

Arctic Climate: Past and Present

Lead Author

Gordon McBean

Contributing Authors

Genrikh Alekseev, Deliang Chen, Eirik Førland, John Fyfe, Pavel Y. Groisman, Roger King, Humfrey Melling, Russell Vose, Paul H. Whitfield

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Summary

The arctic climate is defined by a low amount or absence of sunlight in winter and long days during summer, with significant spatial and temporal variation. The cryosphere is a prominent feature of the Arctic. The sensitivities of snow and ice regimes to small temperature increases and of cold oceans to small changes in salinity are processes that could contribute to unusually large and rapid climate change in the Arctic.

The arctic climate is a complex system with multiple interactions with the global climate system. The phase of the Arctic Oscillation was at its most negative in the 1960s, exhibited a general trend toward a more positive phase from about 1970 to the early 1990s, and has remained mostly positive since. Sea ice is a primary means by which the Arctic exerts leverage on global climate, and sea-ice extent has been decreasing. In terrestrial areas, temperature increases over the past 80 years have increased the frequency of mild winter days, causing changes in aquatic ecosystems; the timing of river-ice breakups; and the frequency and severity of extreme ice jams, floods, and low flows.

The observational database for the Arctic is quite limited, with few long-term stations and a paucity of observations in general, making it difficult to distinguish with confidence between the signals of climate variability and change. Based on the analysis of the climate of the 20th century, it is very probable that the Arctic has warmed over the past century, although the warming has not been uniform. Land stations north of 60° N indicate that the average surface temperature increased by approximately 0.09 °C/decade during the past century, which is greater than the 0.06 °C/decade increase averaged over the Northern Hemisphere. It is not possible to be certain of the variation in mean land-station temperature over the first half of the 20th century because of a scarcity of observations across the Arctic before about 1950. However, it is probable that the past decade was warmer than any other in the period of the instrumental record.

Evidence of polar amplification depends on the time-scale of examination. Over the past 100 years, it is possible that there has been polar amplification, however, over the past 50 years it is probable that polar amplification has occurred.

It is very probable that atmospheric pressure over the Arctic Basin has been dropping, and it is probable that there has been an increase in total precipitation over the past century at the rate of about 1% per decade. Trends in precipitation are hard to assess because it is difficult to measure with precision in the cold arctic environment. It is very probable that snow-cover extent around the periphery of the Arctic has decreased. It is also very probable that there have been decreases in average arctic sea-ice extent over at least the past 40 years and a decrease in multi-year sea-ice extent in the central Arctic.

Reconstruction of arctic climate over the past thousands to millions of years demonstrates that arctic climate can vary substantially. There appears to be no natural impediment to anthropogenic climate change being very significant and greater in the Arctic than the change at the global scale. Especially during past cold periods, there have been times when temperature transitions have been quite rapid – from a few to several degrees change over a century.

2.1. Introduction

The Arctic is the northern polar component of the global climate system. The global climate system has been thoroughly examined in the recent reports of the Intergovernmental Panel on Climate Change (IPCC, 2001a,b,c), which include discussion of the impacts of climate change in the Arctic (IPCC, 2001a). Arctic climate is characterized by a low amount or absence of sunlight in winter and long days during summer. Although these solar inputs are a dominant influence, arctic climate exhibits significant spatial and temporal variability. As a result, the Arctic is a collection of regional climates with different ecological and physical climatic characteristics.

The cryosphere is a prominent feature of the Arctic, present as snow, ice sheets, glaciers, sea ice, and permafrost. The physical properties of snow and ice include high reflectivity, low thermal conductivity, and the high latent heat required to convert ice to liquid water; these contribute significantly to the regional character of arctic climate.

The arctic climate interacts with the climates of more southern latitudes through the atmosphere, oceans, and rivers. Because of these regionally diverse features, an exact geographic definition of the Arctic is not appropriate and this chapter focuses on the northernmost areas (usually north of 60° N), while acknowledging interactions with more southerly areas.

The observational database for the Arctic is quite limited, with few long-term stations and a paucity of observations in general. The combination of a sparse observational dataset and high variability makes it difficult to distinguish with confidence between the signals of climate variability and change.

With respect to the polar regions, the Intergovernmental Panel on Climate Change (IPCC, 2001a) stated:

Changes in climate that have already taken place are manifested in the decrease in extent and thickness of Arctic sea ice, permafrost thawing, coastal erosion, changes in ice sheets and ice shelves, and altered distribution and abundance of species in polar regions (high confidence).

Climate change in polar regions is expected to be among the largest and most rapid of any region on the Earth,

and will cause major physical, ecological, sociological, and economic impacts, especially in the Arctic, Antarctic Peninsula, and Southern Ocean (high confidence).

Polar regions contain important drivers of climate change. Once triggered, they may continue for centuries, long after greenhouse gas concentrations are stabilized, and cause irreversible impacts on ice sheets, global ocean circulation, and sea-level rise (medium confidence).

