaphene and HCH (hexachlorocyclohexane) to seawater, freshwater systems and terrestrial environments (McMichael et al., 2003; AMAP, 2009). On the other hand, HCH might be removed faster than it accumulates through evaporation (Macdonald et al., 2005) and increased microbial activity in warmer water can reduce the half-life of HCH (Harner et al., 2000). An increased demand for oil and gas worldwide combined with increased access to Arctic resources is likely to increase transportation and resource development dramatically over the next 20 years, and will lead to an increased potential for spills, the appearance of invasive species via migration or originating from vessel ballast water, and of contaminant movement into the Arctic (Berner et al., 2005; Occhipinti-Ambrogi, 2007; AMAP, 2009).

Contaminant exposure in human populations is harder to predict. Côté and Williams (2008) highlighted the increased amount ('load') of chemicals from oil development in Arctic Canada that could increase the load of chemicals carried by ocean currents and accumulated in the fat of animals consumed by people. Concerns about organochlorine compounds and heavy metals in traditional/local foods are significant among Arctic peoples. The lack of infrastructure, the cold conditions and the potential for freezing of contaminants can complicate clean-up actions and are a risk to the environment. However, while oil, drilling fluids and drill cuttings can contain mercury, cadmium and lead in varying amounts and occasionally some persistent

organic pollutants (POPs) and radionuclides, quantities are very small and most wastes in modern operations are now 'contained' (e.g., through the use of on-site tanks, holding ponds, down-hole dumping) (AMAP, 2009). Past oil and gas activities in parts of Arctic Russia may have led to some health impacts; however, peer reviewed documentation is unavailable (AMAP, 2009). While human health can suffer from pollution and social disruption, the revenues of oil and gas activity can improve overall health care and well-being (AMAP, 2007, 2009).

Potentially increasing levels of contaminants in marine mammals and other Arctic species can also contaminate food webs relied upon by Arctic peoples (DFO, 2006). This can impact Arctic peoples' diets through biomagnification which has already resulted in an estimated ten-fold higher intake by Arctic residents of toxins in food (Berner et al., 2005). In addition, higher temperatures may allow certain disease vectors and microorganisms to survive longer over the winter causing changes in wildlife health, a concern where traditional/ local foods are consumed (Berner et al., 2005). However, the location, population size and availability of marine food species will change as temperatures rise and sea ice retreats, and the structure of the food web might also change as a result of invasive species (AMAP, 2009). These changes could increase or decrease human exposure depending on the concentrations of lipophilic compounds in food species and the extent to which different food species are consumed. Such issues may also affect marine mammal health, reproduction and populations, and thus also the diet of Arctic people. The cultural services of typically used resources may be lost, and to replace cultural associations with the practice of harvests and the sharing of foods is likely to be more difficult than to replace the actual nutrients of the foods themselves (Hovelsrud et al., 2008).

9.4.5.2.3. Search and rescue, environmental stewardship

Declining sea ice will improve marine access to the Arctic, increasing the length of the summer navigation season, and allowing more offshore oil and gas activity and commercial fisheries. All have economic consequences for the region and beyond as well as potential environmental impacts on the marine ecosystem and fisheries that may affect Arctic communities. These changes will combine with an increase in accidents, risk of oil/fuel/cargo spills (see Table 9.6 for an overview of accidents between 1995 and 2004), disturbance to wildlife, chronic habitat degradation through noise, bottom disturbance, pollution, and destabilization of 'fast ice' associated with marine traffic (Furgal and Prowse, 2008; AMAP, 2009). Increased human activity in general creates a risk of synergistic and cumulative effects, leading to more rapid impacts and potentially to reaching critical thresholds sooner, leaving less time for effective conservation action (Huntington et al., 2007).

Between 1995 and 2004, the majority of accidents in Arctic waters concerned fishing vessels and bulk or cargo vessels (77% and 19% respectively), and occurred mostly in the Bering and Barents seas (Table 9.6). The retreat of sea ice and subsequent increase in activity is also expected to lead to an increase in environmental accidents, with oil spills the greatest threat (AMAP, 2007). Oil tankers entail particular risks, since Arctic ecosystems are exceedingly fragile, oil degrades and dissipates

Table 9.6. Number of accidents and incidents in the circumpolar north during the period 1995 to 2004 as well as the type of vessel involved in the accident. Source: Accident Incident database http://arcticdata.is, accessed 18 February 2010.

Area / Type of vessel	Vessels lost	Fatalities	Fuel spill	Ice damage
Arctic Archipelago				
Bulk carrier / Cargo ship	0	0	1	8
Government vessel / Tug / Barge	0	0	0	2
Total	0	0	1	10
Arctic Ocean				
Government vessel	0	0	0	1
Baffin Bay/Davis Straight				
Bulk carrier / Cargo ship	0	0	0	4
Tanker	0	0	0	2
Fishing vessel	0	0	0	2
Passenger ship	0	0	0	1
Total	0	0	0	9
Barents Sea				
Fishing vessel	2	12	1	0
Bulk carrier / Cargo ship	0	0	2	0
Total	2	12	3	0
East Bering Sea				
Fishing vessel	3	22	6	0
Bulk carrier / Cargo ship	2	8	4	1
Passenger ship	0	0	1	0
Total	5	30	11	1
Hudson Bay				
Bulk carrier / Cargo ship	1	4	0	3
Oil/Gas service & supply / Tanker	0	0	2	1
Total	1	4	2	4
Norwegian Sea				
Fishing vessel	0	0	2	0
Cargo / Passenger ship	0	0	2	0
Tanker	0	0	1	0
Total	0	0	5	0
West Bering Sea				
Fishing vessel	3	21	0	0
Other				
Cargo / Container ship	1	2	1	0
Fishing vessel	1	1	0	3
Government vessel	0	0	0	1
Tug / Barge	1	3	1	0
Total	3	6	2	4
Grand total	14	73	24	28

very slowly at low temperatures, and long distances would render clean-up efforts expensive and time-consuming (Barber et al., 2008b; Byers, 2009; Figure 9.41).

A Global Maritime Distress and Safety System currently exists, and a Search and Rescue (SAR) convention has been in place since 1979, encouraging states to establish SAR regions, the pooling of facilities, establishment of common procedures, and training and liaison visits (Jensen, 2008). International regulation of vessel-source pollution is primarily organized by the IMO. Canada and the Russian Federation rely in addition on Article 234 of UNCLOS – entitled 'Ice-covered Areas' – for prescribing standards that are more stringent than generally accepted international rules and standards (Molenaar and Corell, 2009a). The Arctic states have also adopted several relevant bilateral and regional instruments on monitoring, contingency planning and preparedness for pollution incidents (e.g., The Protocol on Preparedness, Response and Co-operation to pollution Incidents by Hazardous and Noxious Substances, 2000); although not all Arctic states are parties to every agreement. The relevant IMO instruments primarily have a global scope of application and therefore apply to the entire Arctic marine area. The only IMO instrument that is specifically tailored to the Arctic is the non-legally binding (and currently under revision) IMO Arctic Shipping Guidelines (Molenaar and Corell, 2009a).

With the exception of northern Norway and regions in northwestern Russia (Hoel, 2009; Denisov and Mikhaylichenko, 2009), most Arctic marine areas are still missing modern navigation charts, port services, communications, salvage, SAR capability, adequate emergency prevention, preparedness and response and environmental response, and comprehensive weather and sea-ice information. The provision of salvage in Arctic waters will encounter difficulties due to remoteness, harsh climatic conditions, darkness for half of the year, relatively few ports that could provide even temporary repairs or a safe haven, lack of experience of salvors in ice-covered waters, the likely requirement for icebreaking and Polar Class vessels, lack of dedicated salvage depots, and the particular fragility of the Arctic environment (VanDerZwaag and Chircop, 2008). In addition, increased storminess will complicate SAR actions in the Arctic (Kolstad and Bracegirdle, 2008).

Although navigation of the Arctic may be possible before 2050 and hydrocarbon development may increase with increased accessibility, more investment in infrastructure and SAR services and more studies of impacts on environment and local communities are needed to ensure safe shipping and hydrocarbon exploitation with minimal environmental and community impacts (Chircop, 2009; Seidlitz, 2009; Ho, 2010). Rescue operations both by ship and by air need a larger geographical basis than is currently available to support predicted operation levels, and the International Convention on Oil Pollution Preparedness, Response and Co-operation, 1990 (OPRC) demands in addition on-board and on-land emergency response installations (including responding to ice damage) and calls for the establishment of oil spill combating equipment, training, and detailed plans for dealing with pollution incidents. Improved mapping, surveillance and monitoring (of shipping traffic and icebergs), disaster management (groundings/sinkings) and training of local SAR personnel are of key importance (OPRC, 1990; Chircop, 2009).
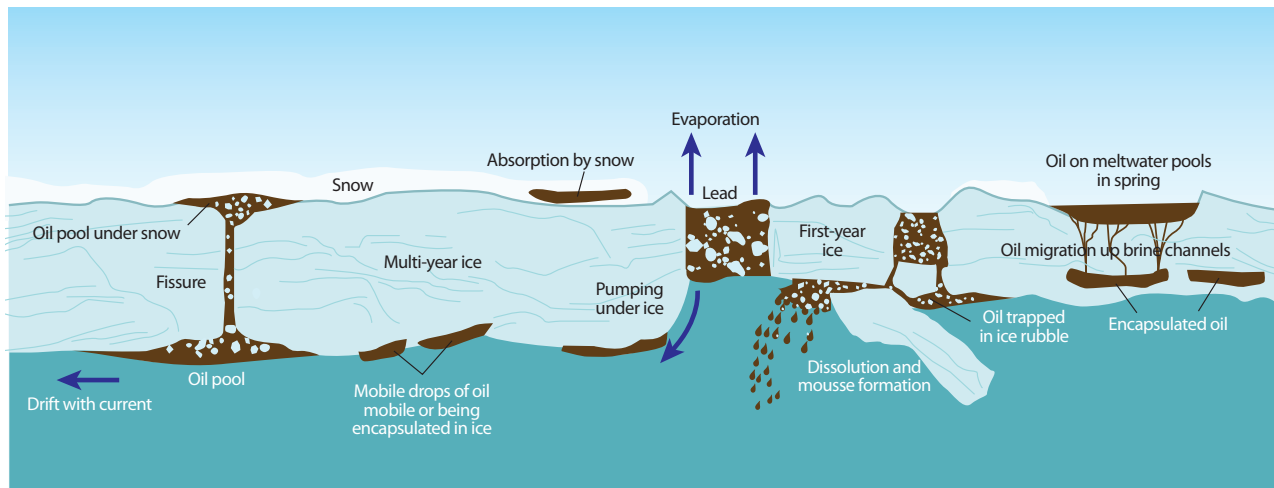
Figure 9.41. Schematic illustration of the behavior of oil in ice-covered water. Source: AMAP (2007; after Bobra and Fingas, 1986).

9.4.6. Governance

The institutional and governance dimensions of projected sea-ice change in the Arctic are profound, and must be addressed in connection with a number of cross-cutting issues affecting the region (e.g., Armitage et al., 2007; Huebert and Yeager, 2008; Young, 2009a). Oil, natural gas and mineral developments, airborne pollutants, sovereignty claims and security concerns, the socio-economic dimensions of renewable resource (fisheries, wildlife) extraction, and demands of northerners to have a greater role in decision-making as they adapt to changes all involve interrelated, multi-scale and complex issues of governance. While cross-cutting governance issues pertaining to the whole cryosphere are discussed in more detail in Chapter 10, particular sea-ice concerns are illustrated in Tables 9.5, 9.7 and Figure 9.38). Internationally, sea-ice change and its consequences will place additional pressure on the existing frameworks for cooperation developed by Arctic nations (see Young, 2005; Stokke and Honneland, 2007). The Arctic Council has been identified as a critical actor in terms of environmental governance (Young, 2005; Huebert and Yeager, 2008) in large part by enhancing the knowledge regarding the consequences of change in the Arctic.

A key governance mechanism is UNCLOS (current version, 1994). This agreement is the primary mechanism for dispute resolution related to rights of exploitation and navigational routes in international waters and provides a legally binding framework to resolve state-based conflicts (Huebert and Yeager, 2008; Stokke, 2009). In brief, it defines key territorial boundaries among coastal nations into zones that can be regulated and exploited. Uncertainty exists as to whether current agreements and governance regimes are adequate to respond to outcomes of rapid sea-ice change and its social, economic and political consequences. For some, it requires a new regional 'Arctic Treaty', which may ensure the sovereign rights of Arctic nations but with greater flexibility to respond to the outcomes of change (see Huebert and Yeager, 2008; Rothwell, 2009). The five Arctic Ocean boundary nations' 'Ilulissat Declaration' (May 2008), appears to be an effort to stem the rush for Arctic territory. While noting that UNCLOS is the correct mechanism to resolve Arctic Ocean interest conflicts it also notes that 'national implementation' of the mechanism's provisions is the means by which to address concerns. For others, existing arrangements (i.e., UNCLOS) already provide a foundation for resolving Arctic development challenges catalyzed by sea-ice change (Stokke, 2009; Young, 2009a,b).

As sea ice retreats, Arctic countries' efforts to secure sovereign claims to Arctic territory raise questions regarding potential militarization, fishery rights and natural resource development. One current debate among Arctic nations is the 'natural prolongation' of land territory defined as the continental shelf. The nations do not rule out an Arctic governed by adaptive flexible institutions designed to enhance equitable access to resources, but signal a top-down sovereignty-based solution to the diminishing sea-ice problem rather than solutions that might come from the sub-national needs of sea-ice service users across the Arctic as a whole. The consequences of sea-ice change for local communities (indigenous and non-indigenous) raise governance issues concerning involvement of local peoples in non-renewable resource extraction and the co-management of natural resources (Ayles et al., 2007; Kofinas et al., 2007; Agrawal, 2008; Armitage et al., 2009; Berkes, 2009; see also Chapter 10 for detailed discussion).

Several conclusions can be drawn. First, there are clear limits to any Arctic governance regime (regulatory or soft-law), because the drivers of sea-ice change occur largely outside the region and fall beyond the jurisdiction of any one Arctic state. Second, there is a need for effective interplay across societal levels and scales between Arctic institutions and existing issue-specific regimes (Stokke and Honneland, 2007; Young, 2009b). Finally, there is a clear need for further monitoring and the use of monitoring data in knowledge co-production and shared learning. Effective monitoring connects with increasingly well-established community-based approaches in which local resource users and others collaborate with scientists and managers in sharing observations of change (Riedlinger and Berkes, 2001; Laidler, 2006; Bravo, 2009).

9.4.7. Adaptation

ACIA predicted unprecedented climate change for Arctic regions, and changes are being documented by instrumental records and local and indigenous observations (e.g., Huntington and Fox, 2005; McBean et al., 2005; Gearheard et al., 2006; Hovelsrud and Smit, 2010). The Arctic Human Development Report (AHDR,

Table 9.7. Cross-cutting issues of sea-ice change and the consequences for peoples and human activities.

Consequence of sea-ice change	Consequences for peoples and human activities
Increased marine and air access	Requires new transportation systems and routes which will pass through areas currently used for subsistence harvesting or otherwise impact natural systems people depend on for provisioning services
	Higher accessibility will make resource development of oil and gas, fisheries, freshwater, and forests linked by rivers to sea ice less expensive
	Indigenous and rural Arctic peoples will experience disruption to cultural patterns and practices of well-being, cultural transmission, and access to land and sea. These disruptions will be perceived differently by different people
	New areas may become attractive for fishing. For some of the Arctic High Seas waters there is not yet an international conservation and management regime in place. This might lead to unregulated fisheries (McIlgorm et al., 2009)
	Increased access and use will require better weather and ice forecasting; more comprehensive charting and related infrastructure development and updating
Increased and cheaper access across the Arctic	Conservation measures may be harder to enforce (see AHDR, 2004: resource governance chapter)
Increased variability in the physical environment	Requires increased preparedness (e.g., weather prediction accompanied by better monitoring/ forecasting) and increased need for engineering to account for a hyper-variable environment

2004) demonstrated that Arctic peoples are susceptible to changing environmental conditions, and are already having to adapt. Among locally relevant climate related changes are projected increases in temperature and precipitation, reductions in sea-ice extent, and increases in the frequency and magnitude of hazardous conditions, including those associated with permafrost thaw, sea-ice stability, and increasing exposure to storms along the Arctic coast (Couture et al., 2002; Johannessen et al., 2004; Kattsov and Källén, 2005; Christensen et al., 2007; Barber et al., 2008b; Sou and Flato, 2009; Zhou et al., 2009). These changes have major implications for ecosystems and for people's livelihoods and wellbeing, and occur in the context of ongoing social, cultural, economic, and political transformations in northern communities (Fenge, 2001; Ford and Smit, 2004; Anisimov et al., 2007; Rattenbury et al., 2009).

Current observations show that sea-ice change has major impacts on the residents of the circumpolar region and that there is need for adaptation to changing conditions, both to new opportunities and challenges. The consequences of changing sea-ice conditions for local communities occur across a number of dimensions, including: resource availability; resource management policy and regulations; non-local or conflicting interests in resource extraction; limited economic opportunity and limited access to markets; demographics; attitudes and perceptions of change; local-global linkages; infrastructure; threats to cultural identity and well-being; transfer of local and traditional knowledge; and flexibility in economic and livelihood terms, and enabling institutions. These are in many cases found in combination and may also combine across scale and sectors. Such combinations may act as facilitators or constraints on adaptation (Hovelsrud and Smit, 2010). The complexity of local adaptation to sea-ice change is further illustrated by the different concerns within a community, within and between sectors or stakeholder groups; these concerns involve both direct and indirect consequences of sea-ice change for a local community and that it is a challenge for politicians and policy-makers to plan under uncertainty and across borders.

The differential vulnerability to sea-ice change in the Arctic is consistent with findings from other regions (Smit and Wandel, 2006; Leichenko and O'Brien, 2009). In an Inuit community for example, groups vulnerable to sea-ice change include those dependent on hunting for livelihoods and food, young Inuit who use the ice for hunting and travel but do not have the depth of understanding to do so safely, and households with limited financial means to afford to adapt. Many adaptations take advantage of new conditions and minimize impacts by using new technology, changing resource use patterns, and/or switching to the consumption of store-bought foods and in many cases this involves financial investment. The ability of households to make this investment is constrained by limited financial capacity, the national tax systems, and other institutional constraints. There are also groups in a community unlikely to be affected by changing ice conditions, including those who do not hunt, are not dependent on products from the hunt, and/or have access to financial resources to adapt.

While research has documented significant vulnerabilities in Arctic communities to sea-ice change, recent work has also noted emerging adaptive capacity which is moderating some of the negative impacts. Social learning, responsive local institutions, livelihood flexibility, and adaptive management have emerged as critical to adaption to changing conditions, and which are helping reduce the dangers of using the ice in a changing climate (Ford, 2009b; Ford et al., 2009; Hovelsrud and Smit, 2010).

Adaptation to sea-ice change is predominantly a local concern, because the changes are felt at the local level first and by peoples and communities whose livelihoods and well-being are closely linked to the local environment and weather. However, other human activities also need to adapt. The development of the IMO 'Polar Code' for ships in the Arctic is one example of adaptation occurring at the international level. The level of awareness of the need for adaptation, and the development of adaptation strategies to Arctic conditions on behalf of extractive industries will have consequences for local vulnerability. Without proper guidelines and regulations in place for industrial development and resource use such as commercial fisheries, cruise traffic and shipping to adapt to changing ice conditions, local communities become more vulnerable.

Adaptation and development of adaptation strategies occur along several dimensions of uncertainty. First, there is uncertainty in the global and regional circulation models used to project future climate change. This is due both to uncertainty in the natural climate variability and in how the climate will behave in the future as a result of human influences. Second, there is uncertainty in determining the different drivers of change (climatic, social, economic, political) and in their interactions and combined impacts on human society. Third, there is uncertainty in understanding the adaptive capacity of a community, a region, a nation, and internationally. Politicians and decision makers will need to consider various levels of uncertainty with respect to adaptation to climate change. Even though existing knowledge is in many respects incomplete there is enough knowledge to understand current and future trends and changes for policy development, and for improved management of fisheries and marine ecosystems (Brander, 2010).

In summary, local focus and traditional/local knowledge are essential in studies *on* and *for* adaptation. There are variations in adaptation needs and strategies between sectors and different levels within the local community. Local understanding of change and adaptation needs is generalizable through the transfer of practical knowledge, and to help increase the scientific knowledge base. More and integrated knowledge about adaptation is needed across scientific disciplines and local/traditional knowledge, and between scientists and decision-makers and on the international arena.

9.4.8. Feedback of human activity to sea-ice change

Sea ice may be viewed as a social-ecological system with a variety of processes coupling the physical, biological, and social components. While a reduction in sea ice will lead to increased accessibility in the Arctic, the substantial increase in human activities may also impact the sea-ice services derived from the ice cover. Comprehensive, overarching surveys of activities and the magnitude of their effects on the sea-ice cover are comparatively few, and typically confined to specific topics, such as the impact of airborne pollutants on sea-ice ecosystems (e.g., Macdonald et al., 2000; Macdonald, 2007; AMAP, 2009). However, local effects have often been examined in greater detail in the context of environmental impact statements that are part of proposed industrial development in Arctic regions (e.g., IMPVA, 2008).

At the local level, feedbacks are mostly expected in the context of modifications of the coastline that are conducive to enhanced entrainment of sediments into sea ice. Both seasonal ice formation and ice melt at the local and regional level are strongly impacted by the influx of freshwater from terrestrial runoff (e.g., Macdonald et al., 2000). Moreover, large-scale river discharge and potential modifications, such as through damming of large rivers can impact stratification and hence impact ice production rates (Cattle, 1985). River discharge has been shown to be affected by human activities in several areas of the Arctic (e.g., Ye et al., 2005) and dams or water withdrawal for irrigation may significantly change river discharge and thereby have significant impacts on the sea-ice regime. Modification of freshwater discharge and its impact on ice formation could promote enhanced ice formation in

the Arctic with impacts that might reverberate at the global level if, for example, discharge from the large Siberian rivers was reduced substantially.

The distribution of shoals and shallow areas is also a key constraint in stabilizing the landfast ice cover (Reimnitz et al., 1994; Mahoney et al., 2007). Hence, coastal development such as dredging of harbors or other areas to mitigate the effects of coastal retreat, or the construction of artificial gravel or ice islands in the context of offshore oil and gas development (Instanes et al., 2005), can have major impacts on the ice cover in the coastal zone. It may also greatly reduce primary production in the sea ice (Gradinger et al., 2009b) by increasing the amount of sediment entrained into sea ice, either through enhanced resuspension as a direct or indirect result of offshore activities or through reduction in ice stability enhancing sediment entrainment in a mobile coastal ice environment (Eicken et al., 2005). While these impacts are only likely to be important on a very confined, local scale in the immediate vicinity of coastal development sites, they can significantly affect use of the ice cover at this scale, both by industry and by coastal communities.

While local effects such as coastal change and river discharge affect landfast ice, human emissions of greenhouse gases, dust and black carbon may result in further reductions in albedo affecting the ice cover on a much wider scale (Eicken et al., 2005; Huntington et al., 2007). Hydrocarbon exploration, production and transportation cause considerable emissions to air and are important variables in regional Arctic emission inventories of greenhouse gases used in climate scenarios. The geographic location of emissions of CO_2 has no impact on the warming potential. For particles (black carbon) on the other hand, the location of the emissions is important and increased activity in the Arctic results in an increase in local emissions, with concurrent effects on the surface albedo (see Chapter 2). Sea-ice retreat will also lead to a shift in patterns and timing of shipping and icebreaker activity (Chircop, 2009). There will be a general increase in ship traffic in new areas, and Arctic marine shipping may adversely impact sea ice through operational discharges and emissions and navigation impacts on sea ice (Molenaar and Corell, 2009a). However, it is possible that changing ice types and a reduced ice-cover season will limit icebreaking needs.

With activities in the Arctic increasing and adding to global impacts on sea ice, feedback effects are likely to play an increasingly important role in the further reduction of sea ice, especially on the local-scale at the Arctic coasts.

Acknowledgments

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10. Arctic Societies, Cultures, and Peoples in a Changing Cryosphere

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Contents

Key Findings

- The changing cryosphere causes physical and biological changes with consequences for livelihoods and living conditions.

- While indigenous peoples and local Arctic communities will be challenged to maintain their way of life, opportunities including increased shipping and commercial development of renewable and non-renewable resources are expected to benefit mainly non-Arctic actors.

- Local, regional, and international challenges include maintaining traditional land use and sea use; hunting, harvesting, and herding activities; infrastructure; search and rescue; pollution; human health; and indigenous and local rights and involvement.

- Coastal erosion, thawing permafrost, and changing sea-ice conditions (such as thickness, extent, and age), when combined with non-cryospheric drivers of change (such as increased economic activity, socio-economic development, demographics, governance, and the health and well-being of the Arctic society and people) will result in multi-faceted and cascading effects.

- Indigenous peoples and other local residents are resilient and active rather than passive actors, but the rate of cryospheric changes will continue to challenge current adaptive capacity.

- Major uncertainties remain in terms of the direct and indirect impacts of cryospheric change on human society, in and beyond the Arctic, and in terms of the timescale of changes and adaptations.

- Understanding of cryospheric change is needed at different scales. Local projections are required by local residents and external groups or industries with economic interests to adapt to changes, and regional understanding of the role of the cryosphere in climate forcing must be considered in developing global mitigation responses.

10.1. Introduction

The aim of this chapter is to identify and link the key effects of cryospheric change with key aspects of human society and activities in the Arctic. Ice, water, and snow are integral to living conditions and livelihood activities throughout the region and in recent years major changes in the Arctic cryosphere have been documented (see the preceding chapters of this report). These comprise a significant component of overall climate change impacts. Impacts of cryospheric change, albeit laden with uncertainty, occur over broad spatial and temporal scales and across local, regional, national, and international societal levels. In addition to the effects of a changing cryosphere, human society in the Arctic is also exposed to non-climatic factors such as the extractive industries, demographic changes, technological developments, and other impacts of globalization processes.

These changes are, in turn driven by complex linkages between changing cryospheric conditions, global resource demands, and societal changes. It is within this wider context of change that the consequences of cryospheric change for human society and activities must be understood. In this chapter, 'human society' refers to large and small, indigenous and non-indigenous communities; Arctic migrants (i.e., seasonal workers); industrial sectors (i.e., fisheries, tourism, mining, petroleum); international relations; and governance. The cross-scale approach (linking local, regional, national, and international scales) to understanding how cryospheric and societal conditions combine to shape adaptation and vulnerability provides the foundation for this chapter.

The Arctic Climate Impact Assessment (ACIA, 2005), which provides the baseline for the present assessment emphasized that a focus on relevant cross-cutting issues and multiple stressors that shape vulnerability and resilience to change were needed in future assessments. ACIA concluded that future assessments of economic consequences of climate change should include all Arctic residents, include integrative vulnerability studies, and increase the relevance for Arctic residents through local focus. One of the major findings on the polar regions in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4) (IPCC, 2007) is that changes in the terrestrial cryosphere, hydrology, permafrost, and sea ice are increasingly affecting biophysical systems and having cascading effects on Arctic societies. Furthermore, the AR4 found that cryospheric changes will have costs and benefits for Arctic socio-economic systems and activities. The resilience shown historically by Arctic indigenous peoples is now severely tested, and substantial investments are needed to adapt or relocate physical structures and communities and to enhance adaptive capacity. The benefits, dependent on local conditions, include reduced heating costs, increased agricultural and forestry opportunities, more navigable northern sea routes, and greater marine access to resources.

The present assessment has identified the following findings as the most significant since ACIA (2005): (i) there have been dramatic changes in the cryosphere in this decade; (ii) the rate at which these changes have taken place is more rapid than previously observed; and (iii) the effects of changes in the cryosphere substantively alter living conditions in the Arctic and have significant implications for individuals and households at

local, regional, national, and global scales (Chapter 1). Current local-scale observations of change (by indigenous and non-indigenous residents) include a significant thinning of sea ice and freshwater ice, a shortening of the winter ice season, a reduction in snow cover, changes in the distributions of wildlife and plant species, thawing permafrost, and increased coastal erosion (e.g., Tyler et al., 2007; Forbes, 2008; Furgal and Prowse, 2008; Ford et al., 2009; Andrachuk and Pearce, 2010; Forbes et al., 2010; Hovelsrud et al. 2010b). Such observations are corroborated by many scientific findings (see Chapters 4 to 9).

The previous chapters describe how cryospheric changes have affected and will affect people in the North. Changes in snow cover and snow duration; permafrost; sea-ice extent, duration, and age; timing of freeze-up and break-up of ice on rivers and lakes; and melting of glaciers and ice caps will create challenges as well as opportunities for people both inside and outside the Arctic. While giving the Arctic region a more defined role in the world economy, cryospheric changes will pose challenges for individuals and local communities, for national and international governance, safety, search and rescue operations, resource management, and demands on technological advances across virtually all sectors of human endeavor in the Arctic.

Sea-ice retreat creates longer open-water seasons and will be generally positive for the transportation sector. These changes may have positive spinoffs for the North, creating employment opportunities and revenues and providing economic opportunities in developing the potential renewable and non-renewable resources in the region. The projected trend of increased tourism, including cruise-ship tourism will create opportunities for entrepreneurs and operators in the region, provided the human skills and financial resources are available. A changing snow regime may provide an increased hydropower potential, and thus opportunities for industrial development in remote areas with demand for more power (e.g., mines). The changing permafrost may offer opportunities for some agricultural activities if it thaws completely.

Drawing on the most recent data about present and future conditions, on current understanding of the complex linkages between environmental and socio-economic conditions, and on the information provided in the preceding chapters, this chapter covers the livelihood activities, sectors, and challenges and opportunities most relevant when considering a changing cryosphere. A detailed assessment of all sectors at all scales for all regions of the Arctic is beyond the scope of this chapter.

10.2. Living in the North

- Key shifts in physical access afforded by the cryosphere will affect individuals, communities, and sectors in the North through multiple pathways with the direction and magnitude of the overall effects dependent upon spatial and temporal scales.

- Accessibility shifts combined with societal changes will affect the economy, health and well-being, and quality of life of all northerners. These shifts will result in far-reaching effects at regional and national levels and beyond the Arctic.

Changes in the cryosphere are expected to have significant impacts on the living conditions and livelihood activities within and outside the Arctic. In addition, such changes will affect seasonal residents and non-residents with interests in the region. The relationship between a changing cryosphere and local impacts is complex, mediated by the interaction of multiple biophysical and human stressors operating over different spatial and temporal scales. For example, risk assessment of using sea ice for travel and hunting must be linked with reduced time to hunt, diminishing land and sea skills and knowledge among younger generations, decreased reliability of traditional knowledge as conditions change, and use of new technology. The overall risk must be balanced by the individual against the need for harvesting (i.e., food security). Moreover, from the individual perspective of a coastal community resident, earlier and increased marine access afforded by sea-ice decline may result in lower costs for supplies, equipment, and non-traditional/non-local foods.

Thus, the sum of such effects on livelihoods, living conditions, and health and well-being is the key issue. This integration of effects from all cryospheric (and other) changes is important across individual, community, regional, and national levels in the Arctic. Although the changing cryosphere will first and foremost impact peoples (indigenous and non-indigenous) and sectors in the High Arctic, significant consequences for the southern Arctic will also occur through, for example, thawing permafrost, changing snow conditions, and changes in frozen river and lake ice (e.g., Anisimov et al., 2007; Furgal and Prowse, 2008).

10.2.1. Arctic populations and economies

The Arctic covers 11% of the global surface area. Demographic estimates vary with definitions of Arctic geographical extent and seasonal migration. Depending on the geographical definition, population figures vary between 4 million (AHDR, 2004) and 9.9 million (Duhaime and Caron, 2009) (see Figure 10.1). Estimates of indigenous populations range from 400 000 to roughly 1.3 million.

Between 2000 and 2005, the Arctic population declined by nearly 1%, with the greatest decline in the Russian Arctic and the smallest decline in Arctic Sweden and Norway. In all other Arctic regions, populations have increased (Mäenpää, 2009). Seasonal variation due to work-related migration is particularly applicable in the Russian Federation, which accounts for roughly 75% of all Arctic inhabitants, 7.1 million (down from 7.9 million in 1995) (Glomsrød et al., 2009), and by far the largest share of Arctic land area.

The Arctic is characterized by a dispersed settlement pattern with few large cities. About one-third of Arctic residents live in settlements with a population size of less than 5000 and one-tenth of the Arctic population lives in one of the five largest cities (> 100 000 inhabitants; see Figure 10.1) (Bogoyavlensky and Siggner, 2004). The Arctic is an exporter of raw materials and energy and an importer of final goods and services; in 2005, the Arctic area was responsible for generating 0.5% of the global gross domestic product (GDP) (Mäenpää, 2009).

The gender balance varies throughout the Arctic regions. With the exception of the Russian Arctic regions, overall there are more men than women in the Arctic (Bogoyavlensky and Siggner, 2004; Duhaime and Caron, 2009). This is partly a consequence of greater migration by men than women into

a

Unalaska

Kodiak
Bethel
Nome
Anchorage
Anadyr
Fairbanks
Whitehorse
Kotzebue
Alaska
(USA)
Barrow
Inuvik
Yellowknife
CANADA
Churchill
Tiksi
RUSSIA
Norilsk
Dudinka   Igarka
Novy Urengoy
Nadym
Salekhard
Vorkuta
Moosonee
Qaanaaq
Greenland
(Denmark)
Svalbard
(Norway)
Longyearbyen
Naryan Mar
Iqaluit
Aasiaat   Ilulissat
Sisimiut   Maniitsoq
Nuuk
Ittoqqortoormiit
Hammerfest
Tromsø   Murmansk
Narvik   Apatity
Kiruna   Arkhangelsk
Goose Bay
Tasiilaq
Qaqortoq
Bodø   Rovaniemi
ICELAND   FINLAND
Reykjavik
Tórshavn
SWEDEN
Faroe Islands
(Denmark)   NORWAY

Arctic population centres

- Settlement
- 2500 - 10 000
- 10 000 - 30 000
- 30 000 - 60 000
- 60 000 - 100 000
- 100 000 - 360 000

b

Population of Arctic regions

- 1 000 000
- 500 000
- 250 000
- 100 000
- 50 000
- 20 000

Indigenous population of Arctic regions, % of total

- < 5
- 5-10
- 10-25
- 25-50
- > 50
- non-indigenous

Alaska
USA
Chukotka
Yukon
CANADA
Northwest
Territories
Sakha
Nunavut
RUSSIA
Taymir
Krasnoyarsk
Yamalo-
Nenets
Nunavik
(Quebec)
Nenets
GREENLAND
Komi
Labrador
Finnmark   Murmansk
Troms   Lappi   Arkhangelsk
ICELAND   Nordland   Norrbotten
Faroe   FINLAND
Islands   SWEDEN
NORWAY

Figure 10.1. Population and economy of the Arctic. (a) Settlements in the Arctic and their population (~2010), (b) total and indigenous population distribution (figure adapted from Nymand Larsen et al., 2010 and Nordregio, www.nordregio.se), (c) the distribution of Arctic peoples subdivided according to language families (adapted from Winfried K. Dallmann, Norwegian Polar Institute, 2004), (d) Gross Product (Gross Domestic Product or Gross Regional Product) within the circumpolar Arctic by sector and country (based on data from Glomsrød et al., 2004).

Arctic regions seeking jobs primarily in the construction and extraction industries. In addition, there is a tendency among younger women in parts of the Arctic to leave to take advantage of education and career opportunities outside the Arctic (e.g., Rauhut et al., 2008; Poppel, 2010).

Since their arrival in the Arctic, the indigenous peoples have lived from the renewable resources of the sea (fish, marine mammals), the land (land mammals, birds, berries), and freshwaters (fish). Although not considered in official Arctic gross regional product (GRP) assessments, the contributions from commercial and subsistence fishing, hunting, and herding activities are documented as playing a significant role in the mixed cash-subsistence economies (e.g., Rasmussen, 2005a; Poppel, 2006; Kruse et al., 2008; Aslaksen et al., 2009; Poppel and Kruse, 2009). Alaskan harvest surveys during the 1990s found an average harvest of wildlife (fish and game) of 375 pounds (~170 kilograms) per rural resident (Wolfe, 2000), and more recent surveys indicate that subsistence fishing and hunting are still a vital part of the economy in many Alaskan communities (Krieg et al., 2009). The importance of subsistence activities can be measured in different ways. The Survey of Living Conditions in the Arctic (SLiCA: Poppel et al., 2007; Poppel and Kruse, 2009) offers opportunities to analyze economic, integrative, nutritional, social, cultural, and identity aspects of subsistence. One of the SLiCA findings was an indication of the importance of traditional/local food to the diet: two out of three Inuit households report that half or more of the household's food consumption is traditional/local food items. The SLiCA report also documented that traditional/local food is still shared between households. In Greenland, Arctic Canada, and northern Alaska, on average 75% of the Inuit households had shared traditional/local food with other households during the previous 12 months (Poppel et al., 2007).

Renewable harvesting is linked to the subsistence-based informal economy and to the market economy. In 2002, the total catch of fish in the Arctic region represented around 10% (7.26 million tonnes) of the global catch of wild fish (Lindholt, 2006). In addition, in 2002 the Arctic region's catch of shrimp and snow crab made up 5.3% (355 000 tonnes) of the total world crustacean catch (Lindholt, 2006).

The GRP is a measure of the total value of goods and services in a given territory. Comparing the contributions from the primary sector (exploitation of natural resources), the secondary sector (manufacturing and construction), and the tertiary sector (public and private services) to GRP, as well as to the overall Arctic GDP, the tertiary sector is the most dominant (Figure 10.1). Only in Arctic Russia is the primary sector more dominant (Glomsrød et al., 2009). This means that, apart from the Russian Arctic, public services such as education, health, and social work account for a significant part of all economic activities (Duhaime, 2004). The largest economies in the Arctic belong to Alaska (United States) and Russia, mainly due to mining and petroleum activity. Regions where more traditional subsistence activities, such as hunting and fishing, play a more dominant role (e.g., Greenland and northern Canada) have much lower GRP. Similarly, reindeer herding in Russia and Scandinavia is of substantial importance to the livelihoods and lifestyles of reindeer herders like the Sámi and the Nenets but does not contribute greatly to GRP in these regions.
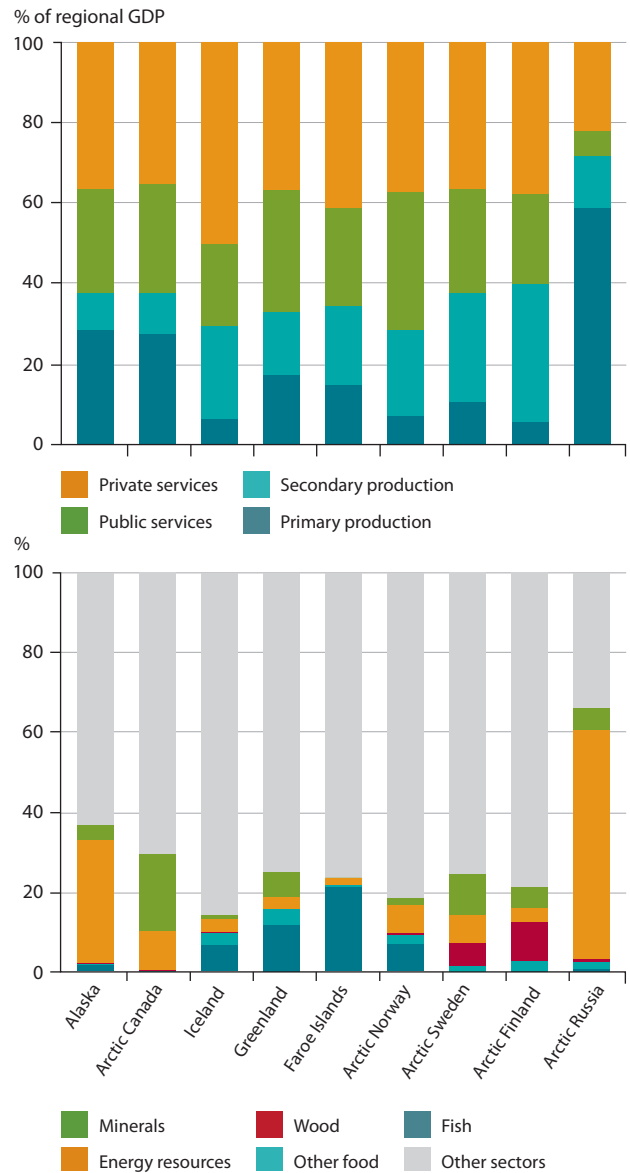
Figure 10.2. Composition and relative size of the main industries (upper) and natural resource-based industries (lower) in the Arctic in 2005. Source: Glomsrød et al. (2009).

The abundance of renewable and non-renewable resources, and the generated share of GDP, varies from region to region (Figure 10.2). In 2005, fishing was the most important resource-based activity in the Greenlandic, Icelandic, and Faroese economies; oil and gas extraction was by far the largest industry in Alaska and Arctic Russia; and minerals (particularly diamonds) contributed the most to GDP for the Canadian Arctic. Fish and energy resources added most value to the Norwegian GDP. In Sweden, the greatest value was added by mineral extraction, and in Finland by timber (Glomsrød et al., 2009). In 2007, Arctic oil and gas activities amounted to one tenth of the world's total oil production and one-quarter of the world's total gas production. Of this, 80% of the oil and 99% of the gas were extracted within the Russian Arctic (AMAP, 2007). The Arctic is also rich in minerals other than oil. The Arctic region contains a large variety of mineral resources: coal, iron, and ferro-alloy minerals (e.g., nickel, cobalt, tungsten), non-ferrous minerals, and precious metals (e.g., zinc, lead, palladium, platinum), and industrial minerals (e.g., diamonds, apatite) (Lindholt, 2006).

The GRP assessments of regional economies do not show the distribution of income within the region and do not provide an exact measure of the total disposable income to be shared among households and individuals in the region. This is because resource rents and returns to the fixed capital may be transferred out of the region and because other mechanisms such as state transfers can contribute to household disposable income. Generally, the net result for all Arctic regions shows that the disposable income of households per capita is lower than GRP per capita (Glomsrød et al., 2009), which supports the hypothesis that the Arctic is a net exporter of raw materials and energy.

For many Arctic indigenous peoples, the cryosphere is fundamental to their cultures and identities in myriad ways. In general terms, the cryosphere has traditionally been used as a platform for travelling, and for livelihood activities such as herding, hunting, and fishing (e.g., Nuttall et al., 2005; Tyler et al., 2007; Ford and Furgal, 2009). The cryosphere also plays a fundamental role in the life and work of non-indigenous residents, through non-renewable resource extraction; hunting, fishing and recreational activities; and extensive use of seasonal ice roads across wet tundra, rivers, and lakes. These roads provide key transportation routes for relatively inexpensive transport of heavy equipment, foodstuffs, and other supplies for residents and industry. Accessibility afforded by these roads reduces the cost of living in the North (see Chapters 5 and 6; Furgal and Prowse, 2008).

Arctic residents rely on their extensive traditional and local knowledge (see description in Section 10.2.2), and on their observations of the environment and weather when making decisions on when, where, and how to build, travel, or harvest. Many Inuit elders and experienced hunters have noted changing sea-ice conditions over the past decade, including later ice freeze-up and earlier break-up, a longer ice-free open-water period in summer, and altered ice dynamics. It has been reported that Inuit observations often correspond to the instrumental record (see Ford et al., 2009; Laidler et al., 2009), and the value of integrating scientific, local, and indigenous knowledge is increasingly acknowledged (e.g., Gearheard et al., 2006; Hovelsrud and Winsnes, 2006; Weatherhead et al., 2010).

The documentation and incorporation of traditional and local knowledge is currently central to research on community adaptation to climate change (e.g., Hovelsrud and Smit, 2010).

10.2.2. Livelihood activities

Livelihood activities, which historically have been closely connected to the natural (primarily renewable) resource base, are also influenced by globalization (e.g., Keskitalo, 2008). Traditional subsistence economies without some form of cash flow no longer exist in the Arctic, but there are still many smaller indigenous communities with mixed cash-subsistence economies (Figure 10.3) in which hunting, harvesting, and herding are combined with predominantly service and administrative jobs (e.g., Hovelsrud, 1999; Duhaime, 2004; Poppel and Kruse, 2009). Shifts toward wage-based economies may erode traditional knowledge among younger generations. In turn, the decreasing ability to 'live off the land' contributes to further shifts. Traditional knowledge is gained through experiences in the natural environment and is based on oral tradition reflecting the past, present, and future. It is passed down through generations and pertains to the environment and a human's activities in it as a holistic relationship. The accumulated knowledge from the past must always be fine-tuned to fit the present (e.g., Hovelsrud and Winsnes, 2006). Hence, traditional knowledge is dynamic and not statically situated in the past. Subsistence activities are adapted to seasonal and interannual variations in, among others, snow and ice, but the current dramatic and rapid changes of the cryosphere further challenge their continuation and in combination with other drivers will most likely contribute to the shift toward wage-based economies.

Many of the smaller communities, relying on harvesting, are vulnerable to changes in the natural resource base and respond by diversifying the targeted species, supplementing with wage jobs and a social safety net of sharing and cooperation. Flexibility in the ability to harvest different species of fish or game and in cash-earning activities is critical to reducing vulnerabilities to change and for ensuring well-being, societal integrity, and human security (Rasmussen, 2005b; Poppel, 2006; Poppel and Kruse; 2009; Hovelsrud and Smit, 2010).

Continuum

Subsistence   Cash

Figure 10.3. There are no longer any societies in the Arctic that only engage in subsistence activities without a flow of cash. All Arctic households and economies have some level of cash flow. The level of cash flow varies by region, season, and household.

10.2.2.1. Reindeer husbandry

Reindeer husbandry is an important livelihood activity for the Sámi across Fennoscandia, as well as for several indigenous peoples in Arctic Russia. Topography, local variation in pasture condition, and the response of snowpack dynamics to weather have implications for the complex relationship between snow, ice, and reindeer pasture conditions (e.g., Kumpula and Colpaert, 2003, 2007; Roturier and Roué, 2009). Reindeer herders employ traditional knowledge to respond to varying pasture conditions (e.g., Tyler et al., 2007). Such herding strategies are shaped by factors such as season, snow type, temperature, landscape, and the physical condition of the animals. Strategies are also influenced by previous experience of unusual weather conditions, such as ice layers on the surface of the snow. Snow types range from those influenced by weather (new and light or icy types) to those transformed by reindeer grazing, digging, and trampling. These snow types are well documented by Sámi (e.g., Magga, 2006), as is their effect on herding practice (e.g., Roturier and Roué, 2009).

Reindeer pasture will be directly affected by the changing snow, permafrost, and lake and river ice conditions (Anisimov et al., 2007). Increased frequency of rain-on-snow and mid-winter melt events in some regions of the Arctic, associated with earlier spring snowmelt and coverage, results in an earlier start of the growing season (Schmidt et al., 2006; Meltofte et al., 2008). This has a generally positive impact on the herbivores' pastures (Forchhammer et al., 2005), particularly inland (Karlsen et al., 2007). However, the earlier break-up of lake and river ice, which accompanies earlier snowmelt, presents challenges for reindeer migration from winter to summer pastures, particularly for newborn calves (Rybråten and Hovelsrud, 2010).

Among the Nenets, rain-on-snow and winter warming events are known to increase ice layer formation over pastures, affecting reindeer access to pastures (Forbes, 2008; Forbes and Stammler, 2009). Freeze-thaw events may also damage reindeer pastures. In Fennoscandia, reindeer mortality and reduced calf production have tended to be higher in winters with heavy ice crust formation (Helle and Kojola, 2008; Moen, 2008). In contrast, studies from Svalbard found no links between icing and reindeer mortality (Tyler, 2010). Mold formation in the soil, which may result from snow falling on unfrozen ground during mild, rainy autumns, is harmful to the animals and may kill calves (Eriksson, 1976; Pruitt, 1984; Turunen et al, 2009). Snow depth, density, and elevation have additional ramifications for reindeer husbandry. Snow depth, which varies by region (snow cover is projected to generally increase in Arctic Russia and Canada and decrease in the European Arctic; see Chapter 4), also determines the energy that reindeer expend seeking grazing areas, by walking across the snow surface and digging through the snowpack.

Reindeer husbandry involves many hazardous tasks, particularly during reindeer gathering prior to migration or slaughter. During these periods the herders use vehicles (motorcycles, snowmobiles, helicopters, airplanes, boats) to gather the reindeer, often working long days (Daerga et al., 2004). Changes to snow conditions may negatively affect these tasks, increasing pressure on time and encouraging risk taking, thus increasing health and safety concerns.

Arctic winter precipitation is likely to increase and depending on the conditions (i.e., rain *vs* snow and the frequency of freezing rain events) will have implications for the future success of reindeer herding. An increase in snow density may be negative in some areas if combined with increased wind, leading to heavier snow packing. Less multi-annual snow at high elevations in summer will decrease the possibilities for reindeer to evade flies and mosquitoes by moving to higher altitude. Although the direct consequences of cryospheric change may seem predominantly negative for reindeer husbandry, the broader context of socio-economic as well as market and political conditions must also be considered.

10.2.2.2. Fishing and hunting

Many inhabitants of the High Arctic, both indigenous and non-indigenous, live near the coasts where sea-ice variability, resulting from dynamic and thermodynamic processes, is a dominant feature of the environment. Across the Arctic, livelihood activities are adapted to local sea-ice conditions and the seasonal and behavioral patterns of the species harvested. Changing sea-ice conditions affect the distribution of these fish, marine mammals, and seabirds species. Existing fish stocks show horizontal and vertical distributional changes as their seasonal migration is related to the formation and melting of the sea ice (Sherman et al., 2009; Sundby and Nakken, 2008; West and Hovelsrud, 2010).

Arctic hunting activities are particularly sensitive to snow, ice, and permafrost conditions (Ford and Furgal, 2009). Marine mammals, particularly seals, have been essential for survival for many coastal indigenous peoples in the Arctic, and loss of hunting grounds, changes to species' availability (see Wiig et al., 2008), and seasonal hunting patterns present challenges for local communities. Access to key hunting areas has been constrained, particularly in spring and autumn, resulting in shortages of traditional/local food items such as seal, walrus, and caribou. However, sea-ice refugia are expected to persist in the Canadian and Greenland archipelago and secure the survival of ice-related species such as narwhal, and some seal and whale species (Armitage, 2005; Gearheard et al., 2006; Hovelsrud et al., 2008; Laidre et al., 2008; Moore and Huntington, 2008; Ford, 2009; Wenzel, 2009). In cold weather, the reindeer in Greenland move toward the coast, where they become more accessible to hunters; while in warmer weather, the animals move toward cooler conditions closer to the ice margin in order to reduce insect harassment. This significantly increases the distance the hunters will have to walk to search for the animals and to carry the harvest back to their boats.

In parts of the Canadian Arctic, hunters are required to travel farther to access hunting areas (Ford et al., 2010a; Pearce et al., 2010). In some instances, travel has had to change temporally as well as spatially, and sometimes simply to another location and not necessarily further away (e.g., SIKU project: Krupnik et al., 2010). In Greenland, a number of local and regional sea-ice travel routes can no longer be used, and hunters traveling by dog team are forced to travel overland, which is more time consuming and can be more dangerous than sea-ice travel under typical conditions (Goldhar et al., 2010).

Hunting and fishing have important symbolic and cultural values in many indigenous communities and are important to

Box 10.1. Traditional beliefs and health

For indigenous peoples, the cryosphere is fundamental to their cultures and identities, and its changes provide the context for current changes in their personal relations, health, and well-being. The concept of well-being is complex with physical, mental, emotional, and spiritual aspects. The complex interrelation is a theme explored by many indigenous cultures.

There often seems to be a discrepancy between the indigenous feeling of well-being and that defined by social scientists. This also has profound implications for conventional Western views on health, where the focus on individual health problems (abuse, maladjustment, lifestyle symptoms) contrasts with that of the collective or community-based approach by many indigenous peoples.

For Inuit, health is deeply connected to social behavior and its relationship with diseases. In the traditional Inuit belief system, Inuit disease is a consequence of transgression against the social or spiritual orders. At the same time, 'common sense empiricism' coexists within the Inuit belief system: snow blindness, for example, is common-sense pathology and does not require a metaphysical explanation. While biomedicine is better informed about the workings of the body, traditional Inuit understanding of the relations between person, family, social group, community, and the environment represents a global view of health that is not shared by biomedicine (Bowd, 2005). This is an important area to consider in future studies of Inuit health and well-being.

nutrition and health; individual, household, and community economies; identity; and cultural, spiritual, and social well-being (see Box 10.1) (Huntington et al., 1998; AHDR, 2004; Poppel, 2006; Hovelsrud et al., 2008; Poppel and Kruse, 2009). Although the overall commercial value of seal hunting is limited, the subsistence harvest of seals, other marine mammals and fish continues to play important roles for indigenous communities (e.g., Rasmussen, 2005a). The SLiCA project indicates that four out of ten Inuit households harvest about 50% or more of the meat consumed (Poppel et al., 2007; Poppel and Kruse, 2009). Thus, subsistence and informal economic activities still provide a significant part of their diet, as well as their health and culture. Seal meat, for example, is an important part of the diet, health, and culture for many Arctic indigenous peoples. It is also important for feeding sled dogs and critical for hunting, fishing, and tourism (Nuttall et al., 2005; Rasmussen, 2005a; Ford et al., 2010a; Pearce et al., 2010). Because of extensive food-sharing networks, changing ice conditions also challenge the cultural and spiritual identity of and nutritional quality for community members who do not rely on hunting and/or do not have access to financial resources to adapt (e.g., Goldhar and Ford, 2010).

10.2.2.3. Agriculture

Agriculture in the Arctic has not changed significantly since the ACIA report (Juday et al., 2005) (see Glomsrød et al., 2009, for a regional comparison of the contribution of agriculture to the regional GDP). Agriculture is a relatively small industry in most of the Arctic in terms of value added (see Figure 10.2, category

'other food'; Juday et al., 2005). Still the sector constitutes an important foundation for settlements in the Eurasian part of the Arctic and in southern Greenland (Tholstrup and Rasmussen, 2009). Arctic agriculture is a mixture of large-scale commercial production and subsistence agricultural production with a focus on cool season forage crops (e.g., alfalfa, red clover, timothy), vegetables (e.g., beans, peas, potatoes), and small grains (e.g., wheat, canola). Livestock production is concentrated on cattle, goats, lamb, sheep, pigs, poultry, and reindeer. Subsistence agriculture, in combination with other food production activities such as fishing and hunting, is present throughout the Arctic, albeit not in great numbers. Agricultural development is constrained by lack of wider market access resulting from remoteness, lack of infrastructure, and sparse population (Juday et al., 2005; Hovelsrud et al., 2010a). Except for reindeer husbandry, crop production is generally more sensitive to cryospheric conditions than livestock production (Juday et al., 2005), but with considerable regional variation within the Arctic. Duration of snow cover, freeze-thaw events, icing, and permafrost constitute the primary cryospheric limitations to Arctic agricultural production.

Although regional differences exist, a changing cryosphere is expected to have predominantly positive impacts on agriculture in the Arctic. In areas with a reduction in snow cover, the growing season may be extended (Torvanger et al., 2004; Falloon and Betts, 2010; Grønlund, 2009; Tholstrup and Rasmussen, 2009). Earlier snowmelt will also lengthen the growing season and reduce the likelihood of fungal damage to crops. A stable snow cover of at least 120 days is required for fungus to develop and harm crops. On the other hand, plants that are well adapted to Arctic conditions may not prove adaptable to a longer growing season (e.g., Aaheim et al., 2009). Ice crust formation and rain-on-snow events (i.e., generally increased variability in precipitation regimes) may also affect agricultural production (Grønlund, 2009). Less snow may expose plants to wind and frost damage through reduced insulation (Bjerke and Tømmervik, 2008; Grønlund, 2009).

Rain-on-snow events and melting and refreezing of snow may result in frost damage, with potentially major economic consequences (Putkonen and Roe, 2003; Aaheim et al., 2009; Grønlund, 2009). Similarly, increased runoff from glaciers, melting snow, and presence or absence of permafrost can affect drainage capacity and result in frost damage if refreezing occurs (e.g., Grønlund, 2009). In general, projections of increasingly mild weather (see Chapter 3) are expected to decrease the frequency of frost damage events. On the other hand, increased precipitation and runoff combined with episodes of freezing and thawing could considerably increase soil erosion in agricultural fields (Øygarden, 2003; O'Brien et al., 2006; Grønlund 2009).

Changes in nutrient flow and distribution resulting from increased runoff can positively affect agriculture, but can also enhance contaminant transfer to rivers, lakes, and the sea (Falloon and Betts, 2010). Other negative impacts of excess water include soil waterlogging, anaerobicity, and reduced plant growth (Bradley et al., 2005). Farming operations on wet soils may cause compaction damage (Finlayson et al., 2002; Webb et al., 2005; Montanarella, 2007) or agricultural machinery may simply not be adapted to wet soil conditions (Eitzinger et al., 2007).

10.2.2.4. **Forestry**

Large tracts of boreal forest cover sub-Arctic and low Arctic areas of the North (e.g., ~57 million hectares in the Yukon, Nunavut, and Northwest Territories of Canada) (Furgal and Prowse, 2008). Similar to marine and freshwater ecosystems, a healthy terrestrial ecosystem is essential to cultural, spiritual, and social well-being of indigenous and local peoples of this area. Moreover, large economic benefits are derived from merchantable timber. Increased freeze-thaw events may cause damage to forests and timber, and increased fungal or insect attacks may destroy drying timber and reduce economic profit (Aaheim et al., 2009). In addition and depending on the region, heavier snowfalls and icing may crush tree canopies and hinder new growth (Keskitalo, 2008). Reindeer grazing has potentially two opposing effects on forest expansion. The removal of lichen may allow birch seeds to germinate and sprout successfully, while reindeer grazing on birch seedlings, shoots, and leaves may hinder the advance of the treeline (Houle and Filion, 2003; Tømmervik et al., 2004, 2009; Dalen and Hofgaard, 2005).

In general, snow is an important factor in damage to forests (Kilpelainen et al., 2010). In Finland, where forestry is of great economic importance, the risk of snow damage to forests is projected to decrease by about 50% toward the end of the 21st century. The effects of cryospheric change on forests will interact synergistically with other non-climatic changes to ultimately result in greater impacts on forestry.

10.2.2.5. **Tourism**

Over the years, access to the Arctic has increased through improved transport technologies, such as ice-strengthened ships, air transport, and specialized vehicles. As a result, Arctic tourism has developed substantially, in terms of the number of tourists as well as geographical and seasonal reach (UNEP, 2007; Stewart and Draper, 2006a). The sector is diverse, aimed at the mass market, and includes sports fishing and hunting, ecotourism, adventure tourism, dogsledding, and cultural and heritage tourism. Iconic Arctic species such as the polar bear draw tourists northward (e.g., Forgione and Martin, 2008; Dickson et al., 2009), but also hold considerable value for many people who will never visit the region.

Tourism is an integral part of local economies, individual and household incomes, revenues, and financial capital for northern infrastructure and has become an alternative source of income for many local communities and gateway cities, enabling a positive interaction between new economic opportunities and traditional activities (Rasmussen, 2005b; UNEP, 2007; Tholstrup and Rasmussen, 2009). Tourism activities include the production and sale of indigenous handicrafts and artwork such as carvings, weavings, and prints. Sport hunting of polar bears is important to economies for some Inuit communities, whereas in other areas such hunting is allowed for subsistence use only (e.g., Hovelsrud et al., 2008).

As provided through the examples above, cryospheric change will result in major changes in livelihood activities throughout the North from the individual to the national level.

10.2.3. **Human health and well-being**

Changing cryospheric conditions directly and indirectly affect the health of Arctic inhabitants. Cryospheric conditions directly affect work, travel, and safety. Indirectly, higher temperatures in combination with altered ultraviolet (UV) levels will affect wildlife (composition and abundance) and increase the risk of zoonotic diseases (Berner et al., 2005; AMAP, 2009).

Climate change has been shown to propagate infectious disease agents into the Arctic and further modifications are expected; a more accessible Arctic will increase human exposure and vulnerability to new and emerging infectious diseases (Parkinson, 2008).

The release of antibacterial pharmaceuticals from human settlements may also weaken human resistance under cold Arctic conditions (Kallenborn et al., 2008). Infrastructure essential for community sanitation and drinking water provision may be degraded or destroyed through the action of thawing permafrost or extreme weather events, such as flooding or storms. This may lead to increased possibility of food- and water-borne diseases and respiratory infections (Parkinson and Butler, 2005).

Melting ice and thawing permafrost may also release persistent environmental legacy pollutants (such as POPs and metals) from past human activities at lower latitudes that have been carried into the Arctic in the prevailing winds and ocean currents and stored in the cryosphere (AMAP, 2009). Contaminants may be released from local sources such as sea ice, permafrost, ice sheets, degrading glacial snow, and from distant sources (Gregor and Gummer, 1989; Pinglot et al., 1999; Hermanson et al., 2005). These can negatively affect the immune system in animals and humans (Kraemer et al., 2005; Hansen et al., 2008) as well as degrade the quality of traditional/local foods. Cycling rates or entry points of contaminants into biotic systems may also change, thus affecting the health and productivity of ecosystems. This may reduce the quality of traditional/local foods as well as commercial products (Chapter 11, Section 11.3).

A decrease in freshwater ice (river, lake, permafrost) and longer open-water seasons mean greater freshwater availability, which as a consequence of climate change also will be warmer; both of these factors will foster zoonotic diseases. Higher ambient temperatures, for example, may result in an increase in some temperature-sensitive food-borne diseases, such as gastroenteritis. In addition, small changes in temperature and humidity will affect the distribution and behavior of many vector-borne diseases such as West Nile virus (e.g., Petersen and Roehrig, 2001; Parkinson and Butler, 2005). Higher temperatures also benefit bacteria, parasites, and insects (Bradley et al., 2005) and may drive increased propagation of zoonotic pathogens (e.g., *Giardia*, *Toxoplasma*, and *Echinococcus*) in water- and food-borne pathways.

Food contamination problems may become particularly acute where permafrost 'ice houses' are used for local and community storage of foodstuffs. In this case, thawing permafrost will negatively affect communities relying on traditional/local foods, thereby having consequences for human health and well-being (Parkinson and Evengård, 2009). Although significant for all coastal communities, subsistence harvesting remains very important in places with few jobs and little cash income as traditional/local foods provide a valuable

source of energy, nutrients, and minerals (Nuttall et al., 2005; Rasmussen, 2005a; Poppel and Kruse, 2009).

Changing sea- and river-ice conditions affect travel safety and inhibit access to key hunting, fishing, and herding areas already constrained by changing natural resource use. The health and economy of some Arctic inhabitants are already affected by changes in the timing of hunting, fishing, and herding activities. A reduction in land-fast ice combined with increased open-water area makes sea ice and fog conditions less predictable, making coastal travel treacherous and communication between communities more difficult (Barber et al., 2008). Traditional travel routes to hunting and fishing grounds or for visiting family are in many places no longer safe, forcing people to travel longer distances, which compromises safety (Ford et al., 2007; Hovelsrud et al., 2008).

Although traditional knowledge is flexible and highly adapted to major fluctuations in weather and resource base, the current rate and magnitude of change challenges the relevance of such knowledge and affects the health and well-being of older generations (e.g., Krupnik and Jolly, 2002; Weatherhead et al., 2010). The significant changes to travel routes and ice conditions therefore present challenges to traditional knowledge, which has implications for cultural and social relations (Ford et al., 2006a,b, 2008; Hovelsrud et al., 2008; Tremblay et al., 2008). Younger generations employ a combination of modern technology, traditional knowledge, and different methods and equipment to ensure a safe and sufficient harvest. Although technological innovations may improve safety, they are costly and are susceptible to failure. They are therefore often seen as an addition, not a replacement for traditional knowledge (Tremblay et al., 2008; Ford et al., 2009; Ford and Furgal, 2009).

10.3. Increased activities and access with local implications

- Reductions in cryospheric components (e.g., ice sheets and sea ice) that presently serve as barriers to access are likely to stimulate increased interest in northern industrial development, particularly in non-renewable resource sectors. This will be offset to some degree by transportation issues (e.g., diminished accessibility by ice roads) that may limit development capacities and rates.

- Consequences include a further shift toward wage-based economies, demands for skilled workers, increased migration, and requirements for enhanced sustainable management, environmental protection, and overall governance.

- Increased demands (e.g., skilled local workers, environmental reviews, management needs, services, and governance) will be required and will further strain a system already struggling to cope with widespread change from multiple factors with limited adaptive capacity.

- Increased challenges to indigenous peoples will have far-reaching effects on language and culture.

- Increased industrial development and related socio-economic and cultural changes in the Arctic will be an inevitable outcome of cryospheric changes driven by climate change. Cascading needs for infrastructure, services, and related development will foster increased energy consumption. Increased production of black carbon ('soot') will result from such activities which, if not properly regulated, will feedback positively to local cryospheric degradation, further exacerbating more rapid climate and cryospheric change.

A variety of factors influence and finally determine decisions whether to locate and develop industrial and other economic activities in the Arctic. These include world market prices, level of costs, and the level of education and training of the local labor force. In addition, unfavorable climate conditions and challenging logistics have undoubtedly been major barriers to increased economic activities in the Arctic region to date. The changing cryosphere is projected to open previously inaccessible or unprofitable areas for large-scale industrial developments, shipping, and tourism with differing impacts on local communities and activities across the Arctic. Direct cryospheric changes will combine with direct and indirect effects from increased industrial activities to result in synergistic and cascading effects. Traditional livelihood activities, such as hunting, fishing and herding are susceptible to increased activities in the region. The regulatory and preparedness aspects of changing access are discussed Section 10.5.

10.3.1. Transport and infrastructure, tourism

Sea ice, snow cover, permafrost, and ice roads on lakes and rivers are vital for transport and low-cost provisioning to remote communities that do not have highway systems. They are also important for industrial areas and developments that cannot use air access or major northward river barge traffic for transporting heavy loads, fuel, and large equipment (Vuglinsky and Gronskaya, 2006; Furgal and Prowse, 2008; Prowse et al., 2009). Dogsleds and snowmobiles are used for hunting and reindeer herding in the Arctic and winter road networks are important for resupplying many communities (e.g., Furgal and Prowse, 2008). Snow cover, frozen rivers, and land-fast sea ice are all critical elements for safe travel, and a reduction in these conditions creates challenges and risks. Access to the sea and transport to, from and within northern regions are determining factors for hunting, fishing, exchange of goods among communities and with regions outside the Arctic, and maintenance of social networks and family ties in the Arctic. Frozen rivers and lakes are also used by reindeer herders during migration. Ice roads contribute significantly to national economies and create a vital lifeline to northern communities and mines (see Box 10.2). Although statistics on ice-road use and volumes are available (Furgal and Prowse, 2008), the importance of ice roads to northern communities has not been scientifically documented (Ford et al., 2008). Similarly, barge traffic along the large northward-flowing rivers that characterize the Arctic contributes significant seasonal capacity for northern transportation.

Enhanced river barges and land-based all-weather roads and railways as well as increased infrastructure to service such transport (e.g., enhanced dredging of rivers) will be needed as an alternative to ice roads to ensure continued access to industrial developments and communities (Prowse et al., 2009) (see also Box 10.2). A six- to nine-week reduction in

Box 10.2. Winter road access to northern communities

Land- and ice-based transportation declines associated with the degradation of cryospheric components is clearly a developing issue of wide concern. This will be exacerbated by increased impetus and activities associated with northern development and until resolved will hamper such development.

The importance of ice roads for access to northern communities is shown by the impact of thawed winter roads in the province of Manitoba, Canada, due to mild weather in March 2010. Typically, the road carries more than 2500 shipments each year to more than 30 000 First Nations people. The stranding of many freight-haulers and local drivers resulted in dwindling construction supplies, rising food and fuel prices and a related rise in unemployment. A state of emergency was declared in 11 communities. First Nations Grand Chief Harper is quoted as saying, "*This has become a human rights issue – no one should be cut off like this. No one should feel this isolated*" (Carlson, 2010).

— Shipping routes   — Pipelines   — Railways   — Roads

Figure 10.4. Arctic transportation routes – roads, shipping, and pipelines.

the ice-road season on the Mackenzie River, for example, could result in a 50% increase in the use of barge-based transport (Lonergan et al., 1993). In areas where open rivers are currently critical for reaching remote regions, such as in Russia, a longer ice-free season is beneficial in economic terms and will provide better access for the increased activities in the Arctic. However, compensatory development will be required due to local reductions in road networks resulting from permafrost degradation. A shorter snow-cover season will reduce snow- and ice-removal costs and improve the mobility and safety of passenger and freight travel through reduced winter hazards (U.S. National Research Council, 2008). In addition, a reduction in snow and ice cover might extend the construction season.

While transport over sea ice, ice roads, and permafrost will become more problematic, sea-ice retreats creating longer open-water seasons will be generally positive for the transportation sector. These changes may have positive spinoffs for the Arctic, as the focus of economies and investors, from inside and outside the Arctic, will be directed to the economic opportunities in developing the potential renewable and non-renewable resources in the region.

The decline in sea-ice extent in the Arctic will increase the opportunity for trans-polar commercial shipping and considerably shorten the route between Europe and the Pacific (see Figure 10.4). Ships will reduce travel distance by 40% by sailing the Northern Sea Route compared with current shipping routes. It is projected that the transit transport from Russia will increase, although seasonal ice will remain an uncertain hazard. A shorter route for ships could save energy, reduce emissions, promote trade, diminish pressure on the main trans-continental navigation channels, and have great economic benefit (Commission of the European Communities, 2008; Molenaar and Corell, 2009). On the other hand, large areas of open warmer water may also lead to increased cloudiness and fog events (Walsh and Chapman, 1998) challenging navigation for shipping and local vessels. An increase in shipping will have significant consequences for local coastal communities, particularly given that marine resupply is likely to
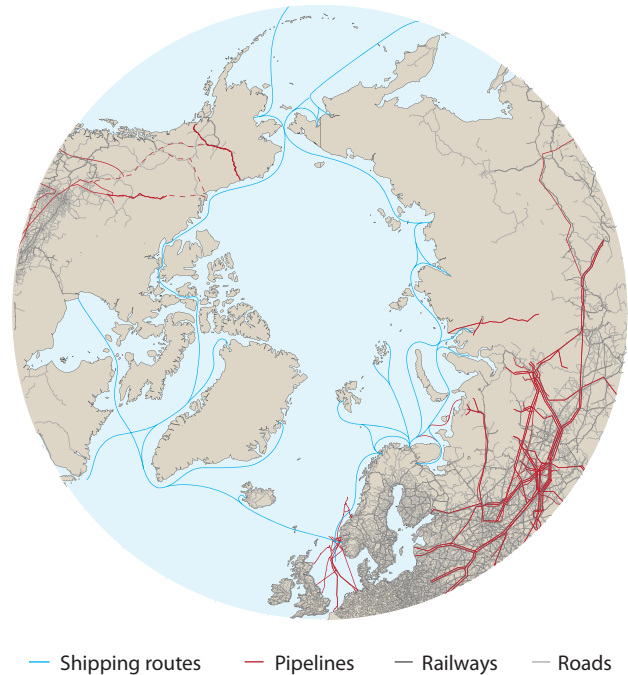
experience easier and longer seasons. Although the Arctic may be navigable before 2050, and routes along the Siberian coast even earlier, more investment is required in infrastructure and the provisioning of marine services to ensure safe and secure ship transit with minimal environmental impact (Ho, 2010). As a result, for the near future, Arctic shipping is expected to be largely to points in the Arctic rather than ocean-to-ocean transits and commercial maritime activity in the Arctic will probably be driven more by economics, dominated by the development of natural resources and tourism, than by diminishing sea ice (see also Chapter 9, Section 9.4.4.1).

An increased need for transport and infrastructure in the Arctic due to developing extractive industries and commercial and tourism related shipping also requires the development of community infrastructure, harbors, search and rescue, and local service sectors. Changes in the frequency and magnitude of iceberg calving may influence access for planned infrastructure development, marine transport, fisheries, and oil and gas exploration and production on Arctic continental shelves, where activity is expected to increase significantly over the coming decades. Adaptation, mitigation, and monitoring techniques will be necessary to minimize the potentially serious impacts. One negative impact expected is disturbance to seals and whales, which contribute significantly to the diet and the informal economy of indigenous peoples throughout the region.

10.3.1.1. **Infrastructure**

Infrastructure includes physical facilities with permanent foundations or the essential elements of a community. It includes schools, hospitals, various types of buildings and structures, and facilities such as roads, railways, airports, harbors, power stations, communication systems, and power, water, and sewage lines. Infrastructure forms the basis for local, regional, and national communication and for economic

Box 10.3. Permafrost threat to infrastructure

While most settlements in the Arctic are relatively small communities, most are located on the coast where permafrost is commonly present. The Arctic population is set to increase (Furgal and Prowse, 2008), and northern Russia already has large cities of over 100 000 inhabitants in permafrost areas. Alaska has a population of about 500 000 in permafrost areas, the main portion (80%) living in areas with a permafrost extent of less than 50% (U.S. Arctic Research Commission Permafrost Task Force, 2003).

Most developments in the Arctic have taken place with an awareness of current permafrost conditions, but projected climate-driven changes in permafrost are likely to affect these and future developments beyond current planning and engineering provisions and can pose a significant challenge to infrastructure, environment, and health in the Arctic.

Coastal Arctic regions have concentrations of industrial facilities associated with oil and gas such as the Prudhoe Bay region in northern Alaska, the Mackenzie Delta in Canada (although production has not yet started there), and the Pechora Basin in Russia. Damage to pipelines may have dramatic environmental effects, especially when oil or gas is released at the coast or at sea. About 35 000 pipeline failures are registered annually along the 350 000-km network of pipelines in western Siberia. Over 20% are probably due to deformations and weakening of foundations induced by permafrost thaw (Anisimov and Reneva, 2006). The world's largest terrestrial oil spill was the 160 000 tonnes of oil-containing liquid in Vozei in 1994, and even 15 years later soil contamination in the surrounding area is still high. The spill was likely to have been caused by uneven ground settlement due to permafrost thaw and thermokarst.

Figure 10.5. Pipeline under construction in the Russian Arctic.

growth. Particular concerns are associated with permafrost thawing and degradation and coastal erosion effects on the stability and maintenance of transportation routes and on industrial development in such regions (see Box 10.3) (e.g., IPCC, 2007). Thawing permafrost may have an impact on air transport through effects on runways (see Chapter 5, Section 5.3.5.1.3). Coasts and infrastructure become more prone to erosion and destruction as the sea ice retreats, permafrost thaws, and the dynamics of freeze-up and break-up events on lakes and rivers cause floods. Damage from ice action and flooding

to installations such as bridges, pipelines, drilling platforms, and hydropower also poses major economic costs.

Thawing permafrost will affect coastal and inland facilities and infrastructure, but costs and risks are more closely linked to the design lifetime of the structure than with thawing permafrost (see Chapter 5, Table 5.5). Nevertheless, many buildings and structures were built assuming a more permanent permafrost layer, which has consequences for their structural stability and safety, especially as rising air temperatures and snow depth exacerbate permafrost thaw (Romanovsky et al., 2010). Proper design and engineering will counteract permafrost changes to a certain extent, but an accelerated thawing of permafrost and rapidly changing conditions may require immediate design changes for high-risk projects such as water-retaining structures, tailings, and storage facilities for hazardous wastes and sewage (see Box 10.3).

Coastal infrastructure will be increasingly exposed to coastal erosion due not only to increased exposure and thawing of coastal permafrost, but also to sea level rise, longer open-water fetch as sea ice declines, and potential flood damage from river-ice jams. Recent research describes how coastal erosion is increasingly threatening Arctic coasts (Martin et al., 2009). In addition to substantial land loss and impacts on infrastructure, such erosion may expose waste dumps at abandoned military and oilfield sites, with potentially severe effects on the health of wildlife and coastal communities. Smaller coastal communities are especially vulnerable to increased contaminant levels from saltwater intrusion as permafrost thaws, coasts erode, and infrastructure is damaged (Berner et al., 2005). In addition, such communities have to deal with damaged infrastructure resulting from permafrost thaw (Anisimov et al., 2007; Andrachuk and Pearce, 2010; Ford et al., 2010a). Infrastructure located in river channels (e.g., gas fields in the Mackenzie River Delta) may be exposed to flood damage from ice jams and backwater flooding (Prowse et al., 2009). In addition, ice instability and related ice damage (e.g., freezing rain events on distribution networks) may increase with a projected increased use of hydropower installations. Especially in coastal mountainous areas, rockfalls and landslides caused by permafrost thaw and increased storminess might cause additional hazards to infrastructure and safety, while some rockfalls may lead to local tsunamis and thereby increase impact (see Chapter 5, section 5.3.5.1.1).

10.3.1.2. **Tourism**

An increasingly important source of income for many communities, tourism also depends on good access. Arctic tourism relies on traditional perceptions of the Arctic environment and expectations about the experience that relate to ice and snow, mountains and tundra, and animals (Pagnan, 2003). With the exception of some cruise lines traveling to the Arctic from Europe and North America, and some Alpine ski operations, tourism in the region consists of small-scale operations (Pagnan, 2003). Tourist activities include traditional hunting and fishing, trophy hunting, expedition-style and destination cruising, dogsledding, and cultural and aboriginal tourism (Notzke, 1999).

The Arctic draws tourists that seek solitude and inaccessible and inhospitable qualities, which may change with reduction in sea ice and opening up of the region.

Current trends indicate that 'climate change' tourism, for example travelling to Greenland or Alaska to view rapidly retreating glaciers, is a growing attraction. A more accessible Arctic may draw greater numbers of another type of tourist who seeks land-based tourism activities, such as sport hunting, winter-based activities, ecological and nature tourism, and retreats and conferences (Stewart and Draper, 2006b). Changes in the predictability of snow conditions, such as winter thaws, increased avalanches, slush torrents and floods will have consequences for winter activities such as dogsledding, skiing, reindeer sledding, and snowmobile touring (e.g., Dawson et al., 2007). Snow conditions also affect ecosystems, biodiversity, and landscape amenity values, and if the scenic values are diminished through environmental changes associated with changes in snow, then tourism will be affected.

Cruise ship tourism is developing rapidly worldwide, and loss of sea ice may open up waterways and opportunities for increased cruise traffic (e.g., Glomsrød et al., 2009), perhaps on the scale presently observed in Svalbard and Alaska (Pagnan, 2003; Johnston, 2006). In addition, visitors on more traditional cruises are becoming more adventurous, seeking out more remote locations such as the Arctic, and it is likely that such cruise tourism will increase independently of sea-ice change (Stewart and Draper, 2006a,b; Snyder and Stonehouse, 2007). On the other hand, a changing cryosphere is altering the character and distribution of the sea ice and increased risks of accidents (e.g., hull penetration by multi-year ice) may have negative implications for cruise tourism, in particular for transits through the Northwest Passage and in the High Arctic (Stewart et al., 2007; see

also Section 10.4.3). Supporting these ventures (e.g., with shipping, infrastructure maintenance, and search and rescue) creates its own sustainability challenge (e.g., Stewart and Draper, 2008).

Changing sea-ice and snow conditions will reduce the opportunities to view snow- and ice-dependent wildlife, such as seals, walruses, and polar bears (Lemelin, 2005; Richardson et al., 2005). The effects of shrinking or disappearing glaciers on tourism in the Arctic are not well documented, but a few studies from the Alps and North America show that hazards related to shrinking glaciers increasingly affect popular mountaineering activities on and around glaciers (Jetté-Nantel and Agrawala, 2007; Smiraglia et al., 2008) and have already limited skiing on glaciated slopes (Abegg et al., 2007). Declining or disappearing glaciers are also found to have a negative impact on the attractiveness for visitors in glaciated natural parks (Scott et al., 2007; Wolf and Orlove, 2008).

10.3.2. Changes in access and availability of renewable and non-renewable resources

Access to Arctic resources will be enhanced by many cryospheric changes even though some changes (e.g., degradation of permafrost and earlier melt of river and lake ice) may reduce accessibility over land. Increased accessibility is particularly relevant for non-renewable resources (e.g., coastal ore deposits and hydrocarbons on shelves) but is also relevant for some renewable resources (e.g., sub-Arctic marine fisheries presently subject to seasonal ice presence) (see Figure 10.6).
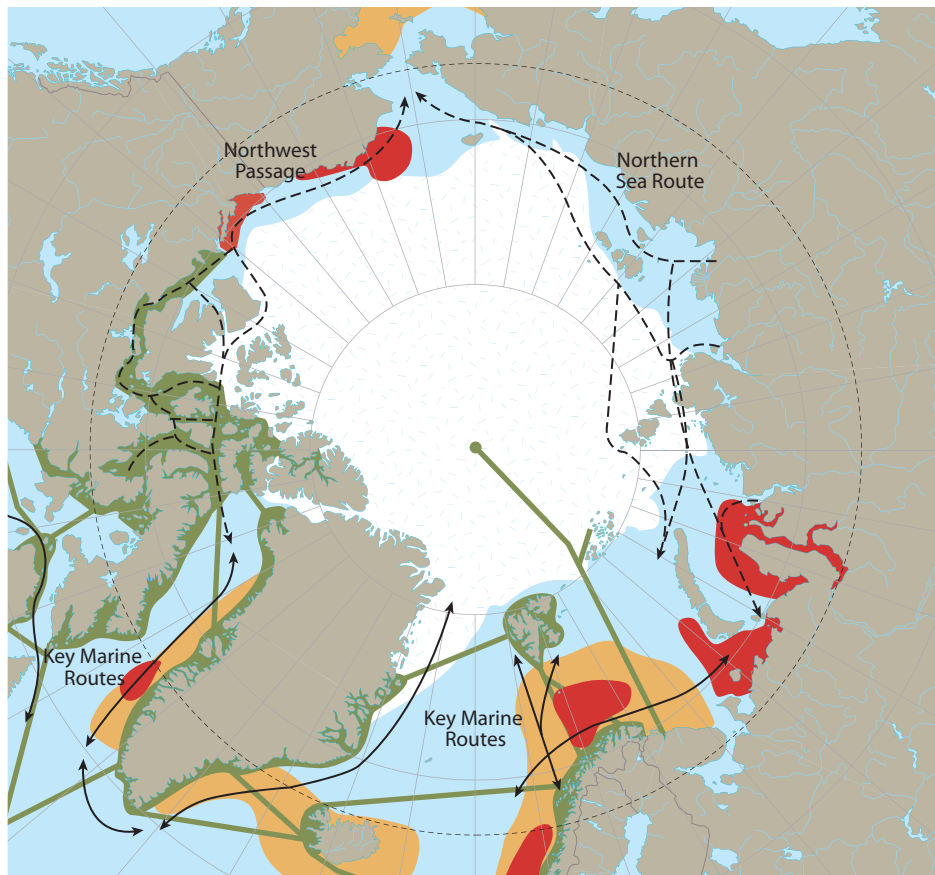
Figure 10.6. Current marine shipping uses in the Arctic.

Probability, %

| Low potential | <10 | 10-30 | 30-50 | 50-100 | 100 |
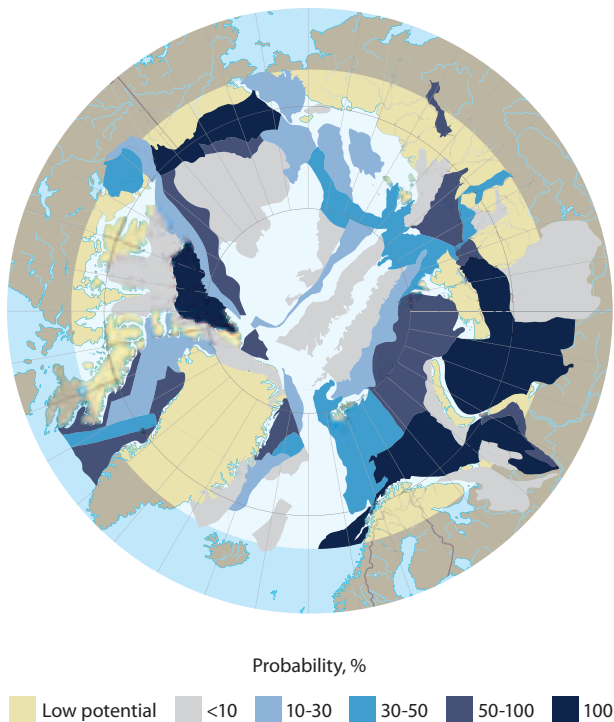
Figure 10.7. Probability of the presence of (at least one) undiscovered oil or gas field (with recoverable resources greater than 50 million barrels of oil equivalent). Source: USGS (2008).

10.3.2.1. **Hydrocarbons and mining**

The Arctic is an important supplier of oil and gas to the global market, and it is likely that climate change will have positive and negative financial impacts on exploration, production, and transportation activities. With a growing global demand for hydrocarbons and minerals, the Arctic stockpile of resources is likely to have an increasingly important role in the global economy. Increased access and activity afforded by cryospheric changes (particularly decreased sea-ice presence) will create challenges for extraction, transport, engineering, search and rescue needs, and responses to accidents.

According to Lindholt (2006), 97% of total Arctic oil and gas production is located in Russia and Alaska. Canada, Greenland, and the Barents Sea are potential locations for future oil and gas production (Figure 10.7). It is estimated that roughly 84% of the undiscovered oil and gas in the Arctic occurs offshore (USGS, 2008). Most of the discoveries remain undeveloped (Gautier et al., 2009), primarily due to remoteness, costs, and technological challenges associated with ice-infested environments (sea ice, icebergs) combined with abundant petroleum extracted at a relatively low cost. With a projected increase in activity, the industry will have to contend with risks of oil spills, spill remediation costs, and stricter environmental regulations. Risk assessments and the marine insurance industry are likely to play a crucial role in industry-related shipping (AMAP, 2007; VanderZwaag et al., 2008; Molenaar and Corell, 2009). Climate change impacts on oil and gas development have so far been minor, but are likely to result in financial costs and benefits in the future (see Section 10.3.1.1.). Reduced sea-ice cover will affect offshore drilling, due to potential for increased wave action and storm surges, and perhaps variable regimes of drifting ice.

The Arctic holds large stores of minerals (Lindholt, 2006) and potentially large reserves of undiscovered sources of raw materials. A reduction in glaciers, ice sheets, and sea ice in and around Greenland, for example, will increase accessibility to new areas with the potential for commercial exploitation of minerals and energy resources. Mining activities in the Arctic are important contributors to national economies and to the global economy. The extraction process itself is not likely to be significantly affected by a changing cryosphere, but the industry will benefit from improved marine transportation conditions to production sites and markets. Any expansion of hydrocarbon and mining activities is likely to require further expansion of air, marine, and land transportation systems and all related regulatory activities.

Impacts of hydrocarbon and mining development on people, communities, and governments can be positive and negative (AMAP, 2007). Such impacts could drive social and economic change, create revenue and employment opportunities, affect households as workers may be away for long periods, and foster increased immigration to the Arctic as jobs are filled by employees from outside the region.

10.3.2.2. **Hydropower and wind power generation**

Expansion of industry and population in some areas of the Arctic has already increased electricity demands (Furgal and Prowse, 2008), and this trend is expected to continue as the hydropower potential may itself attract industries with high energy consumption (e.g., aluminum smelters) that need abundant, cheap, and 'green' (having no greenhouse gas emissions) energy. However, dams and reservoirs essential to hydroelectric production may carry their own suite of environmental effects (e.g., altered flow regimes and contaminant liberation from flooded sediments), which may be enhanced by cryospheric change. The drive to increase renewable sources of energy has led to the rapid increase in the development of large-scale windmill facilities in the Arctic.

Production of hydroelectric power is important in several Arctic countries and still has a vast potential with large as yet unregulated northern rivers (e.g., Prowse et al., 2004, 2009; Chapter 6, Section 6.5.3.2 and Chapter 7, Section 7.7.3.3) and future projections expecting an increase of inflow (Prowse et al., 2009). Some of the current hydroelectric power plants' capacities and designs are seasonally constrained by the effects of river ice, but this could markedly change under future cryospheric conditions. The location of the mineral production process depends on costs, including energy costs. Access to hydropower or alternatives close to extraction sites might be profitable for mining companies; however, susceptibility of over-land distribution networks to cryospheric impacts must be factored into developmental scenarios. Increased hydroelectric capacity, in turn, might result in cascading development effects as more jobs are created, more labor is needed, and facilities are enhanced for increased workforces. At the same time, the adaptive capacity of many Arctic communities will be challenged as they are confronted by many significant changes over a short period.

The availability of resources for hydropower is expected to increase significantly in some areas of the Arctic due to increased precipitation, melt, and runoff. For example, the

hydropower potential in northern Europe is expected to grow by more than 25% by 2050 (Jochem and Schade, 2009) and by up to 30% by the 2070s (IPCC, 2007). Projected increases in winter rain and snowfall and increased freeze and thaw cycles are expected to lead to enhanced winter snowmelt and decline in winter storage and hence to more even runoff over the year (Gode et al., 2007; Furgal and Prowse, 2008). In Swedish sub-Arctic Suova, the increase in runoff due to increased precipitation is projected to be as much as 53% by 2100 (Gode et al., 2007). In addition, glacial runoff (which is significant for hydropower generation in Scandinavia, Iceland, Greenland, western Canada, and Alaska) is projected to increase. The increased potential for hydroelectric power in Greenland, owing to more meltwater from the ice-sheet margin, is already included in the new Greenlandic economic strategies. However, only a few studies have tried to assess how cryospheric change may affect glacial runoff with subsequent consequences for hydropower production. Bergström et al. (2007) found that increased glacial runoff would contribute to increased hydropower production in the Nordic countries until the latter half of the 21st century, when maximum runoff is reached for most of the catchment studied. In the long term, the glacial mass will be reduced and the runoff will decrease.

A key factor is the stability of the water supply to the new power plants, which depends on the freeze-up conditions (Beltaos and Prowse, 2009). In addition, increased amounts of silt and sand in glacial meltwater may fill dammed lakes and affect the viability of the dams, eventually requiring additional reconstruction in order to maintain the water flow needed (see Chapter 7, Section 7.7.3.3). Alternatively, the bedrock topography exposed to the retreating ice may cause new meltwater drainage patterns, leading to a partial or complete drying out of the water supply to the hydropower plants. Research from glaciated catchments in the western United States and in the Alps highlights that existing hydropower dam infrastructure might not be able to utilize the increased runoff caused by climate change (e.g., Barnett et al., 2005). It is reasonable to assume that this too could be the case for hydropower production in glaciated catchments in the Arctic.

The increased availability of runoff in some regions may require expansion of reservoir capacities. Other challenges result from irregular resource availability that may lead to irregular production and increase potential problems linked to ice formation, break-up, and jamming in rivers downstream of hydropower outlets. This must also be linked to the potential changes in energy demand and changes in energy prices in the face of future climate, which will influence production and thereby influence future ice problems of the hydropower industry.

10.3.2.3. **Reindeer husbandry**

For the past 30 to 40 years, long-range reindeer migration patterns have been restricted in Fennoscandia (Riseth 2003) and the Nenets-Yamal area due to the ongoing development of oil and gas extraction and transport infrastructure (Forbes et al., 2009). The construction and operation of these plants require a relatively widespread network of roads, buildings, and power lines, and many of the Scandinavian facilities have been sited within important reindeer pasture areas. Aversion or avoidance of certain reindeer pastures can lead to less optimal range use, complications with herding, increased costs, and reduced production. Such developments may negatively affect the foraging behavior of reindeer and must be seen in relation to changing snow conditions and land use. The rate and magnitude of changing snow-cover conditions are expected to vary significantly due to regional differences, between coasts and inland, and at different elevations (Brown and Mote, 2009). These cryospheric changes combined with the encroaching industrial developments will have significant implications for reindeer herders' choice of pastures.

10.3.2.4. **Land use**

Land use such as forestry and agriculture relies on productive ecosystems and efficient access during key periods. An increasing Arctic population will increase pressure on resource bases on such land use (e.g., Forbes, 2008; Keskitalo, 2008; Tømmervik et al., 2009). Permafrost underlies much of the area; thus its degradation along with reductions in length of the winter season may result in drainage and drying of land with consequences for fire regimes (i.e., enhancement) and services realized from the ecosystem (e.g., reduced harvests due to reduced access via winter roads). Such impacts will be highly regionalized, and regional variations in precipitation, particularly changes in snowfall, will exacerbate or mitigate such effects.

Increased productivity in boreal forests is expected, but reduced accessibility to timber outside frost periods will be a threat to the forestry industry (Lindner et al., 2010). Delayed access to felled timber, due to earlier snowmelt, may have a considerable impact on the quality of the timber due to insect damage and aging (Keskitalo, 2008; Øseth, 2010).

10.3.2.5. **Fisheries**

A changing cryosphere will have consequences for access to coastal and offshore fisheries. Fisheries are important throughout the Arctic (Glomsrød et al., 2009). Reduced sea ice has impacts on the abundance, migration, seasonal distribution, and composition of key commercial fish stocks (e.g., Atlantic cod and herring, with the resulting change in access) (see Chapter 9, Section 9.3.3.3; IPCC, 2007; Stenevik and Sundby, 2007; Brander, 2010). Key stocks will increase and are likely to shift their summer distribution northward in response to sea-ice loss and a projected increase in primary production (Chapter 9, Section 9.3.4.1). The potential for an eastward colonization of new spawning areas also exists in the Barents Sea area, and this may result in wider geographical redistribution of the stocks (Stenevik and Sundby, 2007). Currently, projections of how biophysical changes will influence commercial fisheries resources, in terms of specific stocks and species, are limited and uncertain (e.g., Loeng et al., 2005; Loeng, 2008), but the widely predicted northward shift of boreal species has already been observed in the North Pacific and the Bering Sea and Barents Sea (see Chapter 9, Section 9.3.4.3). While effects on species and location will vary across the Arctic, a northward shift will affect coastal and offshore fisheries. Fish stock abundance is shaped through a combination of human and environmental factors, and it is difficult to assess how cryospheric changes will affect future fisheries as changes in access, technology, and

species composition occur (e.g., Mueter and Litzow, 2008).

A general expectation of climate change for much of the Arctic is that increased precipitation will add to the volume of northward-flowing rivers, but the connection to cryosphere degradation is limited to degradation over the longer term of ice caps, glaciers, and the Greenland Ice Sheet. Both processes will contribute to increased freshwater input to the Arctic Ocean, which will be exacerbated by sea-ice loss. Freshening of marine surface layers will enhance stratification and reduce wind-induced mixing. This decreases nutrient transfer from deeper waters, thereby influencing amounts and types of primary production (see Chapter 9, Sections 9.3 and 9.4 and references therein) and, in turn, affecting production pathways to commercially viable fish stocks.

In Greenland and Alaska, a reduction in sea ice combined with increased glacial runoff and river discharges, altered sedimentation rates, and rising sea temperature is causing changes in nutrient transfer and production and will increasingly affect coastal marine ecosystems (Martin et al., 2009; see also Chapter 9, Section 9.4.4.3). With a freshwater budget shift, Greenland halibut is likely to move from deep waters to shelf areas, requiring changes to the fishery (Loeng et al., 2005; Côté and Williams, 2008), and in Alaska the fisheries depend on the strength of the Alaska Coastal Current. The result may be smaller stocks of Greenland halibut and northern shrimp in the Atlantic area and realignment of entire ecosystem pathways in the Pacific, both of which are fundamentally important for fisheries and marine mammals. Other species may enter the area as a result, offering new opportunities. Conversely, new species could lead to competition and colonization of habitat, diseases, new predators, and parasites (Furgal and Prowse, 2008).

For offshore industrial fisheries, a decline in sea ice-dependent fish species but an overall increase in pelagic productivity in the marginal seas and improved access to the Arctic Ocean are likely to be beneficial, provided minimal mismatch occurs between species and productivity pathways. Drifting sea ice and icebergs are an obvious problem for offshore fishing operations, although with ongoing warming a reduction in the density of spring-summer drift ice would be considered a positive effect. Reduced sea ice may also provide a larger seasonal window for offshore fishing and access to new possibilities for coastal fisheries, while increased storminess due to more open water may be a challenge.

Altered access to the Arctic is likely to trigger development and infrastructure needs (e.g., non-renewable resource development and compensatory activities for resupply as ice roads disappear). These, in turn, result in cascading effects on local infrastructure, services, and Arctic residents. An immediate consequence of the above is that demand presently outstrips capacity; as climate change effects generally and cryospheric change effects in particular occur, ensuing demands will vastly increase. The obvious adaptive response to such demand is increased population (through immigration and seasonal workers), further accelerating socio-economic change and exacerbating local development and change scenarios and demands. Increased demands will in turn increase demand for local energy (heating, electrical service, hydrocarbon use, etc.) to support community needs, increased industrial service and transportation needs, and non-renewable resource extraction. In turn, increased activities will foster increased emissions of greenhouse gases.

10.4. Adaptation across scales

- Changing cryospheric conditions require adaptation by local residents and businesses and by organizations and stakeholders with interests in the Arctic.

- Adaptive actions are being taken regardless of uncertainty in cryospheric and societal future conditions.

- Social and economic challenges stemming from a changing cryosphere may be alleviated through policy development and governance.

Arctic communities have generally adapted well to high natural variability in climate and the resource base (e.g., Huntington et al., 2007a,b; Forbes and Stammler, 2009; Wenzel, 2009; Ford et al., 2010b). It is generally argued that current adaptive efforts are strained by the magnitude and rate of climatic change, and the IPCC (2007) found that the resilience historically shown by Arctic indigenous peoples is being severely tested. Furthermore, the complex inter-linkages (including combined and cascading effects) between societal and environmental change represent unprecedented challenges to the current adaptive capacity (e.g., West and Hovelsrud, 2010). Globalization also plays a role in shaping local vulnerability, adaptive capacity, and resiliency (see Box 10.4). It is increasingly common to discuss two types of adaptation: reactive and proactive (e.g., Crabbé and Robin, 2006; Füssel, 2007; Amundsen et al., 2010). Reactive adaptation is in response to current climate and risk management practices, while proactive adaptation is the planning and implementation of adaptive measures to future climate change and unprecedented events (Füssel, 2007). Studies indicate that reactive adaptation is the most common form, triggered by extreme weather events (e.g., Næss et al., 2005; Penning-Rowsell et al., 2006; Amundsen et al., 2010). Studies of reactive adaptation can be used to analyze the capacity of an individual, institution, or community to adapt to current conditions and past extreme events, which can then inform analysis of the adaptive capacity to future climate change and

Box 10.4. Vulnerability, exposure sensitivity, and adaptive capacity

Vulnerability refers to the manner and degree to which a community is susceptible to conditions that directly or indirectly affect the well-being or sustainability of the community. This includes the sensitivity of the ecosystem of which the community is part or on which the community depends. Use of this term does not presume that communities are particularly vulnerable – some may have relatively few or no vulnerabilities.

Vulnerability is a function of exposure sensitivity and adaptive capacity (Adger and Kelly, 1999; Smit and Pilifosova, 2001; Turner et al., 2003; Ford and Smit, 2004; Wisner et al., 2004; Kofinas, 2005; Keskitalo, 2008). Adaptive capacity is closely related to resilience and reflects an individual's or community's ability to cope with, adjust to, or recover from an exposure sensitivity. It is reflected in the community's management of current and past stresses, its ability to anticipate and plan for future change, and its resilience to perturbations.

unprecedented events (Ford et al., 2006a,b; Rasmussen, 2005b; Smit et al., 2010). However, reactive adaptation is insufficient as a long-term strategy to meet many of the effects of cryospheric change. Thus, proactive adaptation and strategies are urgently needed in order to provide the basis for adequate preparedness. These are especially relevant for aspects requiring lead times for implementation, such as infrastructure development.

10.4.1. **Adaptation and adaptive capacity**

Cryospheric changes are felt first and foremost at the local level and by peoples and communities whose livelihoods and well-being are closely linked to the natural environment. While adaptation is clearly a local concern, institutions across societal scales (municipal, regional, international), sectors, and nations operating in the Arctic must also adapt (Hovelsrud and Smit, 2010). Current adaptations are spatially and temporally diverse and connect local, national, and international levels. Table 10.1 presents an overview of the cross-cutting societal implications, adaptation strategies, and challenges related to adaptation to cryospheric change identified in this chapter.

Social learning, responsive local institutions, livelihood flexibility and diversification, and adaptive management have emerged as critical elements for adaptation (Rasmussen, 2005b; Poppel and Kruse, 2009; Ford, 2009; Ford et al., 2009; Hovelsrud and Smit, 2010; Hovelsrud et al., 2010b). Good governance across societal levels (see Section 10.5) is critical for facilitating such adaptations and for strengthening adaptive capacity. It is important to remember that adaptation to changing cryospheric conditions occurs in the context of, and is inextricably linked to societal change. In addition, it is necessary to distinguish between adaptations to slower versus more rapid changes. The latter may require proactive measures and emergency plans to counteract the consequences of, among other things, extreme events for which anticipatory planning is inadequate. As adaptation is predominantly a local process, local projections of climate and cryospheric conditions are increasingly required from the forecasting communities. The development of and input to local projections will benefit from increased cooperation between scientists and local and indigenous communities. In addition to institutional guidelines or policies, economic and market pressures, technology, and the potential for diversification determine the adaptive capacity of a community or sector (e.g., Keskitalo, 2010).

Table 10.1 illustrates how local impacts create cross-cutting challenges for adaptation and governance. For example, increased industrial activities combined with changing cryospheric conditions create new challenges with local consequences. The development of adaptive strategies to Arctic conditions by industrial sectors will affect local vulnerability. If proper adaptive guidelines and regulations that anticipate future climate scenarios are not in place for industrial development and resource use (such as commercial fisheries, cruise traffic, and shipping), then local communities become more vulnerable to changing ice conditions. The development of the International Maritime Organization (IMO) international 'Polar Code' (IMO, 2009; see also Chapter 9, Section 9.4.4.2) for ships in the Arctic serves as a proactive form of adaptation to changing ocean conditions by setting standards for safety and reporting. The Polar Code forms the first internationally recognized guidelines for the construction

and operation of ships in ice-covered Arctic waters. The IMO guidelines aim to promote the safety of navigation and to prevent pollution from ship operations in Arctic ice-covered waters and to further regulate its development in a clear action to avoid damage to the marine environment. Further guidelines for operations and activities in the Arctic will be necessary to strengthen the adaptive capacity of the operators and actors.

Another example of cross-cutting adaptation, listed in Table 10.1, pertains to thawing permafrost. Currently technology is available to adapt infrastructure to the imminent problems related to thawing permafrost. Permafrost thawing and warming may lead to traffic-related challenges (Woo et al., 2007), and in a warming climate road systems may have to be upgraded more rapidly to counteract the increased rate in permafrost thaw, although systems on permafrost already have a short lifetime and climate change may not noticeably increase maintenance (see Chapter 5, Section 5.3.5.1.3). Future adaptation to reduced transport possibilities on ice roads involves enhanced surface flooding or spray-ice layering and modification of transport schedules to concentrate on the coldest part of winter (Prowse et al., 2009). Where possible, such permafrost adaptations will ensure that local communities and industrial activities can be maintained.

Table 10.1 further illustrates that at the local level, current adaptations moderate some of the regional negative impacts of shorter seasonal duration of land-fast ice and poorer ice quality. For example, adaptations to early spring break-up in the western Canadian Arctic include changing modes of transportation from snowmobiles to boats (Ford et al., 2010a; Pearce et al., 2010). Others increase their adaptive capacity by being flexible and use new technology, change resource use patterns, or reduce reliance on harvested foods (Hovelsrud et al., 2008; Andrachuk and Pearce, 2010; Ford et al., 2010a; Goldhar and Ford, 2010; Goldhar et al., 2010). While there is great uncertainty concerning the type and rate of changes, the trend to an increasingly warmer Arctic is expected to continue, with temperatures projected to rise by 3 to 7 °C toward 2100. The greatest changes are expected to take place in autumn and winter over areas with sea-ice loss (see Chapter 3). It is likely that hunting patterns and strategies and traveling activities will be increasingly affected, compromising safety, food security, and cultures and requiring new adaptive strategies (e.g., Hovelsrud and Smit, 2010). New adaptations may require monetary investments straining resources further (see also Chapter 9, Section 9.4.4.1).

Another local adaptation example comes from reindeer husbandry in which understanding the sensitivity of the reindeer grazing system to changing snow conditions is integral to reindeer herding cultures. Detailed environmental knowledge about how hard a snow or ice layer (bottom crust) must be to hinder foraging in winter is more important for the quality of pastures than snow accumulation (Kumpula and Colpaert, 2003). When ice layers, especially bottom crusts, restrict access to pastures stronger animals may still be able to dig through the ice layers that weaker animals cannot penetrate. Hence, as an adaptation many reindeer herders maintain a few strong castrated males to break through the hard layers (Reinert et al., 2008). Another adaptation option, albeit costly, is to supplement the natural diet with food pellets to alleviate poor grazing conditions.

Coastal erosion in communities in low-lying coastal zones has required adaptive measures such as protection of

Table 10.1. Cross-cutting challenges and adaptation.

Cryospheric change (exposure sensitivity)	Implications and challenges	Adaptation strategies and governance implications	Source
Sea ice			
Shorter duration of land-fast sea ice	Land-fast ice travel and hunting hampered	Changes in transport mode and new equipment such as GPS and satellite phone required	Barber et al., 2008
Poorer ice quality	Safety of sea ice-based hunting by indigenous groups compromised	Shift to open-sea hunting and fishing, could require changes in regulations	Pearce et al., 2010; Poppel, 2006; Ford et al., 2010a
Longer and larger open-water season at the coast	Increased fetch (combined with increased storminess) erodes coastline	Shoreline stabilization and engineering options, eventually relocation; government agency with authority to relocate communities	Forbes et al., 2008; Perovich and Richter-Menge, 2009
Reduced sea ice	Changes in fish stocks and fisheries, increased access for fisheries	Fishery regulations must change	Stenevik and Sundby, 2007; Brander, 2010
Reduced sea ice	Increased traffic at sea in the future	IMO standards for ships that sail in the Arctic	Commission of the European Communities, 2008
Reduced sea ice	Increased industrial activity in the future, oil and gas exploration	Higher costs of security provision, stricter environmental regulations	AMAP, 2007
Permafrost			
Thawing of permafrost – increased coastal erosion	Threatens coastal settlements	Shoreline stabilization and engineering options, eventually relocation; government agency with authority to relocate communities	Johnson et al., 2003; Bronen, 2008; Martin et al., 2009
Thawing of permafrost – increased runoff	Release of legacy pollutants that affect the food chain and have negative health effects	Increased surveillance of contaminants in traditional/local foods; investigation of health and ecosystem effects on biota	Kraemer et al., 2005; Hansen et al., 2008
Thawing of permafrost – drainage of forested areas	Tree death caused by drought; increased forest fire occurrence	Development of increased surveillance and response capabilities to fires; altered management regimes to enhance harvesting of 'at risk' forests	Furgal and Prowse, 2008
Thawing of permafrost	Damage to poorly engineered and constructed infrastructure	Proper design and engineering will to a certain extent counteract permafrost changes	Berner et al., 2005; Anisimov et al., 2007
Glaciers and ice caps			
Increased calving	Threat to planned infrastructure development, marine transport, fisheries, and oil and gas exploration	Improved mapping, surveillance, and monitoring	OPRC, 1990; Chircop, 2009; Molenaar and Corell, 2009
Glacier retreat	Increased hydropower production	Increased water storage capacity	Bergström et al., 2007
Glacier retreat	Reduced attractiveness to tourists		Scott et al., 2007; Wolf and Orlove, 2008
Snow			
Increased frost and thaw and ice cover on tundra	Reindeer access to fodder altered, increased mortality	Traditional Ecological Knowledge – use traditionally castrated male reindeer to break the ice crust; flexibility in choice of pastures	Tyler et al., 2007; Kumpula and Colpaert, 2003
Increased growing season	Increased productivity in forestry and agriculture		Lie et al., 2008
Less snow and shorter winter season	Reduced opportunities for winter tourism		Dawson, et al., 2007
Increased amount of snow	Increased hydropower production	Increased water storage capacity	Gode et al., 2007; Furgal and Prowse, 2008
Increased amount of snow	Damage to forests	Implications for fire regimes; see above	Kilpelainen et al., 2010
Increase in winter thaw	Increased tree death for certain conifers	Implications for fire regimes; see above	Kilpelainen et al., 2010
Milder winters	Increased damage to forests from insects and pests	Implications for fire regimes; see above	Heliövaara and Peltonen, 1999
Increased snowfall and higher winter temperatures	Increase in avalanches	Increased forecasting and surveillance of hazards	Allard et al., 2007
Earlier snowmelt	Drought in dryer areas during spring and summer		Grønlund, 2009
Milder winters	Less frost damage on pastures		Grønlund, 2009
Rivers and lakes			
Reduction in ice duration and ice thickness	Reduced operation of ice roads	Concentrate on the coldest period of winter, eventually construction of land-based road or rail networks	Prowse et al., 2009
Reduction in ice duration	Increased barge transport on rivers and lakes	Development and enhancement of infrastructure to support such (e.g., dredging)	Lonergan et al., 1993; Woo et al., 2007

shorelines and buildings. Such measures include planning and consideration of relocation of communities due to coastal erosion and flooding (Furgal and Prowse, 2008). Adaptation through relocation is a difficult process because proper institutions and financing are lacking (Bronen, 2008), and original locations may have been chosen for very good and still valid reasons. Here the local adaptive capacity is closely linked to institutional capacity and policy.

10.4.2. Challenges and opportunities

Challenges and opportunities shaped by cryospheric changes emerge between and across societal scales and along different temporal scales (see Table 10.1). Some changes pose challenges to some individuals and sectors, and opportunities for others. Indigenous peoples in particular are likely to have to devise new adaptation strategies to alleviate the impacts of a changing cryosphere on the environment, health, and traditional practices. Traditional knowledge, being dynamic and adjustable to a broad range of variability in the environment and natural resources, is an important component of adaptive capacity and will be instrumental in developing such adaptations. If traditional harvesting activities are compromised by changing sea-ice conditions, opportunities to harvest new species with familiar technology may arise. On the other hand, existing technology may not be sufficient to exploit new species. Such challenges may be met by changing harvesting methods from sea ice to open water by investing in boats. In turn, this may lead to other economic opportunities, such as some forms of tourism (e.g., Stewart et al., 2007) that may increasingly shift northward as a result of a changing cryosphere. According to one study, based on one global circulation model, tourism in Canada and Russia could potentially increase by 25% through global warming of only 1 °C (Hamilton et al., 2005; Stern, 2006). The projected trend of increased tourism will provide opportunities for entrepreneurs and operators in the region, provided the human skills and financial resources are available. While continued warming will challenge the use of ice roads for transportation, the resulting open-water networks may provide opportunities for increased barging. Heavy transport to land-locked locations will face major challenges in the absence of ice roads and with increasing permafrost thaw and subsidence (Prowse et al., 2009) and may require adaptive measures such as construction of land-based road or rail networks.

The potential impacts of changes on communities in different parts of the Arctic are diverse. For example, hunters in the northern parts of Greenland feel challenged by thin and less reliable sea ice during winter, while others with more focus on fisheries view these changes as an opportunity to expand fishing activities. The local socio-economic situation and politics determine whether the effects of changing sea ice on relevant fish species will present challenges or opportunities at the local level.

An increasingly ice-free Arctic Ocean offers a number of opportunities for naval and maritime operations in terms of shipping and resource extraction. However, significant ice- and weather-related challenges remain. Operators must adapt to seasonal drift ice and icebergs, annual variations in ice cover, continued local sea-ice formation, and icings on ships and installations due to increased storm surges and wind (Molenaar and Corell, 2009). Free-drifting icebergs are particularly challenging and can lead to shutdown of offshore hydrocarbon production, be a threat to shipping and cause major economic losses. The relatively shallow water depth along the Siberian coast, for example, limits the size and freight capacity of ships (Johannessen and Pettersson, 2008), reducing the opportunities of an ice-free passage. Other shipping-related challenges include limited search and rescue capabilities, infrastructure, and uncertainties about future trade patterns. Challenges for oil and gas facilities include employing appropriate technology, developing risk assessments, and meeting regulatory requirements and increasing societal expectations related to environmental performance and social benefits (VanderZwaag et al., 2008; Efthymiou and Winkler, 2009).

A changing snow regime provides opportunities for some Arctic societies in terms of increased hydropower potential, a more even distribution of water supply and less flood damage, and reduced snow removal costs. Increased potential for hydropower will provide opportunities for industrial development in remote areas with demand for more power (e.g., mines). Conversely, reduced hydropower potential because of drought may challenge development and increase demands for other sources of energy (e.g., wind, coal, diesel). In addition, drought in forests because of reduced meltwater in spring may represent a challenge for the forest industry, and melting snowcaps may reduce attractiveness for tourists. For local and indigenous peoples a changing snow regime will challenge traditional resource management.