The arctic climate is a complex system and has multiple interactions with the global climate system. The sensitivities of snow and ice regimes to small temperature increases and of cold oceans to small changes in salinity, both of which can lead to subsequent amplification of the signal, are processes that could contribute to unusually large and rapid climate change in the Arctic. The Arctic Oscillation (AO) is an important feature of the arctic atmosphere and its connections with global climate (section 2.2). The phase of the AO was at its most negative in the 1960s, but from about 1970 to the early 1990s there was a general trend toward a more positive phase and it has remained mostly positive since. It is possible that this is the result of increased radiative forcing due to anthropogenic greenhouse gas (GHG) emissions, but it is also possible that it is a result of variations in sea surface temperatures. The Arctic Ocean (section 2.3) forms the core of the Arctic. Sea ice is the defining characteristic of the marine Arctic and is the primary means by which the Arctic exerts leverage on global climate. This leverage occurs through mediation of the exchange of radiation, sensible heat, and momentum between the atmosphere and the ocean. Terrestrial hydrology (section 2.4) and arctic climate are intricately linked. In terrestrial areas, temperature increases over the past 80 years have increased the frequency of mild winter days, causing changes in the timing of river-ice breakups; in the frequency and severity of extreme ice jams, floods, and low flows; and in aquatic ecosystems. The increased frequency of mild winter days has also affected transportation and hydroelectric generation.

There are both positive and negative feedback processes in the Arctic, occurring over a range of timescales. Positive feedbacks include snow and ice albedo feedback; reduction in the duration of time that sea ice insulates the atmosphere from the Arctic Ocean; and permafrost–methane hydrate feedbacks. Negative feedbacks can result from increased freshwater input from arctic watersheds, which makes the upper ocean more stably stratified and hence reduces temperature increases near the air–sea interface; reductions in the intensity of the thermohaline circulation that brings heat to the Arctic; and a possible vegetation–carbon dioxide (CO₂) feedback that has the potential to promote vegetation growth, resulting in a reduced albedo due to more vegetation covering the tundra. Polar amplification (greater temperature increases in the Arctic compared to the earth as a whole) is a result of the collective effect of these feedbacks and other process-

es. The Arctic is connected to the global climate, being influenced by it and vice versa (section 2.5).

Based on the analysis of the climate of the 20th century (section 2.6), it is very probable² that arctic temperatures have increased over the past century, although the increase has not been spatially or temporally uniform. The average surface temperature in the Arctic increased by approximately 0.09 °C/decade during the past century, which is 50% greater than the 0.06 °C/decade increase observed over the entire Northern Hemisphere (IPCC, 2001b). Probably as a result of natural variations, the Arctic may have been as warm in the 1930s as in the 1990s, although the spatial pattern of the warming was quite different and may have been primarily an artifact of the station distribution.

Evidence of polar amplification depends on the time-scale of examination. Over the past 100 years, it is possible that there has been polar amplification, however, over the past 50 years it is probable that polar amplification has occurred.

It is very probable that atmospheric pressure over the Arctic Basin has been dropping, and it is probable that there has been an increase in total precipitation over the past century at the rate of about 1% per decade. Trends in precipitation are hard to assess because precipitation is difficult to measure with precision in the cold arctic environment. It is very probable that snow-cover extent around the periphery of the Arctic has decreased. It is also very probable that there have been decreases in average arctic sea-ice extent over at least the past 40 years and a decrease in multi-year sea-ice extent in the central Arctic.

Reconstruction of arctic climate over thousands to millions of years demonstrates that the arctic climate has varied substantially. There appears to be no natural impediment to anthropogenic climate change being very significant and greater in the Arctic than the change on the global scale. Section 2.7.2 examines the variability of arctic climate during the Quaternary Period (the past 1.6 million years) with a focus on the past 20 000 years. Arctic temperature variability during the Quaternary Period has been greater than the global average. Especially during past cold periods, there have been times when the variability and transitions in temperature have been quite rapid – from a few to several degrees change over a century. There have also been decadal-scale variations due to changes in the thermohaline circulation, with marked regional variations.

2.2. Arctic atmosphere

The arctic atmosphere is highly influenced by the overall hemispheric circulation, and should be regarded in this general context. This section examines Northern Hemisphere circulation using the National Centers for

²In this chapter, when describing changes in arctic climate, the words possible, probable, and very probable are used to indicate the level of confidence the authors have that the change really did occur, recognizing the limitations of the observing system and paleoclimatic reconstructions of arctic climate.

Environmental Prediction/National Center for Atmospheric Research reanalyses for the period from 1952 to 2003 (updated from Kalnay et al., 1996). Section 2.2.1 describes the main climatological features, while section 2.2.2 discusses the two major modes of variability: the AO (and its counterpart, the North Atlantic Oscillation) and the Pacific Decadal Oscillation. Because much of the observed change in the Arctic appears to be related to patterns of atmospheric circulation, it is important that these modes of atmospheric variability be described.

2.2.1. Climatology

Atmospheric circulation and weather are closely linked to surface pressure. Figure 2.1a shows the Northern Hemisphere seasonal mean patterns of sea-level pressure in winter and summer. The primary features of sea-level pressure in winter include the oceanic Aleutian and Icelandic Lows, and the continental Siberian High with its extension into the Arctic (the Beaufort High). The sea-level pressure distribution in summer is dominated by subtropical highs in the eastern Pacific and Atlantic Oceans, with relatively weak gradients in polar and sub-polar regions. The seasonal cycle of sea-level pressure over the mid-latitude oceans exhibits a summer maximum and winter minimum. By contrast, the seasonal cycle of sea-level pressure over the Arctic and subarctic exhibits a maximum in late spring, a minimum in winter, and a weak secondary maximum in late autumn. The climatological patterns and seasonal cycle of sea-level pressure are largely determined by the regular passage of migratory cyclones and anticyclones, which are associated with storminess and settled periods, respectively. Areas of significant winter cyclonic activity (storm

tracks) are found in the North Pacific and North Atlantic. These disturbances carry heat, momentum, and moisture into the Arctic, and have a significant influence on high-latitude climate.