Industries and developers will need to invest in stabilizing current infrastructure to meet the engineering, design, and connectivity challenges of thawing permafrost. Local residents may find it challenging that their transportation routes and perhaps local use areas (e.g., lakes for fishing) are disrupted. The changing permafrost offers few opportunities, apart from some agricultural activities.

10.4.3. Hazards and safety

A changing cryosphere poses new hazards associated with the risks of on-ice travel, ranging from that for commercial traffic on ice roads to the use of backcountry lake- and river-ice networks for travel and access to traditional food sources by Arctic residents (e.g., Ford et al., 2008). Current observations show that ice conditions are already unpredictable and changing in such a way that previous understanding of local environments, travel routes, and safety skills may be challenged and insufficient. Changing conditions and increasing concerns about hunter safety create a need for costly new technology (e.g., GPS, satellite phone), although it is noted that such devices may be susceptible to failure and must be an addition to, not a replacement for, traditional or local knowledge (Tremblay et al., 2008; Ford and Furgal, 2009; Ford et al., 2009). The changing sea-ice conditions increase risk and safety in two ways. First, and already a challenge, hunters will have to travel increasingly farther or to different and unknown locations to access the animal populations (Ford et al., 2007; Hovelsrud et al., 2008; Tremblay et al., 2008). Second, a reduction in land-fast ice in the western Canadian Arctic combined with more open water make the sea-ice conditions less predictable and result in unpredictable fog events, making coastal travel treacherous (Barber et al., 2008).

Hazards associated with decreasing Arctic glaciers include rockslides, debris flows, ice avalanches, glacial mudflows, and outburst floods (see Chapter 7). In combination with changes in the permafrost on surrounding slopes, the hazard increases. One result of the destabilization is floods (*jøkulhlaups*) from ice-dammed and sub-glacial lakes, especially in volcanic areas such as in Iceland, Alaska, and Kamchatka, where hazards to people and infrastructure are well documented. Glacial lakes that pose a hazard to people and infrastructure are commonly drained by pipes. Retreating glaciers could lead to an increase in possible hazardous jøkulhlaups, but little is currently known about increased risks to communities in the Arctic. Compared to other glaciated regions such as the Himalayas, the Alps, or the Andes, glaciated regions in the Arctic are generally sparsely populated and the impacts of these hazards on communities and infrastructure are probably limited. In the Himalayas, increased frequency of jøkulhlaups may lead to severe damage and loss of lives downstream (UNEP, 2007; Agrawala, 2008). A few potentially dangerous glacial lakes are reported in Norway (Elverhøy et al., 2002) and Canada (Geertsema and Clague, 2005). In general, the hazard from glacial retreat to communities in the Arctic is limited (for a review of hazards see Chapter 7, Sections 7.5 and 7.7). In addition to retreating glaciers, greater snow accumulation and changes to ice-layer production at high elevations may increase the frequency of snow avalanches and slush torrents and thereby become a greater hazard for people.

Although summer sea-ice cover is projected to disappear toward 2020, seasonal variations in sea-ice extent and thickness will remain. The reduction in sea ice will not only open up seaways, but will also create new hazards for shipping and offshore oilrigs (Figure 10.8). Regional differences in first-year ice and multi-year ice may lead to an increase in drifting multi-year ice and icebergs, creating choke points in narrow channels leading to significant navigation hazards (Wilson et al., 2004). This may affect all forms of shipping, including cruise tourism, particularly for transit through the Northwest Passage and in the High Arctic (Stewart et al., 2007).

An overarching issue for all activities in the Arctic is a general lack of preparedness, across local, national, and international levels, to respond to increased pollution, accidents, and search and rescue needs as access to ice-free areas increases. The new waterways are not sufficiently charted, and there is a need for iceberg and environmental monitoring and for infrastructure (ports, navigation aids, supplies, spares), which is currently limited or lacking, to support high levels of marine traffic (see also Chapter 9, Section 9.4.4.2).

10.4.4. Technological developments as an adaptation

Adaptation by technological developments will be necessary in a changing cryosphere. Infrastructure and buildings constructed on ice-rich permafrost cause permafrost thaw (e.g., Nelson et al., 2002; Anisimov et al., 2007) and this is discussed in Chapter 5 as a major reason for structural damage to buildings (see Section 5.4). The challenge is particularly large in Russia, where many housing and industrial facilities in the northern and southern Russian Arctic, especially in the Sakha Republic, were built on permafrost. Techniques to reduce warming and thawing, such as heat pumps, convection embankments, thermosyphons, winter-ventilated ducts, and passive cooling systems are already common practice in North America, Scandinavia, and Russia (e.g., Instanes et al., 2005). Permafrost engineers must address the problem of preserving infrastructure under projected future climate conditions. One solution is to construct new buildings at the rate at which existing buildings are damaged and abandoned. For existing projects such as the Norman Wells pipeline along the Mackenzie River in the Northwest Territories, Canada, monitoring programs are important not only to facilitate the mitigation of environmental impacts associated with the project (Naviq and AMEC, 2007; Prowse et al., 2009), but also to mitigate the effects of ongoing climate change. As a way of adapting, the Canadian government has enhanced the permafrost-monitoring program in the Mackenzie region to ensure impacts related to hydrocarbon resource development are minimized. For high-risk projects such as hazardous waste storage, the environmental protection agencies can apply the principle of 'perpetual' design, which implies engineering to meet a wide range of uncertainties in the future.

Figure 10.8. Icebergs constitute a hazard to Arctic offshore oil operations. A tug boat stands by to tug an iceberg away from a drilling area. Photo: Will Rose / Greenpeace.

Shipping in Arctic waters requires adaptive technology that can handle treacherous and unpredictable ice conditions, such as ice-strengthened hulls (as required by the IMO 'Polar Code' guidelines). Different solutions for securing safe and cost-efficient transport have been developed or are under development (IMO, 2009; Ho, 2010). These include using dedicated ice-strengthened vessels or ice breakers for the ice-covered areas and applying ship-to-ship transfer in open water (e.g., for commodities such as liquefied natural gas), satellite navigation systems for improved maritime surveillance, and double acting ships with the bow optimized for open-water conditions and the stern designed for icebreaking (Commission of the European Communities, 2008; Molenaar and Corell, 2009). Increased shipping will, however, require more accurate hydrographic charting, enhanced infrastructure to support shipping, better weather forecasting, and enhanced capacity to respond to emergencies and accidents.

Extractive industries view adaptation to Arctic conditions as an opportunity to develop new technology to improve efficiency and reduce costs. Advances have already been made in current focal areas of Arctic sub-sea processing, compression and power transmission, drilling and pumping technologies, and communication technology. Other focal areas are ice-load and ice-scour prediction; offshore Arctic drilling systems; Arctic tanker technology and emergency, evacuation, and spill response systems and technology; and cost-effective logistics.

10.4.5. Adaptation to extreme weather events

With a changing cryosphere, an increased frequency of extreme events is projected. Observations of such increases are already being reported (e.g., Oozeva et al., 2004; Nickels et al., 2005; Laidler and Elee, 2006). In particular, sea-ice retreat is likely to result in more severe weather in future (Kolstad and Bracegirdle, 2008). Increased weather variability is having profound societal and cultural impacts on residents across the Canadian Arctic (e.g., Furgal and Seguin, 2006; Riewe and Oakes, 2006). Traditional sea-ice use and commercial activities such as fisheries, extractive industries, and shipping will both be affected by more extreme weather in the future. Strong winds affecting coastal sea ice and creating open-water areas have consequences for travel and harvesting activities. Open water also leads to the formation of low clouds and fog, which produce freezing drizzle and significant icing on aircraft and vessels (Hanesiak et al., 2010), and requires innovation of drilling procedures as well as procedures for other stages of resource development. Currently, the Northeast Atlantic (the Greenland, Iceland, Norwegian, and Barents Seas) is particularly prone to marine cold-air outbreaks (MCAOs).

The frequency of severe snowfalls and extreme precipitation events is also projected to increase over large areas of the Arctic (Christensen et al., 2007). Snowpack density changes over time and as a result of climate change with consequences for buildings and infrastructure (e.g., Strasser, 2008). It is no longer enough to calculate snow load for buildings based on historical mean snow density (Newark et al., 1989). Similarly, projected increased avalanche occurrences will require adaptive measures.

Wind action may cause significant ice ride-up and pile-ups, which could extend more than 100 m inland (Taylor, 1978;

Kovacs and Sodhi, 1980). Ice pile-up in extreme events has been reported as high as 36 m (Sverdrup, 1904) and 50 m (Taylor, 1978). Pile-ups occur quickly (in about 30 minutes) and have caused fatalities and significant damage to infrastructure and roads, even overtopping dunes (Forbes et al., 2004). Because of the highly episodic and rare occurrence of damaging ice motion, little has been done to mitigate ice ride-up and pile-up hazards, little effort has been devoted to forecasting, and many communities where damage has occurred may not have anticipated the hazard before it occurred.

Sea-ice decline combined with increased frequency of storm events increases exposure to coastal erosion. Increased open-water fetch and waves are major contributors to coastal erosion, which is exacerbated by permafrost degradation along coastal tundra areas. With sea ice forming later in the season, the coast is more exposed to the projected rise in the number of autumn storms and storms with longer fetch. The complex interactions between declining sea ice and other consequences of climate change (e.g., rising sea level, shifting river discharges, runoff, altered sedimentation rates in coastal areas) will increasingly affect Arctic coasts, coastal infrastructure, coastal marine ecosystems, and potentially human resource use (Forbes et al., 2008; Martin et al., 2009; Perovich and Richter-Menge, 2009; see also Chapters 5 and 9). Adaptation measures to such events include development of protective infrastructure, often at great cost, and relocation of settlements and existing infrastructure.

10.5. Governance

- Cryospheric change and the resulting changes across all aspects of human endeavor in the Arctic require effective and inclusive governance at all levels including:

  ○ local-scale attention to sustainable management, regulation of key activities, mitigative, and adaptive solutions;

  ○ regional-scale promotion of effective and integrated management of all development with appropriate enforcement;

  ○ national-scale enactment of appropriate laws and development of regulatory criteria; and

  ○ international-scale development of global conventions regulating climate forcers and international standards for activities.

- Existing governance structures face challenges across scales and major effort is required to develop and implement new governance structures or refashion older regimes.

Governance is key to solving societal challenges, across local, national, and international scales, emerging from a changing cryosphere. Cryospheric changes create new, albeit differentiated by region, forms of access to the Arctic affecting the adaptive capacity and resilience of local communities, national management, and international regulations. Human rights issues, including indigenous rights (Declaration on the Rights of Indigenous Peoples, 2007), may be challenged and resolution mechanisms required. Inclusive governance will be necessitated at all societal and international, regional, national, and local administrative levels. Governance challenges resulting

from a changing Arctic cryosphere must be seen and addressed in connection with other issues affecting the region (e.g., ACIA, 2005; Armitage et al., 2007; Huebert and Yeager, 2008; Young, 2009a). Oil, natural gas, and mineral developments; airborne pollutants; sovereignty claims and security concerns; the socio-economic dimensions of renewable resource (e.g., fisheries and wildlife) extraction; and demands of northerners to have a greater role in decision-making as they adapt to changes all involve interrelated, multi-scale, and complex issues of governance. Figure 10.9 shows marine jurisdiction in the Arctic.

10.5.1. Multi-level governance institutions and co-management

Cryospheric changes will place additional pressure on existing frameworks for cooperation developed by Arctic nations (see Young, 2005; Stokke and Honneland, 2007). Efforts to secure sovereignty claims to Arctic territory raise militarization issues in relation to natural resource development. Although the Arctic states have not ruled out a region governed by adaptive flexible institutions designed to enhance equitable access to resources, top-down, sovereignty-based approaches to declining sea ice and resource competition are signaled (e.g., The Ilulissat Declaration, 2008). Such approaches, however, may be in opposition to solutions derived from sub-national needs of sea-ice service users across the Arctic.

The 1982 United Nations Convention on the Law of the Sea (UNCLOS) (current version, 1994) constitutes the primary mechanism for dispute resolution related to rights of exploitation and navigational routes in international waters and provides a legally binding framework to resolve state-based conflicts (Huebert and Yeager, 2008; Stokke, 2009). An additional 'Arctic Treaty' has been advocated, which may ensure the sovereign rights of Arctic nations, but with greater

flexibility to respond to the outcomes of cryospheric change (see Huebert and Yeager, 2008; Rothwell, 2009). However, the Ilulissat Declaration (May 2008), signed by the five Arctic Ocean boundary nations, supported the view that UNCLOS provided the necessary framework for resolving potential territorial disagreements and development challenges (Young, 2009a,b; Stokke, 2009).

This indicates an ongoing commitment to the current national level Arctic actors to engage in nation-to-nation debates over the future of the Arctic rather than promoting any new international legal strategies or institutions (Lovecraft and Eicken, 2009), maintaining the predominant position of nation states in shaping Arctic policy. The declaration thus implies that sub-national or regional interests (e.g., indigenous groups, non-governmental organizations, and local and regional governments) are not an independent part of the 'orderly' settlement of any possible overlapping claims.

The Arctic Council, a non-treaty, intergovernmental, ministerial-level forum, addresses issues of common concern, in particular on sustainable development and environmental protection, and has been identified as a critical actor in terms of environmental governance (Young, 2005; Huebert and Yeager, 2008). This role is enacted through the enhancement of the knowledge base regarding the consequences of change in the Arctic. However, as issues regarding the Arctic, especially shipping issues, may impact other stakeholders who are not members of the Arctic Council, a more inclusive approach may be required (Ho, 2010). An example of such could be to use, for example, the Antarctic Treaty as a basis or example for multi-lateral use of the region.

Effective co-management arrangements, defined here as the sharing of power and decision-making authority between state-based agencies, communities, and other actors, can play
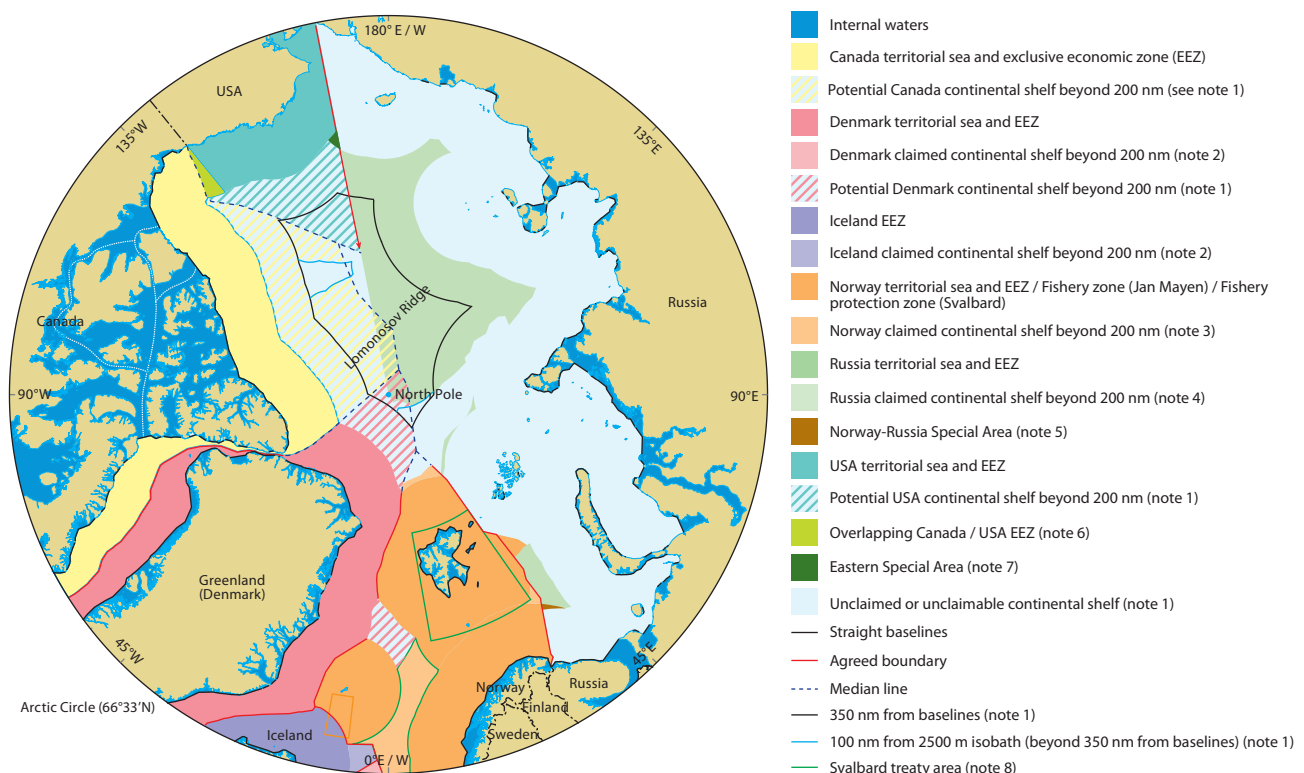

Figure 10.9. Marine jurisdiction in the Arctic. Source: International Boundaries Research Unit, Durham University. Briefing notes referred to in the key are available at www.durham.ac.uk/ibru/resources/arctic.

a central role in fostering the co-production and sharing of knowledge about change and how best to respond to different types of change (Ayles et al., 2007; Kofinas et al., 2007; Armitage et al., 2009; Berkes, 2009). Co-management arrangements are also shown to foster vertical and horizontal institutional linkages that connect local communities with higher order decision makers, thereby fostering multi-level governance (Agrawal, 2008; Berkes, 2009). However, co-management regimes must be carefully designed to foster fit between local users' (usually indigenous communities) concerns and national or sub-national government priorities in order to promote long-term sustainability (Robards and Lovecraft, 2010). In the absence of fully participatory implementation, co-management mismatches can arise and hinder resource users' ability to develop adaptive capacity and governments' ability to collect accurate resource data.

Local institutions have a critical role in structuring the impacts and vulnerabilities of different groups to change, in reconciling individual and collective responses to change by providing incentives and disincentives to collaborate, and facilitating the delivery of resources (information, financial resources) to help various actors cope with change and build capacity to adapt to future changes (Agrawal, 2008).

Applying inclusive, collaborative processes, in which indigenous participants from communities and regions and non-indigenous stakeholders (e.g., scientists) work together, has been shown to produce valuable and integrative insights and to contribute to policy development and governance processes that reduce vulnerability and facilitate adaptation (Riedlinger and Berkes, 2001; Laidler, 2006; Keskitalo and Kulyasova, 2009; Ford et al., 2010b; Hovelsrud and Smit, 2010). This reflects a wider process whereby several Arctic states have shifted more political power and administrative authority to northern peoples (e.g., through land claims), while indigenous peoples are obtaining more representation on such bodies as the Arctic Council (Huebert and Yeager, 2008; Ho, 2010).

10.5.2. Industrial development and regulation

As the cryospheric changes increase, the difference between Arctic residents' view of change and industry's view of opportunities may become more pronounced. Future development of the possibilities for expanded industrial activities in the Arctic is likely to attract diverse stakeholders, and rules of engagement with northern peoples will need to be rethought. The United Nations Declaration on the Rights of Indigenous Peoples (2008) provides the basis for such discussions (see also the United Nations International Labour Organization Convention 169).

Hydrocarbon discoveries may change the economic conditions and way of life of local inhabitants. This depends foremost on the geographical location of these resources (see also Figure 10.7), but also on the involvement of the local population, on the distribution of revenues at a national level, and on the global market for these resources. Addressing the regulatory and stakeholder issues, including the study of impacts on the environment and local communities, should be a key focus area for industries and communities and governments (Seidlitz, 2009).

In 2007, the Arctic Monitoring and Assessment Programme (AMAP) undertook an assessment of oil and gas activities in the Arctic. Having found that oil and gas activity is a major driver of social and economic change, and will most likely increase, the report concluded that governance, regulatory systems, and international standards are important aspects of the performance of the oil and gas industry in the Arctic (AMAP, 2007).

Trans-Arctic marine shipping is expected to be an important driver of Arctic development (Commission of the European Communities, 2008; Molenaar and Corell, 2009), the regulation of which may require local institutional engagement, especially as shipping expansion may impact on stakeholders that are not members of the Arctic Council (Ho, 2010). In some Arctic waters, indigenous peoples have been the sole seasonal users of an Arctic waterway for perhaps hundreds of years. Fisheries and oil and gas activities are to a certain degree 'competing over space' at sea, which will continue and perhaps become more acute in a global warming and cryospheric change context. It will therefore be important to enhance marine safety and ensure marine environmental protection through proper rules and regulations.

Responses to non-renewable resource extraction often include industry applied and broader education and training programs. However, increased demand for local labor has implications for competing local businesses and industry, and concern has been raised about the transferability of industry-specific skills. As oil and gas and mining activities move through their lifecycle and specific jobs shift to different geographical areas, job retention may require willingness to move, separating local residents from their homelands and extended families. Thus, existing and anticipated developmental possibilities can still form barriers for local involvement in non-renewable resource development.

10.6. Cumulative effects and synergies in and beyond the Arctic

- The effects of climate change on various cryospheric components may accumulate both additively and synergistically (i.e., multiplicatively) to result in substantive direct and indirect impacts on humans in the Arctic. These will affect different human groups and sectors of society in different ways.

- Cryospheric change effects will further interact both additively and synergistically with other factors affecting human development in the Arctic (e.g., population growth, industrialization, transition from subsistence to cash and from informal to formal economies, and renewable resource production), resulting in rapid and unprecedented changes for peoples across the Arctic.

- Mitigative and adaptive strategies, including effective scalable and rapid responses to challenges, appropriate anticipatory management protocols for human endeavors, and relevant governance, are all required to meet the challenges of such change.

The combination of changes to the cryosphere with other climatic changes (e.g., temperature rise and sea level rise) and

non-climatic drivers (e.g., increased industrial activity, socio-economic development, and changing demographic patterns, governance, and health and well-being of the Arctic societies and people) will result in multifaceted and cascading effects.

The impacts of individual cryospheric changes on society are clearly numerous. However, a systematic assessment of cumulative effects is still missing. The extent and direction of changes are not easily identified, and it is the combination of many factors that creates the opportunities and challenges locally and beyond the Arctic. Accurate projections of complex cryospheric-societal interactions are not possible to make on theoretical grounds (see also Box 10.5 and Figure 10.10). In addition, the consequences of such interactions depend on context and location (e.g., Huntington et al., 2007b). In some cases, adaptive measures may have unexpected negative consequences and may even exacerbate the problem (maladaptation). This section recognizes the limitations of current knowledge, but through available examples ventures to discuss some possible cumulative effects and synergies.

10.6.1. Cryospheric drivers of change

A number of cumulative effects of a changing cryosphere emerge from the different chapters of this assessment report, and this section highlights a few of these interactions by describing how changes in cryospheric elements or other climatic drivers of change may exacerbate or counteract each other.

Thawing permafrost is linked to snow conditions, and sea-ice reduction may expose coastal permafrost to greater wave action and fetch. Coastal areas with saline and discontinuous permafrost are particularly exposed to the combination of increased wave action, sea level rise, and thermal erosion, and presently no simple engineering solutions exist (see Chapter 5). This combination will lead to increased coastal erosion, with consequences for infrastructure and increased contaminant load (e.g., mercury) (see Outridge et al., 2008). In addition, cumulative effects on permafrost (temperature, precipitation, storm frequency and magnitude, and slow, down-slope movements over time) will add to its thawing rate and will potentially increase impacts on settlements, roads and railways, and infrastructure such as sanitation structures.

Melting sea ice combined with thawing permafrost will release contaminants trapped in the frozen ground and ice (also discussed in Chapter 11, Section 11.3). Melting sea ice will also open waterways and increase the pathways for transport of contaminants into the Arctic Ocean (see also Chapter 9, Sections 9.1 and 9.2). Due to air and ocean currents carrying pollutants into the Arctic from their source regions at lower

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Box 10.5. Uncertainty

Projections of the consequences of cryospheric change for human society are impeded by the accumulation of uncertainties across several levels. Models that themselves are only an estimate of changes may be hampered by a lack of observations of physical variables, extrapolations of other measurements, deductions of parameters from other variables, measurement errors, biases, and preferential locations of measurement stations. In addition to inheriting the uncertainty of physical models, models of climate impacts have their own uncertainties:

- The climate system's natural variability is unpredictable, including interannual to multi-decadal variations. These natural variations can mask (or exacerbate) anthropogenically driven changes over annual to decadal scales, yet they can have effects as large as anthropogenic sources.

- High uncertainty exists in the greenhouse gas forcings (e.g., IPCC scenarios A2, A1B, B1).

- Uncertainties also exist with respect to how aspects of cryospheric change may act to exacerbate theoretical expectations (e.g., methane production and release from degrading permafrost). The effects of these uncertainties will increase over time as the consequences of mitigation strategies are realized. Natural variability is a major reason for across-model simulations (projections) for particular time periods – most models capture natural variations, but their timing will not coincide with those of other models (or the real world) if these variations are natural or internal.

Uncertainties in measurements, modeling, and understanding the physical changes of the cryosphere are amplified when projecting their effects on ecosystems upon which people rely; that is, responses of complex biological systems are in themselves inherently unpredictable, which increases uncertainty of the status of future ecosystem services potentially available from them. Further amplification of uncertainty results when considering the consequences of such changes upon humans and their societies. Figure 10.10 illustrates how changes in cryospheric components may enhance or counterbalance impacts, but also shows how the level of uncertainty about impacts grows with the level of integration of cryospheric components.

Societal, political, and economic influences operating independently of climate change further increase the basic uncertainty inherent in the system. For example, future trade patterns and commodity prices, infrastructure development and technological advances, future greenhouse gas emissions, environmental policies, and how climate will behave as a result of human influence are unknown factors playing a role in future vulnerability to cryospheric and related climate changes relevant to Arctic peoples. Similarly, the nature and scope of adaptive activities, new ways of 'doing business', and many future options are unknowable at present, except perhaps in the most general context. Uncertainty from all sources also increases over a temporal horizon; that is, the more distant in the future the projection is made, the greater the inherent uncertainty in outcomes. Despite the fact that these factors compromise projections of the future, they do not negate the need for adequate preparedness.

While biogeophysical and modeling uncertainties can be estimated, the increasing complexities and levels of uncertainty limit the possibilities to quantify the extent of impacts of cryospheric changes on human society (von Below and Persson, 2008). Anticipating, adapting to, and managing climate and cryospheric changes within these uncertainties are required in a timely manner. Although limited, there is currently sufficient knowledge to understand the direction and trends of the changes for developing appropriate policies to meet the challenges.
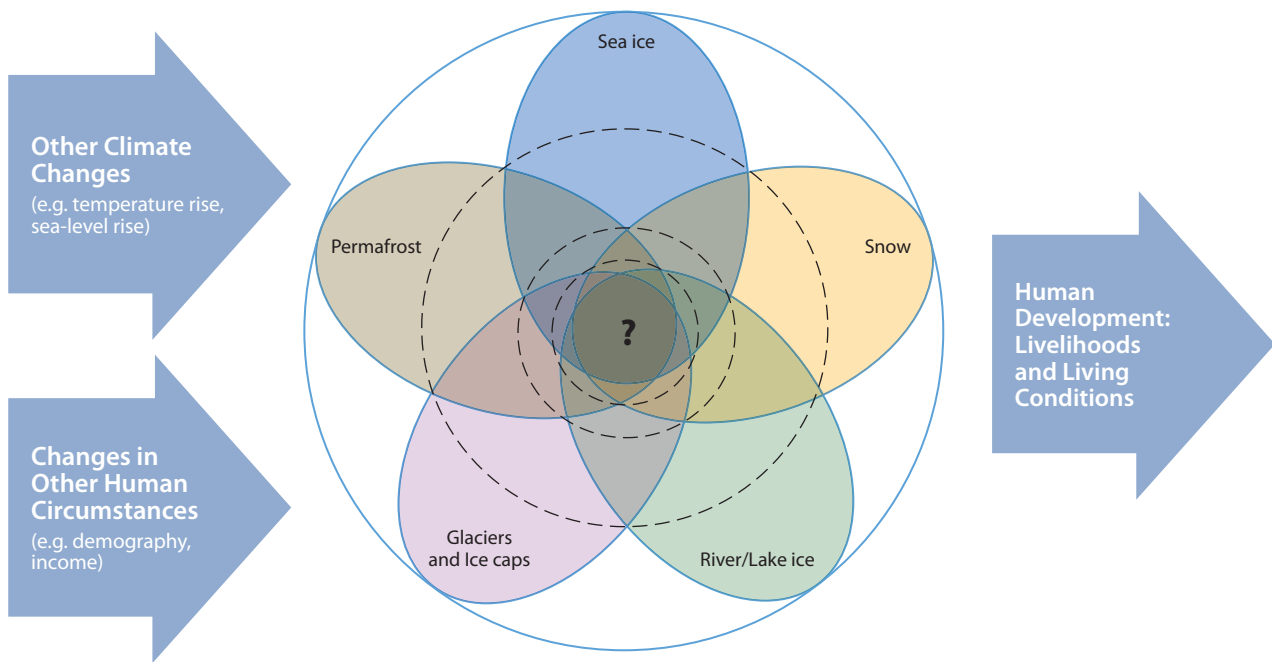
Figure 10.10. Uncertainty. Changes for each cryospheric component will have physical, biological, and societal consequences. Integrative aspects (indicated by increasing overlap of components toward the center of the figure) at different temporal and spatial scales may counterbalance or enhance certain consequences. At the same time, the degree of uncertainty of consequences increases with increasing overlap of cryospheric components (toward the center of the figure, indicated by dashed circles).

latitudes, the Arctic acts as a sink for various types of pollution (Macdonald et al., 2005; AMAP, 2009), including particulate aerosols such as black carbon (Box 10.6). Several studies indicate that airborne pollution leads to accelerated melting of sea ice, and a proposed industrial development in Arctic regions may enhance this effect (e.g., Shindell and Faluvegi, 2009). Accordingly, melting sea ice and thawing permafrost may release great amounts of stored pollutants and adversely affect the health of Arctic residents. Decisions about spatial planning, land use and transportation routes need to incorporate the potential increased risk of slope failure, rockfalls, pollutant release, and coastal erosion to ensure safe communities.

Changes in snow, ice, and permafrost conditions affecting wildlife and fish species important for human livelihoods

Box 10.6. Black carbon

Black carbon (carbonaceous aerosols) results from incomplete combustion of fossil fuels and other organic material (e.g., forest fires) and is a potent, although short-lived, climate-forcing agent. Black carbon is only one of a number of aerosols that may have contributed significantly to Arctic surface temperature increases. Shindell and Faluvegi (2009) estimated that black carbon was responsible for 1.09 ± 0.81 °C of the contribution to warming by aerosols in general (1.48 ± 0.28 °C) between 1976 and 2007. In contrast to carbon dioxide (CO_2), whether delivered from local sources or through transport from longer distances the effects of black carbon are mostly local in that it directly affects the albedo of local snow and ice surfaces (Ramanathan and Carmichael, 2008; Shindell et al., 2008). Deposition of black carbon onto high-albedo surfaces such as snow and ice reduces albedo and enhances radiative absorption, thus fostering melting. Much black carbon deposition occurs in the immediate area of its production, but significant amounts are transported northward from southern sources in the Northern Hemisphere (Hegg et al., 2009). These sources include industrial and agricultural combustion, forest fires, desertification, and general greenhouse gas production. Local northern sources include localized development (e.g., mines, shipping, forest fires). Mitigative measures such as filtering

of soot and increased efficiency associated with fossil fuel combustion are highly effective in reducing the production of black carbon. As a result, as industrial development increases in the North and transportation pathways shift with cryospheric change, regulation of black carbon production from all sources, especially those resulting locally from activities in the Arctic, is required at all scales (local, regional, national, pan-Arctic). Such regulation will benefit from increased compliance and enforcement at all scales and across all sectors; however, initiation of environmental standards is required limiting black carbon production during northern development. Similarly, reduction of black carbon production from southern sources, thereby limiting its transport to the Arctic, is an immediate mitigative measure that could yield substantive benefits by slowing many aspects of cryosphere degradation. There is, however, a potential positive feedback effect linking cryospheric processes promoting drying of sub-Arctic forest and tundra environments (e.g., permafrost degradation, drainage) that, in turn, might foster increased wildfire frequency. Increased fire occurrence will, of course, produce more black carbon and accelerate melt of snow and ice. Thus, increased black carbon production resulting from cryospheric degradation (and all consequences of such) is an example of an integrative forcer linked to cumulative effects of climate change generally.

are likely to exacerbate consequences of societal change and modernization. That is, direct effects from development (e.g., landscape impacts of industrial developments) may affect the northern renewable resource sector, which in turn will also integrate impacts from cryospheric changes (see also Chapter 11, Section 11.4 and the ecological impacts sections in each chapter).

10.6.2. Non-cryospheric drivers of change

Cryospheric change is just one of many factors that shape vulnerability and adaptation to changes in Arctic communities and sectors. Such changes occur in the context of many other factors including resource accessibility, allocation, and extraction policies; limited economic opportunities and markets; access constraints; demographics; attitudes and perceptions of change; bidirectional local-to-global linkages; infrastructure; threats to cultural identity and well-being; transfer of local and traditional knowledge; economic and livelihood flexibility; and enabling institutions. These aspects are rarely independent of each other and frequently combine across scales and sectors (Hovelsrud et al., 2010b). In many cases, socio-economic changes are likely to have greater immediate significance than cryospheric changes, which in turn will affect the ability to adapt.

In general, the effects of industrial development and infrastructural encroachment are experienced by reindeer herders throughout the Arctic as being far more significant than those of cryospheric change (Stammler, 2005; Tyler et al., 2007; Forbes, 2008; Forbes et al., 2009). For example, reindeer herders in the Nenets Autonomous Okrug and the Yamal Nenets Autonomous Okrug consider unbridled oil and gas development on their territories to be a far greater threat to their resilience than extreme weather events (Nuttall

et al., 2005; Forbes, 2008; Forbes et al., 2009). The placement of infrastructure in pasture areas has raised alarm among reindeer herders who fear detrimental effects from windmill parks and mines on reindeer foraging behavior and efficiency, land access and quality, and consequently their livelihoods. Increased industrial activity challenges migration and grazing on traditional pastures (Stammler, 2005; Forbes et al., 2009), leading to areas of unused and underused grazing areas (Caballero et al., 2007; Forbes and Stammler, 2009). Avoidance of certain reindeer pastures can lead to a reduction in optimal range use, leading to complications with herding, increased costs, and reduced production.

Another example is the projected increase in tourism. While there are potential economic benefits for local communities from increased tourism, such an increase may also negatively affect those communities. Given international pressure for sustainable management of wildlife, indigenous communities that depend on hunting, for example, polar bears (for tourism or subsistence) may have to adopt alternative livelihoods (e.g., Hovelsrud et al., 2008). Further expansion of tourist seasons may result in extended use of infrastructure and longer duration of employment and income benefits. How Arctic communities allow their natural and cultural resources to be used by large numbers of visitors and whether or not this is viewed as a disruption depends on the attitude and resources of the community. Impacts may be mitigated through collaborative management with tour operators and in some instances, tax revenues and special fees may offset local costs.

In a review of projects from the Human Dimension of the Arctic System initiative, Huntington et al. (2007b) stated that human activities can greatly amplify the effects of climatic variability and change on Arctic societies (Figure 10.11). This is
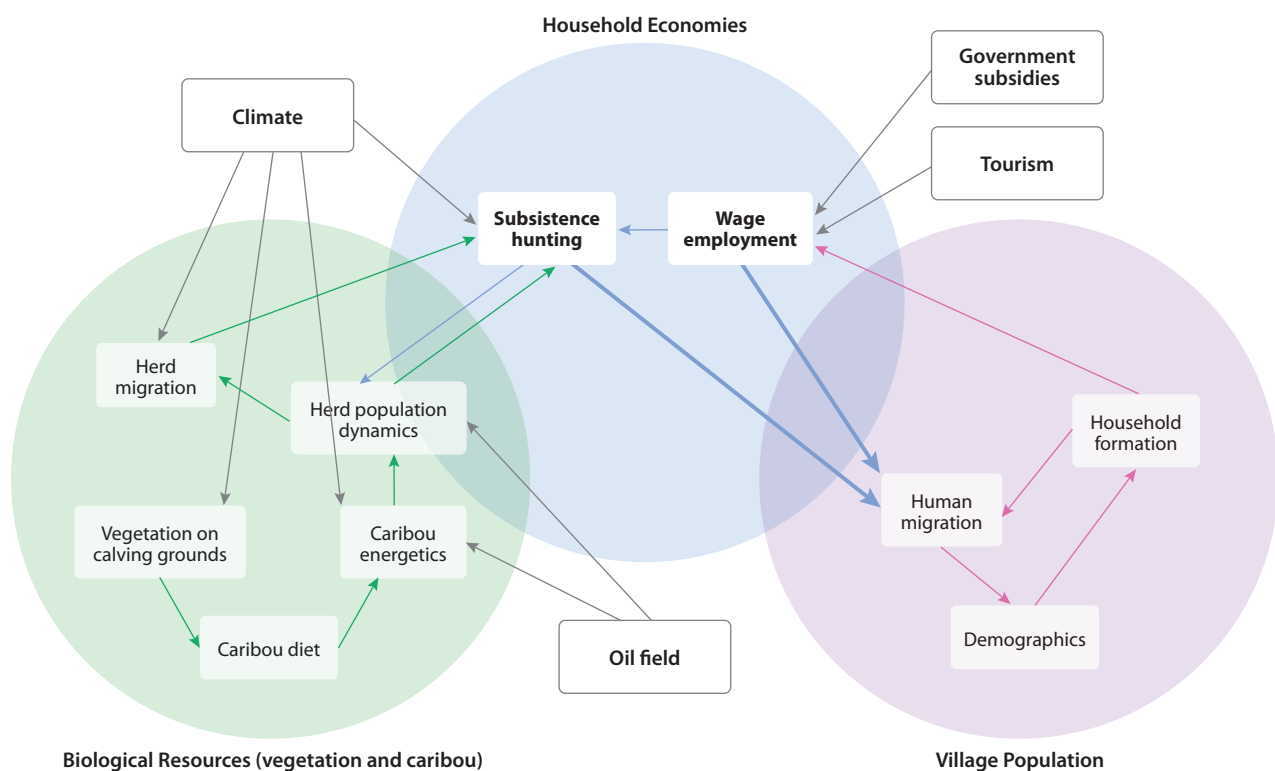

Figure 10.11. Example of complexity of cumulative effects. Source: Huntington et al. (2007b).

exemplified by a study of the cod stock collapse in the late 1950s, which resulted from the combined effects of environmental change and high fishing pressure. When combined, these effects were more severe than the environment or fishing might have caused alone. Similarly, in a study of coastal erosion in Barrow, Alaska, public health was greatly improved through the use of 'utilidors' (utility corridors) for drinking water and sewage. However, two of the pump stations that pump sewage into the lagoon are close to the shore and exposed to major storms, which leaves the entire system at risk of flooding. This made Barrow more vulnerable to coastal erosion and flooding, even if sea ice had not been decreasing (for both studies see Huntington et al., 2007b for references). The latter example also shows how adaptations may turn out to be counterproductive and exemplifies the complexity of and current limits in understanding of the issue.

Ongoing demographic changes may have similar or even more wide-ranging consequences than changes in the cryosphere (or climate). Gender perspectives of development and generation shifts are two important demographic factors. Traditional gender roles regarding who the breadwinner is in a family are changing. In some indigenous communities for example, there is an increasing shift away from the traditional male 'hunter' toward female wage earners, and the new generation may seek other forms of employment than the traditional resource-use activities. Out-migration is another critical demographic factor with consequences for livelihood activities (Rauhut et al., 2008).

Globalization and world markets are also important players in shaping change in the Arctic. However, within this wider milieu the immediate and ongoing additive effects of cryospheric change will compound the effects of changes in the Arctic. In Greenland (Chapter 8), the impacts of climate change have reinforced ongoing socio-economic changes in the same way that climate change has contributed to outlining possible development paths. On the other hand, current adaptations to climate change and policy are predominantly reactive rather than proactive. Cryospheric changes may demand a more proactive adaptation approach in the future to ensure sustainable livelihoods in the Arctic.

10.6.3. **Beyond the Arctic**

There are important linkages between the Arctic cryosphere and the rest of the world; these include physical (e.g., Arctic climate system feedbacks globally), chemical (e.g., pollutant transport to the Arctic), biological (e.g., migratory biota and ecosystem connections), and societal (e.g., tourism, resource extraction, management, politics) linkages. While global climate change is exacerbating Arctic cryospheric change, these changes will in turn cascade to far-reaching subsequent effects of global consequence. The physical feedback connections are discussed extensively in other chapters: for example, Chapter 5 (Section 5.3.4), Chapter 6 (Section 6.3.1), Chapter 8 (Section 8.3.3.3), Chapter 9 (Section 9.2) and Chapter 11 (Section 11.1). The cold Arctic region typically acts as a sink for heat, greenhouse gases, particulate aerosols, and contaminants and performs fundamental regulatory functions for global climate systems. Degraded cryospheric components will increasingly act less so with highly uncertain consequences; however, hemispheric- and global-scale effects are likely. In some cases, these will be responses to Arctic change that affect the globe (i.e., simple outputs), whereas in others they will constitute feedback effects which may further alter global processes (e.g., climate system change, weather pattern change, shifting from sink to source for some greenhouse gases, possibly both CO_2 and methane). Such changes are very likely to affect human society beyond the Arctic in some context over the short (i.e., decadal) and long (i.e., century and beyond) term.

Cumulative effects with consequences for society on a broader scale include the feedback from changing snow conditions and surface albedo and the release of methane by thawing terrestrial and sub-sea permafrost to the climate system (e.g., see Chapter 4, Section 4.4.4 and Chapter 5, Section 5.3.4) leading to accelerated global warming. Such amplification needs to be included when developing mitigation efforts and in updated models projecting future climate change. In contrast, the increased period of open water on lakes and increased growing season for vegetation will lead to increased evaporation and transpiration, drying of the landscape, and a negative feedback to climate, especially when combined with drainage of tundra water bodies fostered by permafrost degradation, although the balance between these two processes is highly uncertain (see Chapter 4).

Changes in the Arctic cryosphere coupled with other climate-driven changes have hemispheric and global-scale societal effects, which may include the following:

- Enhanced ice outflow (icebergs and ice export) leading to shipping hazards in the North Atlantic.

- Alteration of oceanic and riverine heat and freshwater transport to Arctic environments and alteration of oceanic circulation patterns, affecting ecosystems and impacting fisheries and hunting activities.

- Alteration of the structure and functional relationships within and services received from terrestrial, freshwater, and marine Arctic and sub-Arctic ecosystems, including possible loss of iconic species and altered biodiversity.

- Increased activities in the Arctic leading to increased risk of pollution and increased shipping bringing noise pollution and ballast water that contains contaminants and invasive species (Berner et al., 2005; Occhipinti-Ambrogi, 2007; AMAP, 2009).

- Uptake of pollutants via microorganisms into the Arctic food web (e.g., fisheries) that have local and global health impacts.

- Economic opportunities that may provide impetus to national and global economies but must be developed in a sustainable manner.

- Significant contribution to global sea level rise and thus follow-on effects for low-lying coastal regions throughout the world.

10.6.4. **'Winners and losers'**

The concept of 'winners and losers' is complex, context- and location-specific, varies with time, and pertains to multiple factors of change (cryospheric and societal). Industries, which may be located outside the Arctic, will generally gain from cryospheric change, but this is contingent on global and local economic and infrastructure developments. Likewise, while

some local indigenous communities may be viewed as 'losers', some communities may be 'winners' from emerging industries (through jobs, revenues, health care etc.) or changes in wildlife or fish stocks. Yet others may not benefit from industrial development in their region and climatic or environmental changes will add to the challenges (Huntington et al., 2007a,b). In addition, while a community may 'win' in some respects (e.g., economic and health), it may 'lose' in others (e.g., loss of traditional livelihood and culture). Some argue that the notion of 'winners and losers' is not applicable in the Arctic context. There is a need to balance the potential for long-term economic benefits with the shorter-term costs that an unpredictable and changing climate forces cultures to bear in terms of livelihood and social identity.

10.7. Gaps in knowledge and future needs

- Cryospheric degradation will result in cascading effects on Arctic environments and peoples; their livelihoods, living conditions, and quality of life; and the regional climate systems. These effects will have global consequences.

- The precise ways that people and communities will be affected need further study, particularly with regard to adaptation, changes in behavior, health, and resource development.

- Effective observations and monitoring across geographical and temporal scales, sectors, and disciplines are all required in a linked interdisciplinary manner to provide measurement of change, assess efficacy of mitigative and adaptive strategies, and to prioritize needs for governance and other appropriate responses.

The impacts of Arctic change generally are occurring sooner, more rapidly, and more broadly than anticipated only a short time ago. The changes are exacerbated by cryospheric changes and demand substantive and rapid responses from all levels of government within nations, the development of international conventions and protocols, and effective and rapid action through international governance mechanisms. Responses at all levels must include adaptive and mitigative actions as well as appropriate regulatory and compliance actions to protect cultural integrity, promote human health, and ensure sustainable economic well-being. Integrative and adaptive forecasts that can help identify effective responses are required in order to facilitate the early development of policies. Better and more inclusive (especially of Arctic residents) ways of doing business are required. How to do so across scales is a major gap.

10.7.1. Research needs

As this chapter has illustrated, much can be said about the ways that a changing cryosphere affects or is likely to affect people in and beyond the Arctic. As should also be clear, there remains a great deal that is unknown or uncertain, both in terms of the trajectory of cryospheric change and in terms of human response and adaptation. Research is needed in two general areas: to better understand how people *can* adapt to a changing cryosphere and to learn from the ways that people

are adapting already. While this section provides examples of important research areas, an exhaustive list is not possible given the range of topics and ways to consider response and adaptation in human societies.

The capacity to adapt and respond effectively to cryospheric change can vary greatly among individuals, communities, and regions. Similarly, vulnerability to changes can differ even between communities that have many similarities. Research in this area is needed to better understand the underlying conditions and processes that shape vulnerability and adaptive capacity, and to better understand the aspirations and priorities of the people and communities in question in order to identify adaptation goals and pathways.

In a similar vein, little is known about the processes and consequences of migration, relocation, and resettlement of Arctic communities. Some have already occurred for various reasons, and more may be anticipated due to cryospheric change such as coastal and riverbank erosion, loss of permafrost and sea level rise. This question involves local dimensions (in terms of individual experiences) and regional and national dimensions (in terms of the ability of governments to support the construction and establishment of new community sites if needed). Such changes must be considered against the backdrop of urbanization around the Arctic.

The loss of sea ice in particular has generated considerable international interest in Arctic natural resources as well as the strategic importance of the region. While there is a great deal of speculation, the global significance of Arctic shipping routes remains uncertain, and the prospects for exploiting many resources depend greatly on worldwide market forces, not just on physical access. The Arctic remains remote and cold, and the development of minerals, petroleum, and fisheries faces many challenges. As such resources are developed, further research is needed to determine optimal ways of doing so to meet regional and national goals such as financial well-being, political independence, and long-term sustainability.

Coastal and riverbank erosion, although not new phenomena in the Arctic, appear to be increasing. Comprehensive assessments have not yet been carried out on the vulnerability of Arctic communities and infrastructure to erosion or on the implications for the placement of infrastructure related to resource extraction. This issue relates in part to that of migration, relocation, and resettlement, in that the ability of national governments to respond to erosion depends in part on the scale of the problem. One community relocation may be possible, but dozens may cost more money than is available, reducing the number of possible responses as time goes by. Rising sea level will also play a role in the long-term effectiveness of any measures taken in the short term.

Ice is not only a barrier to commercial transportation and development. Ice roads allow a relatively inexpensive, low-impact means of getting equipment and supplies to remote locations in the Arctic. They are also used in some regions to connect communities that do not have year-round road access. However, the use of ice roads obviously depends on lasting, reliable ice. As the climate warms and if weather becomes more variable, the reliability and duration of ice roads may diminish. This will raise costs for businesses and communities where simple alternatives to the seasonal ice roads may not exist. More research is needed to examine the economic and social roles

that ice roads play in the Arctic, the vulnerability of ice roads to a changing cryosphere, and the prospects for alternatives.

The impacts of a changing cryosphere on human health have not been examined in detail, although the effects of low air temperatures, low humidity, and other features of Arctic winters have been studied. A changing cryosphere may remobilize contaminants and other air and water pollutants, increasing the availability of these substances to humans either directly or indirectly through the food web. In addition, retreat of glaciers and declining snow cover may lead to more dust in the air from wind events, adding another health stressor. More research is needed on the likely health effects of a changing cryosphere, in part to help health care professionals recognize and respond to new trends in health.

The safety and well-being of Arctic residents may also be affected by a changing cryosphere. Thinner ice, different conditions on land and at sea, and different patterns of fish and wildlife distribution may increase risks, especially if the production of new knowledge about the changing conditions cannot keep pace with the rate of change. More study is needed on risk perception, risk avoidance, and knowledge production in Arctic communities to help determine how to foster effective adaptation. This work should also examine how instrumental and modeling data can be provided to Arctic residents to help them make decisions about their daily activities.

As previously discussed, cryospheric changes are not the only drivers of change, and it is important that research is initiated on the multifaceted processes where impacts of cryospheric changes intertwine with consequences of industrial development in the Arctic. In some areas, hydrocarbon and mining developments are a greater threat to indigenous resilience than a changing climate (Nuttall et al., 2005; Forbes, 2008; Forbes and Stammler, 2009). In order to understand the impact of such developments on indigenous peoples, it is necessary to investigate environmental and societal conditions in combination with consequences of developments and across local, national, and international scales. To date, technology and the use of best practices have reduced environmental impacts, but additional risks may occur as conditions change or new areas are explored and developed. More generally, it is important to assess the combined impacts of cryospheric, industrial, and societal changes on the livelihoods, living conditions, and quality of life of the indigenous peoples and other Arctic residents.

10.7.2. Observation, monitoring, modeling, and downscaling needs

A key contributor to effective adaptation and response is reliable information about the consequences of cryospheric change for a locale. While environmental conditions cannot be predicted accurately, information on likely future trends and events is nonetheless useful for planning. In addition, real-time information can help reduce risks from hazards such as icebergs, floods, and severe storms. A major challenge is to provide such information on temporal and spatial scales relevant to the users. The areas described below are examples of information needs, but as with the description of research needs, the list is not exhaustive.

Establishing baselines of present conditions in order to monitor changes in snow, permafrost, treeline, and sea, river, and lake ice is especially important. In addition, there is a particular need to monitor the effects of cryospheric changes on socio-economic conditions and biodiversity. This will increase understanding of linkages between, and allow for improved forecasting of future changes. Moreover, by mapping how local and indigenous communities and sectors adapt to change, it is possible to create transferable knowledge. Several monitoring networks are already in place: the Sustaining Arctic Observing Networks (SAON) is a newly established network, endorsed by the Arctic Council, which focuses broadly on observing changes in environmental and physical systems. The network will also include observations of social and economic conditions.

Of key importance are improved mapping, surveillance, and monitoring of drifting ice and icebergs, annual variations in ice cover, and continuing local sea-ice formation along with the frequency of storm surges, wave heights, and wind. A number of programs have been developed (e.g., Arctic Report Card, Polar View). Such monitoring is relevant for ship traffic, industrial developments, and local hunting routes (OPRC, 1990; Chircop, 2009). Medium- and long-range forecasts of ice break-up and freeze-up will be an important aid for companies in deciding on the seasonal window for operation, whereas near real-time monitoring will facilitate safe and productive local travel as well as hunting and fishing. For hunting activities that are particularly sensitive to snow and ice conditions (Ford and Furgal, 2009), knowledge of changes in the timing and amount of snow is important. Monitoring snow cover, winter thaw, and rain will also provide information relevant for hydropower and risks related to drought, snow load, avalanches, and floods. The rate of change in snow conditions requires adaptation by reindeer herders and others relying on terrestrial conditions for resource harvesting and travelling. Traditional knowledge of reindeer, pastures, and snow conditions is essential for successful adaptation (e.g., Tyler et al., 2007). In addition, people in the Arctic need information on qualified changes in snow conditions, such as maximum accumulations for snow load calculations, water supply, flood management, avalanche risk, and snow clearing for permafrost stabilization to adapt (Allard et al., 2007).

Monitoring existing structures built on permafrost, follow-up of historical data, and back calculation of known failures are important to gain a better understanding of the ways such structures are affected. Permafrost also plays a role in preserving ancient and recent records of human culture and past biota that are of considerable value to archaeologists and paleontologists. As permafrost thaws, these unique records will be destroyed, and there is a need to survey areas of thawing permafrost to secure the archives of the past. The need for monitoring is not restricted to cryospheric elements, but extends to monitoring the consequences of increased shipping, tourism, and industrial activity such as pollution, emissions, navigation impacts on marine species, and introduction of non-indigenous organisms through ballast water exchanges.

In addition to monitoring, there is a need for new and improved projections of future trends. In particular, to continue developing downscaling techniques to map future cryospheric changes relevant at local and regional levels. Adaptation is a response to local changes in cryospheric conditions and their impacts, and downscaled projections of future conditions that

capture locally relevant changes help adaptation (Hovelsrud et al., 2010a). For such downscaling efforts to be successful, it is necessary to involve local people, indigenous as well as non-indigenous, to identify the most relevant cryosphere (or climate) elements. Downscaled projections can be combined with spatial use of land, lakes, rivers, and sea ice to provide useful information for the users (Hovelsrud et al., 2010a; Rybråten and Hovelsrud, 2010). Projections of specific parameters will need to cover geographical and seasonal extent and coverage (such as location, packing, and thickness of ice) and ideally on timescales ranging from hourly to months. In addition, research is needed to better understand how to connect science to stakeholders for better decision-making in governance.

Incorporating local and traditional knowledge in the interpretation of satellite-derived snow- and ice-cover products and in the development of community-based monitoring programs will provide local data for scenario development and enhance community involvement and decision-making. Local focus and traditional and local knowledge are essential in studies on and for adaptation (Krupnik et al., 2010; Smit et al., 2010) (see Appendix 10.1 for a list of relevant projects under the International Polar Year (IPY) 2007/08). Local understanding of change and adaptation needs can be generalized through the transfer of practical knowledge and can help increase the scientific knowledge base. More and integrated knowledge about adaptation is needed across scientific disciplines and local and traditional knowledge, between scientists and decision makers, and on the international arena (Hovelsrud et al., 2010a).

Not all needs call for creating new and more data; much information on local impacts currently exists in local records and is compiled in unpublished reports and documents in municipalities and infrastructure maintenance authorities (e.g., roads, railways). It is often difficult to locate such data and information, which is critical for assessing the socio-economic impacts of changing conditions. Such information is highly relevant to monitoring and observing change and is an important source for developing adaptation strategies. It is therefore important to catalogue this type of information and make it available. In addition, current monitoring and observation efforts would benefit from including identification of user needs in order to provide useful information for communities and industries for ensuring safety and for developing environmentally sound activities (Hovelsrud and Smit, 2010; Krupnik et al., 2010).

10.7.3. Meeting governance and technology needs

With an expected increase in human activities such as fisheries, tourism, and industrial development follows an increased demand for seasonal ice-breaking of harbors and routes.

Increases in shipping and marine-based development activities will require increases in environmental regulation, safety preparedness and response, and improvements in marine charting, weather, and ice forecasting. Improved disaster management (groundings, sinkings) and training of local search and rescue personnel are of key importance (OPRC, 1990; Chircop, 2009).

Retreating glaciers could lead to an increase in possible hazardous jøkulhlaups, but little is currently known about increased risk to communities in the Arctic. There is limited knowledge about hazards from the distribution, frequencies, and impacts of such events in the Arctic. For increased avalanche potential, adaptation strategies could build on existing measures from other regions; however, regional assessments and development of effective forecasting of risk is required. In a changing cryosphere, adaptation by technological developments will be necessary. Infrastructure must be designed to withstand projected future climate conditions. Regardless of the solutions, monitoring programs will be important not only for facilitating the mitigation of environmental impacts but also for mitigating the effects of ongoing climate change.

A changing cryosphere will affect wildlife habitat, population sizes, and migratory behavior (see Chapter 9, Section 9.3). To minimize further impact, careful consideration of cruise tourism transport schedules and hunting and fisheries conditions are of critical importance to ensuring the cultural and economic well-being of Arctic peoples (UNEP, 2007). In addition, there are environmental, economic, social, and cultural impacts from garbage, waste, and pollution problems that require regulation.