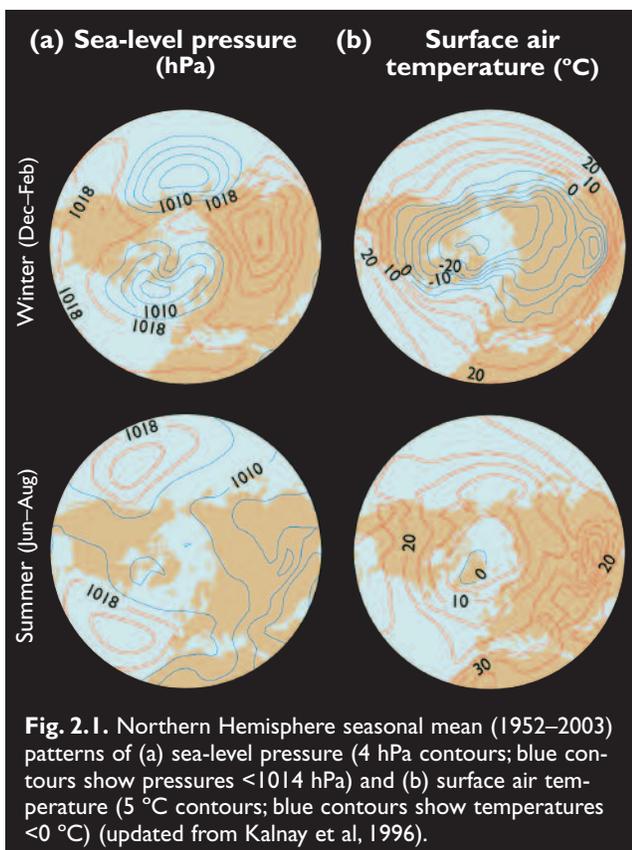
The Arctic is affected by extremes of solar radiation. The amount of solar radiation received in summer is relatively high due to long periods of daylight, but its absorption is kept low by the high albedo of snow and ice. The amount of solar radiation received in winter is low to non-existent. Figure 2.1b shows the seasonal mean patterns of surface air temperature. The Arctic is obviously a very cold region of the Northern Hemisphere, especially in winter when the seasonal mean temperature falls well below -20°C . Temperature inversions, when warm air overlies a cold surface, are common in the Arctic. At night, especially on calm and clear nights, the ground cools more rapidly than the adjacent air because the ground is a much better emitter of infrared radiation than the air. The arctic winter is dominated by temperature inversions, due to the long nights and extensive infrared radiation losses. Arctic summers have fewer and weaker temperature inversions. On the hemispheric scale, there exist large north–south gradients of atmospheric temperature (and moisture). In winter, the continental landmasses are generally colder than the adjacent oceanic waters, owing to the influence of warm surface currents on the western boundaries of the Atlantic and Pacific Oceans.

2.2.2. Variability modes

2.2.2.1. Arctic/North Atlantic Oscillation

The North Atlantic Oscillation (NAO) has long been recognized as a major mode of atmospheric variability over the extratropical ocean between North America and Europe. The NAO describes co-variability in sea-level pressure between the Icelandic Low and the Azores High. When both are strong (higher than normal pressure in the Azores High and lower than normal pressure in the Icelandic Low), the NAO index is positive. When both are weak, the index is negative. The NAO is hence also a measure of the meridional gradient in sea-level pressure over the North Atlantic, and the strength of the westerlies in the intervening mid-latitudes. The NAO is most obvious during winter but can be identified at any time of the year. As the 20th century drew to a close, a series of papers were published (e.g., Thompson et al., 2000) arguing that the NAO should be considered as a regional manifestation of a more basic annular mode of sea-level pressure variability, which has come to be known as the Arctic Oscillation (AO). The AO is defined as the leading mode of variability from a linear principal component analysis of Northern Hemisphere sea-level pressure. It emerges as a robust pattern dominating both the intra-seasonal (e.g., month-to-month) and inter-annual variability in sea-level pressure.

Whether or not the AO is in fact a more fundamental mode than the NAO is a matter of debate. For example, Deser (2000) concluded that the correlation between the



Pacific and Azores high-pressure areas was not significant, and that the AO cannot therefore be viewed as reflecting such a teleconnection. Ambaum et al. (2001) found that even the correlation between the Pacific and Icelandic–Arctic low-pressure centers was not significant. They argue that the AO is mainly a reflection of similar behavior in the Pacific and Atlantic basins. Regardless, the AO and NAO time series are very highly correlated, and for most applications (including this assessment), either paradigm can be used. Before proceeding with a description of the AO/NAO, two cautionary points must be mentioned. First, while the AO/NAO is obviously dominant, it explains only a fraction (i.e., 20 to 30%) of the total variability in sea-level pressure. Second, because the AO/NAO index is derived from a linear statistical tool, it cannot describe more general nonlinear variability. Monahan et al. (2003) have shown that hemispheric variability is significantly nonlinear, and the AO provides only the optimal linear approximation of this variability.

Figure 2.2a shows the AO/NAO time series obtained using monthly mean sea-level pressure for all months in the “extended winter” (November to April). There is considerable month-to-month and year-to-year variability, as well as variability on longer timescales. The AO/NAO index was at its most negative in the 1960s. From about 1970 to the early 1990s, there was a general increasing trend, and the AO index was more positive than negative throughout the 1990s. The physical origins of these long-term changes are the subject of considerable debate. Fyfe et al. (1999) and Shindell et al. (1999) have shown that positive AO trends can be obtained from global climate models using scenarios of increasing radiative forcing due to rising GHG concentrations. Rodwell et al. (1999) and Hoerling et al. (2001) have shown similar positive trends using global climate models run with fixed radiative forcing and observed annually varying sea surface temperatures. Rodwell et al. (1999) argued that slowly varying sea surface temperatures in the North Atlantic are locally communicated to the atmosphere through evaporation, precipitation, and atmospheric heating processes. On the other hand, Hoerling et al. (2001) suggested that changes in tropical sea surface temperatures, especially in the Indian and Pacific Oceans, may be more important than changes in sea surface temperatures in the North Atlantic. They postulated that changes in the tropical ocean alter the pattern and magnitude of tropical rainfall and atmospheric heating, which in turn produce positive AO/NAO trends. Regardless of the causes, it must be noted that AO/NAO trends do not necessarily reflect a change in the variability mode itself. As demonstrated by Fyfe (2003), the AO/NAO trends are a reflection of a more general change in the background, or “mean”, state with respect to which the modes are defined.

Figure 2.2b shows the sea-level pressure anomaly pattern associated with the AO/NAO time series, as derived from a principal components analysis. The pattern shows negative anomalies over the polar and sub-polar latitudes, and positive anomalies over the mid-latitudes. The anomaly center in the North Atlantic,

while strongest in the vicinity of the Icelandic Low, extends with strength well into the Arctic Basin. Not surprisingly, these anomalies are directly related to fluctuations in cyclone frequency. Serreze et al. (1997) noted a strong poleward shift in cyclone activity during the positive phase of the AO/NAO, and an equatorward shift during the negative phase. In the region corresponding to the climatological center of the Icelandic Low, cyclone events are more than twice as common during the positive AO/NAO extremes than during negative extremes. Systems found in this region during the positive phase are also significantly deeper than are their negative AO/NAO counterparts. McCabe et al. (2001) noted a general poleward shift in Northern Hemisphere cyclone activity starting around 1989, coincident with the positive trend in the AO/NAO time series. Figure 2.2c shows the pattern of surface air temperature anomalies associated with the AO/NAO time series. Negative surface air temperature anomalies centered in Davis Strait are consistent with southeasterly advection of cold arctic air by the AO/NAO-related winds. Easterly advection of warmer air, also linked to AO/NAO-related winds, accounts for the pattern of positive anomalies in surface air temperature over Eurasia.

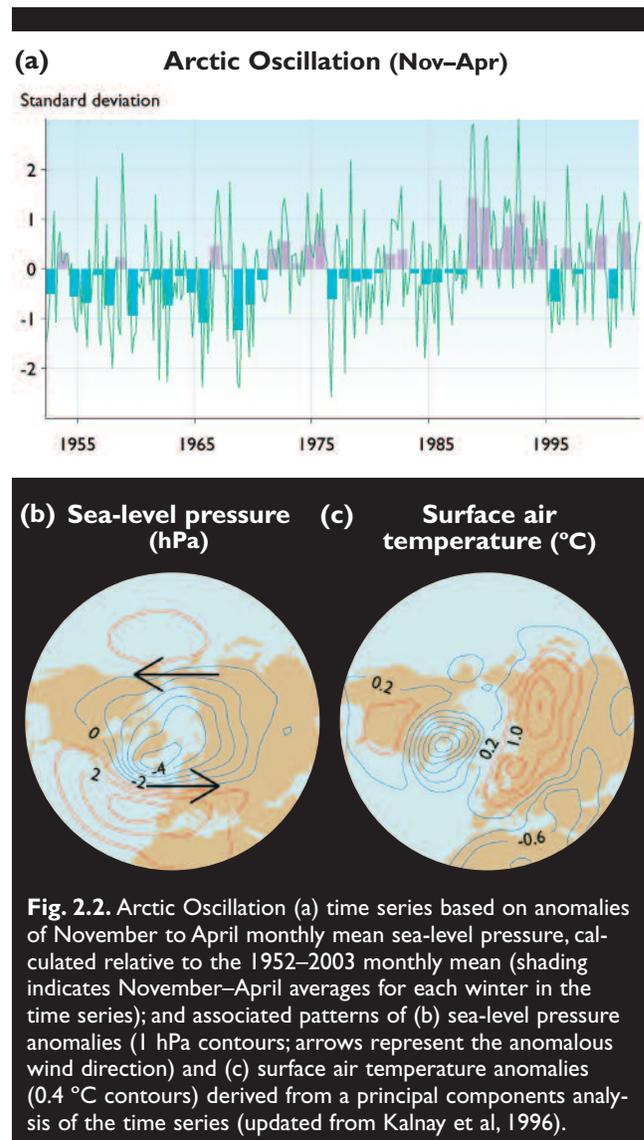


Fig. 2.2. Arctic Oscillation (a) time series based on anomalies of November to April monthly mean sea-level pressure, calculated relative to the 1952–2003 monthly mean (shading indicates November–April averages for each winter in the time series); and associated patterns of (b) sea-level pressure anomalies (1 hPa contours; arrows represent the anomalous wind direction) and (c) surface air temperature anomalies (0.4 °C contours) derived from a principal components analysis of the time series (updated from Kalnay et al, 1996).