There are clear limits to any Arctic governance regime (regulatory or soft-law), because the drivers of cryospheric change occur largely outside the region and fall beyond the jurisdiction of any one Arctic state. Current changes signal the need for effective interplay across societal levels and scales between Arctic institutions and existing regimes (Stokke and Honneland, 2007; Young, 2009b). In addition, there is a clear need for further monitoring and the use of monitoring data in knowledge co-production and shared learning (Riedlinger and Berkes, 2001; Laidler, 2006; Bravo, 2009).

The wide-ranging changes affecting all aspects of the Arctic, when combined with the fundamental and far-reaching added changes resulting from cryospheric shifts, will result in many challenges associated with all aspects of development. Managing such changes sustainably represents a further challenge particularly in the context of developmental cascades. To achieve orderly development, effective governance strategies must address not only the changes anticipated and responsive adaptive or mitigation measures, but also future opportunities. Appropriate regional-scale management bodies may be required to integrate and assess the totality of shifting environments (cryosphere and climate) and developmental plans over varying timescales. Representation of indigenous peoples and other Arctic residents on such decision-making bodies is imperative.

Appendix 10.1. Relevant projects under the International Polar Year

International Polar Year (IPY) projects related to people in the Arctic, based on the IPY honeycomb chart and project database and selected from 194 projects containing the keywords 'human' or 'people'. The IPY in combination with other international research programs will continue to provide a wealth of knowledge on Arctic climate change and impacts.

Topic	IPY-id	Acronym	Full project name
Dynamic Social Strategies	6		Dynamic Social Strategies in Arctic Environments: Long-term Perspectives on Movement and Communication
Historical Exploitation of Polar Areas	10	LASHIPA	Large Scale Historical Industrial Exploitation of Polar Areas
USNPS Environmental Change	21		U.S. National Park Service Understanding environmental change and its biological, physical, social, subsistence, and cultural effects in national parks and protected areas of Alaska, Chukotka, and the Yukon, through research, monitoring, education, and outreach
History of International Polar Years	27		Changing Trends in Polar Research as Reflected in the History of the International Polar Years
Sami in Literature	30		Representations of Sami in Nineteenth Century Polar Literature: The Arctic 'Other'
Monitoring Development in Traditional Indigenous Land Use Areas	46	MODIL-NAO	Traditional Indigenous Land Use Areas in the Nenets Autonomous Okrug, Northwest Russia
Arctic Change	48	ISAC	International Study of Arctic Change
The Linguistic and Cultural Heritage Electronic Network	82	LICHEN	LICHEN: The Linguistic and Cultural Heritage Electronic Network
History of IPY Field Stations	100		Polar Field Stations and IPY History: Culture, Heritage, Governance (1882 – Present)
Northern Climate Variability	120	NORCLIM	Northern High Latitude Climate variability during the past 2000 years: implications for human settlement
Language, Literature, & Media	123		Glocalization – Language, Literature, and Media among Inuit and Sami people
Community Adaptation & Vulnerability	157	CAVIAR	Community Adaptation and Vulnerability in Arctic Regions
Rangifer Monitoring	162	CARMA	Starting the clock for the CARMA Network: Impacts on Human-Rangifer Systems in the Circumarctic
Inuit, Narwhals, & Tusks	164		Inuit and Scientific Descriptions of the Narwhal, Connecting Parallel Perceptions: Interdisciplinary Studies of the Narwhal with a Focus on Tusk Function
Sea Ice Knowledge & Use	166	SIKU	Sea Ice Knowledge and Use: Assessing Arctic Environmental and Social Change
Human Health Initiative	167	AHHI	Arctic Human Health Initiative
Community Resilience & Diversity	183		Arctic Resiliency and Diversity: Community Response to Change
Food Safety & Wildlife Health	186		Engaging communities in the monitoring of zoonoses, country food safety and wildlife health
Exchange Local Knowledge	187	ELOKA	Exchange for Local Observations and Knowledge of the Arctic (formerly APOC (Arctic Peoples Observations Center), #358)
Northern Material Culture	201	NORMA	Northern Material Culture through International Polar Year Collections, Then and Now: In the Footsteps of Murdoch and Turner
Protecting Traditional Knowledge	206		Legal and Constitutional Frameworks for Protecting Traditional Ecological Knowledge in Northern Canada
Global Change, Social Challenges	210		Global Change – Social Challenges Processes of socio-economic changes in the Circumpolar North, with focus on gender and inter- and intra-generational relations
Political Economy of Development	227		The Political Economy of Northern Development
Bering Sea Community Monitoring	247	BSSN	Bering Sea Sub-Network of Community-Based Environmental Monitoring, Observation and Information Stations
Community-based Research Alliance	248	AICEMI	Arctic Indigenous Community-based Monitoring and Information Stations Network: Arctic Community-based Research Alliance
Conservation Hunting	259		Conservation Hunting in the Arctic: An Analysis of Constraints and Opportunities
Impacts of Ecosystem Disturbances	275		Polar Disturbance and Ecosystem Services: Links between Climate and Human Well-being
Initial Colonization	276	ICOL	Initial Human Colonization of Arctic in Changing Palaeoenvironments

Northern Genealogies	285		Northern Genealogies: Development of an ethnodemographic informational system on the peoples of Siberia and the Russian North
Impacts of Oil & Gas Activity	310	GAPS	The Impacts of Oil and Gas Activity on Peoples in the Arctic Using a Multiple Securities Perspective
Fishery Ecosystems	325	ArcOD	Marine and estuarine ecosystems in the eastern, central, and western Canadian Arctic
Land Rights & Resources	337	CLUE	Dynamics of Circumpolar Land Use and Ethnicity
Economy of the North	355	ECONOR	The Economy of the North
Food Surveillance System	384		Integrity of the Traditional Food System and Environmental Health in the Circumpolar North
Survey of Living Conditions	386	SLiCA-RAAS	Survey of Living Conditions in the Arctic, SLiCA – Remote Access Analysis System
Reindeer Herding & Climate Change	399	EALÁT	Reindeer Herders Vulnerability Network Study: Reindeer Pastoralism in a Changing Climate
Monitoring Human-Rangifer Migrations	408	NOMAD	Social-science migrating field station: monitoring the Human-Rangifer link by following herd migration
Land & Coastal Resources	411	ARD	Norwegian and Russian Arctic Resources: Prospects for Social and Economic Development
Integrated Tools for Communities	431	ARCTEC	ARCTEC: A Cumulative Effects Toolbox for Northern Ecological and Social Systems
Well-being of Arctic Indigenous People	433		Pressures and Impacts on the Health and Well-being of Indigenous People of the Arctic: Invitational International Symposium and Symposium Publication
Cultural Heritage in Ice	435		Culturally and Scientifically Significant Materials Recovered from Melting Ice and Cryosols: Recovery, Research, Stabilization, and Community Education
Relocation & Resettlement in the North	436	MOVE	Moved by the State: Perspectives on Relocation and Resettlement in the Circumpolar North
People, Wilderness, & Tourism	448		People and wilderness resources in Arctic. Is local subsistence harvest and exclusive wilderness tourism a road to sustainable well-being or a source of conflict
Sustainable Development	456	PASDAC	Practical Applications for Sustainable Development in Arctic Communities
Arctic Social Indicators	462	ASI	Arctic Social Indicators
Modeling and Observation	40	DAMOCLES	Developing Arctic Modeling and Observing Capabilities for Long-term Environmental Studies
International Study of Arctic Change	48	ISAC	International Study of Arctic Change
Cryos – State of the Cryosphere	105	Cryos	State and Fate of Polar Cryosphere

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11. Cross-cutting Scientific Issues

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Contents

This chapter presents an integrated assessment of cross-cutting scientific issues that affect multiple components of the cryosphere. These include assessments of feedback mechanisms and global sea-level rise; ways in which cryospheric change affects the transport and bioavailability of contaminants within the Arctic region; ways in which cryospheric change affects Arctic ecosystems; and systems available for observing changes in the Arctic cryosphere.

Key Findings

- In the Arctic cryosphere-climate system more feedbacks result in further warming than result in cooling or reduced rates of warming.

- The overall net effect of all the feedbacks on regional and global climate is difficult to assess because of the variability in the spatial and temporal scales over which the feedbacks operate.

- Arctic mountain glaciers, ice caps, and the Greenland Ice Sheet contributed over 40% of the global sea-level rise of around 3 mm per year observed between 2003 and 2008.

- In the future, global sea level is projected to rise by 0.9 to 1.6 m by 2100 (relative to the level in 2000) and Arctic ice loss will make a substantial contribution to this.

- Exposure of Arctic residents and top predators to contaminants released after many years from melting snow and ice and which accumulate in food chains, could further increase.

- Changes in the Arctic cryosphere cause fundamental changes to the characteristics of Arctic ecosystems and in some cases loss of entire habitats.

- Extreme climate events are likely to continue to have severe effects on many ecosystem types.

- There are large gaps in snow and ice observation networks and in knowledge of cryospheric processes. Closing these gaps will increase the potential for understanding, assessing, predicting, mitigating and adapting to climate change.

11.1. Synthesis of feedbacks and interactions: From the cryosphere to the climate system – effects over various spatial and temporal scales

Summary

Changes in Arctic climate are a result of complex interactions between the cryosphere, atmosphere, ocean and biosphere. The present assessment has substantiated and quantified many of the linkages, processes, interactions and feedbacks between the cryosphere and climate across a range of spatial and temporal scales revealing that:

- In the cryosphere-climate system more feedbacks are identified as positive and result in further warming than are identified as negative and result in cooling (or reduced rates of warming).

- Several positive feedbacks are finite: warming due to reduced albedo will no longer increase in areas that have totally lost snow or ice cover.

- Feedbacks operate at different spatial scales. Many of the feedbacks, such as those operating through albedo and evapotranspiration, will have significant local effects that together could result in a global impact. Some processes such as carbon dioxide (CO_2) emissions are likely to have very small global effects but uncertainty is high. Others, such as subsea methane (CH_4) emissions, could have large global effects.

- Some cryospheric processes in the Arctic have teleconnections with other regions, for example, the loss of sea ice north of Eurasia may result in a cooling effect over eastern Asia, and changes in snow cover affect atmospheric circulation. Conversely, major changes in the cryosphere have been largely a result of large-scale processes, particularly atmospheric and oceanic circulation.

- There are also interactions between the elements of the cryosphere acting through the atmosphere and ocean.

- The cryospheric components of the Arctic play a pivotal role in freshwater generation, its intra-Arctic storage, and routing to the North Atlantic where it can produce an important feedback to regional and global climate. With continued climate warming it is highly likely that the cryospheric components will play an increasingly important role.

- Terrestrial snow cover, sea ice and permafrost are involved in multiple temporal and spatial feedback regimes whereas land ice is involved in fewer feedbacks.

- The overall net effect of all the feedbacks is difficult to assess because of the variability in the spatial and temporal scales over which they operate.

- General circulation models (GCMs) do not include all major feedbacks, in part because they do not include all processes that lead to feedbacks (e.g., freshwater runoff from glaciers and ice sheets). Furthermore, the feedbacks may not be accurately parameterized in the models. The lack of full coupling between surface dynamics and the atmosphere is a major gap in current GCMs.

11.1.1. Introduction

Feedback processes are a phenomenon that has been known for more than a century. They are responses to a driving mechanism that subsequently accelerate (positive feedback) or retard (negative feedback) the original driving process. At the end of the 19th century, Arrhenius (Bolin, 2007) described the classic feedback whereby increased air temperature leads to an increase in the amount of water vapor in the atmosphere, which in turn leads to additional atmospheric warming. More recently, Curry et al. (1995) detailed the sea ice-albedo feedback mechanism, including snow cover, melt ponds, and leads. Overpeck et al. (2005) took a broader approach to the analysis of Arctic feedbacks by including all major components of Arctic climate: sea ice, permafrost, terrestrial ice, thermohaline circulation, terrestrial biomass, and marine primary production, as well as population and economic production. Francis et al. (2009) used a similar presentational approach to distil the system into its fundamental physical and human parts, document the key relationships between those parts, and identify the feedback loops. Both Overpeck et al. (2005) and Francis et al. (2009) examined the changes in feedback loops under the scenario of continued warming and a greatly reduced permanent sea-ice cover.

With the vast amount of recently published information on observed and projected trends in the cryosphere, there is a need to re-evaluate the feedbacks between the changing cryosphere and climate warming. There are many feedbacks from the cryosphere to the climate system: some are direct, but others are complex and indirect; for example, climate effects on the cryosphere that then affect ecosystems that in turn affect the climate system. Some of the complexity and variety of interactions between the atmosphere, ocean, and cryosphere is illustrated in Figure 11.1. Further complexity arises owing to scaling issues: individual feedbacks operate over different time scales and their effects can vary from local to global spatial scales. Furthermore, some feedbacks to climate warming are negative and result in reduced rates of climate warming or even cooling, while others are positive and lead to enhanced warming. Consequently, although some feedback mechanisms have been known for a long time, an overall assessment of the net effect of many different potential feedbacks on climate warming has not yet been achieved.

This section summarizes the feedbacks associated with a changing Arctic cryosphere presented elsewhere in this report, describes other feedbacks and interactions that span the various cryospheric components, and provides a preliminary assessment of their relative magnitudes. However, the calculation of the net effects of all feedbacks requires complex modeling that remains a priority for future research.

11.1.2. Feedbacks

11.1.2.1. Greenhouse gases

11.1.2.1.1. Carbon dioxide (land, freshwater, marine)

Carbon dioxide is a greenhouse gas with a radiative forcing of 1.66 W/m². The adjustment time of CO_2 in the atmosphere is approximately 100 years (Forster et al., 2007). It is exchanged between the atmosphere and the biosphere
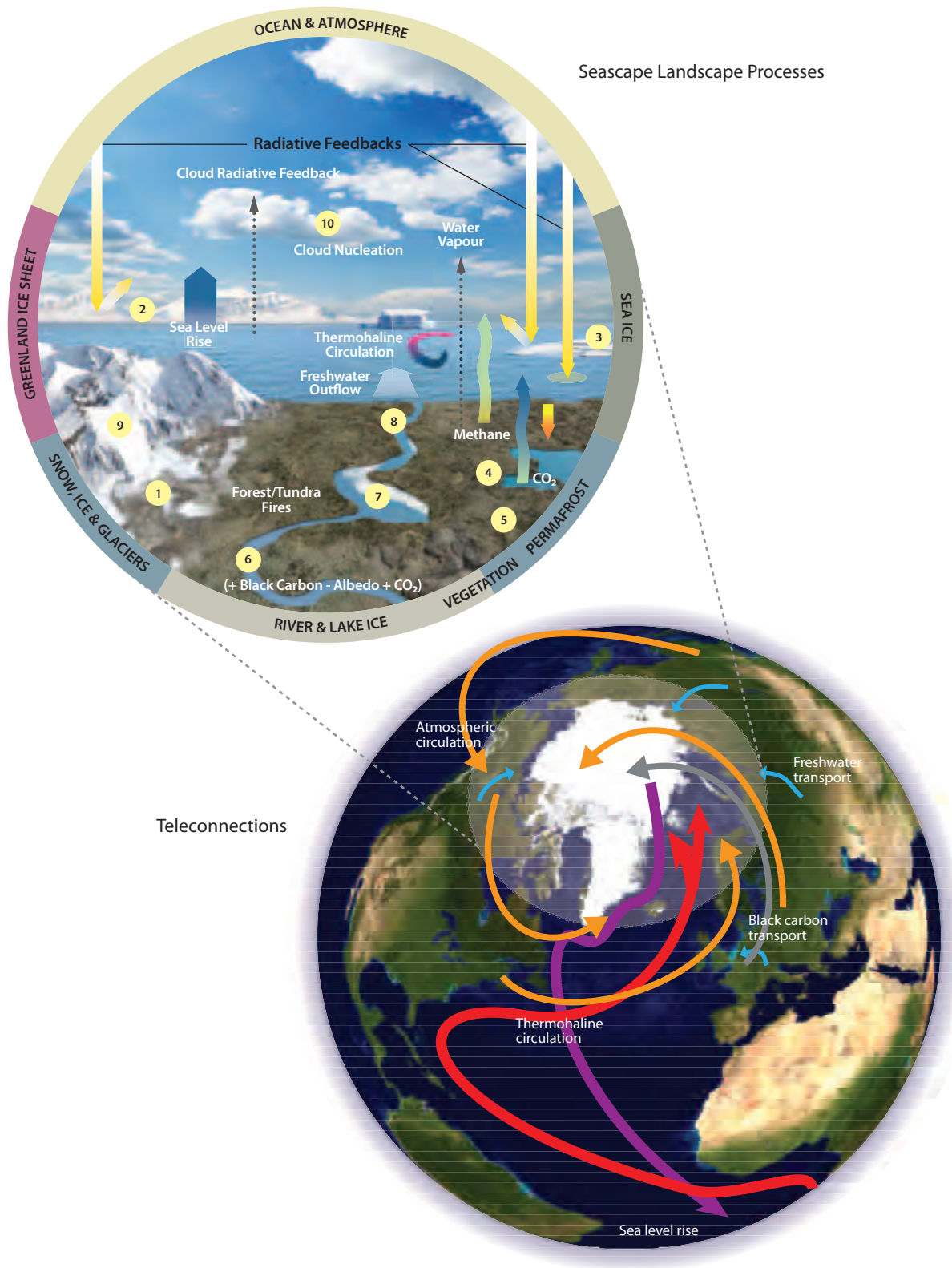
Figure 11.1. Local, regional, and global feedbacks and processes related to changes in the Arctic cryosphere. The numbered yellow circles refer to impacts and processes: (1) Melting and retreating snow cover increases radiation absorption, a radiative feedback. (2) Melting of large ice sheets contributes to sea-level rise and the freshwater flux with potential effects on thermohaline circulation and global climate. (3) Retreating sea ice contributes to increased radiative absorption (ice-albedo feedback) and heat and moisture fluxes to the atmosphere, which impact cloud cover. (4) As permafrost degrades, CH_4 production may increase. With wetland drying, CO_2 emissions may increase, and the atmosphere warms over time. (5) Thawing permafrost changes geomorphic / geochemical processes and fluxes. (6) Increasing precipitation plus melting snow and ice increases river flow and changes the freshwater flux. (7) Shrinking lake-ice cover has ecological impacts generally leading to greater productivity but negatively impacting surface transport. (8) Changes in the magnitude and timing of snowmelt runoff and river-ice processes have both positive and negative impacts. (9) Retreating glaciers initially increase runoff but lower flows eventually result as ice masses diminish. (10) Changes in cloud cover affect the surface radiation budget, which impacts sea ice, in turn affecting cloud cover (cloud-radiation feedback). Source: after Prowse (2009).

through the processes of photosynthetic capture by green plants, short-term autotrophic respiration by plants, and long-term heterotrophic respiration of dead plant material by microbes. A mismatch, particularly in wet areas, between the rate of fixation of atmospheric CO_2 and its release from the biosphere has lead to a net accumulation of carbon in the Arctic for thousands of years with considerable amounts preserved in permafrost (see Chapter 5). Around 44% of the world's near-surface labile soil carbon is found in Arctic soils (McGuire et al., 2009; see also references in Chapter 5), which is equivalent to about twice as much carbon as is currently present in the global atmosphere (Anisimov et al., 2007; McGuire et al., 2009; Tarnocai et al., 2009). This carbon store is sensitive to climate warming.

The major cryospheric controls on carbon *capture* from the atmosphere are the duration of the snow-free period that determines when photosynthesis can occur efficiently, the timing of the snow-free season (earlier snow melt allows plants to harvest high levels of solar radiation whereas solar angles are low and radiation is reduced at the end of the snow-free period; Callaghan et al., 2004a) and damage to vegetation from extreme winter warming events that thaw snow briefly in mid-winter (Bokhorst et al., 2009). Euskirchen et al. (2006) estimated increased carbon drawdown at about 9.5 $gC/m^2/y$ for each day that the growing season was projected to increase between 2001 and 2100. Recent analyses from visible satellite imagery of changes in snow-cover extent show that decreases are more marked in the spring period than in the period of snow-cover onset, and that in May-June there has been an average decrease of 18% over the 1966 to 2008 period, thereby extending the snow-free season (see Chapter 4).

The major cryospheric controls on carbon *release* from the soil are thawing permafrost (see Chapter 5) and soil warming during extreme events such as fire. Experimental warming of soils (Dorrepaal et al., 2009) and observations of areas that are experiencing different degrees of warming (Lee et al., 2009; Schuur et al., 2009; Vogel et al., 2009) show that the Arctic is already losing some of the carbon that has been stored for thousands of years. Extrapolations from measured carbon emission rates suggest that 4.5 to 6.0 kg/ m^2 (or 9.5–13%) of the soil organic matter carbon pool could be lost on a century time scale (Schuur et al., 2009). Currently, widespread permafrost thawing throughout the Arctic has not been reported, although permafrost temperatures are generally rising. However, recent projections of permafrost temperatures show thawing to be widespread throughout the southern regions of the Arctic by 2090 (see Chapter 5).

Overall, models suggest that the Arctic will remain a weak sink of carbon during warming, mainly because of increased growth and extent of more productive vegetation in the future (Callaghan et al., 2004b; Sitch et al., 2007; McGuire et al., 2009). However, both current and future carbon sink activity can be reversed by short-term disturbances such as forest fires and insect pest outbreaks, during long-term changes. The long-term changes include thawing permafrost and altered hydrology related to permafrost dynamics (see Chapter 5) and changes in snow regime (see Chapter 4) that lead to drying and plant water stress. Also, the models exclude disturbance due to human activities such as deforestation at the treeline. Although the future feedback sign and strength status for CO_2 in the Arctic remain uncertain, it is likely that century-long processes of negative feedbacks from increased CO_2 sequestration will be interspersed with episodic releases following disturbances. Some of these disturbances (such as forest fires) will also intensify the positive feedback arising from black carbon deposition on snow and ice through long-range transport on an annual time-scale.

The Arctic Ocean could be a significant sink for carbon as sea ice retreats and the ocean warms. Also, the oceans are sinks for carbon over geological time scales as dead organic matter accumulates in deep ocean sediments and carbonate in the shells of some plankton species becomes calcareous rocks. Currently, there can be mismatches between algal blooms at the retreating ice edge (now over the continental shelves) and the zooplankton and bacterio-plankton that feed on them: the algae that are not eaten fall onto the continental shelves and are available for the benthos (creatures living on the seabed) to feed on or form sediments. As the ice edge retreats away from the continental shelves during climate warming, more blooms could occur, which might lead to more organic material falling to the deeper ocean bed off the continental shelves and accumulating there. However, this negative feedback to climate warming might be moderated by ocean acidification. The pH of the ocean is currently decreasing as more CO_2 is drawn down from the atmosphere: the slightly more acidic conditions that result affect the species that use calcium to build cell walls causing less calcium to be deposited onto the ocean bed.

Freshwaters (lakes, ponds, rivers) feed back to the atmosphere through processes that occur in their immediate vicinity (e.g., carbon fluxes, albedo, evaporation) and downstream where the waters enter the Arctic Ocean (see Section 11.1.2.3). As temperatures rise, the number of days of ice cover decreases and more heat is absorbed by the open water which in turn increases CO_2 capture through the longer duration of primary production at higher rates. However, many northern lakes and ponds are already supersaturated with carbon and are net sources to the atmosphere (Jonsson et al., 2003). This source is likely to grow following increases in the supply of organic material from permafrost thawing and carbon transport to the lakes from enhanced terrestrial plant production, but will be determined by the balance between drying and water-logging of the land surface which is as yet unknown.

The processes and feedbacks described above are all biological in nature but physical and chemical processes within the sea ice itself have also been suggested to play a role. Anderson et al. (2004) and Rysgaard et al. (2009) found that levels of dissolved inorganic carbon increase with depth below the surface mixed layer of the Arctic Ocean. This is a result of enhanced air-sea exchange of CO_2 caused by sea-ice formation and the rapid transport of high salinity brine out of the surface layer. Similarly, Alekseev and Nagurny (2007) showed that the polar amplification of the annual cycle of CO_2 over the Arctic Ocean is connected to sea-ice growth, where the formation and growth of ice during the winter is accompanied by an increase of CO_2 in the sub-ice water layer and a CO_2 flux into the atmosphere. Model results showed that melting of sea ice exported from the Arctic Ocean into the East Greenland Current and the Nordic Seas increases the seasonal CO_2 uptake in the area by around 50%. Alekseev and Nagurny (2007) speculated that, overall, the Arctic Basin is more likely to be a source than a sink of CO_2 over an annual cycle.

11.1.2.1.2. Methane (wetlands, lakes, sub-sea)

In anaerobic soils, ponds and lakes of the Arctic, microbes produce CH_4 rather than CO_2. Methane is 25 times more powerful than CO_2 as a greenhouse gas over a 100-year time frame with a radiative forcing of 0.48 W/m^2 (Forster et al., 2007), so any increases in CH_4 emissions will be particularly important. Unlike CO_2, CH_4 is difficult to measure over large areas because there are geographical hotspots for CH_4 production and because there are episodic releases of CH_4. For example, recent research (Mastepanov et al., 2008; see also Chapter 5) has shown large releases of CH_4 in the autumn from High Arctic permafrost areas in northeastern Greenland, and thaw lakes in Siberia are hotspots for CH_4 production (Walter et al., 2006, 2007a,b).

In addition, it is difficult to project future releases of CH_4 from land because wetlands on permafrost are drying in some areas (Smith et al., 2005; Riordan et al., 2006; Bunn et al. 2007; Goetz et al., 2007; see also Chapter 5), which would decrease current CH_4 emissions. Conversely, thermokarst pond formation is occurring elsewhere with concomitant increases in CH_4 emissions (Christensen et al., 2004; Ström and Christensen, 2007; Myers-Smith et al., 2008; see also Chapter 5). A relatively small source of CH_4 can result in positive radiative forcing even though most of a large area can be a net sink for carbon if vegetation is productive.

Particularly large sources of CH_4 occur in former soils that were inundated when sea levels rose after the last ice age. These areas are the vast continental shelves of the Arctic. Recent investigations in the Laptev Sea have shown high levels of CH_4 throughout the water column and extending up to 1800 m in the atmosphere (Shakhova et al., 2008a,b, 2010; Westbrook et al., 2009). Overall, a 1% release of CH_4 stored in subsea hydrates would be equivalent to a doubling of atmospheric CO_2 concentration in terms of its radiative effect. Although the uncertainties of the size of the subsea carbon reserves and their stability are great, the potential risk from CH_4 release, some of which is already being documented (Shakhova et al., 2008 a,b, 2010), is sufficiently large to cause concern.

11.1.2.1.3. Nitrous oxide (tundra peatlands)

Very high emissions of the powerful greenhouse gas nitrous oxide (N_2O; radiative forcing of 0.16 W/m^2, Forster et al., 2007) have recently been discovered from peat circles in patterned ground of the discontinuous permafrost zone in Russia (Repo et al., 2009). These peat circles are widespread in certain regions. Elberling et al., (2010) also reported high N_2O emissions of 34 mg $N/m^2/d$ in cores taken from northeastern Greenland and incubated in the laboratory that are equivalent to daily N_2O emissions from tropical forests on a mean annual basis. These discoveries bring Arctic permafrost N_2O sources, previously thought of as of very little importance, to the agenda for global studies of N_2O dynamics and their impacts via feedbacks.

11.1.2.1.4. Water vapor

Water vapor is a greenhouse gas with a radiative forcing of 0.07 W/m^2 (stratospheric water vapor from CH_4; Forster et al., 2007). In the Arctic, it is transported from the surface to the atmosphere via evaporation from open water and wet surfaces, evapotranspiration from vegetation, and sublimation from snow (King et al., 2008) and ice. The feedback effects of the processes generating water vapor are complex in that evaporation leads to local cooling, but the increased atmospheric concentrations of water vapor can lead to warming over a wider area because of mixing and transport of water vapor in the atmosphere (positive feedback). It is further complicated by the possibility of increased atmospheric water vapor enhancing cloud formation, which can have either a warming or cooling effect on the surface depending on the cloud height and time of year (Rouse et al., 1997, also discussed below). Assessing and comparing the local versus wider impacts of water vapor will require additional research.

Direct evaporation from melting snow and ice will decrease as the areas of glaciers, lake, pond and river ice, and the duration of ice and snow cover decrease over much of the Arctic (see Chapters 3, 4 and 5). Sublimation should increase initially as the temperature difference between snow and air increases.

Evapotranspiration is expected to increase in those areas of the Arctic where plant production increases, for example in response to a warmer, earlier and longer snow-free season (Euskirchen et al., 2006). Projections of increased forest growth and extent in the Barents region of the Russian Arctic show that by 2080 summer temperatures will have decreased by 1.5 °C due to evaporative cooling (Göttel et al., 2008). Although this cooling is a local effect, the atmospheric mixing of the water vapor is likely to increase warming over larger geographical scales and other feedbacks are also associated with changes in vegetation, such as albedo.

11.1.2.1.5. Ozone and bromine

An important feature of Arctic atmospheric chemistry is the relationship between sea ice, bromine, and ozone. Releases of bromine gas species from the ocean can result in sudden ozone depletion in the lower troposphere during spring. One possible mechanism of bromine release is sea-salt aerosol production from snow lying on sea ice during blowing snow events. A reduction in Arctic sea ice would reduce the importance of this process in the chemistry of the Arctic atmosphere. In chemistry model simulations, Voulgarakis et al. (2009) found large spring ozone increases (up to 50–60%) over the Arctic, due mainly to a reduction in the impact of bromine chemistry, caused by sea-ice retreat. Tropospheric ozone has a relatively small radiative impact (warming), although the effect is greater over bright surfaces (Shindell et al., 2006). With a declining ice cover potentially giving rise to an increase in tropospheric ozone, the feedback would be positive. The impact of a shrinking sea-ice cover on polar stratospheric ozone was studied by Scinocca et al. (2009). They found that the stratospheric response is springtime polar cooling that is dynamical rather than radiative in origin. In the model simulations, the response lags the onset of the sea-ice loss by about a decade. It is associated with an enhanced weakening of the North Atlantic Meridional Overturning Circulation.

11.1.2.2. Transfer of heat and energy between the cryosphere, atmosphere, and ocean

11.1.2.2.1. Albedo

A continued and increased melt of mountain glaciers and ice caps, the Greenland Ice Sheet, and sea ice; an earlier melt of river and lake ice; and a decrease in snow cover will all impact on the Arctic radiation budget. Albedo, the fraction of the

incident sunlight that is reflected by the surface, will decrease and more solar radiation will be absorbed by the ground or open water. This will lead to additional warming. Furthermore, a decrease in albedo resulting from a decrease in snow cover over land may result in a further change in albedo through a change in the vegetation in response to surface warming (Chapin et al., 2005; Euskirchen et al., 2007, 2009). Permafrost thaw is also expected to lead to changes in albedo as areas dry out or become waterlogged. Indirectly, changes in surface albedo can modify large-scale circulation (Dethloff et al., 2006).

For the Arctic glaciers (including those in Greenland surrounding the ice sheet) projected volume loss ranges between 0.049 to 0.131 m sea-level equivalent (SLE), or 12–32% of their current volume by 2100 depending on the choice of GCM (Radić and Hock, 2010). In the Russian Arctic, there will be differences in the extent of glacial retreat between the Arctic continental climate systems of the mountains and Kamchatka. Any translation of losses of ice volume into decreased extent will reduce albedo due to increases of bare ground.

Projections of sea-ice extent show a continued and possibly accelerating decline, leading to an ice-free Arctic Ocean in summer within this century and possibly even within the next thirty years. The thinning of the ice pack, earlier melt onset, and increased open water will contribute to a reduction in albedo. While a variety of recent studies have demonstrated the importance of increased absorption of solar radiation resulting from less sea ice (and hence a lower albedo surface), some recent research suggests that the ice–albedo feedback is potentially less efficient than previously thought (Winton, 2006, 2008; Bitz, 2008; Graverson and Wang, 2009). For example, the annual sea-ice minimum is reached in September at a time when incoming solar radiation is already rather weak.

The albedo of terrestrial snow-covered surfaces in the Arctic is a major feedback to climate over a large area (Fernandes et al., 2009), and an important feedback to climate globally through atmospheric and oceanic teleconnections. Recent observations show a general decrease in snow-cover extent and duration that has led to decreases in albedo. Projections indicate increases in snow cover in some areas at the highest latitudes, strong decreases in the southern Arctic, and uncertainty in areas between. These changes will generally lead to an overall net reduction in the albedo feedback. Snow properties also affect albedo, for example, grain size, age of the snow cover, and the content of extraneous material such as black carbon from industrial pollution and biomass burning south of the Arctic. In fact, warming resulting from black carbon deposition on snow is equivalent to that resulting from a doubling of atmospheric CO_2 in Eurasia (Flanner et al., 2008; see also Chapter 4).

The reflective properties of snow are strongly modified by vegetation in many and complex ways (Pomeroy et al., 2006; Essery et al., 2008; see also Chapter 4). While snow typically absorbs around 10% of thermal radiation incident on it, black spruce *Picea mariana* can absorb 95% of thermal radiation leading to a significant positive feedback (Juday et al., 2005). The snow feedback will be decreased as vegetation height increases above the snow pack during the ongoing and projected process of shrub and tree range extensions into the Arctic (Sturm et al., 2001; Tape et al., 2006). Although increases in shrub advance in the Alaskan Arctic have not yet resulted in warming, it is predicted that an increase of shrubs could increase summer heating of the atmosphere by 3.7 W/m², which

is equivalent to a doubling of CO_2 or a 2% increase in the solar constant (Chapin et al., 2005). In the Barents region, changes in albedo through forest advance could increase temperatures by 1 °C in spring. This example should be compared with the cooling effect of increased evapotranspiration in summer and the expected cooling (not yet calculated) of the increased draw-down of atmospheric CO_2 giving an extra 2 to 17 kg C/m² by 2080. Together, these projections demonstrate the complexity of feedbacks arising from changes in vegetation: the feedbacks operate in different directions, through different mechanisms and over different periods of time.

11.1.2.2.2. **Cloud feedbacks**

The exact nature of the cloud radiation feedback in a warming climate is uncertain. Clouds both reflect solar radiation and absorb longwave (terrestrial) radiation, the magnitudes of which depend on cloud amount, height, particle phase and size, and thickness. Some satellite studies have shown that wintertime cloud amount in the Arctic appears to have been generally decreasing since the early 1980s, but springtime cloud amount has been increasing. While wintertime clouds in the Arctic have a warming effect, springtime cloud can have either a net warming or cooling effect. The overall radiative impact of these cloud cover changes on the surface is one of increased cooling. Therefore, changes in cloud cover may have actually suppressed Arctic warming to some degree (Wang and Key, 2003).

The influence of changes in cloud cover on sea-ice extent and *vice versa* is an important part of the feedback process, but has not been studied extensively until recently. On the time scale of a single season, changes in cloud amount may have minimal influence on summer sea-ice melt. The negative cloud anomaly (less than average cloud amount) and increased downwelling shortwave radiation flux from June through August 2007 did not appear to contribute substantially to the record ice extent minimum in September. On the other hand, there are clearly interdependencies between trends in cloud cover, surface temperature, and sea ice extent. Over the past few decades, more than 80% of the observed surface warming in the western Arctic Ocean during autumn is attributable to decreasing sea ice. Similarly, over 80% of the winter surface cooling in the central Arctic is a result of changes in cloud cover. In spring, only about half of the surface warming is a result of changes in cloud cover (Liu et al., 2009).

Satellite and reanalysis data have shown that sea-ice retreat is linked to a decrease in low-level cloud amount and an increase in mid-level clouds (Schweiger et al., 2008). This is in contrast to the common notion that a warming ocean surface will increase surface evaporation and lead to more low clouds. While surface evaporation does increase, near surface atmospheric instability increases, so cloud formation occurs in the mid-troposphere rather than nearer the surface. However, the radiative effect of these changes is relatively small, as the cloud radiative forcing changes are compensated for by changes in the lower tropospheric temperature and humidity structure. Because of the seasonal timing of this process, the response of cloud cover to sea-ice loss plays a minor role in regulating the summertime ice-albedo feedback. Its role in the cloud-ice feedback in autumn is potentially larger, although the ice-albedo feedback itself is weaker in autumn due to the

relatively low amount of solar radiation at very high latitudes (Kay and Gettleman, 2009).

Increasing temperatures and levels of ultraviolet (UV)-B radiation affect biological emissions from marine ecosystems (Zepp et al., 1995). The grazing of phytoplankton by zooplankton triggers the release of dimethyl sulfide (DMS), which is emitted to the atmosphere where it acts as condensation nuclei for water droplet formation. In other words, the release of DMS from the surface stimulates cloud formation. Furthermore, UV-B radiation converts some dissolved organic sulfur to carbonyl sulfide and warming of the ocean reduces its solubility, thereby releasing more to the atmosphere where more cloud nuclei are formed. The currently increasing presence of sea-ice leads during the summer melt in the Arctic Ocean increases the biological activity of the region. This may result in an accumulation of organic material, especially in the surface microlayer. Particulate organic matter, including microorganisms, small water-insoluble particles and microaggregates, can form a substantial part of the summer aerosol over the open leads in the central Arctic Ocean, thereby enhancing cloud formation (Matrai et al., 2008).

11.1.2.2.3. Atmospheric circulation

Overall, the atmosphere is a driver of change in the Arctic (Francis et al., 2009). It primarily forces rather than responds to changes in the cryosphere, a good example being changes in atmospheric wind patterns over the past decade that have contributed to recent reductions in summer Arctic sea-ice extent (Ogi et al., 2010). But the increase in late summer open water area has, in turn, directly contributed to a modification of large-scale atmospheric circulation patterns. With a reduction in sea-ice cover in late summer, additional heat that was stored in the ocean is then released to the atmosphere in autumn. The transfer of heat from the surface to the troposphere is more effective as the strength of the normally stable Arctic boundary layer is eroded (Overland and Wang, 2010). In years with reduced sea-ice cover the lower tropospheric thickness is greater. This has a large-scale impact, even into the northern mid-latitudes, as the pressure fields and therefore winds are directly related to the atmospheric thickness. Responses can be complex, as the loss of sea ice north of Eurasia may result in a cooling effect over eastern Asia.

There is also a relationship between snow cover and atmospheric circulation. Recent observational studies show that above-normal winter snow depth over European Russia and a corresponding below-normal snow depth over central Siberia – the east-west snow dipole – are associated with reduced Indian monsoon rainfall and above normal sea-surface temperatures (SSTs) over the eastern and central tropical Pacific Ocean during subsequent winters (Ye and Bao, 2001; Ueda et al., 2003; Peings and Douville, 2010). Similarly, below-normal winter snow depth over European Russia and a corresponding above-normal snow depth over central Siberia are associated with increased monsoon rainfall and below normal SSTs. Some GCMs also show this relationship, but the patterns may differ somewhat from the observations. Forcing from the Himalaya region dominates, where a delayed snowmelt leads to a reduction in surface sensible heating and consequently a weakened latitudinal temperature gradient. The result is less Indian rainfall, particularly during the early part of the summer season.

The large-scale effect of changes in snow cover through the snow albedo feedback (SAF) has recently been examined. For example, Fletcher et al. (2009) demonstrated a non-local influence of SAF on the summertime circulation in the extratropical Northern Hemisphere. In models with stronger SAF, increased land surface warming is associated with large-scale sea-level pressure anomalies over the northern oceans and a poleward intensified subtropical jet. This would result in a change in heat and moisture fluxes into/out of the Arctic, although the feedback on snow cover and the impact on sea ice is not clear.

Accumulation and ablation on ice sheets, ice caps, and glaciers depend on the atmosphere circulation patterns, and atmospheric circulation is, in turn, affected by the presence of, and changes in, these features. Ice sheets and ice caps change the elevation and form of the Earth's surface. This changes the air temperature and deforms the atmospheric circulation, forming a 'wave shadow'. For this reason, the Icelandic Low exists throughout the entire year, with the Greenland Ice Sheet to the west, as opposed to the Aleutian Low, which disappears in summer. Smaller ice caps influence the atmosphere on smaller scales. Due to the cooling effect of the Greenland Ice Sheet, the air temperature drops as much as 5 to 10 °C in the atmospheric layer several hundred meters above it. Over the ice caps (Franz Josef Land, Svalbard, Novaya Zemlya, Severnaya Zemlya) and large mountain glaciers, the cooling is about 2 to 3 °C and spreads upward to 200 to 250 m. This may affect the cyclone trajectories ('storm tracks') and the life cycle of pressure systems (Krenke, 1982). Katabatic winds over the glacier surface have an impact on near surface atmospheric circulation. The intensity of such processes depends on changes in glacier size.

Changes in the patterns of atmospheric circulation due to glacier and ice cap retreat and/or disappearance may change the accumulation and ablation rates on the glaciers. For example, more cyclones in winter leads to more accumulation; more anticyclones in summer leads to more ablation. However, the consequences are currently unknown.

Recent modeling by Lunt et al. (2004) supports the influence of the Greenland Ice Sheet on atmospheric circulation. In their study, the conceptual removal of the Greenland Ice Sheet results in an increase in atmospheric thickness (geopotential height) in that region. The region of cooling in the Barents Sea in December, January and February is most intense at the surface and is associated with substantial growth of the sea ice throughout the year. The cooling appears to be forced from above by the atmosphere, and amplified by the sea ice-albedo effect. The cooling appears to be linked with enhanced equatorward flow east of Greenland, associated with the increased geopotential height over Greenland, which brings cold air from the pole equatorward over the Barents Sea. In addition to changes in the stationary waves in the atmosphere, the conceptual removal of Greenland also affects the transient behavior of the atmosphere. In the no-Greenland case, the intensity of the North Atlantic winter storm tracks is decreased. Associated with this is a corresponding decrease in large-scale precipitation. The decrease in storm track activity is likely to be related to the fact that in the no-Greenland case, the meridional

temperature gradient is decreased over the North Atlantic.

A retreating sea-ice margin may enhance melting over the Greenland Ice Sheet. Rennermalm et al. (2009) explored the spatial and temporal covariance of sea-ice extent and ice sheet surface-melt around Greenland from 1979 to 2007. Significant covariance was found in western Greenland. An examination of wind direction patterns and a lag analysis of ice retreat/advance and surface-melt event timings suggested that a change in sea-ice extent is a potential driver of ice-sheet melt, in that late summer wind directions bring onshore advection of ocean heat, enhanced by reductions in offshore sea ice.

There is a strong linkage between sea-ice loss and terrestrial permafrost temperature. Lawrence et al. (2008) examined how rapid sea-ice loss affects the terrestrial Arctic climate and ground thermal state using a GCM. It was found that the accelerated warming signal penetrates up to 1500 km inland and substantially increases ground heat accumulation. Furthermore, enhanced heat accumulation leads to rapid degradation of warm permafrost and may increase the vulnerability of colder permafrost to degradation under continued warming. In contrast to the relationship between snow cover and the Indian monsoon, which operates on a yearly time scale, the changes in atmospheric circulation impact permafrost on decadal scales.

11.1.2.3. Interactions with the freshwater budget of the Arctic

All of the cryospheric components play, to varying degrees, roles in the freshwater budget of the Arctic. Changes in the cryosphere can affect the strength of the thermohaline circulation in the North Atlantic, where it is called the Atlantic Meridional Overturning Circulation (AMOC), and hence, global climate. Assessing the magnitude of these effects requires not only an understanding of how the thermohaline circulation responds to freshwater inputs, but also the sources, locations, distributions, and pathways of freshwater into and out of the Arctic (Randall et al., 2007).

In general, the Arctic Ocean receives freshwater inputs from direct precipitation, Pacific water via the Bering Strait (referenced to a particular salinity), terrestrial ice masses, and runoff from river basins, which includes large areas of the northern hemisphere continents. Importantly, the major river basins contributing flow to the Arctic Ocean are of a nival regime (runoff is dominated by snowmelt), where the snowmelt freshet is the largest and most important hydrological event of the year. Within the Arctic Ocean, freshwater amounts also change due to losses from evaporation and to the growth (-) and ablation (+) of sea ice. Very large volumes of freshwater can also be 'stored' in deep basins with highly variable residence times. Freshwater export from the Arctic Ocean occurs primarily through Fram Strait and the Canadian Archipelago to the Atlantic Ocean, where it plays a role in the formation of deep water in the Greenland-Iceland-Norwegian (GIN) Seas.

It is the export of freshwater that has been identified within paleo-records as weakening the thermohaline circulation and causing major cooling events over the North Atlantic. For example, an Arctic Ocean pathway fed by freshwater flows from the Mackenzie River-Beaufort Gyre, instead of the previously supposed Great Lakes-St. Lawrence routing, has been recently identified as being responsible for the shutting down of the North Atlantic thermohaline circulation, resulting in major cooling associated with the Younger Dryas Period (Peltier, 2007; Murton et al., 2010).

Since publication of the Arctic Climate Impact Assessment in 2005 (ACIA, 2005), a number of updates have been made concerning the relative size of the overall freshwater budget terms and of the cryospheric components that contribute to them. These are considered in Section 11.1.2.3.2, gleaned from the preceding chapters of this report and related literature. First, some example post-ACIA freshwater budget evaluations are considered.

11.1.2.3.1. Recent budget estimates

Total freshwater inputs to the Arctic Ocean, calculated at about 8500 km³/y, are dominated by river flow (38%), inflow through the Bering Strait (30%), and precipitation-evaporation directly occurring on the Arctic Ocean (24%) (Serreze et al., 2006). Importantly, this estimate is an order of magnitude smaller than the total amount stored in the Arctic Ocean. Freshwater exports from the Arctic Ocean occur principally through the straits of the Canadian Arctic Archipelago (35%) and via Fram Strait as liquid freshwater (26%) and sea ice (25%). For such calculations, the volumes of freshwater are referenced to a mean ocean salinity of 34.8. Also recognizing that a steady state of the freshwater budget should not be expected, Serreze et al. (2006) noted that their values indicate larger freshwater inflow through Bering Strait and larger liquid freshwater outflow through Fram Strait than earlier estimates by others.

Peterson et al. (2006) conducted an analysis of changes in freshwater budget components for a broader 'Arctic region' than Serreze et al. (2006), which included the additional large land-ocean catchment of Hudson Bay in North America as well as the Nordic Seas and North Atlantic subpolar basins. Increasing precipitation-evaporation over the marine environments and larger river flow, probably also tied to increases in high-latitude precipitation, was estimated to have contributed ~20 000 km³ of freshwater to the total region from lows in the 1960s to highs in the 1990s. Notably, the river trend included a decline in flow for the Hudson Bay system. Sea-ice ablation added a slightly smaller amount of ~15 000 km³, and glacial melt added ~2000 km³. Due to the lack of complete mass-balance estimates for the Greenland Ice Sheet, its contributions were excluded from the latter amount but it was noted to have been ~80 km³/y in the 1990s and to have recently increased to ~220 km³/y (Peterson et al., 2006). The most recent estimate of current annual loss of ice from the Greenland Ice Sheet is similar at about 205 ± 50 Gt/y[1] (2005–2006) (see Chapter 8).

11.1.2.3.2. Freshwater budget components and changes

Snowmelt and river discharge

Compared to all other world oceans, the Arctic Ocean receives a disproportionately large amount of river runoff to its total volume via the Lena, Mackenzie, Ob, and Yenisey rivers that are dominantly nival rivers. River flow provides the largest input to the Arctic Ocean freshwater budget (Prowse and Flegg, 2000). Although the seasonality of observed multi-decadal increases

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[1] A mass loss rate of 360 Gt/y is equivalent to a mean sea level rise of ca. 1mm/y.

in total Eurasian river flow to the Arctic Ocean has not been fully examined, Chapters 4 and 5 note that some increases in the magnitude and advances in timing of the snowmelt freshet on northern rivers have been observed and greater changes are expected in the future. This will be due not only to changes in the size and seasonality of snowmelt but also to the effects of thawing permafrost in changing flow pathways and storage. Such changes in the timing and magnitude of flows are important to how river water is distributed, directed and/or stored in the Arctic Ocean (e.g., Cooper et al., 2008; Jones et al., 2008).

Mountain glaciers and ice caps, and the Greenland Ice Sheet

The overall freshwater contribution from the glaciers (328 556 km²) in a broadly defined, pan-Arctic drainage basin, as well as small ice caps around the Greenland Ice Sheet (but not including the ice sheet itself), is ~1700 km³ for the period 1961 to 2001 (Dyurgerov and Carter, 2004). Freshwater contributions from these glaciers varied annually from near zero to just under 200 km³, far less than contributed by the corresponding nine major rivers (5282 km³/y) within the same pan-Arctic region. However, there is a greater 'positive change signal' from the glaciers than river discharge (Dyurgerov and Carter, 2004). Partly included in this total are the glaciers from the Alaska-Yukon region, which contribute to the North Pacific Ocean waters near the principal Arctic Ocean inflow point, the Bering Strait. These glaciers have experienced some of the most rapid wastage. In the case of Alaska, for example, freshwater contributions have increased from 52 ± 15 km³/y over the 1950s to 1990s period to ~96 km³/y from the mid-1990s to 2001 (Arendt et al., 2002), and more recent sampling suggests increased ablation particularly at low elevations. In reference to the Bering Strait to which these glaciers contribute, flow volumes to the Arctic Ocean have been significantly revised upward (i.e., by ~50% to ~2500 km³/y; Woodgate and Aagaard, 2005), although the exact contribution of ablating glaciers to such volume increases remains unknown.

Some freshwater budget analyses contain little discussion about the role of the Greenland Ice Sheet, despite its strategic placement as a freshwater source in the North Atlantic. In addition to freshwater volume, the location of the input may also be important (Randall et al., 2007) and meltwater runoff from the ice sheet is potentially a major source of freshening that has not yet been included in relevant models (Randall et al., 2007).

Sea ice

The Arctic Ocean is a salt- rather than temperature-stratified ocean and hence, sea-ice growth/ablation and ocean dynamics can be greatly modified by changes in freshwater. It is the salt-stratified upper layers that provide the vertical stability to permit formation of an ice cover. Surface freshwater layers, however, also contribute with the halocline to thermally shield sea ice from bottom melt, which can be driven by the large quantities of heat contained in warmer, deeper water originating from the Atlantic Ocean. Carmack (2000) estimated that the volume of ice forming and melting each year amounted to a 1.45 m equivalent of freshwater (depth of freshwater across the ocean) assuming an un-deformed ice cover, and an additional 0.45 to 0.7 m of freshwater if the

mass in ice ridges is also considered. However, sea ice has undergone significant changes in areal coverage and thickness.

For some key episodic losses, sea ice-bottom melt has been linked to solar heating of the upper ocean (Perovich et al., 2008). Bottom melt can also result from the loss of thermal insulation from the warmer Atlantic water provided by the surface layers of freshwater and cold halocline. The stability of these upper layers, particularly with enhanced vertical mixing, has been identified as a 'key wild card' regarding future sea-ice loss (Serreze et al., 2007). Sea ice is also exported through Fram Strait along with sea-ice meltwater, but export of sea-ice meltwater seems to be the least likely to influence thermohaline circulation (Jones et al., 2008). This could play a large role in changing ice conditions.

Ocean storage and pathways

The fate of sea-ice meltwater and other forms of freshwater is not simply direct export because the Arctic Ocean also holds significant freshwater in storage with variable releases. While about one-quarter of the total is held on shelves, the majority is in the Eurasian and Canada basins, the latter being the largest single freshwater storehouse in the Arctic Ocean. This has been ascribed to the deep halocline of Canada Basin, which stores freshwater from sea ice meltwater, meteoric water (ocean precipitation and terrestrial runoff), and low-salinity Pacific water from the Bering Strait. Estimates of this storage (like the other freshwater budget terms) vary in the literature, and are primarily due to changes in import-export to Canada Basin and, in the accuracy and ability to measure its content. Despite variations in its estimated volume (e.g., ~46 000 km³; Carmack, 2000) and 25 600 km³ (Yamamoto-Kawai et al., 2008), it is generally accepted that the largest source of the average annual freshwater input (~3200 km³) to this and other freshwater storehouses in the Arctic Ocean (Yamamoto-Kawai et al., 2009) is river runoff – estimated by Yamamoto-Kawai et al. (2008) for Canada Basin to be 800 km³/y and to be just slightly smaller than the amount removed by sea-ice formation (900 km³/y). They further estimated that the average export of ice and liquid freshwater from Canada Basin contributes ~40% of the freshwater flux to the North Atlantic.

The storage values vary with time due to atmospheric circulation, which can control pathways of freshwater to/from storage basins as well as the storage/release from within the storage basins (i.e., Ekman pumping, which under anticyclonic [cyclonic] circulation stores [releases] freshwater). Most recently, measurements from Canada and Makarov basins indicate that there has been a freshwater storage increase of up to 8500 km³, and by extrapolation, almost 11 000 km³ in all the deep basins of the western Arctic. By contrast, the Eurasian Basin in the eastern Arctic and closer to the main export to the North Atlantic has experienced a loss of about 3300 km³, giving a net gain of 7700 km³ (McPhee et al., 2009). This is a significant increase being about four times the volume associated with the Great Salinity Anomaly (a near-surface pool of fresher-than-usual water tracked in the subpolar gyre currents from around 1968 to 1982, which affected regional climate) and similar in magnitude to the total 1981 to 1995 sea ice attrition (melt plus export) estimated in the above-noted freshwater budget by Peterson et al. (2006).

11.1.2.3.3. **Model projections**

At the time of the Arctic Climate Impact Assessment (ACIA, 2005), there was concern about the intensification of the hydrological cycle at high latitudes and the effect this would have on the AMOC. ACIA noted that projections by atmosphere-ocean GCMs produced varying results with changes in the maximum strength of the AMOC by the end of the 21st century ranging from zero to a reduction of 30–50%. The suite of ACIA-selected GCMs (Kattsov and Källén, 2005) did not include freshwater runoff from melting ice sheets and glaciers. Hence, it was surmised that the model AMOC sensitivity to global climate change might be too weak, although sensitivity experiments indicate freshwater runoff must be several times 'present-day' volumes to appreciably alter the AMOC.

A strong scientific debate remains about the potential significance of freshwater effects on the AMOC (Randall et al., 2007). For example, Holland et al. (2007) noted that a constituent result among models for the period 1950 to 2050 (observations and modeled results) is an acceleration of the hydrological cycle, including increased ocean net-precipitation, river runoff, and net sea-ice melt. They also noted, for liquid water, a larger export to lower latitudes, primarily through Fram Strait, and storage in the Arctic Ocean. By contrast, export and storage of freshwater in the form of sea ice decreases, although there is significant variability in sea-ice budget terms. Largely similar results were reported by Koenigk et al. (2007).

A number of efforts are underway to more accurately define the role of freshwater in AMOC weakening, the Coupled Model Intercomparison Project (CMIP) and Paleoclimate Modelling Intercomparison Project (PMIP) coordinated effort being major examples. All models used in CMIP show that AMOC weakening projected for the 21st century is caused more by changes in the surface heat flux than by freshwater, although its effect on high-latitude stratification plays a contributing role (Arzel et al., 2008). They further noted that interannual exchanges in freshwater between the GIN Seas and the North Atlantic have a major driving influence on the interannual variability of deep convection over the 21st century.

11.1.3. **Synthesis**

This review and analysis has substantiated and quantified many of the linkages between the cryosphere and climate identified by earlier recent studies (e.g., Francis et al., 2009). Figure 11.2 synthesizes the major feedbacks on climate mediated by cryospheric processes, where the impacts of changes in various forcing variables (e.g., greenhouse gases, albedo) and processes (e.g., atmospheric circulation and thermohaline circulation) on elements of the cryosphere are depicted. The surface or near-surface air temperature is the variable primarily used to drive changes in the cryospheric component and then to assess the magnitude of warming or cooling. The box colors indicate the expected future impact of a change in the variable or process on the climate, where red indicates warming and blue indicates reduced rate of warming or even cooling. The colors also indicate the type of feedback, where red is a positive feedback from current climate warming to future temperature and blue is a negative

feedback. The color does not, however, indicate the direction of change between the two variables alone. For example, a decrease in sea-ice cover increases atmospheric water vapor, which in turn may decrease sea-ice cover even further because water vapor is a greenhouse gas. This is a positive feedback, but the change in sea-ice cover and water vapor are in opposite directions. The box size and intensity of the colors indicates, at least qualitatively, the relative magnitude of the impact. For example, warming caused by a decrease in albedo through a loss of sea ice is expected to be greater than that due to changes in glacier extent. The feedbacks presented in Figure 11.2 represent a relatively simple synthesis of many complex processes. This synthesis is intended to stimulate a better understanding of the processes rather than give a definitive analysis.