2.2.2.2. Pacific Decadal Oscillation

The Pacific Decadal Oscillation (PDO) is a major mode of North Pacific climate variability. The PDO is obtained as the leading mode of North Pacific monthly surface temperature. Figure 2.3a shows the PDO time series obtained using monthly mean surface air temperature for all months in the extended winter (November to April). As with the AO/NAO time series, the PDO time series displays considerable month-to-month and year-to-year variability, as well as variability on longer timescales. The PDO was in a negative (cool) phase from 1947 to 1976, while a positive (warm) phase prevailed from 1977 to the mid-1990s (Mantua et al., 1997; Minobe, 1997). Major changes in northeast Pacific marine ecosystems have been correlated with these PDO phase changes. As with the AO/NAO, the physical origins of these long-term changes are currently unknown.

Figures 2.3b and 2.3c show the sea-level pressure and surface air temperature anomalies associated with the PDO time series, as derived from a principal components analysis. The sea-level pressure anomaly pattern is wave-like, with low sea-level pressure anomalies over

the North Pacific and high sea-level pressure anomalies over western North America. At the same time, the surface air temperatures tend to be anomalously cool in the central North Pacific and anomalously warm along the west coast of North America. The PDO circulation anomalies extend well into the troposphere in a form similar to the Pacific North America pattern (another mode of atmospheric variability).

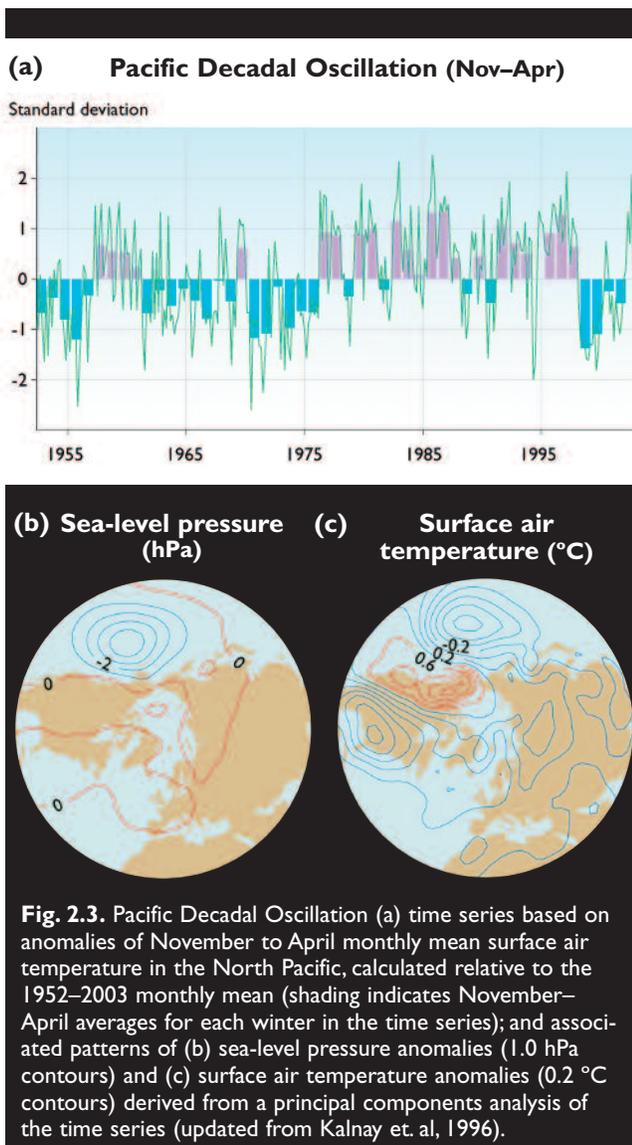
2.3. Marine Arctic

2.3.1. Geography

The Arctic Ocean forms the core of the marine Arctic. Its two principal basins, the Eurasian and Canada, are more than 4000 m deep and almost completely land-locked (Fig. 2.4). Traditionally, the open boundary of the Arctic Ocean has been drawn along the Barents Shelf edge from Norway to Svalbard, across Fram Strait, down the western margin of the Canadian Archipelago and across Bering Strait (Aagaard and Coachman, 1968a). Including the Canadian polar continental shelf (Canadian Archipelago), the total ocean area is 11.5 million km², of which 60% is continental shelf. The shelf ranges in width from about 100 km in the Beaufort Sea (Alaska) to more than 1000 km in the Barents Sea and the Canadian Archipelago. Representative shelf depths off the coasts of Alaska and Siberia are 50 to 100 m, whereas those in the Barents Sea, East Greenland, and northern Canada are 200 to 500 m. A break in the shelf at Fram Strait provides the only deep (2600 m) connection to the global ocean. Alternate routes to the Atlantic via the Canadian Archipelago and the Barents Sea block flow at depths below 220 m while the connection to the Pacific Ocean via Bering Strait is 45 m deep. About 70% of the Arctic Ocean is ice-covered throughout the year.

Like most oceans, the Arctic is stratified, with deep waters that are denser than surface waters. In a stratified ocean, energy must be provided in order to mix surface and deep waters or to force deep-water flow over obstacles. For this reason, seabed topography is an important influence on ocean processes. Sections 6.3 and 9.2.2 contain detailed discussions of the Arctic Ocean and sea ice.