The impacts, interactions, and feedbacks in Figure 11.2 are not necessarily independent. In particular, cloud formation is strongly dependent upon the available water vapor, which may be from local sources or transported from lower latitudes (atmospheric circulation). Additionally, changes in one part of the cryosphere may affect changes in others. For example, thin ice has a very high growth rate but is strongly affected by snow cover. Snow is an efficient insulator, so ice with little snow on its surface can grow much faster than ice with a thick snow cover. This allows for the newly forming sea ice to reach a thickness that might even be larger than that of ice that survived the summer. Although the ice-albedo feedback leads to a decreasing summer sea-ice cover, rapidly growing thin ice leads to a recovery of the sea-ice cover during winter. An additional contributor to this recovery is that often there is little snow on the newly forming sea ice. It is presently unclear how an ice-diminished Arctic will impact on snowfall, so it is difficult to assess to what degree the growth of thin ice may offset the increase in ice melt through the ice-albedo feedback.

Despite the interactions between feedbacks, Figure 11.2 provides a snapshot of how changes in a warming climate affect the cryosphere, illustrating the signs and magnitudes of the feedbacks.

11.1.4. **Uncertainties and recommendations**

Most of the terrestrial feedbacks are poorly quantified over large areas although considerable detail is available for a few specific locations. Generalization of the impacts is difficult at best. *A better observational capacity is needed: an improved coverage of the circum-Arctic area is needed and observations need to be sustained over longer periods. It is particularly necessary to determine the net effects of thawing permafrost and changing snow cover on tundra hydrology – is the net effect a dryer or wetter tundra? The net result will be a major determinant of terrestrial/wetland greenhouse gas emissions.*

Large uncertainties exist in cloud feedbacks and subsea permafrost: *these topics require enhanced and sustained research.*

General circulation models currently include few feedbacks. *Modeling studies including sensitivity analyses need to be intensified to include all the major feedbacks and to calculate the net effects.*

Lack of full coupling between surface dynamics and the atmosphere is a major gap in current GCMs. Chapter 8 *recommends more accurate incorporation of surface albedo and snow microphysics in climate models.*

There remains considerable controversy about the degree

Current warming

Cryospheric component

Forcing mechanism	Carbon dioxide (CO_2)	Methane (CH_4)	Nitrous oxide (N_2O)	Water vapour
Sea Ice	No conclusion on interaction between biological carbon drawdown and physical processes	Less sea ice → ocean warming → greater methane flux to atmosphere which will increase warming over time	Potentially important but currently unknown effects	Less sea ice → greater evaporative surface and greater area for ocean warming which will increase over time
Greenland Ice Sheet	No major effect known	No major effect known	No major effect known	Increasing area of surface melt → increasing evaporation → local cooling but general warming over time
Snow	Decreased snow duration + earlier onset of spring melt → increased primary production (carbon drawdown) → increased cooling over time	Decreased snow-water equivalent → decreased flooding → decreased methane flux → reduced warming	No major effect known	Decreased snow-water equivalent → reduced evaporation → decreased warming
Permafrost	Thawing permafrost → short term damage to vegetation but longer term increased carbon drawdown → increased cooling	Subsea permafrost thaw → increased release of methane → increased warming	Potentially important positive feedback but currently unknown trends	Thawing permafrost → more soil moisture + new lakes + drying lakes → increased evaporation: net balance unknown
Mountain Glaciers and Ice Caps	No major effect known	No major effect known	No major effect known	Increasing area of surface melt → increasing evaporation → local cooling but general warming over time
Lake and River Ice	No major effect known	Decreased ice duration → water column warming → increased methane production and release → warming	No major effect known	Reduced ice duration → increased water temperature + increased wind action → increased evaporation → increased warming

Future warming

Legend:
- Substantial warming effect, increasing over time
- Slight warming effect, increasing/decreasing over time
- Substantially reduced rate of warming (or even cooling), increasing over time
- Slightly reduced rate of warming (or even cooling), increasing/decreasing over time

Figure 11.2. The major feedbacks on climate via the cryosphere. The box colors indicate the expected future impact of a change in the variable or process on the cryosphere (where red indicates warming and blue indicates reduced rates of warming or even cooling). The box size and intensity of the colors indicates, at least qualitatively, the relative magnitude of the impact or levels of understanding (where grey represents uncertainties due to conflicting information etc. and white indicates a knowledge gap). Arrows indicate whether the (warming or cooling) effect is increasing or decreasing over time,

Current warming

Surface albedo	Cloud	Atmospheric circulation	Thermohaline circulation
Less sea ice → lower albedo + less surface for snow accumulation → lower albedo. Increasing warming over time	Less sea ice → more evaporation, heat, and DMS release → less stable boundary layer → more mid-level cloud → increased cooling overall	Less sea ice → more heat to atmosphere → less stable boundary layer + greater tropospheric thickness → change in circulation. Responses vary by region	Less sea ice → less salt rejection → decrease of bottom water formation → less heat from tropics goes to the Arctic → more sea ice
Less ice + more surface water + less snow → lower albedo. Increasing warming over time	Less land ice → more heat and particulate release → less stable boundary layer → more mid-level cloud → increased cooling overall	Removal of GRIS → weakening of Icelandic Low + increased equatorward flow → decreased precipitation + decreased temperature in the Barents Sea	No major effect known
Decreased snow duration and extent → lower albedo. Increasing warming over time	Less snow cover on land → more heat and particulate release → less stable boundary layer → more mid-level cloud → increased cooling overall	Increased snow depth over western Russia + decreased snow depth over Siberia → decreased monsoon precipitation → increased SST over central Pacific	No major effect known
Thawing permafrost → shift in vegetation to shrubs and trees → lower albedo → increased warming	No major effect known	No major effect known	No major effect known
Less ice + more surface water + less snow → lower albedo. Increasing warming over time	Less land ice → more heat and particulate release → less stable boundary layer → more mid-level cloud → increased cooling overall	No major effect known	No major effect known
Less ice + more surface water + less snow → lower albedo. Increasing warming over time	Less freshwater ice → more evaporation and heat → less stable boundary layer → more mid-level cloud → increased cooling overall	No major effect known	No major effect known

Cryospheric component

Future warming

☐ Uncertainty due to conflicting information ☐ Knowledge gap

that is, the acceleration or retardation of the process assuming a time horizon over ~30 yrs. The feedbacks presented in this figure represent a relatively simple synthesis of many complex processes. Possible feedbacks due to black carbon are not included in this diagram as they are not addressed in the SWIPA assessment; for information on black carbon impacts on Arctic climate see AMAP (2011b).

to which current levels of freshwater within the Arctic Ocean can affect the strength of the AMOC. *There is a need to evaluate the potential for the cumulative production and release of large amounts of freshwater from all contributing components, including all components of the cryosphere. Also there is a need to assess whether there are specific circulation patterns and durations that could cause significantly large freshwater production from all terrestrial and marine freshwater sources and their ultimate export to the North Atlantic where they could produce a significant effect on the AMOC and global climate.*

The primary source of uncertainty regarding the atmosphere as a driver of change in the Arctic is precipitation. Future changes in its spatial and seasonal distribution are unclear, as are its effects on the sea-ice mass budget, marine primary productivity, and vegetation. *A better understanding of precipitation effects will only be gained through developing and sustaining a more robust observing network.*

11.2. Sea-level change

11.2.1. Introduction

This discussion of sea-level change (see also Box 11.1) is framed in terms of the impact on global mean sea level of net loss of mass from Arctic mountain glaciers and ice caps, and the Greenland Ice Sheet. Reconstructions of sea-level change using geological evidence show that a sea-level rise of one or more metres per century was not uncommon in the past. It has recently been suggested that such rates occurred during the last interglacial warm period 120 000 years ago when the volume of land ice was similar to that at present

(Berger, 2008; Rohling et al., 2008). Worldwide, the coastal zone changed profoundly during the 20th century, primarily due to a growing population and increasing urbanization. In 1990, 23% of the world's population (or 1.2 billion people) lived within 100 km distance and 100 m elevation of the coast at densities about three times higher than the global average (National Research Council, 2011). Around 200 million people live on coastal floodplains, less than 1 m above current sea level. Sea-level rise is thus one of the major global socio-economic hazards associated with climate change. With coastal development continuing at a rapid pace, society is becoming increasingly vulnerable to sea-level rise and variability. Rising sea levels will contribute to more severe storm surges and flooding, even if hurricane intensities do not increase in response to the warming of the oceans. They will also contribute to the erosion of the world's sandy beaches, most of which have been retreating over the past century. Low-lying islands are also vulnerable to sea-level rise.

11.2.2. Instrumental sea-level record

Since 1870, global sea level has risen by about 0.2 m (Bindoff et al., 2007). During the 20th century, the rate of rise in global mean sea level was in the range 1.0 to 2.0 mm/y (Church et al., 2001), with recent analyses for the past 50 years giving values in the upper part of this range (Holgate and Woodworth, 2004; Church and White, 2006). During this period, the ice sheets of Antarctica and Greenland made a relatively small contribution to sea-level change (a few tenths of a millimetre per year at most; Rignot and Thomas, 2002), so the changes observed were due primarily to the combination of steric changes and global mountain glacier / ice cap melt. Since 1993, sea level has been measured accurately globally using altimeters on satellite platforms. Before then, the data were obtained from tide gauges at coastal stations around the world. Satellite and tide-gauge measurements both show that the rate of sea-level rise has accelerated since 1990 (Figure 11.3). Satellite measurements show that sea level rose at an average rate of 3.3 ± 0.6 mm/y from 1993 to 2008 (Ablain et al., 2009). This decreased to

Box 11.1. Sea-level terminology

Mean sea level (MSL) is defined as the average height of the ocean surface (i.e., the global average of the elevations of the mid-points between local mean high and low tide levels). The term eustatic refers to changes in global mean sea level due to changes in the mass of water in the oceans, but also through tectonic processes which change the size or shape of the ocean basins. The term steric refers to changes in global mean sea level due to changes in water density arising from thermal expansion or compression and/or salinity variations. The local mean sea level (LMSL) refers to the height of the sea with respect to a local benchmark on land, and is determined by averaging the height difference over a period of time long enough to smooth out fluctuations caused by waves and tides. In some regions, vertical movements of the land can be of the same order (mm/y) as sea-level changes, and measurements of local mean sea level must be corrected for this effect. Such land movements can occur because of ongoing isostatic adjustment of the mantle to temporal variations in surface loading resulting from changes in glacier or ice sheet mass. Isostatic adjustments are greatest near the sources of glacier or ice sheet change (where they are referred to as glacio-isostatic adjustments), but combine with other load signals to form a complex global pattern of crustal uplift and depression (National Research Council, 2011).
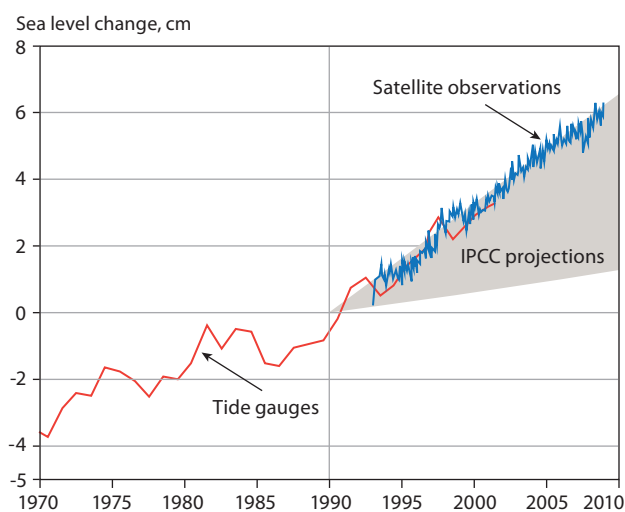

Figure 11.3. Sea-level change during the period 1970 to 2008. Tide gauge data are shown in red (Church and White, 2006) and satellite altimeter data in blue (Cazenave and Llovel, 2010). The grey band shows the IPCC projections reported in its Third Assessment Report (Church et al., 2001) for comparison.
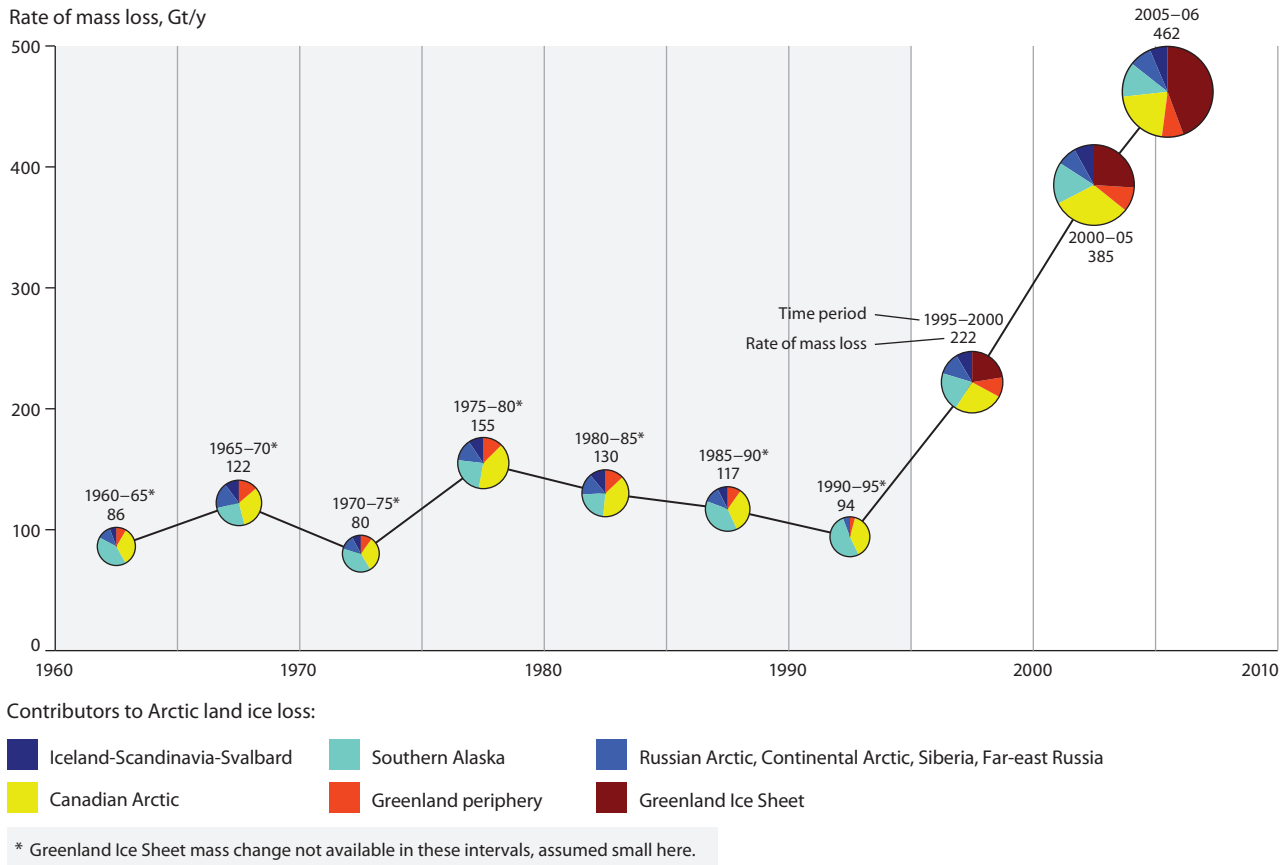
Figure 11.4. Estimated mass losses in all Arctic land ice (mountain glaciers and ice caps, and the Greenland Ice Sheet) from 1960 to 2006. The size of the pie chart indicates the total mass loss from all Arctic land ice, while the respective slices represent the relative contributions from different regions. Except for the most recent estimate, which is only a 1-year net mass loss, estimates are for periods of five mass balance years. Averaging periods and mass loss in Gt/y are given adjacent to each pie chart. 'Greenland periphery' refers to glaciers and ice caps on Greenland that are separate from the main ice sheet. Time-averaged estimates of multi-year geodetic measurements are assigned to all pentads to which each individual measurement refers. Non-ice sheet data are derived from a subset of a global, gridded extrapolation of glaciological and geodetic mass balance measurements (Cogley, 2009). These data are plotted by region in Chapter 7, Figure 7.30. All estimates for periods prior to 1995 do not include the contribution from the Greenland Ice Sheet owing to a lack of reliable data. Estimates for the Greenland Ice Sheet after 1995 are compiled from Chapter 8, Figure 8.19 using the values 50 Gt/y for 1995–2000, 100 Gt/y for 2000–2005, and 205 Gt/y for 2005–2006. Although there are several satellite gravimetric measurements of the mass of land ice on Greenland for the period since 2002 they are excluded from this analysis because the Gravity Recovery and Climate Experiment (GRACE) measures changes in the mass of both the ice sheet and the mountain glaciers and ice caps beyond the ice sheet margin, and changes in these peripheral glaciers are already estimated in the Cogley (2009) pentadal dataset.

2.5 ± 0.6 mm/y for the period 2003 to 2008 (Ablain et al., 2009; Cazenave and Llovel, 2010). Statistical analysis reveals that the rate of rise is correlated with global mean surface temperature (Rahmstorf, 2007; Grinsted et al., 2009). Sea-level rise is a predictable consequence of climate warming for two main reasons: ocean water expands as it heats up, and additional water flows into the oceans from the ice that melts on land.

11.2.3. The role of Arctic mountain glaciers and ice caps, and the Greenland Ice Sheet

The chapters addressing Arctic mountain glaciers and ice caps (Chapter 7) and the Greenland Ice Sheet (Chapter 8) document present knowledge of historical changes in the mass of these ice bodies. Net mass loss from Arctic mountain glaciers and ice caps has increased and corresponds to a rise in mean sea level of 0.5 mm/y in the most recent decade (ca. 1996 to 2006, Dyurgerov et al., 2009) and perhaps even to 0.75 mm/y in the period 2000 to 2005 (Cogley, 2009). The volume of all the world's mountain glaciers and ice caps (all land ice except the Greenland and

Antarctic ice sheets) is estimated to be equivalent to a change in mean sea level of 0.60 ± 0.07 m; the Arctic mountain glaciers and ice caps contain 68% of the total volume of these glaciers and ice caps which is equivalent to 0.41 m of mean sea-level rise (Radić and Hock, 2010). The rate of mass loss from the Greenland Ice Sheet has been increasing rapidly, and its contribution to mean sea-level rise has progressively increased from about 0.14 mm/y (1995–2000), to 0.28 mm/y (2000–2005), to 0.57 mm/y (2005–2006) (averages from Chapter 8, Figure 8.19, assuming that a mass loss rate of 360 Gt/y is equivalent to a rate of rise in global mean sea level of 1mm/y). The most recent estimate of the total mass loss from the Arctic mountain glaciers and ice caps, and the Greenland Ice Sheet (for 2003–2008) thus sums to 1.3 mm/y mean sea-level rise (462 Gt/y) and hence contributes more than half to the most recent estimate of the global rate of mean sea-level rise (2.5 mm/y; Ablain et al., 2009). Both the proportional contribution from the Greenland Ice Sheet and the total contribution from Arctic land ice (i.e., the Greenland Ice Sheet, and the Arctic mountain glaciers and ice caps) to mean sea-level rise have increased strongly since 1995 (Figure 11.4). The uncertainties on the sea-level rise estimates for the

Greenland Ice Sheet and the Arctic mountain glaciers and ice caps vary within the literature, with values of 0.1 to 0.2 mm/y on each of the components resulting in an uncertainty in the sum of the estimated sea-level rise from the two sources combined of 0.23 mm/y (Ablain et al., 2009 Cazenave and Llovel, 2010).

11.2.4. **Additional sources contributing to sea-level rise**

In order to assess the causes of observed sea-level changes and predicted future sea-level rises, other influences on sea-level rise need to be taken into account. These include thermal expansion of the ocean water, mass losses from the Antarctic Ice Sheet and non-Arctic mountain glaciers and ice caps, and change in land water storage. The loss of mass of the global glaciers and ice caps has increased from 0.8 ± 0.1 mm/y (1993–2003) to 1.4 ± 0.25 mm/y (2003–2007) (Cazenave et al., 2008; Cazenave and Llovel, 2010). The Antarctic Ice Sheet is the largest body of ice in the world, and uncertainties in glacial isostatic adjustments in this area are large, making it difficult to obtain accurate estimates of ice sheet mass loss. The most reliable estimates have been determined from satellite observations made during the past decade. The loss of mass from the Antarctic Ice Sheet is observed to be equivalent to 0.2 ± 0.17 mm/y mean sea-level rise in 1993–2003, increasing to 0.5 ± 0.15 mm/y in 2003–2007 (Cazenave et al., 2008; Cazenave and Llovel, 2010). Thermal expansion of the ocean water was the greatest component of observed mean sea-level rise with a value of 1.0 ± 0.3 mm/y in 1993–2003 (Cazenave and Llovel, 2010). In the most recent period the contribution from thermal expansion has decreased to 0.25 ± 0.8 mm/y (Cazenave and Llovel, 2010). The final component of the mean sea-level rise is the change in storage of water on land due to climatic change and human activities. Estimates based on GRACE satellite observations suggest a negative change in terrestrial water storage in the period 2003–2007, contributing -0.2 ± 0.1 mm/y to mean sea-level (Cazenave and Llovel, 2010). A compilation of historical contributions from thermal expansion, the different land ice components, and change in terrestrial water storage is shown in Figure 11.5 and Table 11.1.
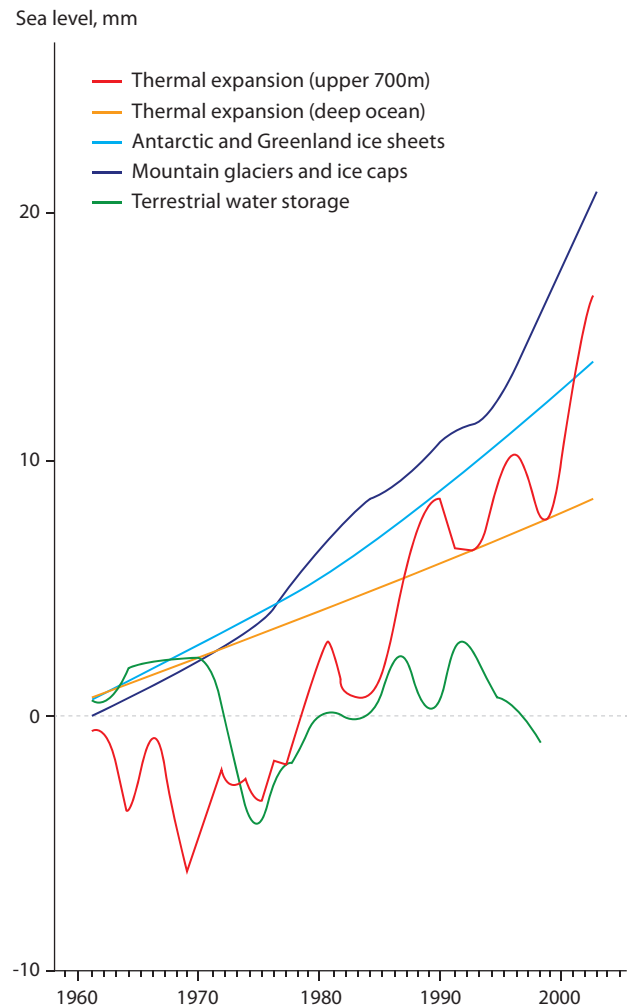
Figure 11.5. Change in total observed mean sea-level rise and its components over the period 1960 to 2003. The components related to the steric sea-level rise are thermal expansion in the upper 700 m and thermal expansion in the deep ocean. The mean sea-level rise from the ice sheets of Antarctica and Greenland, mountain glaciers and ice caps and terrestrial water storage are also shown. Source: based on Domingues et al. (2008).

Table 11.1. Contributions to mean sea-level rise for the periods 1993–2007 and 2003–2007. Quoted errors are one standard deviation. Source: Cazenave and Llovel (2010).

Sea-level rise, mm/y	1993–2007	2003–2007
Observed	3.3 ± 0.4[a]	2.5 ± 0.4[a,b]
Thermal expansion	1.0 ± 0.3[c]	0.25 ± 0.8[d] (Argo)
Ocean mass	2.3 ± 0.5[e]	2.1 ± 0.1[f]
Mountain glaciers and ice caps	1.1 ± 0.25[g]	1.4 ± 0.25[h]
Total ice sheets (Greenland and Antarctic)[j]	0.7 ± 0.2 (0.4 ± 0.15) (0.3 ± 0.15)	1.0 ± 0.2 (0.5 ± 0.15) (0.5 ± 0.15)
Land waters	–	-0.2 ± 0.1[k]
Sum of (2 + 4 + 5 + 6)	2.85 ± 0.35	2.45 ± 0.85
Observed rate minus sum	0.45	-0.05

[a] The observed sea-level rise is GIA corrected (-0.3 mm/y removed); [b] Ablain et al. (2009); [c] mean of Levitus et al. (2009) and Ishii and Kimoto (2009) values; [d] mean of Willis et al. (2008), Cazenave et al. (2009) and Leuliette and Miller (2009) values; [e] observed rate minus thermal expansion; [f] GRACE with a -2 mm/y GIA correction, Cazenave et al. (2009); [g] based on Kaser et al. (2006) and Meier et al. (2007); [h] Cogley (2009); [j] compilation of published results; [k] W. Llovel, K. DoMinh, A. Cazenave, J.F. Cretaux, M. Becker, unpublished manuscript.

11.2.5. **Projections into the future**

Radić and Hock (2011) computed estimates of the mass changes of more than 120 000 mountain glaciers and around 2600 ice caps around the world to 2100 using an elevation-dependent temperature-index mass balance model driven by output from ten GCMs forced by the IPCC A1B emissions scenario. Future volume changes were up-scaled to all glaciers using a regionally differentiated approach. For the Arctic mountain glaciers and ice caps (including those in Greenland surrounding the ice sheet) the projected volume losses due to melt (mass loss by calving is not included) show a large range (51–136 mm mean SLE or a 13–36% reduction in their current volume by 2100) depending on the choice of GCM (Chapter 7, Figure 7.23). The modeled volume reduction by 2100 varies considerably among regions, with the smallest reductions in Greenland (8 ± 4 %) and the largest in Svalbard (54 ± 15 %), with a large spread among models for most regions (Chapter 7, Figure 7.23). Most models show the greatest Arctic sea-level contributions coming from the Canadian Arctic, Alaska (including the glaciers of northwestern Canada), followed by Svalbard and the Russian Arctic. Contributions from Arctic Canada show a very large range; from a mean SLE contribution of <10 mm to the largest projected contribution for any of seven Arctic regions, indicating a large spread in the projected temperature and precipitation fields among the GCMs for this region (Chapter 7, Figure 7.24).

The above projections do not include mass losses due to calving that result from changes in the dynamics of tidewater glaciers. In the polar regions, there is evidence of accelerating flow and increased calving of tidewater glaciers, especially in Greenland and in western Antarctica, but it is not yet possible to predict reliably whether this will become more widespread, or for how long it may be sustained.

Since the IPCC Fourth Assessment, a number of new studies of ice-sheet mass budget have considerably enhanced understanding of the potential for rapid collapse of ice sheets in response to changing climate (Allison et al., 2009). Recent observations have shown that changes in the rate of ice discharge into the ocean can occur far more rapidly than previously suspected (e.g., Rignot, 2006). Currently there is no dynamic ice-sheet model that can predict the response of the Greenland Ice Sheet to a warmer climate on a century time scale. To get around this difficulty, Pfeffer et al. (2008) attempted to assess the rates of sea-level rise that could occur under a range of physically plausible scenarios. Under their 'Low 1' scenario, the sea-level rise contribution from Greenland is estimated by assuming a doubling in ice discharge, and a continued increase in losses by surface melt from both the ice sheet and mountain glaciers and ice caps at present-day rates. In this case, the total contribution to sea-level rise would be about 0.17 m by 2100, with contributions of 0.09 m from ice discharge, and 0.07 m from surface melt, respectively.

In western Antarctica, the Amundsen Coast basin (which includes Pine Island and Thwaites Glaciers), has shown increased mass loss in recent years. These marine-based glaciers contain about 1.5 m SLE of ice, and are particularly sensitive to changes in climate and in ocean circulation and temperature. The average ice velocity in this region is 2 km/y, which is considerably higher than the average velocity for all Antarctic ice streams (0.65 km/y) (Pfeffer et al., 2008). If ice discharge into the oceans from this region continues to increase, the sea-level rise contribution from western Antarctica will be significant. Pfeffer et al. (2008) estimated the potential sea-level rise contribution from Antarctica by 2100 by (i) doubling the velocities of Pine Island and Thwaites Glaciers within the first decade, (ii) accelerating the rate of ice loss by surface mass balance from the Antarctic Peninsula at the present-day rate of surface mass balance change, and (iii) scaling dynamic losses from the Antarctic Peninsula to losses by surface mass balance with a ratio of 1.31 (as observed in Greenland). The resulting projected sea-level rise is 0.15 m by 2100 (this case is also referred to as the 'Low 1' scenario).

Under the 'Low 1' scenario, the Greenland and Antarctic ice sheets would together contribute up to 0.31 m to mean sea-level by 2100 (Pfeffer et al., 2008). Under a more extreme scenario, 'High 1', that envisages more dramatic increases in dynamic losses, contributions to sea-level rise by 2100 of 0.54 m from Greenland (0.47 m from ice dynamics alone) and 0.62 m from Antarctica (all but 0.01 m from ice dynamics) are projected (Pfeffer et al., 2008), a total of nearly 1.2 m.

According to Radić and Hock (2011), the projected multi-model mean sea-level rise contribution from all mountain glaciers and ice caps on Earth is 0.12 ± 0.04 m by 2100. This is larger than an estimated contribution from surface mass balance of 0.08 m (Pfeffer et al., 2008). In the Radić and Hock study, mass loss by calving is neglected. Pfeffer et al. (2008) suggested that ice discharge from fast-flowing tidewater glaciers and ice cap outlets could add between 0.09 m (Scenario 'Low 1') and 0.47 m (Scenario 'High 1') to global mean sea level by 2100. Adding in the steric contribution to projected sea-level rise estimated by the IPCC AR4 Working Group 1 (Meehl et al., 2007) of 0.30 ± 0.15 m by 2100, the sum of land ice contributions to global mean sea level by 2100 would range between 0.79 m (Scenario 'Low 1') and 2.01 m (Scenario 'High 1') (Pfeffer et al., 2008). It should be emphasized, however, that these estimates are not the result of rigorous projections, but of scenarios designed to bracket the potential range of sea-level rise that could occur as a result of changes in ice dynamics.

As an alternative method for projecting the combined contributions of surface mass balance and ice dynamics to 21st century sea-level change, Rahmstorf (2007) and Grinsted et al., (2009) made semi-empirical projections by linking past sea-level changes to observed surface air temperatures (see Figure 11.6). Their statistical projections of sea-level rise for the range of temperature projections associated with the IPCC A1B emissions scenario (2.3 to 4.3 °C temperature increase for 2100) range from 0.97 to 1.56 m above 1990 sea level by 2100. This is consistent with what happened during warming induced by variations in Earth's orbit during the last interglacial period (Overpeck et al., 2006).

Total projected sea-level rise resulting from all sources by 2100 cannot be estimated with high confidence at present. Lower and upper limits have a range of 0.79 to 2.01 m. A range of 0.9 to 1.6 m is considered the more plausible current estimate by the authors of the SWIPA assessment.

Finally, it should be emphasized that sea-level changes are far from equal from region to region as changes in the Earth's rotation and gravity field, ice load, ocean temperatures, ocean circulation, and freshwater supplies also impact the local mean sea level (Cazenave et al., 2008; Cazenave and Llovel, 2010). Between
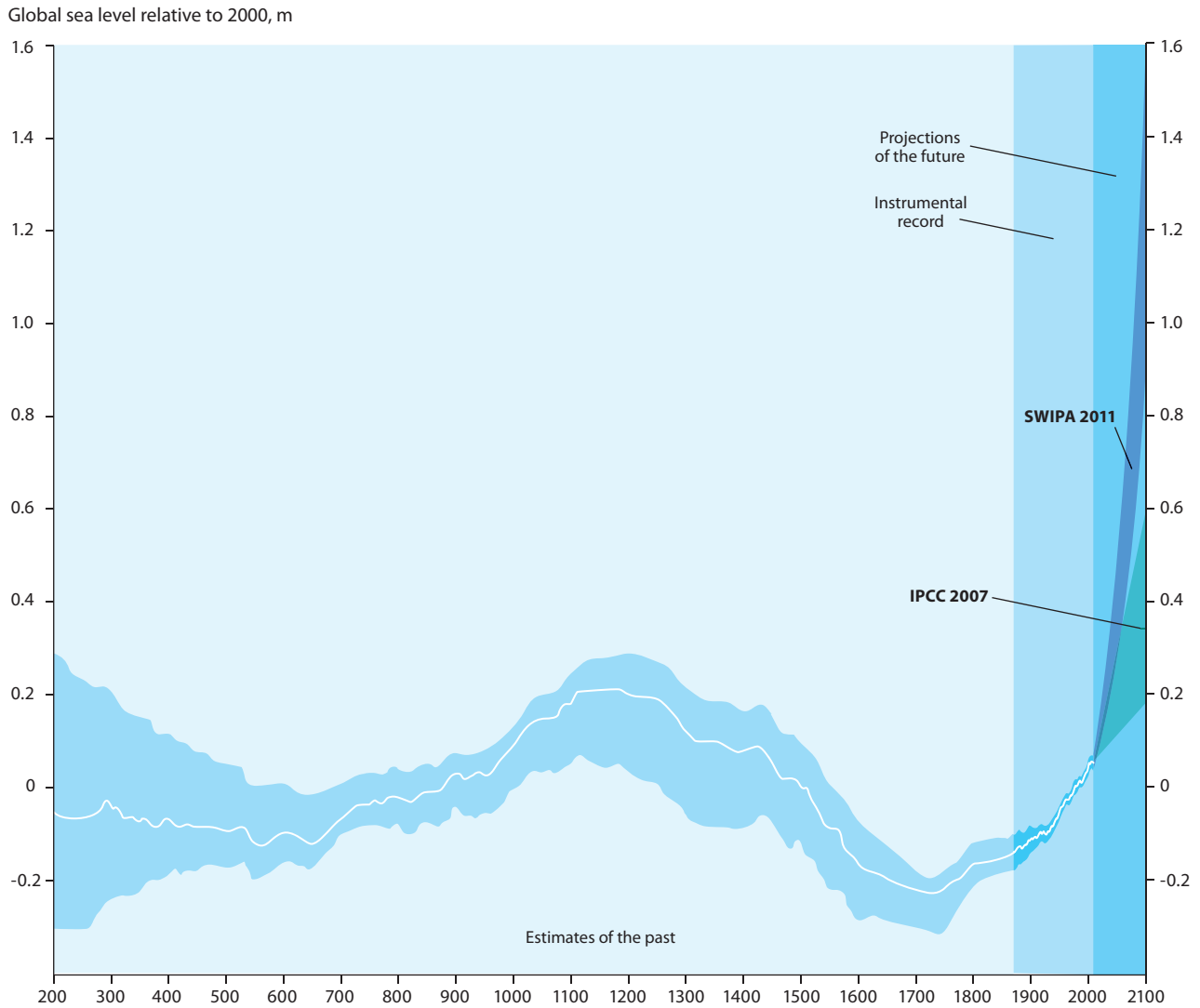
Global sea level relative to 2000, m

Figure 11.6. Semi-empirical estimate of sea-level rise by 2100 (Grinsted et al., 2009), based on projections of future surface temperature change made using climate models forced by the IPCC A1B emissions scenario.

1993 and 2007, the trend in observed sea level varied spatially from -5 mm/y to 15 mm/y. Taking into account the influences of changes in the Earth's rotation, the gravity field and related effects for the period 2000 to 2008, Bamber and Riva (2010) suggested that mass loss from Greenland resulted in local sea-level rise around South America, and local sea-level fall in much of eastern Canada and northwestern Europe. Trends in local mean sea level will also depend on location relative to regions where ice mass loss is occurring due to the isostatic response (Milne et al., 2009). Overall, there is a tendency toward falling local sea level in the vicinity of regions where large decreases in ice mass are occurring.

11.2.6. Impacts of sea-level rise

The impacts of sea-level rise will be felt through an increase in mean sea level as well as through more frequent extreme sea-level events, such as storm surges. Impacts include increased severity and frequency of flooding in low-lying areas, salt water intrusion into surface water bodies and freshwater aquifers, beach erosion, and damage to infrastructure, agriculture, and the environment, including wetlands, intertidal zones, and mangroves, with significant impacts on biodiversity and ecosystem function. Hundreds of millions of people live in low-

lying coastal areas, including heavily populated deltas such as Ganges-Brahmaputra (Bangladesh), the Rhine (Netherlands), and the Mekong (Vietnam) (Nicholls and Cazenave, 2010).

Sea-level rise severely threatens the existence of a number of small island states. They are in danger of becoming completely submerged or facing increased coastal flooding. Several of these states, such as the Maldives in the Indian Ocean, the Bahamas in the Caribbean, and Tuvalu in the Pacific, are topographically flat and at or near sea level. Others are volcanic with only narrow coastal strips that are habitable. Sea-level rise (especially if coupled with storm surges from stronger and longer lasting tropical storms) would have a great impact on the socio-economic condition of small islands and small island states. Adaptation will be challenging in many island settings, and in some places may not be feasible at all.

Changes in ocean circulation due to ocean temperature rise and freshening of sea water may affect regional sea levels, particularly in the northeastern United States, eastern China, and Japan, where sea level is projected to rise more than the global average. This would place large coastal cities, such as Shanghai (China) and New York City (USA) at greater risk of permanent inundation and coastal flooding.

11.3. Contaminant pathways and change in the cryosphere

Summary

Loss of continuous sea-ice cover in Fram Strait has ultimately led to increased re-evaporation of previously trapped hexachlorobenzene (HCB) from open surface waters along the west coast of Spitsbergen (Svalbard) during the winter season. This increased evaporation of HCB is evident in atmospheric monitoring data from the Zeppelin research station on Svalbard. The general loss of sea-ice cover over extensive areas is increasing the potential for widespread re-evaporation of some legacy contaminants and enhanced deposition of others, such as polychlorinated biphenyls (PCBs).

Thawing permafrost is leading to the remobilization of previously stored anthropogenic contaminants into rivers and lakes, and sometimes even into groundwater. Melting of snow fields and glaciers is leading to the release of contaminants deposited over previous decades.

Change in sea-ice cover, hydrology, and organic carbon cycling will result in the redistribution/remobilization of contaminant loads. Such changes in the cryosphere will be accompanied by shifts in ecosystem function and species ranges, which will in turn alter the transfer and biomagnification of contaminants through food webs and, thus, also the exposure of higher trophic levels, including humans, to persistent toxic chemicals.

More extensive areas of open water and longer duration of open water owing to the loss of sea-ice cover will increase a range of human activities in the Arctic (transportation, exploitation of resources, industry, tourism), all of which will increase the risk of additional contaminants entering the Arctic environment from local sources.

11.3.1. Introduction

Contaminant concentrations and transport pathways in the Arctic environment have been monitored for more than two decades within the framework of the Arctic Monitoring and Assessment Programme (AMAP). The extensive AMAP contaminant monitoring program provides a unique opportunity for evaluating the impact of cryospheric change on contaminant fate and distribution within the Arctic environment, by using the contaminant patterns and levels reported as a sentinel for early environmental change in the Arctic cryosphere.

The Arctic was long considered to be pristine, little influenced by anthropogenic contaminants. However, research over the past two decades, strongly motivated by AMAP, has demonstrated the ability of persistent, semi-volatile compounds to travel long distances, with prevailing winds and ocean currents providing effective means to transport contaminants to the Arctic (AMAP, 1998).

This review focuses on the priority contaminants defined by AMAP, which include persistent organic pollutants (POPs) such as PCBs, hexachlorocyclohexane (HCH) isomers, HCB, DDT (dichlorodiphenyltrichloroethane) derivatives, as well as certain metals (e.g., mercury, lead) and radionuclides. These contaminants have been extensively studied for their human and environmental health implications (AMAP, 2009).

Within the Arctic, a number of concentrating processes (e.g., Macdonald et al., 2002) lead to high exposure to classical POPs, especially for marine top predators. The exposures have consequences for the health and wellbeing of those Arctic residents that exploit marine mammals as traditional food resources (Van Oostdam et al., 1999; AMAP 2011a).

Early discussions of contaminant pathways considered the Arctic as a static system of boxes (environmental compartments or 'reservoirs' that contain contaminants, such as the atmosphere, ocean and soil) and arrows (fluxes), where each contaminant could be viewed as either moving between the fixed boxes or stored within them (AMAP, 1998; Macdonald et al., 2000a). Since the early 1990s it has become clear that the Arctic is changing (ACIA, 2005). The present assessment has assembled an impressive body of evidence to show that cryospheric change is pervasive and that the Arctic continues to evolve rapidly (e.g., Stroeve et al., 2007). Given these observations, it is apparent that the size of the boxes, the connections between the boxes, and the directions and sizes of the arrows between them are all subject to change within the cryosphere, and within upstream components in temperate and tropical regions. Surprises in global geochemical systems come from feedbacks (Shindell, 2007). With respect to contaminants and their fate in the Arctic environment, there are three important feedback systems: the hydrological cycle, the cryospheric cycle, and the organic carbon cycle. The Arctic features all three feedbacks and therefore offers a wide range of possible changes to contaminant cycles.

It has long been known that climate warming can affect contaminant pathways simply through inherent physical and chemical properties (e.g., vapor pressure and Henry's Law constants, McKone et al., 1996). With an increase in average global temperature, the time a contaminant spends in the air will increase, and seasonal thermal forcing will have a greater effect on contaminants that undergo multiple 'hops' (i.e., repeated re-mobilization from secondary sources such as ocean surfaces and soils); such contaminants include DDT, PCBs, and mercury (Hg). As the hydrological cycle becomes more active (Stocker and Raible, 2005), contaminant deposition and re-emission will also become more active. On the other hand, the process of global distillation, which forces contaminants out of warm reservoirs in temperate industrial areas and into cold reservoirs at the poles, is likely to decrease because polar amplification will tend to reduce the thermal contrast between the two reservoirs (e.g., Screen and Simmonds, 2010). As highlighted by Macdonald et al. (2003, 2005), thermal forcing is but a small component of the contaminant cycle. Global change in general, and cryospheric change in particular, have a far greater potential to alter contaminant pathways through alteration of water balance (e.g., Smol and Douglas, 2007a), through phase change in water itself (e.g., Kwok et al., 2009), and through consequent change in organic systems (e.g., McGuire et al., 2009). Recent work has begun to reveal the important connections between global change and contaminant pathways in the cryosphere (Outridge et al., 2005; Stern et al., 2005; Garrett and Zhao, 2006; Schindler and Smol, 2006; Eckhardt et al., 2007; Schiedek et al., 2007; Becker et al., 2008, 2009; Gieisz et al., 2008; McKinney et al., 2009; van den Brink et al., 2009; Wang et al., 2010; Bustnes et al., 2010; Carrie et al., 2010).

Every step along the transport and redistribution pathways to and within the Arctic can be affected by climate change because reactivity, adsorption, air-sea exchange, and precipitation all depend on temperature. Some of these processes can be modeled thermodynamically (e.g., partitioning, vapor pressure), but it is clear that crossing the 0 °C isotherm has a unique capacity to alter systems (Macdonald et al., 2005). For example, whether precipitation occurs as rain or snow makes a large difference to contaminant exchange and deposition (Wania and Halsall, 2003; Daly and Wania, 2004; Lei and Wania, 2004; Herbert et al., 2006a,b) as does the presence of ice cover on ocean, lakes, and rivers. In addition to changing the connectivity between reservoirs, changes have already occurred in transport pathways within the atmosphere and ocean (McLaughlin et al., 2002; Macdonald et al., 2005; Shimada et al., 2006). Perhaps one of the greatest ways in which cryospheric change is manifested in contaminant pathways, is by altering the size of storage reservoirs. This is particularly important for classical contaminants that have been loaded heavily into the environment during the anthropocene (e.g., PCBs, DDT, Hg). Thus, the melting of permanent snow and glacial ice that have accumulated contaminant inventories during earlier eras (in many cases for more than 50 years) with high emissions will release these contaminants into the biosphere (Blais et al., 2001; Thies et al., 2007; Wang et al., 2010).

Priorities for Arctic environmental contaminant research with a focus on cryosphere-related processes should aim at an understanding of temporal and spatial distribution patterns including comprehensive source elucidation for selected compounds in ice-associated environments. The latter data will also be necessary to validate multi-compartmental models and identify environmental factors influencing chemical transport processes (e.g., atmospheric, oceanic, riverine and ice-borne transport).

This assessment takes as a point of departure the review by Macdonald et al. (2005) and provides an update on how cryospheric change alters the transport and fate of anthropogenic pollutants in the Arctic, including recommendations for future research priorities. Although climate change also alters the vulnerability of biota to contaminants (e.g., through multiple stressors), the focus here is on pathways. Change in the cryosphere also operates locally including, for example, the release of oil (drilling activities, natural oil seeps, oil sands) or other contaminants by thawing permafrost (increased active layer depth) and the risk of contamination or oil spills from increased shipping and other activities in the Arctic.

11.3.2. Impact of cryospheric change on contaminant pathways

11.3.2.1. Thawing permafrost and melting of glacial ice and snowpack

A decrease in the extent and thickness of permafrost due to warming in the Arctic is expected to release previously deposited contaminants into rivers, lakes and the ocean (Macdonald et al., 2005; Schindler and Smol, 2009; Rydberg et al., 2010). Emissions of naturally occurring potentially toxic substances have also been reported. Currently, permafrost and snow/ice cover prevent diffusion of naturally occurring Radon-222 (^{222}Rn) from rocks and soils into the atmosphere. With warming,

^{222}Rn evasion will increase, as will the deposition of lead-210 (^{210}Pb) to surfaces and the production of polonium-210 (^{210}Po), its granddaughter (Macdonald et al., 2005). Any increase in ^{222}Rn evasion due to warming/permafrost thawing would be accompanied by a widespread increase in the radioactive dose to northern residents.

11.3.2.1.1. Permafrost reduction

Permafrost thawing and the associated increase in seasonally occurring active layers above the permafrost will ultimately allow the release of previously stored anthropogenic contaminants through migration (Dyke, 2001), as well as an increase in microbial transformation. Contaminants previously immobilized in permafrost for many decades will eventually be reactivated and introduced into the Arctic food web. Klaminder et al. (2010) found that thawing permafrost promoted lead (Pb) migration out of soils and into two lakes. Lead was emitted from gasoline combustion processes and continuously transported into the Arctic via the prevailing winds over many decades (from the early 1900s to the 1980s), and is still considered a significant anthropogenic pollutant in the Arctic. The elevated levels of Pb together with higher ^{206}Pb/^{207}Pb ratios observed in sediments deposited after the 1970s showed that the more than 90% reduction in atmospheric deposition of Pb since the 1990s has not yet been realized within the lake due to an increased supply of Pb from drainage basin storage owing to thawing permafrost.

11.3.2.1.2. Changing characteristics of snow cover and glacier meltdown

Processes in the snowpack directly affect concentrations of gases and contaminants in the Arctic troposphere (Wania et al., 1999; Daly and Wania, 2004; Domine et al., 2007; Hansen et al., 2008; Meyer and Wania, 2008; Meyer et al., 2009a,b). The snowpack is a reservoir for semi-volatile organic contaminants, sequestered from the atmosphere during the winter and released either to the atmosphere or to aquatic systems during spring (e.g., Halsall, 2004; Domine et al., 2007). Due to colder temperatures, the High Arctic snowpack has a higher surface area index ($\sim$2500 m^2/m^2) than sub-Arctic regions ($\sim$1000 m^2/m^2), therefore providing a greater capacity to retain sorbed contaminants. Warming of the High Arctic will reduce the snowpack index, thus reducing the efficiency of air-to-surface transfer for more volatile pollutants (Taillandier et al., 2006; Domine et al., 2007). How this change in snowpack structure would be offset against predicted increases in snowfall for some parts of the Arctic is unclear. Nonetheless, surface wastage of permanent snowpack may lead to a considerable pulse of persistent contaminant concentrations into the ambient air (troposphere) or the water, depending on the exact circumstances of melting and the properties of the contaminant in question (see also Chapter 4, Section 4.4.3.4).

Recent studies conclude that ice on glacier surfaces contains considerable amounts of POPs, current-use pesticides and Hg (Donald et al., 1999; Hermanson et al., 2005; Maggi et al., 2006; Bogdal et al., 2009; Faïn et al., 2009; Chapter 7, Section 7.7.3.5). Where there are industrial sources nearby, higher loadings may occur for particular chemicals such as metals (e.g., Chizhova, 2005). Deposited contaminants, once buried beneath the firn, would normally follow the glacier ice, entering

meltwater centuries (from active glaciers) to millennia (from ice caps) after being deposited. Recent accelerated ablation of surface ice during summer is now releasing some of this archive (e.g., Blais et al., 2001; Gieisz et al., 2008) to become available for uptake into local food webs.

11.3.2.1.3. Changing ice cover on lakes and rivers

Reduction in ice cover on lakes may alter contaminant pathways in at least four ways, including: changing the interaction between inflow and lake stratification (Semkin et al., 2005); enhancing the production of sinking organic carbon leading to contaminant burial (Outridge et al., 2007; Carrie et al., 2010; Gantner et al., 2010); increasing photodegradation (e.g., Hammerschmidt and Fitzgerald, 2006); and enhancing air-water exchange (Macdonald et al., 2005). Furthermore, thawing of permafrost in the drainage basin together with a greater frequency of extreme events like floods will augment the input of soil, which may contain previously deposited contaminants (e.g., Klaminder et al., 2010). Fluxes of Hg and dissolved organic carbon, which tend to increase with river flow (Stanley et al., 2002), will also increase in northern rivers if flows become more variable, even if the average flow remains constant. Most of these changes lead to greater supply of contaminants to lakes and greater retention in deep water or sediments. For Hg (Semkin et al., 2005) and perfluorinated compounds (Veillette et al., 2009) it has been shown that over half of the contaminant load in inflowing water shunts through the lake to the outflow if the lake is ice-covered and if there is stratification (<4 °C water) beneath the ice. Loss of ice cover and breakdown of thermal stratification, projected to occur increasingly earlier in the year along with loss of ice, lead to greater mixing of inflowing water into the lake. Furthermore, greater turbidity (and density) of inflowing water, a result of more extreme freshets, may also favor entrainment into affected lakes. Other cryospheric changes, including the melting of permanent snow and ice, may produce temporary increases in loadings from contaminants (POPs, metals, radionuclides) deposited from earlier periods of high use, for some compounds more

than 60 years ago (Klaminder et al., 2008; Faïn et al., 2009; Rydberg et al., 2010).

11.3.2.2. Air-sea exchange

The global distribution of *p,p* -DDT (1,1,1-trichloro-2,2-bis(4-chlorophenyl)ethane) and its major transformation products, as projected by a coupled ocean-atmosphere model (Stemmler and Lammel, 2009), illustrates the potential for surface oceans to store DDT, transport it northward, and re-release it to the atmosphere. The delayed DDT pulse that arrives in northern waters is influenced by ice cover and water temperature, and these parameters ultimately determine whether the DDT will be volatilized and subsequently taken up by Arctic biota.

The efficacy of the air-sea exchange in Arctic water is governed by ambient temperatures (atmosphere and ocean) and ice coverage. The consequences of vastly reduced ice cover over much of the western Arctic Ocean during the past decade (Stroeve et al., 2007) will be increased evasion of (semi)-volatile contaminants presently trapped below the ice. HCHs are especially prone to this change (e.g., Li et al., 2004). High loadings that accumulated in surface waters of the Arctic Ocean during the 1970s and 1980s are now re-entering the atmosphere due to the sharp reductions in emissions and, consequently, reduced air concentrations (Bidleman et al., 2007). Because there is a relatively large inventory of HCH in the ocean, the removal of ice cover allows the oceanic HCH to enter the atmosphere rapidly through evasion. This was demonstrated near Resolute in the Canadian Archipelago by a shift in enantiomeric composition of atmospheric α-HCH from nearly racemic, implying a long-range atmospheric source of industrial HCH, to a microbially-degraded ratio matching that observed in the surface ocean (Figure 11.7). The conclusions drawn for α-HCH do not necessarily apply to other atmospherically transported POPs, especially those that have not achieved equilibrium with the water. For example, in the case of PCBs or endosulfan (which have not achieved equilibrium with the surface Arctic Ocean) the loss of ice would promote more rapid invasion into water (Macdonald
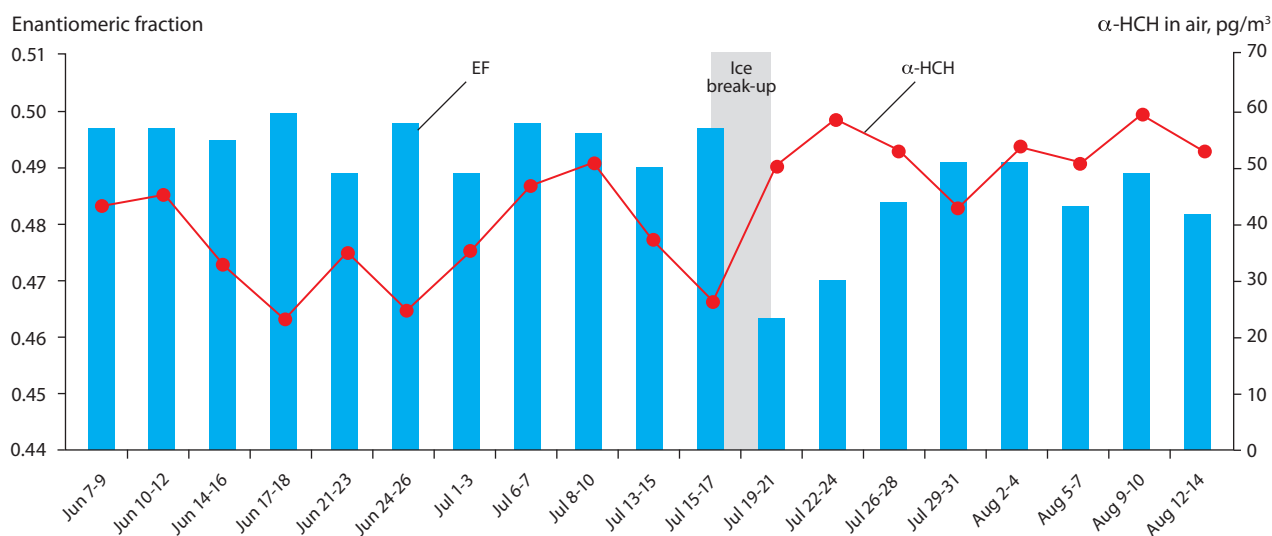

Figure 11.7. Average air concentrations and enantiomer fractions for α-HCH at Resolute Bay, Canadian Archipelago, before and after ice break-up (Jantunen et al., 2008).
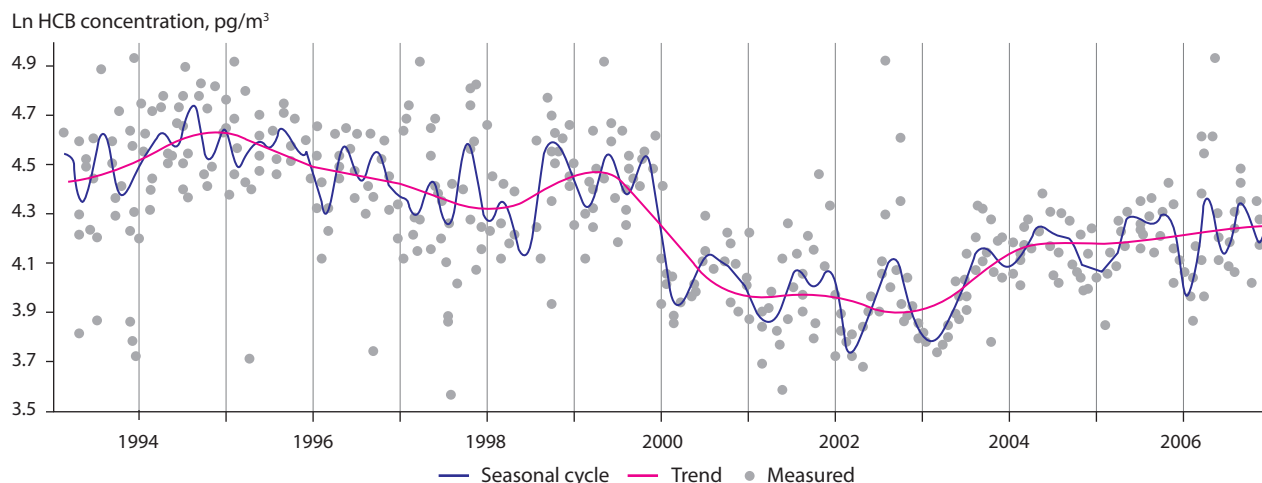
Ln HCB concentration, pg/m³

Figure 11.8. Temporal trend analysis for hexachlorobenzene (HCB) in air at the Zeppelin research station (Spitsbergen) using statistical digital filtration. Note: concentration values are given in a logarithmic scale. Source: Hung et al. (2010).

et al., 2000b; Weber et al., 2006; Gioia et al., 2008).

Within the Arctic, trends and concentrations of POPs in the atmosphere during the past ten years have varied regionally (Hung et al., 2010). For HCB, formerly used as a fungicide but now released to the environment predominantly as a by-product of industrial chemical processes, generally downward trends were reported at all Arctic stations. However, during the past four years, levels at the Zeppelin station (Spitsbergen) have been increasing (Figure 11.8).

Increasing air concentrations of HCB, observed only at the Zeppelin station, may partly be explained by increased evasion from the ocean along the western coast of Spitsbergen (Svalbard, Norway), which has been ice-free throughout winter between 2005 and 2009 (Slubowska-Wodengen et al., 2007; Nilsen et al., 2008). A permanently ice-free state at 80° N is unusual and although the evasion of HCB could be interpreted as a manifestation of regional climate change on the POPs distribution in the Arctic, dramatic loss of ice elsewhere in the Arctic Ocean (e.g., Stroeve et al., 2008) implies that exchange has been widely affected in the Arctic Ocean for most POPs. Alternatively, increased air concentrations of HCB may be due to the presence of HCB as an impurity in currently used pesticides in some European countries (Becker et al., 2009; Lohmann et al., 2009).

The Hg cycle in the Arctic Ocean may also be affected by reduced ice cover. The Hg concentration in the upper ocean is the result of a dynamic balance between Hg deposition and Hg evasion (Outridge et al., 2008). The entry of Hg into the ocean from the atmosphere requires conversion of Hg(0) to Hg(II), which can then deposit into water on particles, in contrast to the evasion of Hg, which occurs predominantly via gaseous exchange of Hg(0). One of the deposition processes for Hg – 'atmospheric mercury depletion events' – involves halogen chemistry in the atmosphere initiated by solar radiation at polar sunrise (Steffen et al., 2008). Here, the balance between first-year ice, snow and open ocean is critical to the deposition process. On the other hand, photoreduction of Hg(II) in the water produces an excess of Hg(0), which then supports net evasion, provided there is open water (Andersson et al., 2008). Because ice cover interacts with invasion and evasion of Hg differently, the loss of ice cover has the capacity to change the

quantity of total Hg held in the upper ocean, and the balance between forms of Hg.

11.3.2.3. Transport by sea ice

Transpolar ice-drift is known to re-distribute contaminants released from the large eastern Arctic rivers in Siberia (Arctic Russia) or contained in sediments over the Siberian shelves (AMAP, 1998). Recent change in the Arctic has increased the areas over which first-year ice is produced (e.g., Kwok et al., 2009) and the amount of open water exposed to storms in the autumn, and is therefore likely to have impacted the entrainment of sediment into ice by suspension freezing, by removing large areas of multi-year ice. Simulations of contaminant transport by ice using a 3D model (Pavlov, 2007) have investigated the potential for contaminants located in the vicinity of river mouths of major rivers flowing into the Arctic Ocean to be transported elsewhere during the 15 years following release (Figure 11.9).

Recent studies (Hakkinen et al., 2008; Pavlov and Pavlova, 2008) have shown an increase in sea-ice drift velocities in the Arctic Ocean over the past three decades, which leads to the intensification of surface currents in the drifting ice. Thus, a higher contamination risk from ice transport is implied, since water masses previously moving along the potential source regions of western industrialized countries are now expected to be reaching the Far North more rapidly as previously described due to the increased velocity of the inflow into the Arctic Ocean. Countering this argument, Karcher et al. (2010) suggested that with the reduction in ice cover, especially in summer, the importance of sea ice will decline as a transport medium for radionuclides from atmospheric fallout or in association with resuspended sediment. Overall, it is considered that the Arctic will tend to trap a greater proportion of its radionuclide inventory within the Arctic.

11.3.2.4. Contaminant change in marginal ice zones

The marginal ice zone (MIZ) comprises a circumpolar band of flaw leads north of the landfast ice. Here, stratification, supported by ice melting, and upwelling create the circumstances for primary production to occur from June to August. The organic

Figure 11.9. Distribution of pollutant concentrations in the surface layer of the Arctic Ocean 15 years after release from sources in the mouths (model assumption) of the major Arctic-flowing rivers: (a) Ob and Yenisey Rivers; (b) Lena River; (c) Kolyma River; and (d) Mackenzie River. The scale shows the percentage of pollutant in relation to its concentration at source. Arrows show the source location. Source: Pavlov (2007).

carbon so produced is presently considered to provide the basis for survival of almost all ice-associated organisms in the Arctic marine ecosystem (Rysgaard and Nielsen, 2006; Sundfjord et al., 2007; Hegseth and Sundfjord, 2008). The consequent lipid production in the lower pelagic ice-associated food web (Soreide et al., 2008; Falk-Petersen et al., 2009) provides a reservoir for lipophilic POPs (bioaccumulation), which then transfer into higher trophic levels (biomagnification). Bioaccumulation and biomagnification together put at risk Arctic top predators such as seabirds and marine mammals (Borgå et al., 2005; Sobek et al., 2006). Reliable productivity makes the MIZ an attractive location for foraging by predators such as seabirds, whales, and polar bears. A change in the location and timing of the MIZ due to climate change therefore has potentially large consequences for where, when, and how much lipid is produced, and whether that lipid supports epontic, pelagic, or benthic communities (e.g.,

Grebmeier et al., 2006). Furthermore, change in food production is often caused by – or invites – invasive species (Hegseth and Sundfjord, 2008; Vermeij and Roopnarine, 2008). Lipophilic contaminants will closely follow the changes in lipid production and transfer in the food web. Marine animals have adapted their reproduction to historic availability of high-quality energy. Trophic mismatch due to change in seasonality, or altered energy balance due to change in forage location or favored species leads to nutritional stress, which is compounded by contaminant release from stored fat.

Increases in surface water temperature and concentrations of particulate organic carbon associated with climate change are predicted to affect the bioaccumulation of hydrophobic organic contaminants in the Arctic marine food web (Borgå et al., 2010). A net reduction in bioaccumulation was predicted for two PCB compounds, driven in the main by their reduced bioavailability

associated with increases in particulate organic carbon in the water column.

It is important to note that change in ecosystems, and consequent change in contaminant pathways within organic systems, can be effected from the bottom up, as described for the MIZ, and from the top down (Macdonald et al., 2005). In the latter, the removal of ice as a habitat for polar bears or walrus would signal a change in food-web structure with consequences for biomagnifying contaminants.

11.3.2.5. Change in foraging behavior

The transport of contaminants within large individuals (whales) or large migratory populations (fish, birds), can represent a pathway of major importance to the biota themselves (Blais et al., 2007) or to their predators, including humans, while being of negligible importance in large-scale budgets or fluxes (Wania, 1998). Cryospheric change has great potential to alter migratory pathways, for example, through invasive species encouraged by climate change (Babaluk et al., 2000), which can import contaminants from outside the Arctic, or through change in feeding locations following altered prey distributions (Gaston et al., 2003), which then alters exposures within the Arctic (Michelutti et al., 2009).

Climate impact on foraging behavior is reported across all trophic levels in marine and terrestrial food webs (Couillard et al., 2008; Laidre et al., 2008; Loseto et al., 2008; Moore, 2008; Moore and Huntington, 2008; Thiemann et al., 2008; Cherry et al., 2009; Corcoran et al., 2009; Falk-Petersen et al., 2009; Sharma et al., 2009; Verreault et al., 2010). Indigenous peoples are potentially subject to some of the greatest changes in contaminant exposure because their diets shift between terrestrial and aquatic foods (Usher, 2002) in response to changes in the distribution of the food (e.g., through migration) or changes in terrain conditions that prevent travel to favored harvesting areas (Macdonald et al., 2005). McKinney et al. (2009) concluded that polar bears from the Canadian Hudson Bay area have a modified exposure to POPs owing to a change in diet over the period 1991 and 2007. The dietary change was directly attributed to an increase in the consumption of open water-associated seal species (relative to ice-associated seal species) during years of earlier sea-ice break-up. Likewise, in the Beaufort Sea, Gaden et al. (2009) found that Hg concentrations in ringed seals depend on ice cover, presumably through the intermediary effect of ice on prey availability. These newly emerging results confirm predictions by Macdonald et al. (2005) that changes in ice conditions cascade into food webs and foraging strategy and thence affect contaminant pathways.

11.3.2.6. Change in mercury methylation

The complexity of the Hg cycle, involving transformations between Hg(0), Hg(II) and alkylated forms of Hg (AMAP, 2011a), provides great potential for change either through physical components of the system (see Section 11.3.2.2) or chemical transformations. Methylmercury (MeHg, CH_3Hg) is the most important form of Hg in relation to toxicity and capacity to biomagnify (Macdonald and Loseto, 2010). The process of methylation is directly linked to the organic carbon cycle such that change in carbon pathways due to cryospheric change (e.g., organic carbon production, carbon flux in the ocean, remineralization) have consequences for MeHg production (Cossa et al., 2009; Sunderland et al., 2009). MeHg

is usually formed by natural methylation processes of Hg(II), as reviewed in several comprehensive studies (Hammerschmidt et al., 2006; Fitzgerald et al., 2007; AMAP, 2011a). Intensification of Hg methylation/demethylation induced by climate change is a crucial aspect of the Hg cycle in all environmental systems and, perhaps, especially the Arctic given the projected changes in hydrology (soil moisture), the cryosphere (ice cover, permafrost) and the organic carbon cycle (remineralization and sulfate reduction).

Recent studies show a wide variation in Hg exposure in predatory animals (fish, whales, seals) and conflicting temporal trends in Hg concentrations in tissue (AMAP, 2011a). These biological trends cannot simply be linked to recent atmospheric Hg trends, which are relatively stable, and must, therefore, result from processes within the biogeochemical cycle in the Arctic (Macdonald and Loseto, 2010). In the case of seals from the Beaufort Sea, Gaden et al. (2009) proposed that changes in ice cover contribute a non-linear trend in tissue Hg caused by shifts in foraging strategy. Likewise, increases in Hg and PCB observed in Mackenzie River burbot *Lota lota* collected near Fort Good Hope (Carrie et al., 2010) have been assigned to increases in primary production that are likely to have been supported by declines in ice cover and/or increases in temperature.

In models, Hg concentrations in fish often respond directly to atmospheric deposition of Hg (Knightes et al., 2009), a finding supported by the results of adding isotope-labeled Hg to the drainage basin of an experimental lake (Harris et al., 2007). Newly deposited Hg took a shortcut route to the biota. Modeled response times for Hg in predatory fish varied across ecosystem types and were highly affected by sediment burial rates and active layer depths when the system was not dominated by inputs from the watershed. These results emphasize the importance of intermediary cryospheric processes – the ones presently undergoing rapid change – in controlling the exposure of aquatic biota to Hg deposition, past or present.

11.3.2.7. Change in weather patterns

The Northern Annular Mode (Wallace and Thompson, 2002) and related indices such as the North Atlantic Oscillation (NAO) and Arctic Oscillation (AO) have been used to understand decadal-scale variability in northern hemispheric wind speed/direction and precipitation (Becker et al., 2008, 2009). As reported in Chapter 2 (see Figure 2.5), during the past decade, the AO has been on average in a neutral phase, but with considerable variation shifting between positive and negative phase.

Air concentrations of HCH measured at the Zeppelin research station on Svalbard correspond well with variations in the AO over the past decade, showing stronger variations in concentration, but generally declining with time (Figure 11.10). This signal could not be found in data generated at the Alert Station on Ellesmere Island, north-eastern Canada. The Zeppelin station is obviously strongly influenced by the marine environment. Thus, this relationship, confined to the ocean-influenced site (Zeppelin), implies that ice cover together with atmospheric pressure patterns (winds) contribute substantially to atmospheric POPs variability. If Shindell et al. (1999) are correct in proposing that climate change may manifest itself in more frequent AO positive states, then such a change implies a shift in the relationship between atmosphere and ocean for some POPs.
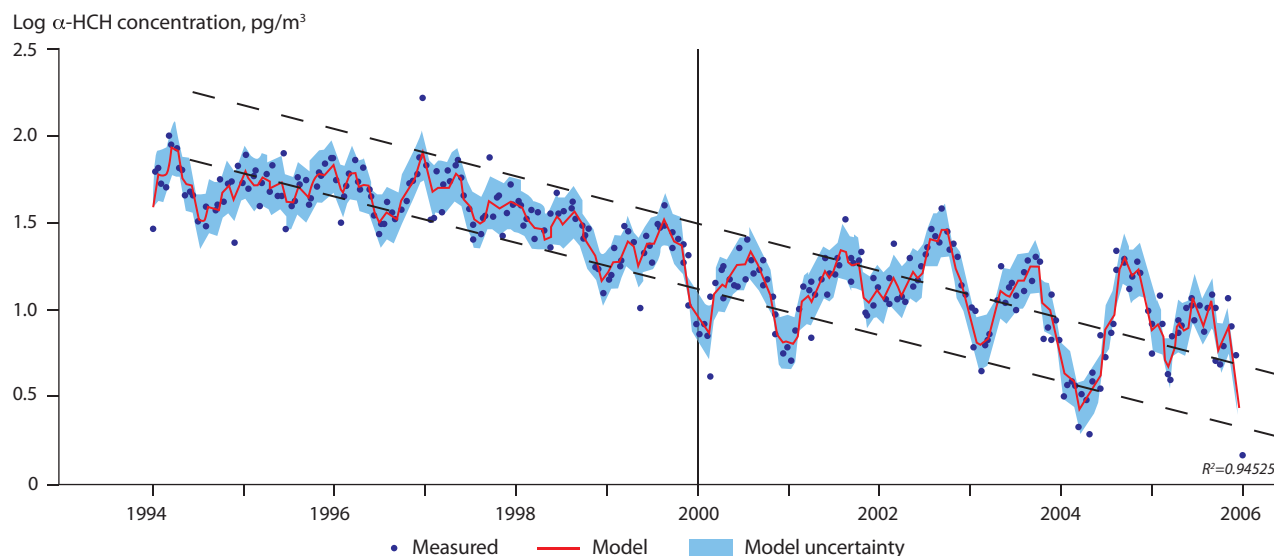
Log α-HCH concentration, pg/m³

Figure 11.10. Influence of the Arctic Oscillation on the atmospheric distribution of α-HCH in Zeppelin air data. The DHR model fit is represented by the R^2 value and uncertainty is to within 95% confidence limits. Dashed lines represent the expected differences between summer and winter concentrations assuming a constant decline. Source: Becker et al. (2008).

The differences in intra-annual distribution patterns indicate that much larger concentration variations between summer and winter months are occurring in years when the AO is negative (after 2000).

11.3.2.8. Invasive species and the Arctic food webs

A nearly ice-free summer is now considered likely for the Arctic Ocean by mid-century (see Chapter 9). This would have considerable consequences for the local marine ecosystems as it will allow the northward movement of more southerly species that will compete with indigenous species for habitats and food resources. The indigenous, more sensitive Arctic species (such as crustaceans and fish), will eventually be marginalized. Invasive species can alter risk from contaminants in several ways, including: alteration of lower food webs, thus changing carbon production and flow (Macdonald et al., 2005); alteration of availability of prey species, thus changing the net energy obtained through foraging (Gaston et al., 2003); and alteration of migratory pathways that supply contaminants through biovectors (Blais et al., 2007). Migration ranges of non-Arctic commercially exploited fish species (Atlantic cod, mackerel, capelin, etc.) that currently approach Arctic waters (Drinkwater, 2005; Feehan et al., 2009; Huse and Ellingsen, 2009; Opdal, 2010) will extend into the Arctic with warming, potentially displacing indigenous species (Macdonald et al., 2003). Other species may also enter the Arctic because thermal or ice barriers have been removed, or because they have been brought in through increased shipping (Occhipinti-Ambrogi, 2007). This alteration to foraging and feeding habits of top predators (e.g., Loseto et al., 2008; Lindstrom et al., 2009; Post et al., 2009) will alter the pathways and biomagnification of POPs and MeHg.

Terrestrial food webs can also be affected by invasive species and shifts in species migration patterns (Feehan et al., 2009; Prowse et al., 2009b; Sharma et al., 2009). Many non-Arctic bird species migrate into Arctic habitats (Henningsen and Alerstam, 2008). The Arctic presently appears poised for invasion (Vermeij and Roopnarine, 2008), which suggests that monitoring and ecosystem studies are urgently required (Fox et al., 2009; McKinney et al., 2009; Michelutti et al., 2009).

11.3.3. Impact of cryospheric change on local sources of contaminants in the Arctic

11.3.3.1. Thawing permafrost

In regions of permafrost in the Arctic, historical disposal of wastes has occurred in the form of sewage lagoons, dump sites at DEW (Distant Early Warning) line installations, solid waste dumps in small communities, and oil drilling sumps (Macdonald et al., 2005). The permafrost itself is a large component of the containment strategy at these sites. With permafrost degradation, however, contaminants contained in these sites can be washed into rivers or the ocean or leached into the local soil and groundwater, possibly contaminating drinking water. Sub-permafrost water resources and surface lake systems, often used as drinking water sources, may also be directly influenced by thawing permafrost (Kokelj et al., 2009; Haldorsen et al., 2010). This will also include remobilized anthropogenic and biogenic contamination (ACIA, 2005).

Disposal of PCBs in landfills and dumpsites has received significant attention at DEW line sites across Arctic Canada (e.g., Brown et al., 2009). Recently, polybrominated diphenylethers (PBDEs, flame retardants present in household items, electrical goods, etc.) have been found at high concentrations in soils around dumpsites in Cambridge Bay and Iqaluit in Arctic Canada (Danon-Schaffer et al., 2008). Permafrost thawing has the potential to release these contaminants into aquatic systems.

In Canada, since the mid-1990s climate change has been incorporated into the design of waste-containment structures that depend on the encapsulation of wastes in permafrost to contain mine tailings or similar waste (Chapter 5, Section 5.4); however, some of the older sumps, such as those used in the past to store process water and drilling fluid from oil and gas exploration activities in Arctic regions, have been found to be leaking (AMAP, 2010). There are around 300 to 400 of these sumps in northern Canada, while around 700 oil and gas production and exploration sites in Alaska have onsite drilling waste disposal pits (AMAP, 2010). There are

no current estimates of the use and number of oil pits in Russia. The use of pits in Alaska has largely been phased out and replaced by down-hole injection of wastes, while old pits containing oil-based drilling mud are being remediated.

Oil and gas pipelines in the Arctic are also vulnerable to permafrost degradation and thawing. Although studies of the permafrost-related performance of pipelines in Alaska and northern Canada have found the integrity of these pipelines to be robust to climate change, the situation in Russia is different (Chapter 5, Section 5.3.5.1.2). Above-ground pipelines in Russian permafrost areas have been affected by the thawing of ice-rich soils and uplift of pipeline foundations due to frost heave. In the 350 000-km long network of pipelines in Siberia, about 31 000 failures are registered each year, more than 20% of which are probably due to deformations and weakening of foundations induced by permafrost thaw. The world's largest terrestrial oil spill occurred in 1994 when multiple breaks occurred in a pipeline in the Komi Republic, contaminating a large area of land and two river waterways. Uneven ground settlement due to permafrost thaw and thermokarst is likely to have contributed to the breakage. Whether recent warming played a role might be difficult to establish, but projected increased thaw in the Arctic will only lead to greater risk in the future.

11.3.3.2. Exploitation and transportation activities in the Arctic

With decreased ice coverage of marginal sea areas of the Arctic for longer periods of the year, shipping routes between Europe, Asia, and North America will open up either via the Northern Sea Route (Russia) or the Canadian Archipelago. Shipping will introduce various contaminants, such as hydrocarbons and antifouling chemicals (e.g., diuron, tributyltin), exhaust fumes from the combustion of ships' fuel, and possibly also non-indigenous species on ships' hulls and in ballast water (Macdonald et al., 2005). There will also be increased risk of accidents that could result in oil spills from tankers. The environmental hazards associated with increased ship movements across the Arctic have been assessed in a review by Huntington (2009). The author stated that during the next several decades, Arctic marine mammals will face threats from six areas of human influence: climate change, environmental contaminants, offshore oil and gas activities, shipping, hunting, and commercial fisheries. Increased tourism (i.e., cruise ships and airborne transportation) is an additional threat for the Arctic environment. The potential contamination risks of increased offshore industrial activities in the Arctic were identified as much as a decade ago (Klungsøyr et al., 1995; Macdonald and Bewers, 1996). The decrease in marginal sea ice will open up the possibility for oil and gas exploration on the Arctic continental shelves of North America and Eurasia, resulting in a risk of oil spills from offshore platforms and the increased tanker traffic carrying oil to market. There are no effective means of cleaning up oil spills in broken ice and responding to such spills in winter would presently be impossible, given limited daylight and harsh weather (AMAP, 2007).

Karcher et al. (2010) highlighted the likely intensification of risks from artificial radionuclides released within the Arctic as a consequence of reduced ice cover. Russia has plans to develop

floating nuclear power systems for potential use within the Arctic; transport of such systems to and from the Arctic and the handling of waste products are of obvious concern. Likewise, the reduced cost of shipping via the Arctic applies equally to shipping nuclear materials and/or the number of ships with nuclear propulsion systems.

Although the ice may clear from substantial areas of the Arctic's marginal seas annually, ice cover will still accumulate in winter and winds will produce pressure ridges and keels. Large open areas where occasional patches of thick ice and icebergs are free to drift in summer might make the Arctic a more dangerous place to navigate.

11.3.4. Future research priorities

With respect to cryospheric change and potential impact on pollution patterns and distribution, the major future challenges are expected to encompass the following five research priorities:

1. The influence of change in ice distributions (areal and seasonal) on the introduction, distribution, accumulation and transformation of anthropogenic pollutants (sources and sinks). This includes sea ice, glacial ice, and permafrost.

2. The potential for new contaminants or re-introduced contaminants to be sequestered in the Arctic by changes in Arctic hydrology, ice cover and vegetation.

3. The potential for complex contaminant mixtures to exert an effect on humans or wildlife as a consequence of nutritional or other stresses brought on by cryospheric changes such as difficulty to access food or change in available food (invasive species, altered migrations).

4. Change in exposure pathways for contaminants owing to industrial development in the Arctic, expansion of Arctic communities and increased shipping through Arctic waters due to more suitable ice conditions for shipping.

5. Increased coordinated observing systems where contaminants are measured within the context of other physical changes (temperature, winds, ice conditions, seasonality, hydrology, etc.).

11.3.5. Future perspectives

The scenario descriptions in the concluding reports of the Arctic Climate Impact Assessment (ACIA, 2005) and the Intergovernmental Panel on Climate Change (IPCC, 2007) allow a first-approximation assessment of the future challenges associated with pollutant transport and distribution related to climate change in the Arctic (Macdonald et al., 2005). However, the scenarios now appear conservative compared to the real-world developments that have taken place in recent years and should, therefore, be updated before comprehensive studies of the effects of climate change on contaminant loads in the Arctic are performed. Technological developments within the field of analytical chemistry leading to new instrumentation that allows the quantitative detection of ultra-trace levels of hitherto unknown contaminants will inevitably lead to the identification and quantification of many new Arctic contaminants. An important perspective to maintain is that action to mitigate climate change might

also help to mitigate contaminants. An example would be the double threat of high atmospheric CO_2 levels and Hg emissions from the combustion of coal where reduction of CO_2 emissions will also lead to reduced Hg emissions. Future scenarios envisaged:

- Increased industrial activity, transport across the Arctic Ocean and offshore activities in the Arctic (due to continuous reductions in sea-ice extent in the central Arctic) will lead to significantly more potential local contaminant sources.

- The lack of sea-ice cover in the central Arctic Ocean will lead to increased summer evasion of some (semi)-volatile POPs (e.g., HCH) and increased deposition of others (e.g., PCBs, currently used pesticides).

- Change in the amount, composition and duration of snow will alter the deposition and transfer efficiency of contaminants.

- Ice-associated fauna will change and new marine species (fish, crustaceans, other invertebrates, etc.) will shift northward into the central Arctic basin, serving as contaminant carriers and altering the food-web structure with consequences for biomagnification of POPs and Hg into top predators. Change in ice and permafrost will alter foraging opportunities for wildlife and humans, thereby altering exposure to biomagnifying POPs and Hg.

- The human population of the Arctic will increase due to new working opportunities (e.g., offshore oil and gas activities, mineral exploitation) and with this will come additional domestic and industrial sources of contaminants. New chemicals will inevitably enter the Arctic environment.

- New analytical instruments will be developed that allow increasing numbers of chemicals to be detected at ever lower detection limits, permitting the development of longer and better time series in virtually all environmental compartments of the Arctic.

11.3.6. Conclusions

Within the Arctic, it is evident that the levels and distributions of contaminants are influenced by cryospheric change from the local to regional scale and that these changes are directly linked to the changing climatology of the Arctic region. With the loss of ice and frozen soil, former contaminant sinks have become sources for legacy contaminants such as POPs and Hg. Exchange between air and water is increasing, causing increased evasion of some contaminants (HCH, Hg(0)) and increased deposition of others (PCBs, endosulfan). Models of atmospheric POPs transport clearly show the influence of climatology (e.g., MacLeod et al., 2005). The expected increase in human activities in the Arctic, will promote the introduction of new chemicals and point sources within the Arctic. Increased human activity in ice-free areas over longer seasons, especially those activities concerned with offshore oil exploitation and shipping, increase the risk of oil spills. Increased thawing of permafrost and soil instability will require greater attention to ensuring the containment of contaminants in sewage lagoons, tailings ponds, and dumps, and the structural stability of pipelines traversing permafrost.

11.4. Impacts of changing snow, water, ice and permafrost on Arctic ecosystems

11.4.1. Introduction

The Arctic region contains a remarkable diversity of habitats and associated ecosystems, from mountain glaciers and ice caps to tundra, inland waters and icy seas. Many of the ecosystems are interdependent, and changes in climate are beginning to have not only direct effects on the ecology of these environments, but also indirect effects through the physical and biological coupling between them (Box 11.2). These impacts will have direct consequences for humans living in the North via changes in ecosystem services, and indirect consequences elsewhere, for example via changes in global biodiversity, biogeography and biogeochemical fluxes. The shifts in Arctic ecosystem structure and dynamics are also resulting in a variety of feedback effects that may accelerate their rate of change (see Section 11.1). This section summarizes information from the preceding chapters to provide a synthesis of how climate-dependent alterations in snow, water, ice and permafrost are affecting and will increasingly affect the ecology of the Arctic, including potential impacts via the linkages between ecosystems.

11.4.2. Snow

The extent, quality and duration of snow cover, exert a controlling effect on ecosystems throughout the Arctic (Chapter 4). There are large interannual fluctuations in each of these snow-related factors, and the observed responses to this natural variability, in addition to the results of controlled snow-manipulation experiments and longer term paleo-ecological records, are providing insights into how Arctic ecosystems may continue to respond to climate forcing. Over the past 40 to 50 years there has been an overall trend of increasing winter precipitation in the Arctic accompanying the warming air temperatures that have been observed, however there is large spatial variability in current and projected future trends, and thus the ecological responses can also be expected to vary among regions. Global climate models predict that there will continue to be large differences in precipitation trends among regions, with decreases in snow water equivalent (SWE) over Scandinavia and Alaska, no change in SWE over the boreal forest region, and increased SWE over northern Siberia (by 15–30%) and the Canadian Arctic Archipelago (Brown and Mote, 2009). There is a consistent projected trend, however, of decreased snow cover duration (SCD) throughout the Arctic, with the greatest changes predicted to be over the more southerly regions (Brown and Mote, 2009). The ecological effects of these pronounced changes may be compounded by a shift toward increasingly frequent episodes of winter melting that cause surface ice crusts over the snowpack, although basal ice layers that also impede animal foraging may become less frequent given the progressively shorter duration of snow cover.

In the terrestrial environment, changes in snow cover have wide-ranging ecological effects on soil, plant and animal communities, as well as downstream effects on lakes, rivers and wetlands (see Section 11.4.3). Snow is a habitat for diverse microbial communities including snow algae and associated

Box 11.2 From land to sea: ecosystem coupling between the Mackenzie River delta and the adjacent Arctic Ocean

Arctic ecosystems will continue to experience the impacts of climate change on snow, water, ice and permafrost, not only in their immediate environment but also via the adjacent ecosystems to which they are physically or biologically coupled. For example, decreasing occurrence of ice jams on the Mackenzie River (Canada) are resulting in less flooding of higher elevation lakes and wetland communities within the flood plain delta (Lesack and Marsh, 2007; see also Chapter 6, Figure 6.20). These ecosystems contribute to nutrient and organic carbon exchange processes that ultimately affect the biogeochemical characteristics of Mackenzie River water discharged to the coastal ocean (Emmerton et al., 2008). This reduced connectivity between lakes and the river may also limit habitat availability for fish populations migrating to and from the sea.
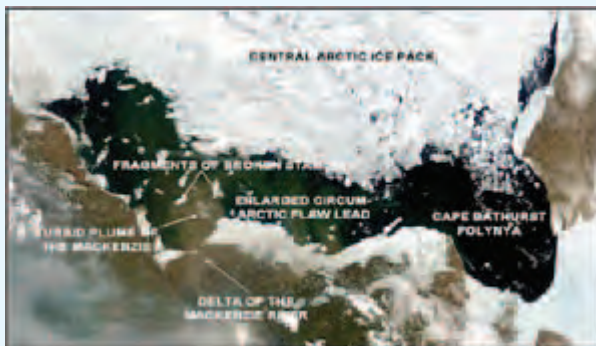
Water, particulates and dissolved materials exported from the terrestrial catchments of large Arctic rivers, including from permafrost soils, vegetation, and wetlands, are initially retained behind ice dams or stamukhi formed by ice convergence in the coastal environment (see also Chapter 6, Figure 6.21). These ephemeral lagoons such as the stamukhi-dammed waters of 'Lake Mackenzie' appear to have distinct microbial communities and are biogeochemical processing sites for terrigenous materials prior to their sedimentation to the marine benthos or discharge to the ocean (Galand et al., 2008). Changes in the duration and magnitude of these ice dams relative to shifts in the timing and magnitude of peak discharge of the river are likely to influence the export of material to coastal Arctic ecosystems. The land-fast ice and dams also form the landward edge of the band of open water that persists throughout the winter, creating the circum-Arctic flaw lead ecosystem.

Figure 11.11. MODIS satellite image taken in June 2004 of the Mackenzie Shelf region. The delta lakes have been flooded by the seasonal peak discharge of the Mackenzie River. The thick ice ridge parallel to the coast (stamukhi) has begun to break up, releasing turbid river water into the sea. Reproduced from Fortier et al. (2008).

bacteria and fungi (Jones et al., 2001; Harding et al., 2011). These organisms provide organic carbon and nutrients to the soil during snow melt. Snow also acts as an insulator for Arctic soils, and increased snow depth (predicted for the High Arctic) may result in warmer winter soil temperatures, increased biogeochemical processing of organic materials, and increased respiration. For example, in the Swedish sub-Arctic an extra 20 to 30 cm depth of snow (added to the 74 and 27 cm of snow over birch and heath vegetation, respectively, at the time of maximum snow accumulation) resulted in a 77–157% increase in ecosystem respiration rates (Larsen et al., 2007). Also, in High Arctic Canada an experimental increase in snow depth over birch hummock tundra from 30 cm (the natural snowpack depth) to 100 cm (achieved with snow fences) resulted in a 60% increase in carbon dioxide (CO_2) efflux rates (Nobrega and Grogan, 2007).

Many Arctic plants, especially lichens and bryophytes, become physiologically active even before the overlying snow completely melts. However, this competitive advantage may be lost with the earlier disappearance of snow, and the longer, warmer growing season may favor the invasion of plants from more southern latitudes. Extreme warming events during winter can result in the sudden loss of protective snow over vegetation and cause a range of damage including mortality of buds and shoots, delayed bud burst in spring, and reduced flowering (Bokhorst et al., 2008). Such effects may result in decreased rates of primary production and may partially offset the increases in productivity associated with shrub expansion across the Arctic (Bokhorst et al., 2009). These thaw events can result in the loss of winter dormancy, which renders the

plants and their roots vulnerable to damage during subsequent freezing of snow-free ground (Beier et al., 2008).

Snow cover has a strong effect on the ecology of Arctic animals, throughout winter for resident species and at the time of arrival for migratory species. For the latter, small changes in the duration of snow cover can translate into large changes in breeding success. For example, studies on pink-footed goose *Anser brachyrhynchus* show that a 10% decrease in snow cover duration can advance egg-laying by 5 to 6 days, resulting in a 20% increase in breeding success (Madsen et al., 2007). The end date of snow cover directly affects the duration of the growing season for plants and therefore the food resources available to Arctic herbivores. For example, reindeer *Rangifer tarandus* calf production in Finland has been shown to increase by almost 1 calf per 100 females for each extra day of the snow-free season (Turunen et al., 2009). Positive effects of longer growing seasons in recent years have also been observed for musk ox *Ovibos moschatus* populations in northeastern Greenland (Forchhammer et al., 2008). On Ellesmere Island, Canada, years of extensive snow cover in the past have resulted in reduced vegetation growth and reduced herbivore populations (musk ox and Arctic hare *Lepus arcticus*), which in turn resulted in reduced populations of Arctic wolves *Canis lupus arctos* that prey on the herbivores (Mech, 2004).

Snow quality also has an effect on animals. The formation of surface and basal ice-crusts can drastically reduce the availability of the underlying vegetation to herbivores (see Section 11.4.3 for examples). The overall hardness of the snowpack can affect the availability of subnivean space for rodents, and the resultant

changes in their populations can have indirect effects on other species used as alternate prey by their shared predators (Kausrud et al., 2008). The nutrient and organic carbon content, acidity, and contaminant load are also likely to influence the ecological effects of Arctic snow, particularly on soils and vegetation, and on downstream ecosystems such as lakes and wetlands, including thermokarst ponds that are extremely abundant throughout the circumpolar North. Growth of brown trout *Salmo trutta* in alpine lakes of western Norway was negatively correlated with spring snow depth (Borgstrom, 2001), presumably as a result of effects on water temperature, nutrient levels and lower-trophic level productivity. Thus, in addition to direct effects on runoff, snow accumulation in concert with lake ice shifts is also likely to alter ecological processes in Arctic lakes.

11.4.3. Water

The availability of liquid water is a fundamental requirement for life, and shifts in freshwater supply through climate change have direct and indirect effects on all Arctic ecosystem types and the ecological linkages between them. Water availability rather than temperature appears to be an increasingly limiting factor for the northward progression of the treeline, with previous trends of positive tree growth now appearing to be weaker, non-existent, or even reversed as a result of temperature-induced drought (Wilmking et al., 2005). These observations imply that the treeline will advance over the course of this century in a non-uniform way, colonizing only those parts of the landscape where there is sufficient moisture (Chapter 4). Many Arctic aquatic ecosystems have begun to show shifts in water balance, in some cases leading to the complete drainage or drying-up of lakes and wetlands. The degradation of permafrost soils in Siberia has resulted in collapse, drainage and complete disappearance of many water bodies (Smith et al., 2005), while elsewhere in the Arctic, permafrost thawing over the past 50 years has created new basins and increased development of shallow water ecosystems (Payette et al., 2004; Walter et al., 2006). These thaw lakes and ponds are important wildlife habitats, as well as biogeochemical 'hot spots' in the landscape where greenhouse gases are produced and ventilated at high rates to the atmosphere (Walter et al., 2006; Laurion et al., 2010). A shift in the precipitation-evaporation balance in parts of the High Arctic has resulted in the complete drying-up of shallow ponds in some rock basins, perhaps for the first time in millennia (Smol and Douglas, 2007a). Late-lying snow banks are an important source of water for wetlands (Woo and Young, 2003). Semi-aquatic ecosystems are therefore vulnerable to the ongoing trend of decreased snow cover duration, and desiccation of wetlands has been observed at a number of Arctic sites (e.g., Vincent et al., 2009).

Warmer Arctic temperatures will be accompanied increasingly often by a shift in precipitation from snow to rain, with many ecological consequences. For example, rain tends to favor broadleaf over conifer tree species, which in turn may decrease carbon storage, increase soil respiration, and reduce the timber potential of the forest (Chapter 4). Sudden rains during winter can result in a rapid loss of protective snow cover that makes the vegetation more vulnerable to freezing damage (see Section 11.4.2). Rainfall over snow cover can also result in the production of surface crusts of ice that prevent herbivores from accessing the underlying vegetation. For example, a rain-on-snow event on Banks Island in the Canadian Arctic resulted in a thick layer of ground ice; these conditions are thought to have resulted in the death of 20 000 musk ox (Rennert et al., 2009). Major declines in Peary caribou *Rangifer tarandus pearyi* in the Canadian Arctic Archipelago have also been attributed to unfavorable snow or ice conditions (Miller and Barry, 2009). The negative effects of ice-crusts associated with rain on snow are well known to reindeer herders in Fennoscandia, and there are increasing frequencies of such events in some regions. The shift from snow to rain will also affect marine and freshwater ecosystems. Snow provides a highly reflective surface that greatly reduces the penetration of light for photosynthesis in the planktonic communities within and beneath ice (Vincent and Belzile, 2003) and rainwater saturation of snow will greatly increase under-ice irradiance, as well as hasten melting of the remaining snow and ice.

Arctic freshwaters provide important migratory routes for fish, and changes in climate may alter the connectivity among lakes and river channels, as well as the physical coupling to the coastal marine ecosystem. Glacier-fed rivers are showing sustained periods of higher discharge associated with the increased duration and intensity of melt under a warmer climate (Chapter 7), and this provides more favorable habitats for some invertebrate and fish species (see Section 11.4.4.2). Increasing discharge has also been observed in the large Siberian rivers, with record flows in 2007 (Shiklomanov and Lammers, 2009). The changing freshwater content of the Arctic Ocean is likely to have a variety of effects. Increasing flow from Arctic rivers may bring increased nutrients and allochthonous (external) organic matter that stimulate processes at the base of marine food webs. However this water, combined with the effects of melting sea ice, may reduce the salinity of the Arctic Ocean mixed layer, thereby intensifying stratification. Over recent years, freshening of the Arctic Ocean surface waters has been accompanied by a shift in phytoplankton toward small-celled species (picophytoplankton, especially a cold-water ecotype of *Micromonas*; Lovejoy et al., 2007) that are more competitive in stratified nutrient-poor conditions. This cellular carbon is much less efficiently transferred within marine food webs than that associated with diatoms and other large cells, and the resultant changes may therefore alter organic carbon export to fish communities and to benthic ecosystems (Li et al., 2009).

11.4.4. Ice

11.4.4.1. Greenland Ice Sheet

The Greenland Ice Sheet has long been recognized as a habitat for microbial ecosystems associated with sediment accumulation and ephemeral meltpools. These cryoconite (cold rock dust) communities are dominated by nitrogen-fixing cyanobacteria that are darkly pigmented with UV photoprotectants (Vincent, 2000 and references therein), and may become more widespread over the ice as the Arctic climate warms. Runoff from the melting ice cap provides substantial water, nutrients and probably organic carbon to coastal environments around Greenland, and increased fluxes may affect marine food webs. Changes in iceberg fluxes from the ends of ocean-terminating glaciers may influence benthic communities through ice scouring, and cause local changes in

the surrounding environment. The effect of increased melting of the Greenland Ice Sheet on the adjacent marine ecosystems remains little explored, and is an important priority for future research (Chapter 8).

11.4.4.2. Glaciers, ice caps and ice shelves

Arctic glaciers provide a habitat for cryoconite microbial communities, as well as water supplies for streams, rivers, wetlands and the coastal ocean. From the 1990s onward, there have been increasing rates of mass loss from glaciers in Alaska, the Canadian Arctic, and Iceland, and this trend is likely to continue over the next few decades (Chapter 7). In the medium-term, ongoing increases in discharge, especially winter flows, may stimulate benthic invertebrate diversity (e.g., Milner et al., 2008) and enhance salmonid populations (e.g., Fleming, 2005). In the longer-term, with glacial shrinkage and declining discharge, there will be change in fluvial habitats, including the opening up of new streams in the landscape caused by large-scale glacial recession (Milner et al., 2008). Higher temperatures are likely in the absence of glacial meltwater inputs, which are likely to favor declines among cold-water salmonids (e.g., Matulla et al., 2007), and the invasion of warm-water taxa from the south. Changes in runoff volume will also influence the coastal environment, with potential effects on currents, nutrient supply, marine productivity and fisheries, and the eventual retreat of tidewater glaciers will result in loss of the productive glacier-front habitat for marine birds and mammals (Chapter 7).

Unusual ice features in the Arctic and their associated microbial ecosystems are likely to become extinct through ongoing climate change (Vincent, 2010). The ancient ice shelves along the northern coastline of Ellesmere Island in High Arctic Canada harbor diverse communities of microbes and micro-invertebrates (Varin et al., 2010), and these shelves are currently experiencing rapid attrition. For example, in the warm summer of 2008, there was a 23% decrease in overall ice shelf habitat, including complete break-out and loss of the Markham Ice Shelf (Vincent et al., 2009). These ice shelves dam bays and fiords to create 'epishelf lakes', an ecosystem type that is well known in Antarctica. Fracturing and break-up of ice shelves can cause a draining of such lake ecosystems, and the loss of their unusual fresh and brackish water communities. Only one epishelf ecosystem is believed to remain in the High Arctic (Milne Fjord, northern Ellesmere Island, Canada), and it is vulnerable to ongoing warming (Veillette et al., 2008). The loss of ice shelf ecosystems will affect biogeochemical processing of material entering the sea from land-based ecosystems, and conversely, the newly opened coastal waters are likely to cause accelerated warming of the adjacent terrestrial environment.

11.4.4.3. Lake ice

Ice covers Arctic lakes for six to 12 months each year and therefore has a major influence on the physical, chemical and biological characteristics of these ecosystems (Chapter 6). There is evidence of earlier break-up and delayed freeze-up at many sites (e.g., Latifovic and Pouliot, 2007), and this trend is predicted to accelerate in the future (Prowse et al.,

2009a). Such changes may favor some fish and invertebrate species by providing more ice-free habitat, and by reducing the period of time that the lakes are sealed from oxygen exchange with the atmosphere. Paleolimnological studies suggest that the shorter duration of snow and ice cover results in higher primary production at sites throughout the Arctic (Melles et al., 2007; Smol and Douglas, 2007b; Antoniades et al., 2009). Such changes have been attributed to improved light conditions for photosynthesis, which in some lakes may be compounded by increased wind-induced mixing and entrainment of nutrient-rich deeper waters into the photic zone, and increased nutrient inputs from their catchment areas in a warmer climate. The longer ice-free period will allow more exposure of the water column to direct solar heating, and in the High Arctic may drive lakes from cold monomictic (one period of mixing a year) situations to dimictic (two periods of mixing a year) conditions. The latter may favor higher productivity in a warm surface mixed-layer, and the development of cladoceran zooplankton communities (Sorvari et al., 2002), leading to increased contaminant transfer to high trophic levels (Chételat and Amyot, 2009). These warmer conditions may force cold-adapted specialists into more restricted habitats, for example Arctic char *Salvelinus alpinus* may be increasingly confined to deep lake habitats (Power et al., 2008). Alternatively more complex responses are possible for some fishes. In mid-latitude lakes (e.g., 60° N, northern boreal) lake trout *S. namaycush* tend to use epilimnetic waters which more closely approximate their thermal preferences thereby promoting growth. Projected effects of earlier ice loss and radiative heating suggest shallower and earlier thermocline development which will decrease the availability of this habitat (Mackenzie-Grieve and Post, 2006), thus stressing such populations. Finally, the increased evaporative losses associated with a longer period of open water may result in decreased lake levels, a loss of shallow water habitat, and in extreme cases complete loss of the aquatic ecosystems (e.g., Smol and Douglas, 2007a).

Changing lake ice will also alter the coupling between these ecosystems and other environments and biota. These climate-dependent changes may influence the through-flow of contaminants from catchments via lakes to the sea (e.g., Semkin et al., 2005). Coupling of lake-ice decline to increased contaminant exposure is supported by correlative links observed in mercury and polychlorinated biphenyl incorporation into burbot, a locally harvested freshwater fish species present throughout many Arctic freshwaters (Carrie et al., 2010). The increasing open water conditions may attract aquatic bird populations and thereby new trophic levels (e.g., Vincent et al., 2009), and may favor the competitive success of less cold-tolerant plant, fish and other organisms invading from the south, perhaps driving some northern species to extinction (e.g., Sharma et al., 2007).

11.4.4.4. River ice

Ice is a major structuring feature of Arctic rivers and their associated flood plain ecosystems, and changes in the duration and magnitude of ice accumulation will have many ecological impacts, not only on these waters but also on the coastal ocean ecosystems into which they discharge (Chapter 6). Ice reduces the winter survival of many river-dwelling organisms (e.g.,

Huusko et al., 2007) and release from this constraint will have a mixture of negative and positive effects on riverine communities. Reductions in ice cover may result in decreased habitat availability for some fish species (e.g., Linnansaari et al., 2009), and mid-winter break-up may be especially deleterious (Wrona et al., 2005). Conversely, with ongoing warming, many Arctic rivers that currently have bed-fast ice may open up as winter habitats for fish, and increasing zones of broken ice or open water during winter will aid the oxygenation of rivers. River ice, especially during break-up, causes substantial bank and channel erosion that not only controls the geomorphology of the fluvial environment and therefore the spatial distribution of habitat types (Prowse et al., 2006), but also affects the timing and magnitude of particulate and dissolved material exported from terrestrial catchments to the sea. Ice scour damage results in tiered bands of different types of riparian vegetation that correspond to break-up frequency and severity (Prowse and Culp, 2008), and reductions in river ice could result in reduced diversity and structure of such vegetation and its associated biota.

The larger Arctic rivers have lake-rich floodplains that are productive sites for aquatic and semi-aquatic plants and animals. These also act as biogeochemical processing regions for river water before its discharge to the sea (e.g., Emmerton et al., 2008), and changes in the flooding regime are likely to affect not only the ecology of the floodplain delta, but also biogeochemical characteristics, primary production and food web processes in the coastal marine ecosystem. Ice-jams result in high spring flood levels, and decreasing ice compounded with reduced spring runoff as a result of less winter snow pack will reduce ice-jam flooding (Beltaos et al., 2006). This could isolate some of the higher elevation floodplain lakes and their biota from the river (Box 11.2). Changes in coastal stamukhi ice will also affect how river water is biogeochemically modified before it is discharged into the marine environment (Box 11.2).

11.4.4.5. Sea ice

Sea ice is a key component of the Arctic Ocean ecosystem, which is currently undergoing rapid change (Chapter 9, Section 9.3). Observations and models indicate that the extent of multi-year ice has diminished greatly over the past few years, and that Arctic seas may be seasonally ice-free within a few decades. This has enormous ecological implications for the structure and functioning of all Arctic Ocean ecosystems. It will also impact strongly on coastal terrestrial ecosystems and the lifestyle of indigenous communities living on the coast.

Sea ice is a habitat for marine mammals, and for diverse communities of other biota that live within or in association with the ice. Changes have already been observed in Arctic marine ecosystems that have been attributed to recent decreases in sea-ice extent and duration. These include: a regime shift in the Bering Sea from an ice-dominated ecosystem, with bottom-feeding birds and mammals that depend on carbon input to the benthos from sea-ice algae, to an open water ecosystem more dominated by plankton and pelagic fish (Grebmeier et al., 2006); satellite evidence of increased phytoplankton biomass in response to more open water and a longer open-water season (Arrigo et al., 2008); and a decline in condition and reproductive

success of polar bears *Ursus maritimus* in western Hudson Bay associated with earlier ice break-up (Regehr et al., 2007). Such changes are likely to accelerate over the course of this century, but there are many uncertainties that make future predictions challenging. Reduced sea ice is also almost certain to lead to increased shipping, tourism and exploration for oil, gas and mineral resources, all of which are likely to have ecological impacts on land as well as in the sea. Also, changes in sea ice will affect hunting and traditional over-ice transport routes that will become increasingly dangerous, even in winter. Finally, the increased extent and duration of open water will affect the temperature and precipitation regimes of coastal ecosystems, including those based on land, as well as coastal erosion processes (Chapter 9, Section 9.4).

Changing sea-ice conditions (less multi-year ice, more seasonal ice, earlier melt, later freeze-up) may lead to increased light availability for photosynthesis and therefore increased biological productivity at all trophic levels (see Chapter 9, Section 9.3). However, these positive effects may be offset by increased freshwater runoff and ice melting that enhances stratification and inhibits the entrainment of nutrients from the deeper layers into the surface waters by mixing, thereby limiting phytoplankton production. Field measurements in Canada Basin have revealed high pCO_2 levels in its surface waters after extensive ice melt had occurred, leading to the prediction that the Arctic Ocean will not become a large photosynthetic sink for atmospheric CO_2 under ice-free conditions (Cai et al., 2010). Also, increased runoff of river water rich in dissolved and particulate organic matter may cause local reductions in underwater light, particularly to the deep population maxima of phytoplankton that can contribute a large fraction of total water-column photosynthesis (Retamal et al., 2008; Martin et al., 2010). The loss of snow-covered sea ice also results in water column biota experiencing much greater ultraviolet (UV) radiation exposure (Vincent and Belzile, 2003), and this high energy waveband may have negative effects on some phytoplankton and other biota. Satellite imagery suggests that there are large differences between different parts of the Arctic Ocean in the response of phytoplankton to decreased ice cover, however ongoing work is required to validate such observations for Arctic seas, where the surface chlorophyll concentrations may not be an accurate guide to total water-column algal biomass.

Sea ice is an important habitat for lipid-rich, Arctic-adapted zooplankton that feed on ice algae and ice edge phytoplankton blooms, and these in turn provide high quality food for seabirds, fish and whales. The algal populations in these habitats also provide an important food source to benthic communities. Ice-associated marine mammals will be especially impacted by changing sea ice. Already, ice-related declines have been reported in hooded seals *Cystophora cristata*, harp seals *Phoca groenlandica* and ringed seals *Phoca hispida* at some locations, and there is also evidence of impacts on the distribution and health status of Pacific walrus *Odobenus rosmarus* (Chapter 9, Section 9.3). Whale species may differ greatly in their sensitivity to ice change depending on their degree of specialization and flexibility in diet. Narwhal *Monodon monoceros* are likely to be strongly impacted by the loss of pack ice in their winter-time benthic feeding regions (Laidre et al., 2008), while the more generalist feeding habits of beluga *Delphinapterus leucas*, grey whales *Eschrichtius robustus* and bowhead whales *Balaena mysticetus* may allow them to adjust more readily to changes in the Arctic marine ecosystem

caused by loss of sea ice (Moore et al., 2010). Polar bears are especially prone to negative consequences of the changing sea-ice conditions, and drastic declines have been predicted (e.g., Durner et al., 2009), including complete extirpation from many areas where they are presently common (Regehr et al., 2010). The ecology of terrestrial species that use sea ice seasonally as foraging habitat (e.g., Arctic fox *Vulpes lagopus*, Gagnon and Berteaux, 2009; Tarroux et al., 2010) will also be strongly modified by sea-ice decline.

Ice-associated zooplankton are important for certain fish species, notably capelin *Mallotus villosus* and Arctic cod (*Boreogadus saida*), which in turn are resources for several commercially important species such as Atlantic cod *Gadus morhua*, as well as whales (e.g., beluga). Bird populations that depend on these zooplankton, such as some auklets, are likely to experience ongoing decline, although some bird species appear to be increasing in population size, possibly as a result of earlier ice break-up (Chapter 9, Section 9.3).

11.4.5. Permafrost

Permafrost, defined as soil, rock or sediment that remains at or below 0 °C for two or more consecutive years, covers vast areas of the Arctic. Unlike sea ice, it does not provide a unique habitat for a dependent biota (with the exception perhaps of extreme cold-adapted microbes in permafrost micro-niches; Gilichinsky et al., 2008), but it does strongly influence the hydrology, biogeochemistry and geomorphology of the landscape and thereby the biological productivity and biodiversity of northern regions (Chapter 5). Permafrost provides an impermeable barrier to drainage, and the resultant high water table favors wetlands, lakes and ponds, and their associated biota (Woo and Young, 1998). Uplift and erosion of permafrost soils cause variations in surface topography and drainage, and the resultant heterogeneity causes a patchy distribution of plants and other organisms across the landscape, allowing increased species diversity. The degradation of permafrost proceeds in a great variety of ways that depend not only on climate but also on geomorphology, soil texture, hydrology and ice characteristics. These diverse degradation pathways in turn affect the nature of the ecological responses, including creation of lakes and wetlands, conversion of forests to fens, and increased drainage with loss of lakes or little change in vegetation (Jorgenson and Osterkamp, 2005).

In many parts of the Arctic, the permafrost has begun to warm and the active layer has deepened. However, the extent of these changes is highly variable regionally, with a deepening of the active layer by several metres or more at some sites and yet no detectable change at others. In Russia, for example, thawing has been pronounced over the past three decades at many sites, especially in the discontinuous permafrost zone, while at some other sites such as the Urengoy area in western Siberia, the permafrost continues to be stable (Romanovsky et al., 2010). Slumping and drainage due to permafrost melting may damage some vegetation, such as the complete destruction of forest ecosystems (Jorgenson and Osterkamp, 2005), but the disturbance also allows new species to colonize and become established (e.g., Lantz et al., 2009), including invasive species from the south. In parts of the Arctic, thermokarst lakes and wetlands are expanding as a result of permafrost thawing and erosion, thereby producing more habitats for aquatic birds and

other animals. Such effects have been observed in northern Quebec (low Arctic Canada), for example, where the southern limit of discontinuous permafrost has moved 130 km northward over the past few decades (Thibault and Payette, 2009) and thaw lakes have become larger and more abundant (Payette et al., 2004). Elsewhere, however, the degradation of permafrost is causing a draining of the landscape and loss of aquatic and semi-aquatic ecosystems. Such changes may result in a more homogeneous northern landscape, with reduced habitat and species diversity. The sub-Arctic palsa mires (bogs containing permanently frozen peat hummocks) of Fennoscandia, for example, are predicted to disappear almost completely by 2050; this loss is likely to have detrimental effects on migrating birds such as waders that feed on the abundant insect life in these habitats (Luoto et al., 2004).

There are complex interactions between permafrost soils and vegetation that affect soil temperature, surface albedo and even a region's sensitivity to fire. By trapping snow and directly insulating the ground with its foliage, plant communities can reduce soil temperatures by several degrees, and permafrost patches may therefore continue to exist beneath undisturbed vegetation despite rising air temperatures. Trees and shrubs have a much lower albedo than snow on tundra, and the conversion of northern landscapes to shrub tundra and forest is likely to cause increased heating; after snowmelt, however, increased evapotranspiration from vegetation can also cause local cooling (Göttel et al., 2008). Finally, increased vegetation biomass combined with drier conditions is likely to increase the frequency of fires on the tundra, with additional impacts on permafrost stability and temperature.

Thawing permafrost may release organic matter, gases, inorganic solutes and microbiota into the surrounding environment, with implications not only for land-based ecosystems, but also for thaw ponds, northern rivers and the coastal ocean. The increasing incidence of fires and other extreme weather events (Post et al., 2009) may also exacerbate permafrost degradation and the transfer of soil materials into waterways. Terrigenous, organic-rich particles released into rivers are a substrate for bacteria and archaea (Garneau et al., 2008), and climate effects on these organisms combined with increased light limitation of phytoplankton may drive Arctic rivers and estuaries toward even greater heterotrophy and net CO_2 efflux in the future (Vallières et al., 2008). Viable microbial populations dating back three million years have been identified in Siberian permafrost (Gilichinsky et al., 2008), and the impacts of release of these ancient genotypes into the present-day biosphere are completely unknown. Globally significant quantities of organic carbon are stored in frozen northern soils (Tarnocai et al., 2009), and some of these compounds released into lakes, rivers and coastal seas may be converted into CO_2 and methane. Sub-Arctic thaw lakes have been found to be stratified throughout most of the year, with anoxic bottom waters despite their shallow depths; these conditions favor microbial processes that convert allochthonous tundra carbon to methane, which is ultimately released to the atmosphere (Laurion et al., 2010). Large quantities of methane are stored in the tundra and in the frozen sediments beneath Arctic seas as gas hydrates and free gas. The continuous efflux of this gas (e.g., into the ocean over the East Siberian Arctic Shelf; Shakhova et al., 2010) is likely to stimulate methanotrophic bacteria and perhaps chemotrophic food webs, but such effects

have been little explored to date, and there is much uncertainty about how much warming will be required to trigger globally significant releases from these vast greenhouse gas reservoirs.