The term “marine Arctic” is used here to denote an area that includes Baffin, Hudson, and James Bays; the Labrador, Greenland, Iceland, Norwegian, and Bering Seas; and the Arctic Ocean. This area encompasses 3.5 million km² of cold, low-salinity surface water and seasonal sea ice that are linked oceanographically to the Arctic Ocean and areas of the North Atlantic and North Pacific Oceans that interact with them. In this region, the increase in density with depth is dominated by an increase in salinity as opposed to a decrease in temperature. The isolated areas of the northern marine cryosphere, namely the Okhotsk and Baltic Seas and the Gulf of St. Lawrence, are not included in this chapter’s definition of “marine Arctic”.



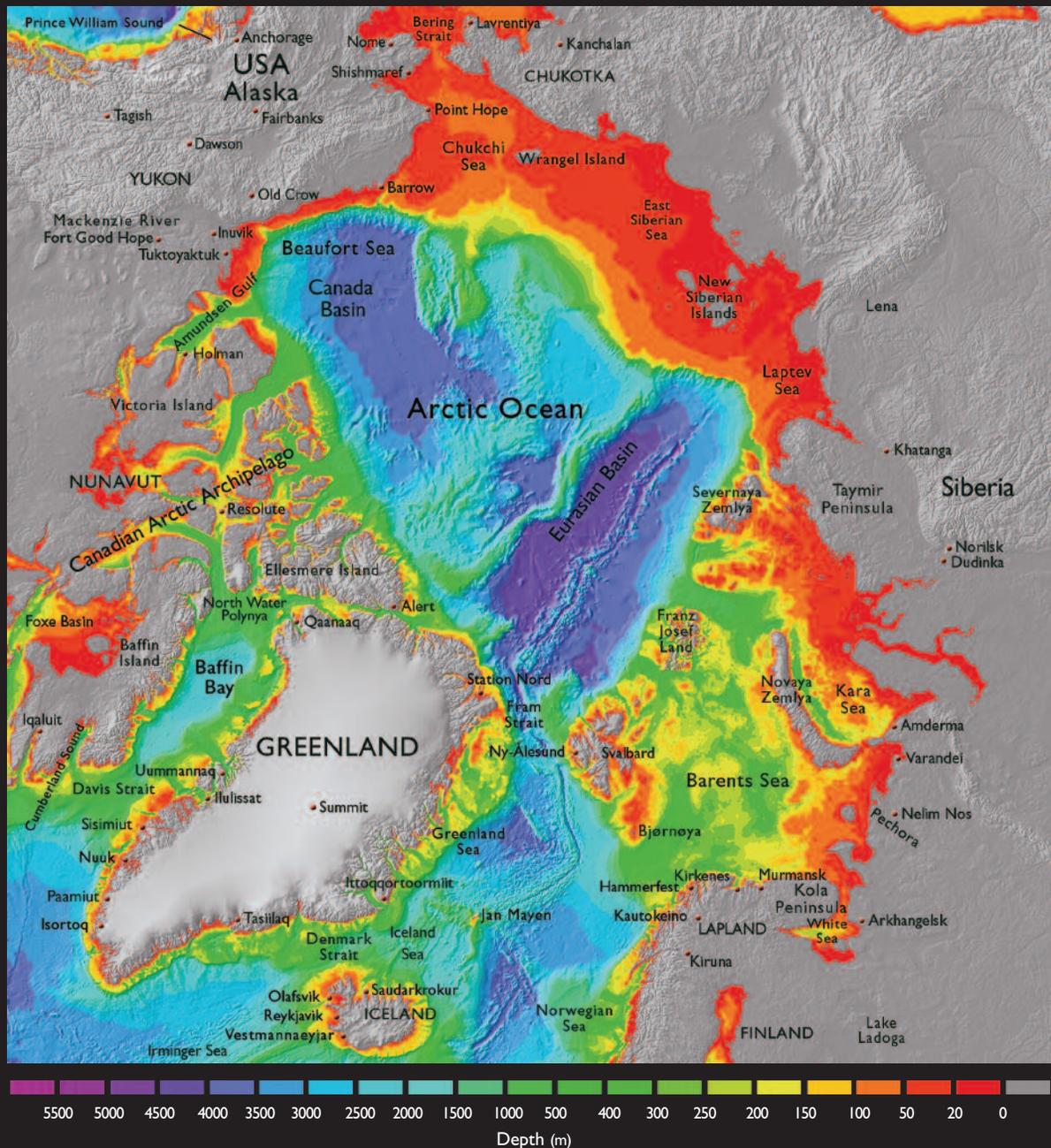


Fig. 2.4. Topographic features of the marine Arctic (International Bathymetric Chart of the Arctic Ocean; <http://www.ngdc.noaa.gov/mgg/bathymetry/arctic/arctic.html>).

2.3.2. Influence of temperate latitudes

Climatic conditions in northern mid-latitudes influence the Arctic Ocean via marine and fluvial inflows as well as atmospheric exchange. The transport of water, heat, and salt by inflows are important elements of the global climate system. Warm inflows have the potential to melt sea ice provided that mixing processes can move heat to the surface. The dominant impediment to mixing is the vertical gradient in salinity at arctic temperatures. Therefore, the presence of sea ice in the marine Arctic is linked to the salt transport by inflows.

Approximately 11% of global river runoff is discharged to the Arctic Ocean, which represents only 5% of global ocean area and 1% of its volume (Shiklomanov et al.,

2000). In recognition of the dramatic effect of freshwater runoff on arctic surface water, the salt budget is commonly discussed in terms of freshwater, even for marine flows. Freshwater content in the marine context is the fictitious fraction of freshwater that dilutes sea-water of standard salinity (e.g., 35) to create the salinity actually observed. For consistency with published literature, this chapter uses the convention of placing “freshwater” in quotes to distinguish the freshwater component of ocean water from the more conventional definition of freshwater.

The Arctic is clearly a shortcut for flow between the Pacific and Atlantic Oceans (Fig. 2.5). A flow of $800\,000\text{ m}^3/\text{s}$ (0.8 Sv) follows this shortcut to the Atlantic via Bering Strait, the channels of the Canadian

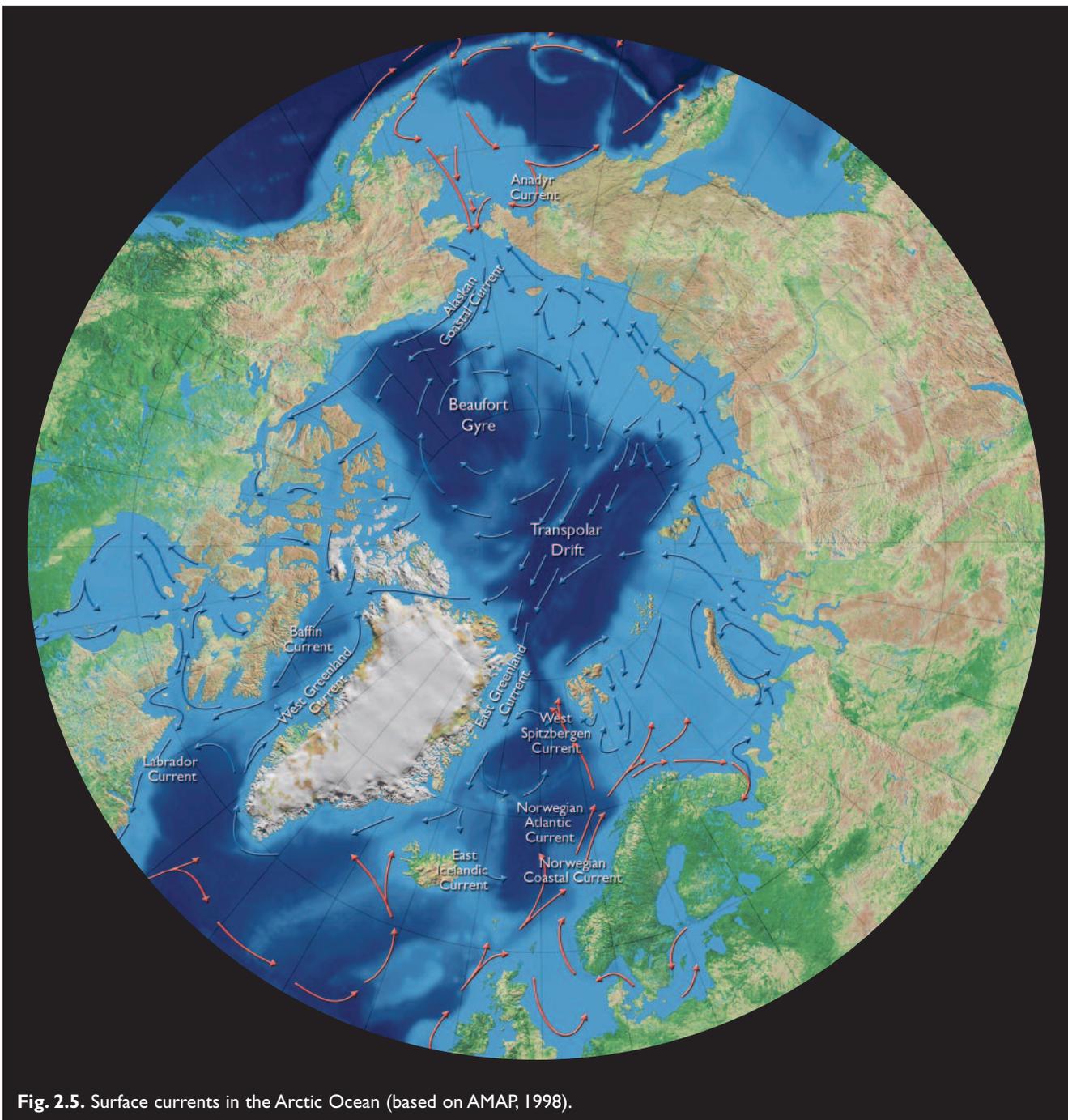


Fig. 2.5. Surface currents in the Arctic Ocean (based on AMAP, 1998).

Archipelago, and Fram Strait (Melling, 2000). The flow is driven by higher sea level (~ 0.5 m) in the North Pacific (Stigebrandt, 1984). The difference in elevation reflects the lower average salinity of the North Pacific, maintained by an excess of precipitation over evaporation relative to the North Atlantic (Wijffels et al., 1992). By returning excess precipitation to the Atlantic, the flow through the Arctic redresses a global-scale hydrologic imbalance created by present-day climate conditions. By transporting heat into the Arctic Ocean at depths less than 100 m, the flow influences the thickness of sea ice in the Canada Basin (Macdonald R. et al., 2002).

Much of the elevation change between the Pacific and the Atlantic occurs in Bering Strait. Operating like a weir in a stream, at its present depth and width the strait hydraulically limits flow to about 1 Sv (Overland and

Roach, 1987). Bering Strait is therefore a control point in the global hydrological cycle, which will allow more through-flow only with an increase in sea level. Similar hydraulic controls may operate with about 0.2 m of hydraulic head at flow constrictions within the Canadian Archipelago. The present “freshwater” flux through Bering Strait is about 0.07 Sv (Aagaard and Carmack, 1989; Fedorova and Yankina, 1964).