11.4.6. Conclusions

The Arctic's changing climate has already had marked impacts on snow, water, ice and permafrost regimes of the region. Such changes will continue to accelerate over the course of this century and will have ongoing impacts on the ecosystem services that are vitally important to northern indigenous communities, such as safe drinking water supplies and fishing, hunting and other harvesting resources. Each of these changes is affecting different Arctic ecosystems in different ways. However, there are also some commonalities. In each ecosystem type, from permafrost to sea ice, there are large variations among Arctic sectors in the pace of change. This heterogeneity is likely to persist in the future, underscoring the urgent need for downscaling climate projections and impact assessments to specific Arctic sectors and localities.

Extreme climate events are likely to continue to have severe effects on many ecosystem types. Examples discussed above include rain on snow in winter giving rise to surface crusts of ice that impede animal feeding, a seasonal thawing of permafrost that breaks winter dormancy and cold-tolerance of plants, and winter break-up of river ice that disrupts fish habitats. Such extremes compounded with the general warming trend is hastening the demise of unique habitats, for example palsa mires in Fennoscandia, ice shelves and epishelf lakes in northern Canada, and multiyear sea ice across Arctic Ocean.

As the Arctic region warms and loses its extreme habitats, it is becoming more homogenous and more suited to generalist species. Specialists such as psychrophilic microbes, polar desert plants, and obligate sea-ice biota including seals, walrus and polar bears will become increasingly limited in habitat range, while invasive generalists from the south are likely to become increasingly prevalent constituents of each type of Arctic ecosystem. The loss of perennial sea ice and increased access to resources in the Arctic will be accompanied by increased human activities, which will impact all ecosystems. Accelerated development of the North will create ongoing challenges for Arctic ecosystem management, and for the traditional lifestyles of Inuit and First Nation communities that have been based on these ecosystems.

Many of the ecosystems referred to here are interdependent, and the overall effects of changing snow, water, ice and permafrost on Arctic ecology cannot be evaluated without considering this connectivity. The melting of glaciers, ice caps and snow and the resultant transfer of water, nutrients and stored contaminants from land to the ocean mean that shifts in the terrestrial cryosphere have strong ecological effects downstream, for example, changes in organic carbon and nutrient export to coastal food webs, and in the freshwater content and stratification of the Arctic Ocean ecosystem. Conversely, changes in sea ice may affect the temperature and precipitation regime of coastal terrestrial ecosystems, and the migrations of fish and birds between marine and freshwater environments. The many forms of ice throughout the Arctic region give rise to diverse, spatially complex habitats for terrestrial and aquatic life (e.g., Box 11.2), and ongoing climate change has begun to alter these ice-dependent ecosystems and the connections among them.

11.5. Observational needs and knowledge gaps for the cryosphere

Summary

Cryospheric observations help us understand, assess, predict, mitigate, and adapt to climate variability and change; improve weather forecasting; help reduce the loss of life and property from disasters; provide a better understanding of environmental factors affecting health; and improve the management of resources. Closing gaps in snow and ice observation networks and in knowledge of cryospheric processes would allow the full potential of snow and ice information to be realized for the benefit of society. This assessment has made the following findings:

- Many Arctic surface-based observation networks for snow, lake and river ice, permafrost, and precipitation have diminished or been completely lost. Many important observations and monitoring efforts are research based and in need of sustained funding.

- There is no pan-Arctic dataset of *in situ* snow measurements. There are few measurements of snow depth on sea ice. Space-based capabilities for snow extent are robust, but methods of estimating snow water equivalent and snow depth are limited.

- Precipitation gauge networks are the most important source of information on high latitude snowfall but have large errors. Satellite methods are challenging but promising.

- While satellites have provided reliable observations of sea-ice extent, concentration, and motion for over 30 years, methods for estimating ice thickness from space are only now being developed. *In situ* measurements of ice thickness are sparse.

- There are numerous boreholes that provide temperatures for permafrost studies but many of the records are discontinuous and short. There is a wealth of historical data extending back 50 to 100 or more years; data rescue efforts are needed.

- Glacier inventories have basic data for less than half of the world's glaciers. Satellite-based inventories are discontinuous. Mass balance measurements over long periods are available for only a small subset of Arctic glaciers.

- Since the Arctic Climate Impact Assessment there has been considerable progress in the ability to estimate the mass balance of ice sheets. Gravity, laser altimetry, and radar data from satellites have become a major source of information. These new observational methods have broadened the range of estimates, revealing that uncertainties may be larger than previously thought.

- Traditional knowledge can provide important value-added content to data products and serves to make the data more relevant to northern users. Community-based observing programs provide a mechanism for two-way knowledge transfers.

- There are considerable uncertainties in modeling cryospheric processes. Permafrost models under-represent ice content and the insulating effect of the organic layer; climate models do not resolve the steep topography of the Greenland Ice Sheet margins; models of snow-vegetation interactions need to be improved; and models that link meteorology to glacier mass balance need to incorporate downscaling techniques and satellite data.

11.5.1. Introduction

Snow and ice observations play a crucial role in numerical weather prediction, climate monitoring, change detection, hazard forecasting, and water resource management. A comprehensive set of observations is required to assess the efficacy of mitigation and adaptive strategies and to prioritize needs for governance. Improved mapping and monitoring of drifting ice and icebergs, local sea-ice formation, annual variations in ice cover, and trends in storm surge characteristics are important to shipping, industrial developments, and local hunting. Forecasts of break-up

and freeze-up of the ice will help industry to determine the seasonal window for operations. Monitoring of snow cover, winter thaw, and rain will provide information relevant for hydropower, drought, snow load, avalanches, and floods. For many hunting activities, knowledge of changes in the timing and amount of snow and ice are important. For maintaining the integrity of infrastructure, monitoring of permafrost characteristics is important.

However, conditions in areas where the cryosphere exists are harsh, and *in situ* observations there are difficult and expensive. Satellite monitoring overcomes some of the logistical obstacles, but satellites are costly and do not yet fully address the range of geophysical variables needed to understand the cryosphere. Furthermore, satellite products will always require ground truth for validation. A comprehensive cryosphere observing system must therefore be a combination of ground-based instrumentation, satellite remote sensing, aircraft measurements, modeling, and data management (Figure 11.12). Surface and airborne observations provide data that cannot currently be measured from space, more
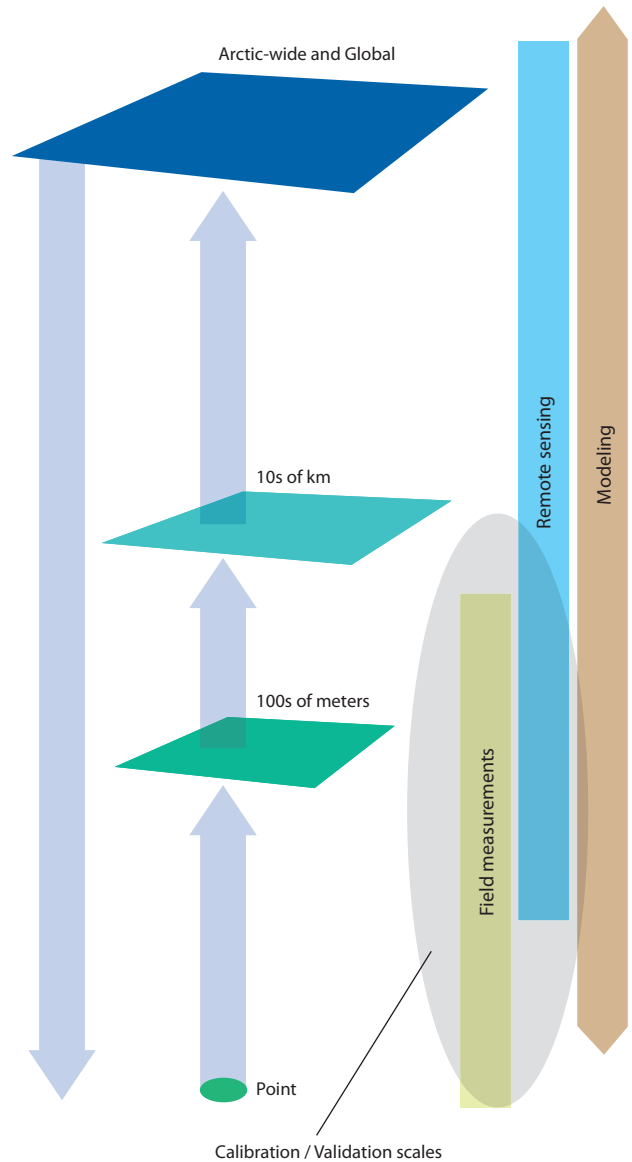
11.12. The various types of observation system (satellite, aircraft, and in situ) and scales of operation. Sources of photos from top down: ESA-AOES Medialab, NASA/Patrick Black, Martin Fortier/ArcticNet, Roland Kallenborn.

detailed information in critical areas, and observations with which to calibrate and validate satellite retrievals. Satellite instruments are essential for delivering sustained, consistent observations of the global Arctic cryosphere and are key to extending local *in situ* measurements. No single all-encompassing sensor exists; rather, the combination and synthesis of data from different yet complementary sensors is essential and underlines the critical importance of maintaining key synergetic elements of the system.

While the previous chapters use snow and ice data to assess changes in the Arctic cryosphere, this section assesses the state of the cryosphere observing system itself. It addresses observations more broadly, synthesizing the material presented earlier and focusing on observational gaps. Table 11.2 provides a summary of the observing systems for snow and ice. It lists the measurement approach for the major variables, describes the status of the networks (operational or research), provides a qualitative assessment of how well each is meeting the measurement requirements, and outlines any major issues. As robust as the satellite and *in situ* measurement networks are, there are many shortcomings in the cryosphere observing system that give rise to sometimes-large uncertainties in assessments such as SWIPA. These are discussed for each cryospheric element in the following sections. The societal impacts of changes in the cryosphere, and hence the importance of robust measurements of snow and ice properties, are described in Chapter 10.

Table 11.2. Observational readiness of snow and ice measurements for the observing system overall. Green satisfies the requirements (≥ 85%); blue meets the requirements most of the time (≥ 70%); yellow meets the requirements some of the time, or only for specific conditions; orange does not meet the requirements. O: operational, R: research, C: commercial, L: long-term (≥ 20 years) record.

|	In situ	Satellite	Major gaps in observations
Ocean			
Ice extent	Coastal radar (R), Ship observations	Passive microwave (O, L)	*In situ* coverage is sparse and incomplete
Ice concentration	Ship observations	Passive microwave (O, L)	Potentially large uncertainties in satellite retrievals in summer
Ice thickness	Ice-profiling sonar on moorings (O), Mass balance buoys (C), Electromagnetic sleds (R)	Optical, Laser and Radar altimeter (R)	Satellite methods are still developing; snow depth on ice is an unknown
Ice motion	Drifting buoys (O, L), Coastal radar (R)	Passive and active microwave (O, L), Optical (R)	Important small-scale motions not captured by satellites; *in situ* measurements sparse
Snow depth on ice	Depth gauge (R)	Passive microwave (R)	Satellite method is limited to first-year ice with potentially large uncertainties; *in situ* data are sparse
Sea level	Tide gauges (O, L), Bottom pressure recorders (C)	Altimeters (R)	
Surface temperature	Drifting buoys (O, L)	Optical (O, L)	Uncertainty in satellite estimates due to cloud cover
Albedo	Radiometers (O, L)	Optical (O, L)	Sparse *in situ* coverage; significant uncertainty
Terrestrial			
Snow cover	Depth gauge (O, L)	Optical (O, L)	*In situ* network is declining
Snow depth	Depth gauge (O, L)	Optical (R)	Satellite method is limited to tall-grass prairie
Snow-water equivalent	Various methods (O)	Passive microwave (R)	*In situ* coverage is sparse
Freshwater ice	Visual observations (O, L)	Optical (R)	Declining observation network
Glacier, ice cap, ice sheet thickness	Seismology (R)	Radar (R), Gravity (R)	Sporadic coverage
Glacier length, area	Surveys (R, L)	Optical (R)	Incomplete coverage
Glacier, ice cap, ice sheet motion	GPS (R)	InSAR, Visible imager (R)	Sporadic coverage
Permafrost: ground temperature	Boreholes (O, L)	(none)	Large parts of the Arctic not covered
Permafrost active-layer thickness	Boreholes, Probes (O, L)	(none)	Large parts of the Arctic not covered
Surface temperature	Thermisters, Thermocouples (O, L)	Optical (O, L)	Satellite method is clear sky only
Albedo	Radiometers (O, L)	Optical (O, L)	Sporadic *in situ* coverage; significant uncertainty

Latitude of August ice edge , °N
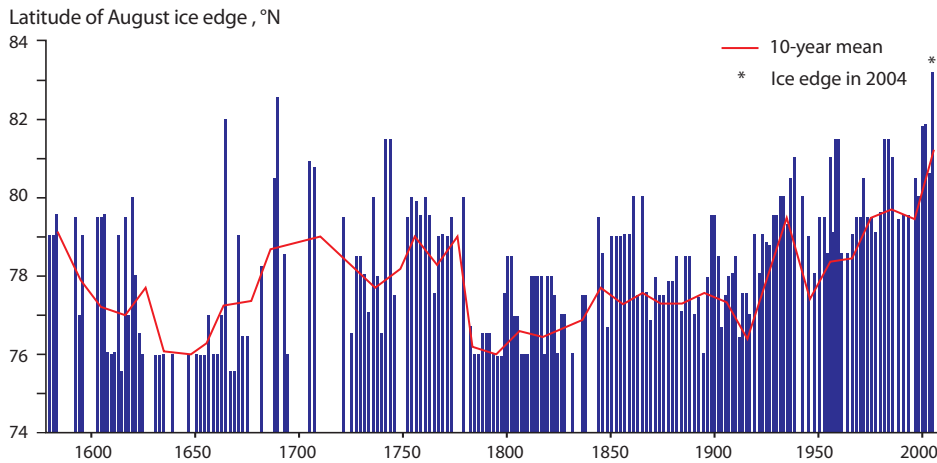
Figure 11.13. Northerly position of the ice edge in August between Svalbard and Franz Josef Land for the period 1553 to 2004 given by its mean latitude in the sector 20–45° E (blue bars). Source: Falk-Petersen et al. (2008).

11.5.2. Sea ice

The first quantifiable observations of sea-ice properties came from ships sailing in and near ice-covered regions as early as the 1500s (Figure 11.13). National ice services began routinely producing ice charts for the Arctic in the 1950s using visual aerial reconnaissance missions and aerial photography. The thickness of ice fragments laid on edge due to the icebreaker movement has been determined using onboard digital video cameras on Russian icebreakers for many years. Submarine sonar observations of ice draft under the Arctic sea ice began in earnest in the 1950s, although the data are sparse and remained classified for decades. Moored upward-looking sonar (ULS) gives thickness and volume flux estimates from below. Electromagnetic (EM) sensors mounted on ships, surface vehicles, helicopters, or aircraft can estimate thickness. Surveys using new airborne and satellite lidar methods can now provide thickness estimates for large regions, although uncertainties, particularly in snow cover, limit accuracy. More recently, near-surface observations have been taken from autonomous underwater vehicles and from unmanned aerial vehicles, although their small size limits the type of sensor that can be carried. Manned Russian 'North Pole drifting stations' have operated on the Arctic sea ice from 1937 to 1991 and since 2003. Mass balance buoys are now being deployed to obtain ice and snow thickness and internal temperatures.

Satellite-borne visible and infrared sensors provide observations of surface albedo and temperature, snow cover, radiative fluxes, ice extent, concentration, motion, and melt, but coverage is limited due to frequent cloud cover, and for visible imagery, sunlight. Satellite passive microwave radiometers provide a 30-year plus record, with near-complete daily coverage, of the polar regions under all sky conditions for concentration and extent, motion, and melt. Unfortunately, the spatial resolution of these products (6 to 25 km) makes it impossible to obtain detailed information on the ice cover, such as deformation, melt-pond and lead formation, and ridging. Synthetic aperture radar (SAR) instruments provide high-resolution information on deformation, leads, ridging, and new ice production, which is important for marine transportation.

Scatterometry provides information at a similar spatial scale to passive microwave data, but can provide better information on perennial ice cover and complementary information on other properties (e.g., melt and snow). Spaceborne laser and radar altimeters can provide indirect estimates of ice thickness,

which is useful for navigation and for modeling ice-atmosphere fluxes. Accurate estimates of ice thickness from altimeters require good knowledge of snow depth and density. More *in situ* measurements of snow depth are needed for better estimates of ice thickness (Chapter 9, Section 9.1.5). The potential to derive snow depth on floating ice by passive microwave remote sensing has been demonstrated by Markus and Cavalieri (2000), but it is only applicable over first-year ice regimes and requires further research. Other satellite-based products for estimating ice thickness and age are currently being developed.

There is no single accepted passive microwave ice concentration product[2], leading to confusion among users. Rigorous evaluation and consolidation of products are needed, together with formal estimates of algorithm uncertainties and errors. This should include improved inter-sensor calibration with longer periods of overlap, rigorous evaluation and intercomparison of current algorithms, and the development of data fusion methods to obtain optimal combined products. Sea-ice observations are not well suited for model assimilation because their error structures are not well known at the grid cell or pixel level.

11.5.3. The Greenland Ice Sheet

The Greenland Ice Sheet has been studied extensively in an effort to quantify the surface mass balance and to estimate its contribution to global sea level. However, *in situ* observations are limited due to the large area and remoteness of the ice sheet. The notable exception is the record from the Greenland Climate Network (GC-Net) of automatic weather stations, operating since 1995 and distributed across the ice sheet. Scientists must therefore rely on remotely sensed observations and estimations from downscaled general circulation models (GCMs) or high-resolution regional climate models (RCMs) to investigate surface mass balance. However, systematic biases in the models relative to the *in situ* record illustrate that the climate of the ice sheet is more or less decoupled from the coastal regions where most of the direct observations that feed the models are made.

Satellites have triggered a quantum leap in observations of

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[2] The term 'product' as used here does not imply an extensive level of validation nor the adherence to any particular standard. The products may be research or operational datasets, but all have undergone at least a basic level of validation.

the ice sheets (Chapter 8). Images acquired by reconnaissance satellites starting in 1962 provide a treasure of historical data for gauging subsequent changes in the positions of shear margins, grounding lines and glacier terminuses (Kim et al., 2001; Zhou and Jezek, 2002). Passive and active microwave observations are used to measure the onset, duration, and extent of ice sheet surface melt. Passive microwave data have also been used to estimate average annual accumulation (Zwally and Giovinetto, 1995). Satellite altimeters are invaluable for measuring ice sheet elevation and also provide information on subglacial water movement (Wingham et al., 2006; Fricker et al., 2007). SAR interferometry (InSAR) provides a powerful means to investigate glacier flow. Satellite gravity instruments provide measurements of ice sheet mass change (Chen et al., 2006; Velicogna, 2009).

In situ measurements are progressing, although at a slow pace due to resource-demanding fieldwork (Chapter 8). Snow pits are used to measure near-surface temperature, snow and firn density, crystalline structure, surface accumulation, and when coupled with GPS measurements, surface velocity. *In situ* seismological experiments are used to measure ice thickness, near-surface density, and the properties of the glacier bed. Surface and airborne radars are used to measure ice thickness, although spatial coverage is limited and there are still large gaps in the ice thickness maps. To efficiently fill these gaps, fixed-wing aircraft, unmanned aerial vehicles, and satellite implementations of ice sounding synthetic aperture radars should be explored. A combined effort in advancing process theory, field methods, coverage, and modeling efforts could facilitate major advances in understanding ice sheet mass balance and help create less uncertain future scenarios of ice mass loss and sea-level rise.

Ice sheet models are insensitive to ocean forcing, in contrast to recent observations of significant velocity variations in marine-based outlet glaciers, which were most likely triggered by ice-ocean interactions and thinning at the grounding line. Gaps in knowledge are mainly related to physical processes of fast-flowing outlet glaciers and ice streams, bed topography, and melting rates under floating glacier tongues. Melt rates on the submarine parts of terminal ice cliffs are also poorly known and often ignored.

Compared with processes that take place at the base of the ice sheet, those that influence the surface mass balance are relatively well understood and considerably easier to observe. Despite this, reconstructions of the individual components that make up the surface mass balance differ significantly. The differences can be as large as the inferred increase in mass loss due to changes in ice dynamics over the past decade. The uncertainties in the surface mass balance are largely a result of the paucity of spatially extensive *in situ* observations. Reducing uncertainties will require a concerted effort to collect more targeted *in situ* data, especially from the percolation and ablation zone. Improvements in process understanding and modeling of key processes such as refreezing, blowing snow, and subgrid-scale effects will also be needed. Meteorological measurements from automatic weather station networks are the key means of estimating near-surface fluxes. These measurements should be maintained and spatially extended to key regions on the Greenland Ice Sheet.

Instead of reducing uncertainty about changes in the mass balance of the Greenland Ice Sheet, new observational methods from space have broadened the range of estimates, revealing that the uncertainties of methods may be larger than previously thought. They also point to the incomplete understanding of the fundamental dynamics of calving glaciers and land-terminating outlets of large ice sheets. Most changes appear in the marginal regions of the ice sheet where data are scarce. Many climate models do not resolve the steep topography of the Greenland Ice Sheet margins, limiting skill in simulating orographic precipitation and ice sheet ablation. New progress in snow physics theory and parameterizations needs to be incorporated into climate models (Chapter 8).

Climate models generally do not include ice sheets and glaciers, which limits their use for projecting sea-level rise. Particular challenges in modeling sea-level rise are the coupling of ice sheet and ice bed and of ice sheet, ice shelf, and ocean. Key uncertainties in predicting Greenland's contribution to sea-level rise are ice dynamics and surface mass balance. The response to warming remains a large uncertainty.

11.5.4. **Terrestrial snow**

Surface-based methods of measuring snow properties include graduated probes, acoustic depth sounders, coring devices, snow pillows (pressure transducers), and excavation of snow pits with detailed sampling. Worldwide, many surface-based snow-observation networks have diminished or have been completely lost. The number of Arctic stations reporting snow depth increased between 1960 and the 1990s, but decreased in many areas over the past decade. The most common observation is snow depth. Far fewer snow-water equivalent (SWE) observations are available, and observations of other snow parameters (e.g., layer snow temperature, layer boundaries, hardness, grain form, grain size, density, strength, and stability) are limited. A coordinated plan for surface-based snow observation networks must be developed, first at the national, then at the international level. Although there are regional snow cover products (Chapter 4, Appendix 4.1), at present there is no pan-Arctic dataset of *in situ* snow measurements, and there is no single archive that has all snow depth data. Data are particularly sparse for the mountain and tundra regions.

Observations of snow extent from visible, near-infrared, and microwave satellite sensors have been widely used since 1966. Space-based capabilities for observing snow depth and SWE are more limited. Microwave sensors appear ideal for this purpose, but current sensors lack optimal combinations of frequencies and spatial resolutions, particularly for mountainous terrain. Long-term SWE data have been derived from passive microwave sensors, although comparative analyses of basin-wide SWE and winter precipitation suggest uncertainties in the determination of snowfall amounts and SWE over large watersheds and regions with different physical characteristics (Yang et al., 2009). SAR observations overcome the resolution problem, but current sensors operate at frequencies too low to be useful for most snowpacks. Priority should be given to research and the development of algorithms and new sensors to measure SWE (high frequency SAR), under a wide range of vegetation conditions.

Energy and mass-balance snow models, now widely used for many applications, are highly nonlinear and include several

interacting variables. These models are under-constrained by mass-oriented observations of SWE or snow depth alone. Snow temperature observations are necessary to constrain the energy states of these models. Snow observation sites are often not representative of surrounding areas. This results in observation biases that can be easily propagated to large regions through modeling and data assimilation. Such biases are common and can be quite large.

Further work is needed to improve models of snow-vegetation interactions. There are dramatic differences in snow model response if vegetation is included. One of the problems is that parameterization of key processes such as snow unloading are either non-existent or based on site-specific data. Integrated multi-sensor data fusion and global analysis systems that blend snow observations from all sources must be improved. The ideal global snow observing system will use observations from all relevant sources in coherent, consistent high-resolution analyses of snow extent, depth, SWE, wetness, and albedo. No current system provides global coverage.

11.5.5. Permafrost and seasonally frozen ground

The Global Terrestrial Network for Permafrost (GTN-P) identifies permafrost thermal state (i.e., ground temperature) and the active layer thickness as key variables for monitoring (WMO, 1997). Many permafrost temperature records are of short duration and discontinuous, although some sites have 20- to 40-year time series. Currently about 400 boreholes are listed in the GTN-P database; around half of these are in Arctic countries (Figure 11.14). There are currently over 100 Arctic sites that report seasonal thaw, soil temperatures, moisture and frost heave, and subsidence.

No formal international network exists for seasonally frozen soil temperature or moisture observations, although several programs make some data available. The Arctic Coastal Dynamics (ACD) program began in 1999, where a network of 25 key sites was identified along the coastline of the entire Arctic Basin (Rachold et al., 2005). Some national networks are extensive. There were more than 800 stations in the former Soviet Union where soil temperature and soil freezing depth were measured on a daily basis (Chudinova et al., 2006), but the number of stations has declined significantly in the past two decades. Canada and the United States have many stations where soil temperature is measured, although not all are Arctic.

Many of the GTN-P sites are not maintained for long-term monitoring. There are significant thematic and regional gaps in the present networks, especially in eastern and central Canada, most ice-free areas in Greenland, and north-central and northeastern Russia. Many sites lack instrumentation for collecting the ancillary data required to analyze the linkages between permafrost and climate. Existing borehole and active layer networks must be expanded. Soil temperature and frost depth measurements should be standard parameters for all cold-region meteorological stations.

Properties of permafrost terrain are currently not directly detected from remote sensing platforms. Zhang et al. (2004) and Duguay et al. (2005) provided a comprehensive overview of satellite remote sensing of permafrost and seasonally frozen ground. Surface indicators of permafrost terrains that are discernable by remote sensing include pingos, thaw lakes and

Figure 11.14. Soil temperature stations and sites in the Northern Hemisphere. Blue and green symbols are sites for which data are available. Red symbols indicate sites for which data are not available. Source: adapted from IGOS (2007).

basins, thaw slumps, thermo-erosional valleys, thermokarst mounds, ice wedge polygons, beaded drainage, slope failures, and rock glaciers. SAR data have been used to map zones of wintertime heat loss that indirectly approximate the distribution of permafrost (Granberg, 1994). Scatterometer data have been used to monitor freeze and thaw cycles (Bartsch et al., 2007). Data from passive microwave sensors can be used to detect surface-soil freeze or thaw status (Zhang and Armstrong, 2001) as well as soil moisture. The application of multi-temporal, basin-scale gravity data for the detection of mass loss from ground ice melting in lowland permafrost regions should be evaluated.

There is a wealth of historical observations extending back in time for 50 to 100 or more years. For example, the former Soviet Union has some sites with daily observations extending back to the 19th century. However, many of the historical national databases lack adequate metadata. While data rescue is labor intensive and can be expensive, it would provide a valuable historical perspective, putting cryosphere-climate interactions into the current context of rapid change (Chapter 5).

There are considerable uncertainties in modeling future permafrost distribution and dynamics. These include an under-representation of the ice content and the organic layer and its importance in insulating permafrost during climatic warming. Permafrost models also fail to adequately represent the disequilibrium that has arisen because some current permafrost is related to past climates. This results in a lag period between a climatic change and a response of the permafrost.

11.5.6. Glaciers and ice caps

For glaciers and ice caps, key uncertainties are the incomplete knowledge of their total mass and the regional variations in mass balance. New datasets on ice thickness and glacier outlines are required to improve estimates of glacier melt. The World Glacier Monitoring Service (WGMS) combines *in situ* measurements of mass balance and length change with global information from remote sensing data (Haeberli, 2006). Systematic compilation of inventories at regional or national scales started in the 1940s based on aerial photography. Of around 160 000 glaciers, about 44% (71 000) are currently

stored in the World Glacier Inventory (WGI) (WGMS, 1989). Many regions are still not included in the WGI although data have been compiled.

High-resolution multispectral optical sensors are the most efficient means for glacier mapping. Repeat inventories are needed at five- to ten-year intervals for global change studies and assessing change of water resources. The Global Land Ice Measurements from Space (GLIMS) Project (Bishop et al., 2004; Kargel et al., 2005) will establish a digital baseline inventory of ice extent during the period 2000 to 2005 for comparison with inventories created at earlier and later times. This will in turn lead to improved information on the global extent of glacier retreats and advances. Enhanced resolution data from satellite scatterometers are used to generate products such as melt onset and freeze-up dates, melt duration, and inferred facies distributions for ice caps and large glaciers. InSAR can be used to map glacier surface topography and provide maps of surface motion. Speckle tracking with visible imagery is also important for motion estimation and is more widespread than InSAR approaches, as interpretation of interferometric velocities is not straightforward. The use of satellite gravity measurements for estimates of mass change for large glaciated regions should be expanded.

Continuous annual mass balance measurements made over relatively long periods from field measurements of accumulation and ablation are only available for about 50 glaciers worldwide (Dyurgerov and Meier, 1997). Many glacier regions have no mass balance observations, which biases the network of mass balance sites strongly toward those glaciers that are easily accessible, particularly smaller glaciers in Europe, Scandinavia, and North America. There are issues with measurement methods being used: such as whether data for winter and summer balances are available or only the annual balance, whether processes of internal accumulation are taken into account in the calculation of annual balance, and whether changes in glacier area over the period of record are taken into account in estimating the annual glacier-wide balance (Chapter 7).

For the most part, ice thickness measurements come from surface-based radars, as borehole studies of ice deformation properties and of subglacial till are rare. However, some of the ice masses of the Arctic are too large to survey with surface radars, so airborne surveys are important. Ice coring in the cold firn of glaciers and ice caps is needed to extend the climate archive obtained by coring on the ice sheets (Chapter 7).

The glacier topography database is fragmentary and of poor quality. Space-based data are needed to improve it. A dedicated InSAR mission for precise mapping of glacier topography is a high priority for determining the evolution of changes. A satellite mission providing information on accumulation should be implemented. Repeat laser altimetry surveying should ideally be extended to all major glaciated regions of the Arctic.

Models that link meteorology to glacier mass balance and dynamic response need to be improved. Downscaling techniques need to be developed for feeding such models with GCM data. Remote sensing data are needed to initialize and validate these models. Water management tools for glacier runoff will also benefit from these developments.

11.5.7. **Freshwater ice**

Freeze-up and break-up dates from lakes and rivers have traditionally been determined from surface observations. Unfortunately, surface-based lake and river ice observations have been steadily declining since the 1980s to a point where networks have almost disappeared in many countries (Figure 11.15). In Canada, freeze-up and break-up observations were reported for only 12 water bodies across the entire country during the 2000/01 season. Some lake and river ice observations from various countries have been compiled into the Global Lake and River Ice Phenology Database

Figure 11.15. Lake ice observations in 1970 (upper) and 1995 (lower). Courtesy of C. Duguay, University of Waterloo.

(GLRID). The database contains records for 748 sites, but the number of sites reporting ice observations has plummeted since the 1980s. Financial cutbacks and the automation of meteorological stations in the vicinity of lake and river ice observation sites are the two main reasons for the drastic decline in the surface-based networks. A set of target regions and lakes and rivers must be identified for future long-term ice monitoring. A major data rescue effort must be undertaken (Chapter 6).

The Canadian Ice Service (CIS) began operational, weekly monitoring of ice extent on large lakes in 1995 using visible/ IR and SAR satellite data. The program started with 34 lakes in 1995 and now has 136 lakes, mostly in Canada with a few in the United States. It is possible to derive dates of complete freeze-over and when water is clear of ice with an accuracy of ± 1 week using this dataset. In 1997, NOAA began generating a daily snow and ice product with the Interactive Multisensor Snow and Ice Mapping System (IMS). The IMS incorporates a wide variety of satellite imagery as well as derived mapped products and surface observations. The coarse resolution of the 24-km product allowed for mapping of ice extent only on the largest lakes of the Northern Hemisphere, but a 4-km resolution product became available starting in 2004.

Lake-ice thickness measurements have primarily been through field observation programs. Like the freeze-up and break-up network, ice thickness networks are sparse, and the number of sites has been steadily declining since the 1980s (Lenormand et al., 2002). Ice thickness has been estimated with some success using numerical ice growth models (see Duguay et al., 2003) and through the synergistic use of optical and SAR data on shallow Arctic lakes (Duguay and Lafleur, 2003). Passive microwave data offer a promising means to obtain lake-ice thickness from very large lakes. It might also be possible to obtain the total thickness of snow and ice on medium to large lakes from snow surface elevation data acquired by spaceborne lasers (Jeffries et al., 2005).

Knowing which lakes and rivers freeze to their bottom (grounded ice) or not (floating ice), and when this occurs during the course of the winter period is critical for mapping and monitoring water availability and fish over-wintering

habitats and for planning winter ice roads. The potential of SAR imagery has been demonstrated for such purposes (Jeffries et al., 2005; Duguay and Lafleur, 2003). Surface measurements of parameters such as the dates of the first appearance of ice, complete freeze-over, beginning of thaw, and when the water body becomes completely free of ice (which are important for transportation) are not currently available.

11.5.8. Solid precipitation

Methods of observing solid precipitation include the precipitation gauge network, satellite remote sensing, and ground radar. Precipitation gauge networks and data are long term and fundamental; they make it possible to define global snowfall and climate regimes as well as changes. Manual and automatic precipitation gauges can measure the water equivalent of snowfall but not snow particle size. Manual gauges are standard at most national networks, although the use of automated systems is growing rapidly. Snow rulers are also used for snowfall observations in national or regional networks. They provide snow depth information only, not SWE. As important as gauge measurements are, quality control can be difficult (Figure 11.16).

Estimating solid precipitation amounts and rates from space is a challenge. The use of satellite data for solid precipitation measurements is limited and is an active area of research. Experimental products are available from passive microwave sounders (Kongoli et al., 2003) and cloud radar (Hiley et al., 2010). Various satellite data have been merged to improve the spatial and temporal coverage of precipitation products (Huffman et al., 1997). The future precipitation radar of the Global Precipitation Mission will observe solid precipitation, although its utility for high latitude snowfall is not clear. Of course, remote sensing of snow-cover extent and SWE data provide information on snowfall and have been widely used for large-scale climate and hydrology analyses.

Satellite retrievals have been blended with gauge data. For instance, the Global Precipitation Climatology Project has provided global coverage since 1987. Examination of hydrological budgets over several global river basins revealed

Figure 11.16. Conditions can make quality control challenging for solid precipitation measurement in the Arctic. Photos courtesy of B. Goodison, World Meteorological Organization.

unrealistically low precipitation in the merged products (Fekete et al., 2004). Over land, the magnitude of these combined products is primarily dominated by gauge observations. The under-estimation of land precipitation in the gauge-based analyses, and thus the merged analyses, is mostly due to the combined effect of a sparse gauge network biased toward low elevations, a lack of consideration of orographic effects, and gauge undercatch of precipitation, especially snowfall in cold and windy conditions.

Although ground-based radar provides information about precipitation at high spatial and temporal resolution, the coverage it provides is limited – much less extensive than the gauge networks. The radar network is particularly sparse and expensive to operate at high latitudes. In complex terrain, mountains often block the radar beam or the radar is located to scan over the top of mountains and not in the valleys. A new innovation is the deployment of a network of redundant low-cost, low-maintenance radars to scan the low levels of the atmosphere.

Problems in current precipitation gauge networks include the sparseness and decline of precipitation observation networks in cold regions, uneven distribution of sites and a resulting bias toward coastal and low-elevation areas, spatial and temporal discontinuities of precipitation measurements induced by changes in observation methods (such as

automation) and by different observation techniques used across national borders, biases in gauge measurements (such as wind-induced undercatch), underestimation of trace and low amounts of precipitation, and blowing snow in the gauges. There is no operational precipitation network over the oceans.

11.5.9. Summary

While there are numerous snow and ice surface measurement sites across the Arctic, operations at existing stations are, in general, not well coordinated. There is a need to improve the coordination of resources provided by national and international agencies responsible for cryospheric observations, and to facilitate the transition of research-based products into sustained monitoring systems. There is also a need to standardize the types and methods of measurements at surface stations, so that a robust and consistent set of snow and ice properties is available across the Arctic.

The satellite observing system for the cryosphere is robust, and missions planned for the next 10 to 20 years will provide even greater capabilities. Figure 11.17 shows the timeline of satellites that are used for monitoring the cryosphere. However, some potential gaps will be detrimental to long-term monitoring. In particular, the current gap in laser altimetry and the potential near-future gap in gravity measurements will impact ice sheet and glacier monitoring and change assessment.
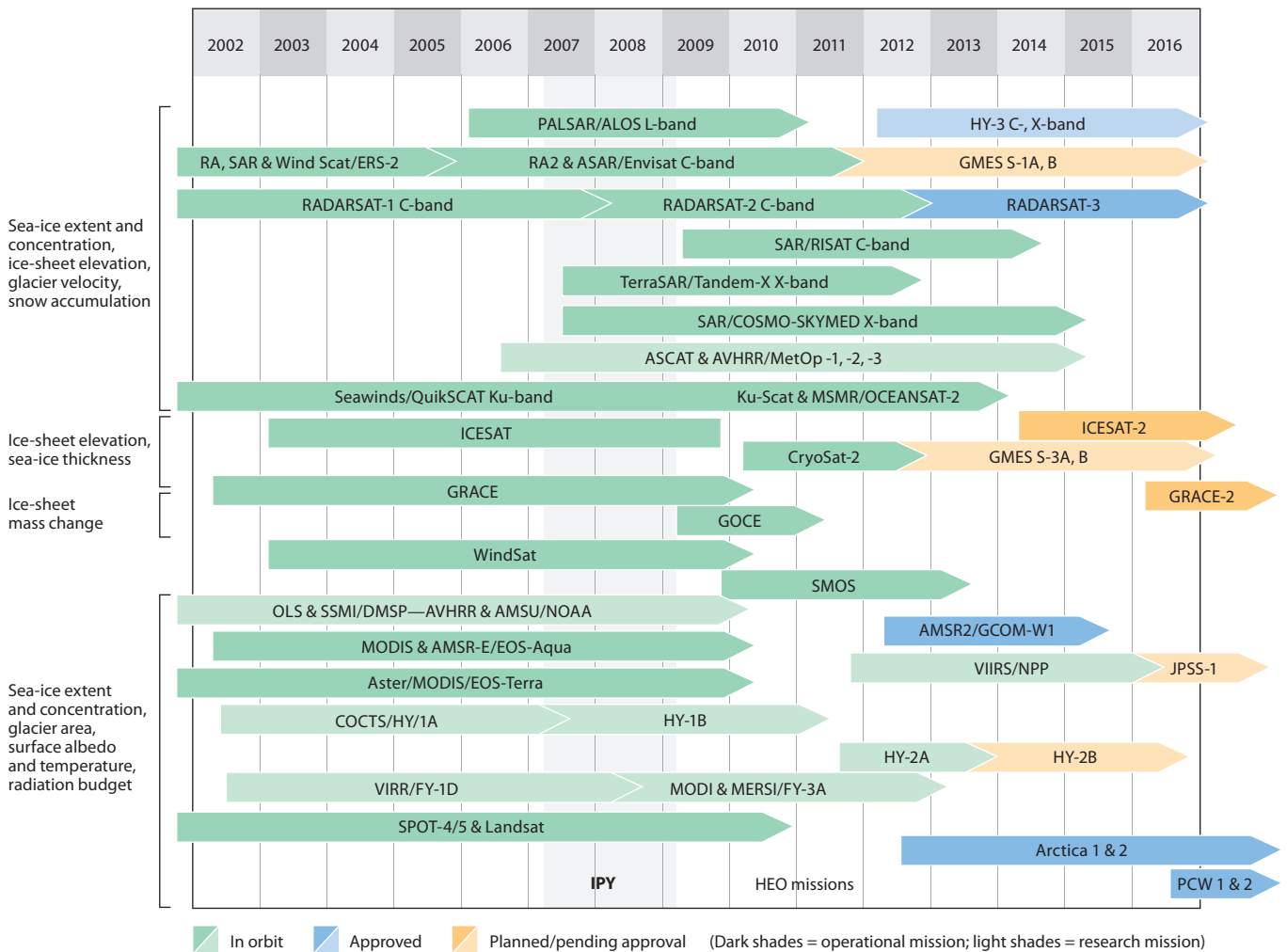

Figure 11.17. Timeline of satellites used for measuring and monitoring the cryosphere as of 1 November 2011. Source: adapted from a figure by M. Drinkwater, European Space Agency.

Additionally, there are some critical parameters that are difficult to measure from space, notably sea-ice thickness, snow water equivalent, and accumulation on glaciers and ice sheets.

Traditional knowledge can provide important value-added content to data products and information and serves to make the data more useful and relevant to northern users (Chapter 10). Community-based observing programs provide a mechanism for two-way knowledge transfer, such as the Gwich'in First Nation International Polar Year (IPY) environmental monitoring project that includes snow cover monitoring along with a wide range of environmental, wildlife, and cultural variables. The ELOKA (Exchange for Local Observations and Knowledge of the Arctic) project will facilitate the collection, preservation, exchange, and use of local observations and knowledge of the Arctic and will foster collaboration between local and international researchers. Similarly, Sámi and scientists in Scandinavia are uniting traditional knowledge and scientific information on snow properties and monitoring (Chapter 4). Traditional knowledge can provide important supplementary information for the assessment of changes in the Arctic cryosphere.

While there are gaps and deficiencies in the cryosphere observing system, the data that do exist are underutilized. The use of cryospheric products in weather and climate models needs to be extended. Surface albedo is a good example, as many models use rudimentary and unrealistic albedo schemes for snow and ice. Data assimilation schemes for snow and ice variables are generally underdeveloped. Sea-ice extent and concentration data are assimilated at several real-time modeling centers, but ice thickness and motion data are not. Reprocessing of data products should be organized and promoted by national and international agencies and organizations.

Acknowledgments

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12. SWIPA Synthesis: Implications of Findings

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Contents

12.1. Introduction and scope of chapter

Climate change is causing significant changes in the Arctic cryosphere, and these changes are expected to continue (Chapters 2 and 3). As reported in Chapters 4 to 9, these changes involve the timing and duration of snow and ice cover and seasonally frozen periods. They may also involve the loss or severe degradation of specific elements of the cryosphere. Observed and projected changes in the cryosphere cross the full spectrum of temporal scales (i.e., seasonal to century levels and beyond). The spatial scales involved similarly range from local to pan-Arctic and global. The changes are having, and will continue to have profound impacts on Arctic environments and human activities in the Arctic (Chapters 2, 10 and 11). Within this general context, four issues are especially important: the nature and magnitude of the changes, their timing relative to the seasonal/annual cycle, the spatial and temporal scales on which they occur, and their consequences for physical and biotic systems. This chapter provides a synthesis of the cryospheric changes documented in the preceding chapters and evaluates their combined impact on the Arctic environment and its inhabitants.

12.2. Ongoing and projected changes in the cryosphere

The increase in annual average temperature over the Arctic since 1980 is twice that over the rest of the world (with the exception of some regions of Antarctica), almost certainly due to polar amplification processes (see Chapter 1, Box 1.3). This has had considerable consequences for the cryosphere, most of which in the Northern Hemisphere is in the Arctic. It is now evident that feedbacks between the climate system and the cryosphere and between the different elements of the cryosphere are significant, and that the changes may be greater overall than projected by either the Arctic Climate Impact Assessment (ACIA) in 2005 or the Intergovernmental Panel on Climate Change (IPCC) in 2007. The current trends in cryospheric change are projected to continue within the Arctic (additional warming of 2 to 7 °C by 2100), with consequences of increasing significance expected for biological and human systems. The following sections address first-order changes (i.e., those in the cryospheric components themselves), feedbacks, and the second and higher-order changes induced by the cryospheric changes (i.e., changes in Arctic ecosystems and in the services these provide to humans).

12.2.1. Changes in the cryosphere

Substantial changes are expected throughout the Arctic cryosphere, driven by warming and in part by feedbacks among the cryospheric components and between the cryosphere and other systems.

Past/Present Climate (Chapter 2): ACIA projections of greater and more rapid warming in the Arctic than in other parts of the world are confirmed for the immediate past, particularly in autumn and spring. Sea-ice reduction is coupled with increased warming – this may represent a large shift in the Arctic climate system as evidenced by the greatest warming being over the Arctic Ocean (rather than on land as previously observed).

Model Projections (Chapter 3): General circulation model (GCM) projections based on scenarios of anthropogenic greenhouse gas (GHG) emissions indicate that the Arctic will continue to warm at a greater rate than other regions of the world over the next several decades. These projections may be regarded as 'best estimates' of future changes, but are subject to uncertainties arising from (i) natural variations in the climate system, (ii) the range of plausible trajectories of GHG concentrations, aerosols and other climate drivers over the next century, and (iii) systematic errors arising from model formulations, particularly the parameterizations of unresolved processes. The net effect of these various factors is a range of uncertainty that can be comparable to the projected change, although the use of ensembles of simulations from multiple models can help to extract the 'signal' of climate change. It must be kept in mind, however, that natural variability can result in occasional short-term trends of cooling during long-term trends of warming.

The overall average increase in surface air temperatures in the Arctic is projected to be 3 to 6 °C in autumn and winter by 2080 (6 to 7 °C in areas of sea-ice loss and 2 to 3 °C over land), relative to a 1957–2006 baseline. Precipitation is projected to increase throughout the Arctic with the greatest increases at the highest latitudes (a 5–40% increase by the end of the century for the IPCC B1 emissions scenario). The projected changes are greatest in winter (e.g., as snowfall) and autumn, and least in summer. Despite overall increased precipitation in the Arctic, some areas may experience drying of the landscapes due, in part, to increased evaporation associated with higher summer temperatures, a longer summer season, and perhaps altered drainage due to permafrost changes in some areas.

Snow (Chapter 4): Snow (structure, quantity, timing, extent, duration) is a fundamental attribute of northern environments. Snow-pack structure is changing, with winter thaws and rain-on-snow events increasing in frequency and with more ice crustings. Satellite imagery indicates that average snow cover extent decreased in summer (May–June) by 18% between 1966 and 2008. Annual snow-cover duration across the terrestrial Arctic (excluding Greenland) decreased by 3.9 days per decade, mostly due to the earlier disappearance of snow in spring (averaging 3.4 days per decade earlier for 1972 to 2009) with decreases projected to continue (by 10–20% over most of the Arctic by 2050). Some of this decline is likely to be due to albedo feedbacks (see Section 12.2.2). Projections suggest a slight increase in snow water equivalent (0–15%) over most of the Arctic by 2100 with high regional variation. In those areas with projected decreases in snow-cover duration and snow-water equivalent, the earliest and largest changes are projected to occur over coastal regions (as has been observed in the recent past).

Permafrost (Chapter 5): Permafrost temperatures have increased throughout the circumpolar area but there is high regional variation in the rates and magnitudes of warming. The thickness of the active layer (the surface layer above permafrost that thaws in summer) has progressively increased over much of Eurasia and central Alaska, but has remained relatively stable in northwestern Arctic Canada and the Alaskan North Slope. Over the past few decades, permafrost has disappeared from several lower latitude sites formerly characterized by discontinuous

permafrost. Projections indicate that by 2100 the area of near-surface permafrost will be reduced by 16–20% in Canada with an additional 9–22% projected to contain taliks (i.e., areas of unfrozen ground) embedded within the permafrost. There will be widespread permafrost degradation over up to 57% of Alaska (850 000 km^2). Eastern areas of Siberia, which are particularly susceptible to climate warming, are showing similar widespread changes that are likely to continue. The consequences of these changes for landforms, ecosystems and infrastructure – such as melt water drainage, soil subsidence, slumping to lakes, and high silt loads in rivers – are likely to be profound (see Section 12.3).

River and Lake Ice (Chapter 6): The seasonal duration of river- and lake-ice cover is declining, primarily due to the earlier occurrence of spring break-ups. Some of the most profound changes have been observed on far-north lakes in recent decades. Models suggest that lake-ice duration on mid-sized lakes in the Northern Hemisphere could be 15 to 50 days shorter by around mid-century (2040–2079) compared to 1960–1999, mainly due to ice break-up occurring 10 to 30 days earlier. Maximum lake-ice thickness has also been modeled to decrease, by 10 to 50 cm. Changes in ice composition reflecting spatial patterns in ice-thickness reduction and snow-cover accumulation are also anticipated. Reduced temperature gradients along large northward-flowing rivers will favor a greater proportion of break-ups by thermal processes rather than mechanical (dynamic) processes, and thus reduced severity of ice-jam flooding. Mid-winter break-ups and associated flooding will become increasingly common in northerly areas. The nature of break-ups will also be affected by changes in the form (rain vs snow) and magnitude of precipitation.

Mountain Glaciers and Ice Caps (Chapter 7): General declines have been observed in mass (and volume) of mountain glaciers and ice caps over the past century, however, the rates of decline appear to have increased substantially since 1995. Mass losses from mountain glaciers and ice caps in the Arctic may have exceeded 150 Gt[1] per year over the past decade, similar to the mass loss from the Greenland Ice Sheet. Half of the observed loss is from glaciers in the Canadian Arctic and southern Alaska. Future projections indicate continuing losses due to warming, with the greatest losses projected for the Canadian Arctic and Alaska, followed by Svalbard and the Russian Arctic. Under the IPCC A1B emissions scenario, the total volume of Arctic glaciers is projected to decline by 13–36% by 2100, corresponding to a rise in sea level of 51–136 mm (but this estimate does not include losses by iceberg calving).

Greenland Ice Sheet (Chapter 8): Up until 1990, the total amount of ice added to and lost each year from the Greenland Ice Sheet was roughly in balance. Mass loss by both meltwater and ice discharge have increased since 1990 and the inputs and losses are no longer in balance. The estimated rate of net mass loss from the Greenland Ice Sheet over the recent past is four times that of earlier estimates (i.e., 205 ± 50 Gt per year for 2005–2006 compared to 50 ± 50 Gt per year for 1995–2000, of which roughly half is due to melting and half due to discharge).

Increased ice discharge is primarily due to accelerated flow in many of the fast-flowing glaciers that terminate on ocean water in the fjords. This speeding up of the ocean-terminating glaciers, a phenomenon that has spread northward along the west coast of Greenland over the past 10 to 15 years, appears to have been triggered by contact with warmer ocean water.

Sea ice (Chapter 9): The average extent of the summer (September) minimum sea-ice cover has declined by 25–30% since estimates based on satellite information began in 1979. The rate of decline increased over the past ten years and is now greater than reported by ACIA. Complete summer sea-ice loss is projected as likely to occur by mid-century. Variability in summer sea-ice extent is projected to increase initially, but then to decrease as ice-free summers are approached. Sea ice is generally thinning and older ice types (5+ years old) are being lost with increasing frequency; first-year ice is projected to dominate in the future. Shifts in the proportion and areal extent of multi-year ice versus first-year ice have immediate physical consequences including a larger area of open ocean during summer and autumn and greater heat absorption. Loss of summer sea ice and increased heat storage in the upper ocean are modifying regional wind patterns with increasing potential for impacts in mid-latitudes as sea ice retreats further in the coming decades. However, trends toward increased cloudiness have suppressed surface warming of the Arctic Ocean to some degree. Projections of annual maximum extent and duration of seasonal sea-ice cover are uncertain and regionally variable (and affected by local bathymetry and weather conditions), but maximum extent and duration are very likely to be less than at present (i.e., with a more northerly distribution, smaller area, shorter duration). Corroboration of local changes, especially for land-fast ice through traditional knowledge and local observations, support these findings (i.e., later freeze-up, earlier break-up, longer ice-free summer, shifted ice dynamics and types). The extent and location of summer/autumn polar pack (i.e., multi-year sea-ice remnants) will be uncertain in future years as the summer ice decline continues, but most models project that pack-ice remnants are very likely to occur along the northwestern margin of the Canadian Archipelago and the north coast of Greenland at least until complete loss.

12.2.2. Feedbacks

Arctic climate change results from complex interactions between the atmosphere, ocean, and individual components of the cryosphere. Many aspects of Arctic climate change are simple responses to a driving force, for example, higher (or lower) air temperatures will alter the ice balance in a particular area. Other changes may involve a feedback whereby a change in one component of the system drives a change in another, which ultimately induces additional change in the original component (see Chapter 11, Section 11.1). Such feedbacks can be positive (i.e., induced change reinforces and exacerbates the original change), whereas others can be negative (i.e., induced change dampens, cancels or reverses the original change)[2]. Feedbacks are important because they may alter rates of change, magnitudes of change, or even directions of change. Owing to

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[1] 1 Gt = 1 gigatonne = 1 billion tonnes, which is slightly greater than 1.1 cubic kilometres of water.

[2] Note – positive and negative here refer to the feedback itself and not the warming or cooling effects in the cryosphere context.

their unpredictable effects and their variable scales (spatial and temporal) of operation, feedbacks also add to the uncertainties of outcomes especially for higher-order consequences of climate and cryospheric change. Appropriate parameterization of processes and their interactions, and thus feedback processes, may be inadequately represented in climate models. This affects projections of future states of the climate system.

Two aspects of feedback are important in the present context: feedbacks between the cryosphere and other climate system components, and feedbacks between different cryospheric elements. Examples are discussed below. The list of major feedbacks is much longer than given here, however, and many remain poorly understood (see also Chapter 11, Section 11.1). A third level of feedback is that between the cryosphere and human systems. This feedback is extremely important but also highly uncertain – feedbacks at this level are discussed in Sections 12.3 to 12.5.

12.2.2.1. Feedbacks between cryospheric and other climate system components

Potential feedbacks between the cryosphere and climate include:

- Snow has an extremely high albedo (surface reflectivity). Albedo of snow and ice decreases as they melt and metamorphose, and/or as particulate matter (dust, soot, volcanic ash) is deposited onto them, and/or as particulate matter is exposed at the surface as it is released from snow and ice by melt. Increased heating of darker land and sea surfaces will occur as snow amounts decrease in time and space which, in turn, will exacerbate melting. This feedback is partly responsible for the accelerated rates of cryospheric change observed over the recent past, albeit with high uncertainty associated with them. The replacement of sea ice by open water over wide areas of the Arctic Ocean has substantially increased heat absorption by the upper ocean. This appears to have increased the rate of melting of the remaining sea ice, to have altered the Arctic climate system itself (heat storage and wind patterns), and to have accelerated changes in other cryospheric components. Moreover, higher ocean temperatures pre-set conditions for seasonally later sea-ice formation, ultimately resulting in a thinner and generally less consolidated sea-ice cover, and influencing the magnitude and timing of ice loss and seasonal melt the following year. This may be exacerbated by increased snowfall (which would insulate and slow seasonal ice formation); however, some effects may be modified by the increased freshening of the system, as less saline waters are more liable to freeze.

- From a total of 48 interactions (six cryospheric components and eight forcing mechanisms), present evaluation (Chapter 11, Section 11.1) shows that more feedbacks between the cryosphere and Arctic climate change have a warming effect (fourteen) than a cooling effect (eleven), and result in accelerated rates of climate change in the Arctic (i.e., 'polar amplification'). Of the remainder, three are uncertain due to conflicting evidence and 20 represent gaps in knowledge.

- Feedbacks operate at different spatial and temporal scales and so understanding their consequences is difficult and associated with high uncertainty. Additional uncertainty arises because effects from some feedbacks are expected to intensify over time, while others are expected to diminish.

- Spatial connections mean that the effects of cryospheric change within the Arctic will have implications for areas outside the Arctic and these effects may be large and possibly counter-intuitive. For example, sea-ice loss north of Eurasia may lead to colder weather in Europe and Siberia during late autumn and early winter (see Chapter 2, Box 2.1). Reduced summer sea ice, increases ocean heat absorption which is released back as the atmosphere cools in autumn. The lower atmosphere thickness is increased which thus increases pressure, resulting in de-stabilization of the Arctic boundary layer. Winter winds flowing out of the Arctic may therefore be more intense and affect lower latitudes to a greater degree than is typical, resulting in winters of greater severity in Europe and Siberia – a counter-intuitive effect of climate warming.

- The production and release of GHGs from Arctic stores or processes is likely to increase as the cryosphere degrades or changes. Changes in permafrost and lake ice, for example, can enhance the production of methane (which has great capacity to induce warming) and nitrous oxide and increase the seasonal window for their release. However, drying of the active soil layer can cause a methane sink in some cases. Overall, processes such as carbon dioxide fluxes are likely to have very small global effects. Others, such as those that may result in the release of subsea methane, could have large global effects, although in both cases the uncertainty is high.

12.2.2.2. Feedbacks among cryospheric components

Feedbacks among cryospheric components include:

- The occurrence of relatively warm open water near floating ice tongues from the Greenland Ice Sheet appears to have increased mass loss from the ice sheet through increased flow rates to the sea and thus increased calving of icebergs, particularly in western Greenland. Similarly, the degradation and complete loss of semi-permanent land-fast sea ice on the northern coast of Ellesmere Island in the Canadian Arctic appears to have contributed to the break-up of the 3000–6000 year old marginal ice shelves in the area, probably due to their exposure to a warmer and more open-water environment.

- Snow insulates underlying environments from atmospheric temperature extremes. As snow extent and duration decrease, and despite the likely increase in snowfall amounts, ice (lake, river, sea) and permafrost environments will probably both be exposed earlier in the season to generally warmer atmospheric conditions. If so, this will exacerbate spring thaws and increase active-layer depth in permafrost. Consequently, decreased integrity of ice and permafrost can be expected due to the insulating properties of snow.

- In the terrestrial environment, permafrost acts as a key element determining hydrology over great areas of the Arctic. As a result, changes in the permafrost will alter drainage and this could lead either to drying or to increased ponding over significant areas. Drying creates conditions

for wildfires in tundra and forested areas, which will be exacerbated by higher summer temperatures and increased evaporation. Fires may contribute to increased emissions of GHGs and other climate forcers.

- Decreased ice on land, rivers, lakes and the sea will result in greater evaporation of water. The likely effect will be an increase in clouds at mid-levels in the atmosphere. Clouds reflect incoming radiation, which means increased amounts of cloud may result in overall cooling of the atmosphere. Increased evaporation also affects precipitation regimes.

These feedbacks, as well as additional feedbacks noted in other chapters, generally exacerbate the effects of changes in the climate system and the cryosphere. Furthermore, as already noted, feedbacks also increase uncertainty in the timing, rates, nature, spatial scales and outcomes associated with climate change. The relevance of these feedbacks and their interactions with Arctic ecosystems are discussed in the following sections.

12.2.3. Effects of cryospheric changes on other aspects of the Arctic environment

Changes in the physical nature of the cryosphere have consequences for other components of the Arctic environment.

Snow: Changes in snow will have profound effects on terrestrial environments and their soil, plant and animal communities, as well as on aquatic environments influenced by terrestrial runoff (wetlands, rivers, lakes, coastal seas). Increased insulation resulting from greater snowfall may increase winter soil temperatures, thus increasing rates of biological processes, such as decomposition and nutrient release. In turn, where water is not limiting, there may be an overall increase in terrestrial productivity manifested by seasonally earlier, more intense and higher levels of production; as well as shifts in community types (e.g., greater shrubiness). These effects may be offset by earlier disappearance of snow, winter thaws, or rain-on-snow events. Winter-resident herbivores, such as rodents, rely on sub-nivean environments that, if affected by adverse events or if frozen, would lead to decreased population abundance. Follow-on responses by resident predators to declining food sources have effects throughout the entire food web. Similarly, consequences for large herbivores, such as musk-ox and caribou/reindeer that forage through the snow in winter, will reflect the combination of many local factors (e.g., winter mortalities due to limited access to forage, increased mortality due to moulds, possibly increased fecundity due to earlier forage availability in spring), but overall the net effect is likely be negative. Effects on peoples who hunt or herd could be substantive, with a significant decline in food security and traditional activities, particularly during winter. Conversely, increased nutrients and earlier production in terrestrial environments may result in greater local productivity with positive consequences for some animal species and people. Water from snow melt transports nutrients to aquatic systems, possibly enhancing local productivity, and if trapped on the landscape tending to counteract drying regimes.

Snow insulates ice and blocks light penetration, whether the ice is on lakes, rivers or the sea. By insulating the underlying ice, increased snowfall will tend to result in the ice being thinner (although shorter snow durations may offset this to some degree, and increased snow loads on thin ice could promote snow-ice formation). Winter nutrient accumulation in the water column combined with light penetration during early polar spring 'jump start' the seasonal production in freshwater and marine ice-covered ecosystems. Greater snowfall, particularly during polar spring will reduce the light available for photosynthesis by phytoplankton below the ice, although this effect may be offset by longer open water conditions. This shading effect may also impede photosynthesis by sea-ice algal communities, which can represent a large fraction of the total annual primary production, and thereby alter marine food webs.

Permafrost: Permafrost affects the landscape drainage patterns, and is responsible for the vast range of tundra wetlands, ponds and small lakes present in the Arctic. Degradation of permafrost alters the nutrient processes in these systems and in many areas has resulted in the drainage and loss of lakes. Tundra water bodies are ecological and biogeochemical hotspots providing key habitats for biota. They are also significant sources of GHGs to the atmosphere, and permafrost degradation may alter GHG export from such areas. Permafrost degradation in terrestrial areas may also result in slope instabilities and subsequent failure; failures affect the receiving ecosystems (lakes, rivers) and also represent hazards to humans and built environments (roads, towns, industrial sites). High uncertainty surrounds effects of permafrost thaw on soil water content thus subsequent effects on ecosystems and on GHG emissions – projections are uncertain but models suggest that the tundra will continue to act as a weak carbon sink at least for the next 100 years (i.e., net balance throughout the Arctic will be slightly toward continued storage of carbon thus slightly ameliorating effects of GHG production). Up to 65% of the Arctic coastline is composed of unconsolidated materials with considerable ice-rich constituents and is therefore at risk due to permafrost degradation; this is particularly acute near the deltas of most major Arctic rivers. Alteration of coastlines due to slumping will be exacerbated by a decline in sea ice, longer open-water fetches, and increased effects of storms. Continental marine shelves and the nearby low-lying deltas of large northern rivers also hold large reserves of carbon, mostly in the form of methane capped by an impermeable layer of permafrost. Degradation of the permafrost may ultimately result in the abrupt release of large quantities of this potent GHG, representing an uncertain future threat.

River and Lake Ice: Second and higher-order effects of changes in freshwater ice (e.g., on gas exchanges, mixing regimes, contaminant capture, exposure to ultraviolet radiation, and overall biological productivity and diversity) will have numerous implications for the ecosystem health of aquatic systems. The effects of changes in river and lake ice are very likely to be closely coupled with nearby land-form changes due to permafrost thaw. Changes in ice-cover dynamics are likely to decrease the frequency and magnitude of spring ice-jam flooding in many areas of the Arctic, affecting the ecology of deltas, which is especially relevant to the overall production in the large Arctic rivers. Ice-jam flooding produced by mid-winter melt and ice break-up represents an increasing hazard

to nearby communities and infrastructure. Predictions of ice-induced flood conditions, however, are confounded by uncertainties about the intensity and magnitude of future snowmelt runoff that affect the dynamics of river-ice break-up processes.

Mountain Glaciers and Ice Caps, and the Greenland Ice Sheet: The consequences of changes in these cryospheric components include exposure of new land and altered drainage patterns, initial increases in runoff followed by eventual declines as glacier area decreases, and water outbursts of several types if glacier lake dams fail. Increased iceberg formation and continued break-up of ice shelves (until they disappear completely from the Arctic) are also expected, and may create hazards in the marine environment. Also, release of materials stored in ice (contaminants, sediments, fine particulates) may affect downstream water quality (generally negatively). Increased water output during the summer melt period is also expected and may represent a positive benefit in some locations, at least in the near term.

Sea Ice: A progressive decline in the area of multi-year sea ice, its seasonal replacement by younger ice, lengthening of ice-free periods, and increased area of open water have profound implications for the Arctic. These physical changes will alter albedo and increase heating and gas exchange in marine waters. These effects will, in turn, have follow-on effects on the climate system, marine ecosystems and human activities. Some ice-associated ecosystems (such as multi-year sea ice and epishelf lakes) are at risk of complete loss; others will change significantly. Phytoplankton production rates in the marine ecosystem are likely to change, but to a variable extent among sites. In general, these rates are likely to increase as a result of sea ice loss and increased exposure to sunlight, although this effect may be offset to some degree by increased water column stability caused by freshwater inputs to the surface layer. In addition, ice-associated biota are likely to decline, while other biota will become more important. Both types of effect will alter structural and functional relationships within the ecosystems. Altered trophic relationships due to changes in pivotal components (ice algae, cryophilic zooplankton, Arctic cod) will induce effects in those species that prey upon them. As it is the higher-trophic level species (seabirds, marine mammals, predatory fishes) that are the most directly relevant to humans, changes in sea ice will ultimately affect human food and exploitation opportunities in the Arctic. Increased extent and duration of open water in the Arctic Ocean will also feed back to climate and affect coastal environments.

Induced Effects and Ecosystem Overview: Physical changes in the environment resulting from changes in cryospheric components induce follow-on effects, many of which ultimately cascade to ecosystems. The rest of this section reports wide-ranging and over-arching induced effects (for further details see Chapter 11, Sections 11.3 and 11.4).

- *Shifts in Terrestrial and Freshwater Ecosystems and Water Balances*: Greater precipitation overall (regardless of how delivered) primarily as snow, increased glacier and ice-mass shrinkage, shorter ice duration, and warmer winters, will all combine to result in an overall increase in the amount, temporal extent, and availability of liquid water in the Arctic

(at least in the short term). Processes that increase drying (evaporation, drainage) will counterbalance this to a varying degree. High spatial variation is also likely, thus local water balances will be highly variable. Effects that increase water in the landscape will occur in the near future and most will continue at least until mid-century, although some will ultimately decline (such as water flow from smaller glaciers). Assuming warming feedbacks continue, mass loss associated with large glaciers and ice sheets will continue for centuries. As previously noted, these effects will also be modified locally by changes in permafrost, effects from ice, and similar linkages among cryospheric elements, all of which will affect local and regional hydrology. Moreover, higher air temperatures in summer will alter evaporation by increasing drying regimes. A change in the water balance will have significant follow-on effects on terrestrial ecosystems by creating new habitats. Responses in the vegetation may include greater productivity, northward shifts in distribution, increased invasions of pest and disease organisms, local extirpations of some endemic Arctic flora, and increased local biotic diversity but with species composition differing from present-day. Changes in the vegetation and productivity will then enable similar shifts in animal diversity within the terrestrial landscapes. Nutrients and other constituents (silt, contaminants) carried by water will affect freshwater ecosystems with similar consequences. Infilling of habitats due to sedimentation, and slumping and de-watering over large areas of tundra will also affect biota (waterfowl, freshwater fishes) that rely on these biological 'hotspots'. In some areas new aquatic and terrestrial habitats will be created as glaciers and ice sheets decay, and freshwater ice is lost particularly where it currently freezes to the bed or exists as perennial forms. Colonization of these new environments represents a microcosm of wider shifts that are likely to be forthcoming for many terrestrial and freshwater ecosystems in the Arctic.

- *Shifts in Marine Ecosystems*: Sea ice affects and controls element cycling in marine ecosystems (thus regulating primary production), water-mass mixing processes (including upwelling of deeper waters), and light penetration. Sea ice also acts as a physical habitat for some marine mammals and seabirds. As a result, shifts in the presence and types of sea ice present will have far-reaching consequences for marine ecosystems in the Arctic. These are likely to include: increased production in the central Arctic basin accompanied by a switch from light- to nutrient-limited conditions; a shift in primary production (and associated secondary production) from a high proportion of ice-based primary production to much greater plankton-based production, particularly over coastal shelves, with consequences for nearshore benthic production (possibly decreased) and unknown consequences for higher trophic levels such as fishes and marine mammals; and, northward shifts in marginal ice zones associated with highly productive sub-Arctic seas and general ocean warming, which are likely to be accompanied by changes in the distribution of marine biota of direct interest to humans (fishable species as well as new predators such as killer whales). Alternatively, the seasonal timing of sea-ice retreat will affect the availability

of ice-associated zooplankton and lipid-rich forage fish – critical resources for higher trophic levels. Negative follow-on consequences for Arctic-adapted biota may ensue. Mismatches in space and time between sea ice and Arctic ecology will have significant consequences for populations. For example, reproductive success in seabird populations is likely to decline in many locations as land-fast ice edges retreat or degrade earlier and as underlying fish quality changes as species distributions shift. Shifts in biodiversity may also result as sea-ice habitats change or are lost (see Section 12.3.5).

- *Contaminants*: Glacial ice and/or snowpacks, once considered semi-permanent sinks for contaminants will eventually become local sources as melting releases stores of long-lived contaminants to Arctic waterbodies, landscapes and the atmosphere (see Chapter 11, Section 11.3). This will increase the likelihood of these contaminants re-entering biological and human systems. Such contaminants include legacy persistent organic pollutants (POPs) and metals and possibly also radioactive elements. At least over the short term (to mid-century), degradation of multi-year sea ice is likely to alter contaminant dynamics. Shifts in snow characteristics are likely to alter the dynamics of volatile contaminants with greater transfer to the atmosphere and/or water possible. Altered landscape hydrology and drying of some landscapes may also release or expose contaminants stored in freshwater sediments. Increased exposure is likely to increase entrainment in seasonal flows and biota, thus encouraging contaminant re-entry into Arctic ecosystems. Accelerated remobilization of legacy POPs is very likely as the Arctic Ocean becomes more exposed and warmer; moreover, decreased ice is likely to increase this, whether or not winds change in the future. Uncertainties with respect to biotic degradation processes for POPs may alter rates; this is a significant gap in understanding. Altered ice dynamics will alter the mercury cycle in the Arctic Ocean. Enhanced entrainment and transport of contaminants is likely throughout the Arctic Ocean via greater amounts of seasonal first-year sea ice. Shifts in marginal ice zones, the sites of extensive biological production supporting most *in situ* trophic productivity in the Arctic marine system, will induce shifts in lipid pathways and transfers. As a result, contaminants preferentially stored and transferred in lipids are very likely to affect higher trophic levels (Arctic top predators such as seabirds, whales, and polar bears). Changes in foraging behaviors are also likely with altered trophic patterns and loss of key habitats such as sea ice. Follow-on health effects of contaminants on biota, as well as on humans relying on these organisms are possible but details are uncertain at this time. Colonizing and invasive species that are typically highly migratory will increase in the Arctic as thermal and ecological barriers decline; the proximate effect will be to alter contaminant pathways in unknown ways. In addition to the cryospheric linkages outlined above, human activities and the changing climate will also influence contaminants in uncertain ways, for example, increased human access and development may increase the potential for contaminant release.

12.3. Integrated consequences of cryospheric-driven changes on Arctic societies

The lifestyles and activities of humans in the Arctic are very closely associated with the environment; that is, the Arctic in general and the cryosphere in particular exert great and close control over when, what and how activities can occur (Chapter 9, Section 9.4 and Chapter 10). This is true irrespective of whether one lives on the land with a traditional lifestyle and a primarily subsistence-based livelihood, whether livelihood is primarily wage-based and thus life is oriented more toward urban centers, or whether one is a seasonal worker in the North. These forms of livelihood have different associations with the Arctic landscapes and seascapes, and cryospheric changes will affect each in different ways. Also, the nature and consequences of cryospheric change will differ depending on whether the area occupied is primarily inland or coastal. The various sectors from which livelihoods are derived will be exposed to different consequences of cryospheric change. The consequences include a combination of direct effects (access, hazards) and indirect effects (shifted or lost ecosystem services requiring adaptation or replacement), both of which sum to an overall effect. Finally, changes also have non-material consequences as they affect the overall experience, perception, cultural value and well-being of people living in close relation to the Arctic cryosphere.

12.3.1. Changes to access in the Arctic environment

Ice in its many forms represents the solid phase of water present throughout the Arctic cryosphere. Ice thus represents a barrier (or impediment) as well as a benefit to virtually all human activities in the Arctic. As a result, changes in ice alter aspects of human accessibility to local and regional land and aquatic areas, directly affecting how these are utilized. Altered accessibility has far-reaching social, cultural and economic consequences.

- River and lake ice, and permafrost in terrestrial landscapes offer relatively inexpensive seasonal transportation routes via ice roads between southern centers of supply and northern communities and industrial sites. Thus, changes in these cryospheric components (such as shorter seasons, altered routes, weaker or thinner ice) affect transportation options and result in increased costs for overland transport. Few viable adaptation options exist, especially for communities far from coasts or large river systems. Permanent all-weather roads or railways may ultimately be necessary at significant cost to regional and/or national economies. On the other hand, cryospheric changes may also result in increased opportunities for water-based transport (longer seasons, more water from increased precipitation), but these are mostly restricted to northward-flowing large rivers, large lakes, and coastal areas.

- Land-fast sea ice provides local transportation options when it is present, and negative effects result from its earlier seasonal degradation. Activities on land-fast sea ice include hunter travel to floe edges for wildlife harvesting, inter-community travel, and industrial activities. Land-fast sea ice also provides a stabilizing and protective feature for

coastlines and near-shore activities such as hydrocarbon development. Its degradation will require changes in how these areas are used and the duration of their use. Altered shorelines will also affect communities (see Section 12.3.4). Alternatively, the presence of land-fast sea ice inhibits summer marine transportation, often requires local ice-breaking activities, and affects the re-supply of communities and coastal shipping. Earlier seasonal disappearance is likely to benefit such activities. Shifts in seasonal sea ice will also permit earlier and later, and perhaps increased, overall ice-breaking to allow for longer operational shipping seasons. Ice-breaking activities affect sea-ice habitats and ecosystems and may also affect local human activities.

- Pack ice, especially when highly consolidated and at its greatest seasonal extent during late winter, acts as a barrier to Arctic marine transportation and so restricts access by humans to many areas over much of the year. Decreased duration and areal extent of pack ice will thus facilitate marine transportation. Open areas in pack ice (polynyas and flaw leads) are key habitats for ice-associated Arctic biota but also represent access points for ice-capable shipping, and thus potential conflicts arise between habitat use by biota and economic endeavors.

- At the level of the individual (in contrast to the previous points which address the community level), over large areas of the Arctic access involves travel both on land and over local infrastructure such as roads. Similar to the consequences outlined above, changes in terrestrial and freshwater cryospheric elements will affect local accessibility. Permafrost degradation, changes in lake and river ice, glacier or ice sheet retreat, and changes in snow regimes will all affect the integrity of local landscapes and thus their suitability for overland travel. Poorer travel conditions increase hazard and may require a change in lifestyles, areas and timing of use, and traditional practices for indigenous groups. Changes are also likely to be required in local land use practices associated with industrial activities (e.g., seismic exploration). Where built infrastructure is present (roads, bridges, airstrips) cryospheric changes will affect local usage patterns and may require increased maintenance schedules or re-construction.

- Sea-ice affects marine vessel traffic whether this is destinational shipping to/from the Arctic, marine fishing, ship-based tourism, research cruises, military activities, coastal monitoring, or ultimately transpolar shipping. Altered sea-ice regimes for land-fast and pack ice will affect access associated with these activities. Although decreased sea ice can generally be expected to lengthen seasonal periods of access, ice regimes are expected to be highly variable in space and time. Moreover, larger areas of open water allow for increased consequences of wind and storms, and perhaps increased fog and icing events. Winds and storms also move large masses of unconsolidated ice, which represent a significant but highly dynamic local issue for shipping. Thus, while access may overall be longer or better, it may be offset by more rigorous operational conditions, greater probability for accidents, and associated requirements for safety and infrastructure. Also, feedbacks of human activities on the cryosphere from increased shipping may occur with increased production and deposition of contaminants and perhaps also engine soot in sensitive northern ice environments. Issues associated with Arctic shipping consist of those within territorial jurisdictions as well as those associated with the offshore environment. Given the connectivity among Arctic marine areas due to currents and present-day ice movements, accidents (particularly hydrocarbon spills) have the potential to affect large areas for long periods of time. Thus, cooperative pan-Arctic solutions to these issues are required.

- Overall reductions in sea ice will also allow for access to areas that were previously accessible only with great difficulty. Increased access may stimulate industrial activities (mines, sea-bed activities, hydrocarbon development), thus increasing local economic output, shipping, and further alteration of local lifestyles. These areas are likely to be those that persist for the longest, and thus within the next few decades may become remnant refuges for Arctic endemic biota. Modeled projections indicate that remnant multi-year sea ice is most likely to become associated with the northwestern margin of the Canadian Arctic Archipelago and northern Greenland, although substantive mobility may also occur. Action will be required to ensure the sustainable uses of such areas.

- Although limited at present, future glacial retreat, especially in Greenland, will expose new land allowing greater access to non-renewable resources. Economic benefits derived from increased access may be offset by the longer travel times required for hunters to reach wildlife populations. Thus, the overall balance of these changes to human activities is a dynamic consequence of potential positive and negative effects that cross many scales.

Altered accessibility within, to and from the Arctic is an inevitable outcome of cryospheric change and will affect many aspects of human activities in this area, albeit with uncertain timings. Even more uncertain are the local consequences of effects, whether these will be beneficial or detrimental, and whether the overall balance of effects will be negative or positive at a particular level or to a particular group in Arctic society, as well as collectively across all levels. The ability of humans to adapt to the challenges and opportunities associated with cryospheric change in the Arctic depends to a large degree on non-cryospheric factors and drivers of change. It is entirely possible that the overall balance of effects at one level may be negative (e.g., personal travel on sea ice) while at another level the effects may be positive (e.g., community re-supply and costs of purchased goods). Moreover, there may be local differences in adaptive capacity and involvement that will determine whether a change will be largely positive or largely negative. It is the sum of the effects that affect local cultural, economic and societal viability, and thus detailed examinations are required to better understand adaptation options to changes in the cryosphere.

12.3.2. Humans, biodiversity changes and land-use activities

Humans in the Arctic rely on ecosystem services and utilization of terrestrial and aquatic ecosystems for security of their local food supply, health and well-being, integrity of cultures and

lifestyles, income, and economic development. Cryospheric changes generally affect all these aspects of life, however, the nature and potential effects of these changes, and whether they represent potential threats or benefits, depend upon the economic sub-sector to which the individual belongs. Examples of some possible consequences follow, but this list is by no means comprehensive.

- Changes in land-fast sea ice have affected travel to/from the floe edge to hunt, with subsequent effects on safety, food security and personal economies of indigenous peoples. Such effects are locally variable; for example, generally negative in Nunavut, Canada but positive in Greenland (where a switch to boats has increased local harvests of some marine mammals). These effects are likely to continue and become more acute, however, longer ice-free seasons may benefit new approaches to hunting, traveling and local fishing but this is likely to require financial and technological capacity building. Changes in traditional harvesting practices that are closely connected to indigenous cultures will also be required. Moreover, current knowledge (i.e., traditional knowledge) must be expanded to accommodate cryospheric change. Increased reliance on some forms of technology (e.g., personal GPS devices) may aid activities in changing environments, but these do have limitations.

- Terrestrial cryospheric changes also affect travel to hunting, fishing and harvesting sites. Moreover, these activities often rely on large land mammals, such as caribou, that also travel as migratory herds. Migration routes and timing may be altered due to cryospheric changes and a mismatch may occur in traditional timing or place, or in hunter or herder access to areas for these activities. Routes of migratory land animals which cross water bodies increase the likelihood of habitat contraction and catastrophic losses of large numbers as cryospheric changes occur. In the case of herding activities, adjustments in the timing of seasonal activities are likely to be required. These adjustments are also management issues and so require adaptation in governance and regulation.

- Changes in snowpack conditions will affect accessibility, and the forage base and its quality for reindeer and other land herbivores. Follow-on effects are likely to be negative overall for reindeer husbandry, herd survival and well-being, as well as for harvesting wild populations, and thus engender further effects on the cultural, economic and lifestyle significance of this activity.

- As sea ice retreats, shifts in commercial or locally harvested fish species may occur and fishing activities may require adjustment. These will be highly regional in nature and may be exacerbated by unrealistic expectations regarding the biomass and sustainability of Arctic fisheries. In freshwater ecosystems, shifts in productivity and colonization may increase overall fisheries production particularly in the more southerly areas of the Arctic, although this is likely to be at the expense of some Arctic-adapted species.

- Large shifts in biodiversity are expected as cryospheric changes continue (see also Section 12.3.5). These include the following changes (most of which have negative outcomes

relative to the present situation): loss/extirpation of some organisms; changes and re-structuring of communities; loss of some unique habitats and communities; and colonization by invasive or pest species. Alternatively, some shifts in biodiversity offer opportunities, including: changes in ranges; higher productivity within some existing ecosystems; and new colonizing species some of which may be harvestable. The outcomes of biodiversity shifts are all highly uncertain, but it is reasonable to conclude that most Arctic ecosystems and communities will change substantively, ultimately with significant effects on humans. Also, given the rate and scope of cryospheric changes in the Arctic, it may be decades before ecosystems are sufficiently re-equilibrated and resilient to permit large-scale commercial exploitation. In the interim, some aspects of local subsistence resource harvesting may be seriously challenged.

- Cryospheric changes will also affect aspects of land use other than travel. Local hydropower options will be affected by snow changes (increased snow and more runoff is likely to represent a benefit) but altered lake (reservoir) and river-ice regimes will require adjustment to operations. Similarly, altered glacial mass balances imply short-term flow increases for some glacial-fed rivers and hydropower planning, as is happening in Greenland. Glacier retreat and changes in land drainage patterns, and eventually the complete loss of some glaciers will require planning to ensure water supply remains steady over the lifetime of the facilities.

- Ground subsidence in ice-rich permafrost areas affects travel, infrastructure, and long-term integrity of buildings. In addition to local economic consequences of infrastructure maintenance, repair and replacement, loss or rupture of containment structures or pipelines may have profound environmental effects including release of hydrocarbons or toxic metals into sensitive environments. This is exacerbated by aging and/or poorly inspected infrastructure, and follow-on consequences of permafrost changes such as altered flow regimes possibly facilitating wider movement of spilled materials.

Perhaps the biggest challenge in the future will be to try to maintain environmental and ecosystem integrity in the face of more rapid, yet still uncertain, changes resulting from many causes: substantive cryospheric changes, additional environmental changes due to climate change, and shifts in human populations and the nature of their activities in the Arctic. Uncertainty of outcomes is extremely high, especially at the ecosystem level and is compounded by high variability in system drivers and potential responses. Feedbacks between human systems and the environment, both within and outside the Arctic, further increase the uncertainty of outcomes. These factors will combine to make planning and environmental stewardship extremely difficult.

12.3.3. Industrialization and development

Cryospheric changes can create opportunities for industrial development in the Arctic by increasing accessibility. These opportunities include those associated with non-renewable resources including hydrocarbon development on land and coastal shelves and slopes, new mining possibilities (primarily

on land but potentially also in the sea), and transportation needs and opportunities. Opportunities also exist for renewable resource sectors, including increased potential for forestry and agriculture further north than at present, fisheries and aquaculture potential, and tourism.

Challenges to development opportunities also exist. These include high uncertainty in future states of the Arctic system and thus limited ability for predicting future potential (especially for industries that rely on renewable resources such as water). The overall system is rapidly following a trajectory of high change; high variability exists in the system, and the effects of changes on industrial and other human developmental activities are uncertain. Also, human responses to change are uncertain and the capacities to adapt to such challenges are highly variable across the Arctic. Nevertheless, appropriate planning and regulation are required to ensure that exploitation of new opportunities does not exacerbate changes in the cryosphere.

Aspects and challenges specific to many of these industrial or economic activities have been discussed or touched upon in other sections and will not be re-examined here. However, several have not and these are outlined below.

- **Tourism**: Increased ship-based tourism will be fostered by increased access associated with sea-ice changes, and existing evidence suggests a recent increase in this activity. Expectations of service, safety and positive experiences will be high and many local communities may be challenged to meet the capacity (e.g., in terms of providing local cultural experiences to meet demands). Regional or national governments may also face challenges (e.g., search and rescue needs). As components of the cryosphere continue to degenerate or re-equilibrate, the ice-associated animal species are likely to be reduced or re-distributed into smaller more northerly areas. Thus, access may become more difficult and, as species become rarer, disturbance will be an increasing problem that is likely to require responses to ensure sustainability both of species and their dwindling habitats. At the same time, land-based tourism to observe Arctic wildlife, experience adventures, or see cryospheric structures is a growing industry. As cryospheric degradation continues, however, the potential for this may be altered.

- **Forestry and Agriculture**: Increased precipitation and the resulting increase in local productivity may facilitate growth of harvestable renewable resources. This may also be positively affected by increased warning. Thus, along the southern margins of the Arctic and in many northerly sub-Arctic areas, increased production in renewable resource sectors is likely. Such increases, however, are likely to be highly variable in space and time, and effective increases in productivity may not be realized for decades. Moreover, in some areas drying will increase moisture stress on forests and exacerbate fires. Stressed trees are also more susceptible to pests.

- **Hydrocarbons**: Stores of oil, natural gas, and gas hydrates in the Arctic are significant and many are readily accessible in terrestrial and marine shelf environments, albeit with major technological challenges. Moreover, as access changes additional areas may become exploitable. As a result,

hydrocarbon development, production, and transport to southern markets are very likely to increase substantially over the coming decades. Increased variability and uncertainty in cryospheric environments, the Arctic climate system itself, and long-term adverse consequences of accidental discharges of oil, drilling wastes, or contaminants in cold and ice-associated environments, all represent substantive issues to be addressed.

12.3.4. Hazards

Changes in the Arctic cryosphere either create new hazards or alter exposure to existing hazards. As with most consequences of climate and cryospheric changes, hazards cross a wide spectrum of relevance ranging from immediate physical and direct hazards to those that are less direct and perhaps secondarily derived. Increased potential or realized hazards will require increased personal vigilance and local risk assessments, mitigation where required, and direct responses to the hazards themselves. Moreover, some hazards are predictable or can be assessed through risk-based methods whereas others cannot. Preparedness to meet both types of challenge is required. In addition to the hazards previously discussed, examples across the spectrum of likely hazards include the following.

- Decline of summer sea ice will expose vulnerable shorelines to open water and storm events, thus contributing to significant erosion and flooding in some areas. Sea-ice declines will interact with other cryospheric changes (e.g., permafrost decline) and other climate-driven changes (e.g., increased runoff) to further affect coastlines. The result will be significant consequences for shore-based infrastructure, coastal communities, and some cultural sites. Reinforcement and stabilization of buffering systems on coasts and/or re-location of coastal communities may be required with significant costs incurred by individuals, local and regional governments, or local industries.

- Where increased ice calving occurs directly into the sea, iceberg transport into southern shipping lanes of the Northwest Atlantic represents a hazard, but uncertainties presently preclude projections into the future. This represents a knowledge gap.

- Glacial outburst floods and avalanches are all exacerbated by increased snowfalls, enhanced freeze/thaw cycles, and degradation of cryospheric components such as glaciers, ice and permafrost. Slope instabilities further increase the likelihood of such occurrences. Moreover, effects from such events can be locally widespread and reach distant locations downstream. As a result, infrastructure, towns, and habitation in such areas are at increased risk as the cryosphere continues to degenerate. Appropriate risk assessments and potential remedial actions are required.

- Ice- and permafrost-related travel is inherently risky and longer freeze-up or break-up seasons, perhaps thinner ice, and less stable conditions are likely to increase human strandings or mortalities. Increased loss of equipment, industrial loads, foodstuffs, and perhaps hazardous goods can also be expected. Reduced use of ice and permafrost, reduced loads, alternative transportation routes, enhanced

search and rescue responses, and increased monitoring are required as local shorter-term responses. Longer-term responses may involve developing alternative transportation routes over great stretches of difficult terrain.

- Cultural or archaeological sites on coastlines, and/or lake or river margins are likely to be lost as permafrost degrades. Also, loss or exposure of paleontological remains is likely to occur. In many cases these sites have neither been discovered nor explored, and so their loss represents a significant cultural loss of global relevance. Permafrost changes may also enhance industrial hazards such as containment failures, pipeline ruptures, and infrastructure failures ranging from personal housing to community services. Enhanced monitoring and risk assessment is required; replacement planning requires enhanced engineering protocols, implying greater costs.

- Spring ice break-ups, particularly on large northward-flowing rivers (the Arctic has five of the ten largest rivers in the world and a high number of smaller but still substantive rivers), are projected to become less dynamic (mechanical) in nature. Although decreases in break-up flooding will have negative effects on the ecology of productive deltas, overall the changes are expected to be beneficial to communities and infrastructure in vulnerable locations. If, however, break-up occurs much earlier, particularly during mid-winter, or if enhanced snowmelt runoff drives break-up, increased severity and impacts of ice-jam flooding will result. The potential interactions between these two cryospheric components, river ice and snow, remain to be fully resolved.

- Decreased sea ice and more open water could alter the probability of higher winds and the severity of storms over the Arctic Ocean. Such storms will also result in more energetic motion of remaining ice. These represent uncertain hazards to human endeavors in the Arctic and require better weather prediction for large areas of the Arctic.

12.3.5. Ecosystem and biodiversity losses

Some cryospheric components and their associated biota are unique and irreplaceable once lost or degraded. Ice shelves, most of which fringe the northern margins of the Canadian Arctic Archipelago have persisted for the past 3000 to 5500 years. Substantial areal loss has occurred over the past 100 years, including large changes since 2000 (e.g., 23% loss of total ice shelf area in the warm summer of 2008, relative to 2007). The epishelf lakes retained behind ice shelves are unique polar aquatic ecosystems, but are now almost completely lost from the Arctic, with only one remaining.

Changes in permafrost condition fundamentally change hydrological conditions on land, and unique elements in permafrost areas such as peat mounds with frozen cores and ice-cored hills have been observed to decline. Permafrost degradation in land areas is likely to increase slope instability, while permafrost thawing in coastal areas may increase coastal erosion. Permafrost degradation, both in land and coastal areas, is likely to cause significant ecosystem changes.

Permafrost contains viable life to at least 400 m depth, in some cases preserved for significant periods of geological time (millions of years in northern Siberia). Biological activity

(bacteria, viruses) is evident after thousands of years of existence in permafrost. In many cases in deeper horizons these constitute 'prehistoric' floras of potential significance (e.g., 'new' genes predating humans, possibly suitable for molecular applications; re-emergence of human diseases); permafrost degradation will variously destroy or liberate such organisms.

Multi-year sea ice represents a unique, albeit spatially and temporally highly dynamic habitat. Many Arctic biota are highly 'ice-adapted' (e.g., iconic species such as ringed seal, narwhal, polar bear, ivory gull, walrus; specialized primary producers such as sea-ice diatoms; and ice-associated amphipods and copepods) and thus depend to a large degree on the presence of relevant sea-ice amounts and types. Others are highly 'ice-associated' (e.g., Arctic cod and seabirds such as thick-billed murre). Sea-ice degradation will significantly reduce habitat availability with consequences for population abundance, distribution, and persistence of many of these species. Spatio-temporal mismatches among ecosystem components are also likely to occur, with uncertain consequences for the remainder of the ecosystems. Similarly, although less obvious, semi-unique small-scale ecosystems and biota (e.g., seasonal melt-water ponds on multi-year sea ice) will be adversely affected. The sheer number of these small ecosystems, their intimate connection to productivity pathways to higher biota, and limited understanding all argue for their significance. Colonization by sub-Arctic biota and/or shifts in balance among obligate Arctic cryophiles and Arctic habitat generalists will tend to offset these losses of biodiversity, but will engender wide changes in the ecosystems. Substantive ecosystem and biotic consequences of cryospheric change are likely to involve widespread re-equilibration of existing Arctic ecosystems, and possibly the formation of completely new ecosystems which may take decades to stabilize. Follow-on consequences to humans will result and include conservation issues for Arctic biota, habitats and vanishing ecosystems. Also, there may be degradation or loss of supplies of local foods, traditionally relied upon by indigenous peoples. Tourism represents a renewable resource development opportunity for many areas of the Arctic – loss of key habitats and biota may diminish some of the attractiveness of this activity.

In addition to the above changes, some of which represent globally significant potential reductions or losses, many species, particularly birds migrate to the Arctic seasonally and winter in non-Arctic areas. Cryospheric shifts, particularly drainage of large terrestrial areas and shoreline impacts are likely to have significant impacts on such species. Similarly, changes in hydrology and altered vegetation and productivity regimes affect migrant terrestrial avifauna. The ultimate consequences are uncertain but migratory populations may decline, particularly if habitats are lost. Mismatches in food and key life history events such as nesting success in the Arctic will exacerbate such issues. This adds to the overall conservation needs required in the Arctic and beyond.

12.3.6. Shift in lifestyles, culture and socio-economics

Sustainability of some Arctic ecosystems as used for traditional activities, in particular food harvesting, is uncertain due either to difficulties in access or a declining resource base. Food

gathering and traditional annual cycles provide a basis for the culture of Arctic indigenous peoples. Development, shifts to wage-based economies, human demographic changes, and associated embedded changes all affect traditional lifestyles and culture. Cryospheric changes are likely to exacerbate the magnitude and rate of such changes. Decreased applicability of traditional knowledge, and limits to adaptive capacity and potential responses are likely to be presently strained, and will become more so as further changes occur. Food security is at risk as are culture, health and well-being of peoples with lifestyles connected to the cryosphere. Temporal and spatial mismatches are/will be important in ecosystem contexts and also in human contexts. If prey do not occur where people traditionally hunt, or if they are of lower quality, then it may not be possible to harvest adequate amounts.

12.4. Integrated consequences of cryospheric changes for lower latitudes and the global system

The Arctic is part of the global climate system and is closely linked to key elements through heat exchange and carbon cycling. Thus, in addition to significant effects within the Arctic, change in the Arctic cryosphere will feed back in many ways to affect the entire globe. Such feedbacks have far-reaching implications on the global climate system, sea level, and humans outside Arctic. In addition, the Arctic is economically linked to the entire globe and increased access and activities will have effects locally and within the global context.

12.4.1. Climate system changes

The Arctic is closely linked with the global climate system primarily through heat exchange in the atmosphere and ocean, but also through carbon cycling (e.g., GHGs). Alteration of these processes within the Arctic cryosphere results in their alteration globally, with significant consequences to humans at this scale.

Albedo Effects: Reduced Arctic albedo, particularly for the Arctic Ocean, will increase surface heating within the Arctic but will also generally increase heat retention by the globe. This feeds back within the Arctic to effect more rapid and greater change in the cryosphere. Heating of the Arctic also reduces the temperature gradient between the Arctic and southern areas, and thus may affect the strength of atmospheric and oceanic transport mechanisms.

Patterns of Atmospheric Circulation: Shifts in Arctic sea ice and increased heating will warm the lower atmosphere in the Arctic. This change will affect weather at lower latitudes of the Northern Hemisphere, particularly in winter. However, incomplete understanding of the nature and the closeness of the coupling between sea-ice extent, Arctic lower atmospheric heating, stability of Arctic boundary fronts, and circulation responses at lower latitudes complicate the development of adequate models with which to project future trends. Rapidly changing, perhaps increasing extremes, and highly variable weather responses at lower latitudes may result. Recent 'outbreaks' of cold Arctic air masses (e.g., 2005, 2008, 2009, 2010) over lower latitudes particularly in Eurasia appear to have resulted from weakening of the 'polar vortex' which typically

traps cold Arctic air near the pole. This variability may reflect an early shift toward altered Arctic, and perhaps northern hemispheric, climatic patterns.

Greenhouse Gas Releases: The Arctic is an effective moderator of the global carbon cycle with respect to storage capacity and local processes cycling GHGs. Over the recent geological past the GHG balance has been toward the storage of carbon (the Arctic has been a long-term sink). However, recent changes suggest the storage capacity is declining. Thus, GHGs are not as readily sequestered there and may contribute to global warming. Moreover, in some areas cryospheric changes are resulting in releases of stored GHGs and/or increased processing of GHGs. Permafrost degradation in some areas exacerbates this and its loss may potentially release large stores of trapped GHGs over decadal to centennial time frames. The significance of this feedback to the global system could be to further and very rapidly accelerate global climate change.

Contaminants: In parallel with the lessening of the temperature gradient between the Arctic and more southerly latitudes, the capacity for northward transport of contaminants is reduced, and thus accumulation in southerly areas is likely to increase. Also, it is very likely that the stores of legacy contaminants within the Arctic are being mobilized by cryospheric degradation. Their release from ice, snow, permafrost, glaciers and ice caps and subsequent re-entry into ecosystems appears to be occurring.

12.4.2. Global marine effects

Sea-Level Change: The issue of sea-level change, at local and global scales, is one of the more serious consequences of climate change, primarily due to the scale of potential effects. Sea-level change is a complex phenomenon resulting from many factors. Climate change affects sea level primarily through water mass changes and through density changes due to changes in temperature and salinity. Global sea levels are also affected by mass losses from non-Arctic glaciers and the Antarctic ice masses.

Present (2003–2008) mass losses from Arctic glaciers and the Greenland Ice Sheet contribute a total increase of 1.3 mm per year to the rise in global mean sea level. Increasing contributions from the Greenland Ice Sheet and other Arctic glaciers have occurred since 1995. Contributions from other sources (Antarctic Ice Sheet, non-Arctic glaciers) are added to these. Total projected sea-level rise resulting from all sources by 2100 cannot be estimated with high confidence at present. Lower and upper limits have a range of 0.79 to 2.01 m. A range of 0.9 to 1.6 m is considered the more plausible current estimate by the authors of the SWIPA assessment. Nevertheless, significant effects of the rising sea level will become increasingly evident by mid-century. High regional variation in sea-level rise will result from concurrent changes in other factors that include gravity fields, ocean temperatures, freshening, tidal effects and local isostatic rebound or subsidence of land. Rates will also differ, thus impacts may be highly regionalized and realized over varying time scales.

Mean sea-level rise increases possibilities of coastal flooding, erosion, infrastructure damage, environmental impacts on ecosystems, and saltwater intrusions into groundwater. Such effects may be accompanied or exacerbated by local additional

effects. Ultimate global effects realized at the century scale and beyond include significant inundation of low-lying coasts and possibly complete submergence of small islands in some areas of the globe, although growth of coral atolls may offset this to some degree.

Atlantic Meridional Overturning Circulation (AMOC): Ocean circulation (thermohaline circulation) is a global phenomenon by which heat and water are transported between the polar and equatorial regions. This circulation is powered in part by density differences among water masses due to differences in temperature and salinity. Inflows to the Arctic resulting from this phenomenon consist of surface (lower salinity, warmer) waters entering from the Pacific via the Bering Strait, and surface (warmer) waters entering from the North Atlantic. These inflows are counterbalanced by the outflow of Arctic waters primarily through the Canadian Archipelago and along eastern Greenland, mostly as lower salinity, cold and freshened surface flows. Additional deeper outflows of Arctic Ocean water (high salinity, cold) occur in the northeastern area of the Atlantic. The North Atlantic thus has several mixing zones where warm surface currents from the south interact with cold surface and sub-surface currents originating from the north. Areas of deepwater renewal (Northeast Atlantic – Barents Sea / Fram Strait; Northwest Atlantic – northern Labrador Sea / Southern Greenland) are essential to return water to the south and to the present strength of North Atlantic surface currents (the so-called 'Gulf Stream') which transport heat to northern Europe.

Several changes in the Arctic cryosphere are anticipated to increase freshwater inputs to the Arctic Ocean, thus reducing its salinity (at least in surface waters). These include: increased direct precipitation, possibly increased inputs of low salinity Pacific waters, increased runoff from large Arctic rivers, reduction of ice stores on land, and degradation of perennial sea ice. The increased input of freshwater will have local effects within the Arctic primarily on coastal shelves, many of which are largely associated with increased stratification. Concern has also been raised that increased freshwater outflow from the Arctic (possibly as deeper water masses) will alter the dynamics of ocean circulation and mixing in the North Atlantic. That is, the AMOC may be weakened. The SWIPA assessment found no conclusive evidence to support significant effects of cryospheric change on overturn circulation or deepwater formation. Total increased outputs of freshwater are likely to be at least one order of magnitude less than present flows of the major currents; accordingly, impacts of cryospheric change are likely to have little influence on major ocean current patterns at least for the foreseeable future. This is particularly so in comparison to the direct atmospheric effects of climate change on the strength and patterns of this circulation.

Freshwater Outflow: Most effects of cryospheric change will result in increased presence of freshwater in Arctic environments, much of which will eventually enter the Arctic Ocean. Considerable entrainment of freshwater occurs on marginal shelves and deeper in the Canada Basin, with accompanying shifts in salinity. Recent estimates suggest increased storage of freshwater has occurred in the Canada Basin, while export of freshwater stores from the Eurasian Basin appears to have increased. Large-scale export of freshwater

from the Arctic over geological time has been linked with large-scale cooling of the Northern Hemisphere. Factors that drive such export are poorly understood, although regional climatology is implicated. Thus, if storage turns to export, there may be significant surprises in store for the northern hemisphere climate.

Freshwater entry to coastal shelves and fjords decreases overall salinity and results in a mixed-layer of freshened seawater. Runoff, particularly from ice mass changes in Greenland, together with atmospheric forcing controls the depth of the mixed layer. As modeled freshwater outputs increased, mixed-layer decreased. This is likely to have profound consequences for nearshore productivity, although little research exists regarding likely changes on fjord and shelf ecosystems, or with respect to the shift of those effects to higher trophic levels of direct interest to humans. Also, as larger amounts of freshened water are expected to exit the Arctic seasonally (primarily as surface outflows in Baffin Bay / Davis Strait and along eastern Greenland), downstream effects on productive sub-Arctic seas are uncertain but may be profound. Better understanding of physical and biotic linkages is required for Arctic and sub-Arctic ecosystems, as well as cross-regional physical linkages to 'downstream' ecosystems.

12.4.3. Trade and economic development

Altered access (particularly that associated with changes in sea ice) increases possibilities for transpolar shipping and extraction of renewable and non-renewable resources. Changes in the cryosphere appear to be stimulating increased interest in exploration, development, and perhaps also migration from southern regions to the Arctic. Substantive follow-on effects will ensue if this occurs. Overall, reduction in polar sea ice can theoretically reduce intercontinental shipping distances by up to 40%, thus stimulating trade; realization of this potential is likely within the foreseeable future but benefits may take several decades to occur. Similarly, declines or shifts in some cryospheric components enhance development opportunities (albeit with many challenges remaining), many of which will benefit non-Arctic as well as Arctic residents. Coastal and some inland Arctic communities may realize benefits from increased trade (and in some cases these may offset effects of degraded land-based transportation). Overall, increased transport presents significant issues for resolution, such as development and implementation of a polar shipping code, establishing routes, mapping and development of appropriate services, and accident prevention and response.

12.5. Knowledge gaps and recommendations

12.5.1. Gaps in information and knowledge

Shortfalls in understanding and thus ability to project future states result from insufficient data (such as observational data) and the shortcomings of climate models. Each of the preceding chapters of this report noted discipline-specific gaps and issues, the resolution of which is required to ensure more accurate understanding in the future. In addition, an integrative section (Chapter 11, Section 11.5) focuses on many over-arching issues.

Some gaps are also explicitly or implicitly (through inferred conclusions) noted above. The following deficiencies are noted in the knowledge base.

- Although satellite observations of the cryosphere have expanded in type and quantity over the past decade, observational networks are still limited throughout the Arctic for many components, their key drivers and induced effects, and their linkages to human systems. This is particularly true for direct observations. Many important observations and monitoring efforts are research-based and in need of sustained funding. Coordination and common data linkages among observational networks are presently limited.

- There are considerable uncertainties in modeling some cryospheric processes. Permafrost models under-represent ice content and the insulating effect of the organic layer; climate models do not resolve the steep topography of the Greenland Ice Sheet margins; models of snow-vegetation interactions need to be improved; and models that link meteorology to glacier mass balance need to incorporate down-scaling techniques and satellite data.

- The effects of clouds on altering the nature and rates of change in the climate system, and particularly in the Arctic cryosphere, are very poorly known, and at present cannot be effectively modeled by many GCMs. Increased open water in a cold environment is likely to increase local cloudiness; how this may link to local cryospheric changes and thence to the global climate system is uncertain.

- Increased atmospheric carbon dioxide concentrations along with other GHGs fuel climate warming. They also have additional effects, for example carbon dioxide is sequestered within aquatic systems where acidification processes occur. Gas exchange between the atmosphere and waterbodies will increase as concentrations and temperatures rise and as duration and extent of ice covers decline. There is a need to better understand effects and model future situations, particularly for the Arctic Ocean.

- Cryospheric degradation will result in cascading effects on Arctic environments and peoples, their livelihoods, living conditions and quality of life, and the regional climate systems. It will also result in global consequences. The precise ways in which people and communities will be affected needs further study, particularly with regard to adaptation, changes in behavior, health, and resource development.

- Effective observations and monitoring across geographic and temporal scales, sectors and disciplines are all required in a linked inter-disciplinary fashion to provide measurement of change, assess efficacy of mitigative and adaptive strategies, and to prioritize needs for governance and other appropriate responses.

- Climate-driven cryospheric change is only one of many facets of overall Arctic change. Consequences of all drivers of Arctic change, including those in the cryosphere, need to be integrated into an overall understanding of the nature, rates and potential costs/benefits. This integration needs to include cumulative effects of multiple drivers, stressors or forcers on the Arctic system, and needs to cross the

physical, ecological and human systems present in the Arctic, and within the latter to cross the various sectors of human society.

12.5.2. Recommendations

The focus for the SWIPA assessment report has been to summarize the evidence for recent cryospheric change in the Arctic, and the effects and consequences of this change for the entire Arctic system. The objective of the SWIPA program was not to focus on the ultimate cause of the cryospheric change (see Chapter 1). Instead, the syntheses embodied in the preceding chapters of this Science Report highlight the present state of knowledge within the various disciplines, and the follow-on effects of present and ongoing change on Arctic ecosystems and humans. The following recommendations aim to increase that knowledge base and facilitate its use.

Near-term needs

- Increase the nature, spatial and temporal resolution, and availability of data from all observational platforms, all continuing into the future – including data for physical, ecological and human elements of the overall system and with a specific focus on feedback linkages within the system (e.g., enhanced methane releases due to cryospheric change) and other gaps.

- Modeling approaches provide insight into future possibilities, but efforts are required to reduce uncertainties in projections and enhance their spatial resolution, to parameterize and incorporate important feedbacks; and to downscale model results to appropriate regional levels.

- Enhance existing efforts to understand causal linkages across all aspects of the Arctic system, both among cryospheric components and between the cryosphere and other components of the Arctic system.

- Integrate and foster seamless availability of data products associated with the cryosphere, the consequences of change, and additional potential drivers.

- Identify and resolve impediments to inter-disciplinary linkages within and among cryospheric components, thereby fostering cross-disciplinary and integrated studies of changes in this system and their consequences for the Arctic.

- Link existing or develop new quantitative models to provide an integrated perspective of climate and cryospheric changes for the Arctic.

- Assess risks, or develop approaches to do so, of cryospheric changes for ecological and human systems in the Arctic across all relevant scales to allow effective planning and remediation. Identify shortfalls in design or approaches presently in place to address future changes in the cryosphere.

- Assess the consequences of cryospheric change for other physical and ecological systems (e.g., enhanced gas exchange driving acidification of Arctic aquatic systems).

- Evaluate roles and cumulative effects of potential drivers of continued cryospheric change across spatial scales to better understand causation and interactions, thus informing mitigation efforts.

Medium-term needs

- Identify and resolve impediments to linking all drivers and consequences of Arctic change into an overall understanding ranging from the physical to human levels of the system.

- Link or develop quantitative and qualitative approaches to enable cross-disciplinary analyses of change, causation and consequences.

- Initiate regional assessments focused on inter-disciplinary integration of recent changes, drivers of change and consequences; and develop projections of future possible outcomes to better inform planning and preparedness.

Long-term needs

- Develop and implement an overall strategy to assess Arctic change from all causes on a regular basis within which major proximate drivers such as cryospheric changes can be assessed.

- Integrate regional and sectoral assessments into regular Arctic-wide assessments.

12.6. Conclusion

The findings summarized in the SWIPA Science Report provide insights into the scope and nature of cryospheric changes in the Arctic. These, in turn, in combination with widespread climate variability, over large spatial and temporal scales, and with human adaptive capacity will ultimately determine whether particular changes in the cryosphere are viewed as opportunities or challenges. The overwhelming consequence appears to be a highly variable and uncertain future for the Arctic. This is perhaps the main challenge resulting from cryospheric change in the Arctic. Strategies and approaches that incorporate such uncertainty as a basic consideration underpin effective solutions. Thus, past trends, expectations of smooth transitions, expectations of slow rates of change, and expectations of low frequencies of extreme events appear to be becoming less trustworthy predictors of future situations and risks. As a consequence, planning for surprises or unanticipated events is increasingly important.

Acronyms and abbreviations

AAR	Accumulation area ratio
ACIA	Arctic Climate Impact Assessment
ALT	Active-layer thickness
AMAP	Arctic Monitoring and Assessment Programme
AMOC	Atlantic Meridional Overturning Circulation
AO	Arctic Oscillation
AOGCM	Atmosphere-Ocean General Circulation Model
ATM	Airborne Topographic Mapper
BC	Black carbon ('soot')
BP	Before present
BVOC	Biogenic volatile organic compounds
C	Carbon
CAD	Canadian dollar
CALM	Circumpolar Active Layer Monitoring
CH_4	Methane
CMIP3	Coupled Model Intercomparison Project phase 3
CO_2	Carbon dioxide
DIC	Dissolved inorganic carbon
DMI	Danish Meteorological Institute
DO	Dissolved oxygen
DOC	Dissolved organic carbon
ECMWF	European Centre for Medium-Range Weather Forecasts
EEZ	Exclusive economic zone
ELA	Equilibrium-line altitude
ENSO	El Niño / Southern Oscillation
ERA-40	ECMWF Re-Analysis covering the ~40 year period 1957-2002
ESA	European Space Agency
FYI	First-year ice
GCM	General circulation model
GDP	Gross domestic product
GHG	Greenhouse gas
GLRIP	Global Lake and River Ice Phenology dataset
GPR	Ground-penetrating radar
GPS	Global Positioning System
GRACE	Gravity Recovery and Climate Experiment
GRP	Gross regional product
Gt	Gigatonne
HCB	Hexachlorobenzene
HCH	Hexachlorocyclohexane
Hg	Mercury
IPCC	Intergovernmental Panel on Climate Change
IPCC AR4	IPCC fourth assessment
IPY	International Polar Year
ky	Thousand years
MAGT	Mean annual ground temperature
MeHg	Methylmercury
MIZ	Marginal ice zone
MSL	Mean sea level
MYI	Multi-year ice
N_2O	Nitrous oxide
NAM	Northern Annular Mode
NAO	North Atlantic Oscillation
NASA	National Aeronautics and Space Administration
NCAR	National Center for Atmospheric Research
NCEP	National Centers for Environmental Prediction
NMVOC	Non-methane volatile organic compound
NO_2	Nitrogen dioxide
NOAA	National Oceanic and Atmospheric Administration
NO_X	Nitrogen oxides
NSIDC	National Snow and Ice Data Center
pbb	Parts per billion
PCB	Polychlorinated biphenyl
PDD	Positive degree day
PDO	Pacific Decadal Oscillation
PNA	Pacific North America
POC	Particulate organic carbon
POP	Persistent organic pollutant
ppt	Parts per thousand
RCM	Regional climate model
SAR	Synthetic aperture radar
SAT	Surface air temperature
SCD	Snow-cover duration
SLE	Sea-level equivalent
SLiCA	Survey of Living Conditions in the Arctic
SLP	Sea-level pressure
SLR	Sea-level rise
SMB	Surface mass balance
SO_4	Sulfate
SRALT	Satellite radar altimetry
SST	Sea-surface temperature
Sv	Sverdrup (1 Sv = 10^6 m³)
SWE	Snow-water equivalent
SWEmax	Maximum monthly SWE
SWIPA	Snow, Water, Ice and Permafrost in the Arctic
TEK	Traditional ecological knowledge
ULS	Upward-looking sonar
UNCLOS	United Nations Convention on the Law of the Sea
USD	US dollar
w.e.	Water equivalent
WMO	World Meteorological Organization

ARCTIC COUNCIL

AMAP
Arctic Monitoring and
Assessment Programme

AMAP – IASC – IASSA – WCRP-CliC