The Bering inflow of “freshwater” destined for the Atlantic is augmented from other sources, namely rivers draining into the Arctic Ocean, precipitation over ocean areas, and sea ice. The total influx to the marine Arctic from rivers is 0.18 Sv (Shiklomanov et al., 2000), about 2.5 times the “freshwater” flux of the Pacific inflow through Bering Strait. This estimate includes runoff from Greenland, the Canadian Archipelago, and the water-

sheds of the Yukon River (carried through Bering Strait by the Alaskan Coastal Current), Hudson Bay, and James Bay. The average annual precipitation minus evaporation north of 60° N is 0.16 m/yr (Barry and Serreze, 2000), corresponding to a freshwater flux of 0.049 Sv over marine areas. The combined rate of freshwater supply to the marine Arctic is 0.3 Sv.

Sea ice has a high “freshwater” content, since it loses 80% of its salt upon freezing and all but about 3% through subsequent thermal weathering. Although about 10% of sea-ice area is exported annually from the Arctic Ocean through Fram Strait, this is not a “freshwater” export from the marine Arctic, since the boundary is defined as the edge of sea ice at its maximum extent.

Freezing segregates the upper ocean into brackish surface (ice) and salty deeper components that circulate differently within the marine Arctic. The melting of sea ice delivers freshwater to the surface of the ocean near the boundary of the marine Arctic. The flux of sea ice southward through Fram Strait is known to be about 0.09 Sv (Vinje, 2001), but the southward flux of seasonal sea ice formed outside the Arctic Ocean in the Barents, Bering, and Labrador Seas; the Canadian Archipelago; Hudson and Baffin Bays; and East Greenland is not known.

The inflows to the marine Arctic maintain a large reservoir of “freshwater” (i.e., diluted seawater and brackish sea ice). Aagaard and Carmack (1989) estimated the volume of “freshwater” stored within the Arctic Ocean to be 80 000 km³. A rough estimate suggests that there is an additional reservoir of approximately 50 000 km³ in the marginal seas described in the previous paragraph. The total reservoir of “freshwater” equals the accumulation of inflow over about 15 years.

The “freshwater” reservoir feeds two boundary currents that flow into the western North Atlantic – the East Greenland Current and the Labrador Current (Aagaard and Coachman, 1968a,b). The former enters the Greenland Sea via Fram Strait and the latter enters the Labrador Sea via Davis Strait, gathering a contribution from Hudson Bay via Hudson Strait.

Northbound streams of warm saline water, the Norwegian Atlantic Current and the West Greenland Current, counter the flow of low-salinity water toward the Atlantic. The Norwegian Atlantic Current branches into the West Spitzbergen Current and the Barents Sea through-flow. The former passes through Fram Strait with a temperature near 3 °C and follows the continental slope eastward at depths of 200 to 800 m as the Fram Strait Branch (Gorshkov, 1980). The latter, cooled to less than 0 °C and freshened by arctic surface waters, enters the Arctic Ocean at depths of 800 to 1500 m in the eastern Barents Sea (Schauer et al., 2002). The West Greenland Current carries 3 °C seawater to northern Baffin Bay, where it mixes with arctic outflow and joins the south-flowing Baffin Current (Melling et al., 2001).

The inflows via the West Spitzbergen Current and Barents Sea through-flow are each about 1 to 2 Sv. The West Greenland Current transports less than 0.5 Sv. The associated fluxes of “freshwater” are small because salinity is close to 35. All fluxes vary appreciably from year to year.

The Fram Strait and Barents Sea branches are important marine sources of heat and the most significant sources of salt for arctic waters subjected to continuous dilution. The heat loss to the atmosphere in the ice-free northeastern Greenland Sea averages 200 W/m² (Khrol, 1992). The average heat loss from the Arctic Ocean is 6 W/m² of which 2 W/m² comes from the Atlantic-derived water. The impact of the incoming oceanic heat on sea ice is spatially non-uniform because the upper-ocean stability varies with the distribution of freshwater storage and ice cover.

2.3.3. Arctic Ocean

The two branches of Atlantic inflow interleave at depths of 200 to 2000 m in the Arctic Ocean because of their high salinity, which makes them denser than surface waters despite their higher temperature. They circulate counter-clockwise around the basin in narrow (50 km) streams confined to the continental slope by the Coriolis Effect. The streams split where the slope meets mid-ocean ridges, creating branches that circulate counter-clockwise around the sub-basins (Rudels et al., 1994). The delivery of new Atlantic water to the interior of basins is slow (i.e., decades).

The boundary currents eventually return cooler, fresher, denser water to the North Atlantic via Fram Strait (Greenland side) and the Nordic Seas. The circuit time varies with routing. The role of arctic outflow in deep convection within the Greenland Sea and in the global thermohaline circulation is discussed in section 9.2.3. In the present climate, Atlantic-derived waters in the Arctic Ocean occur at depths too great to pass through the Canadian Archipelago.

Inflow from the North Pacific is less saline and circulates at a shallower depth than Atlantic inflow. It spreads north from Bering Strait to dominate the upper ocean of the western Arctic – the Chukchi and Beaufort Seas, Canada Basin, and the Canadian Archipelago. An oceanic front presently located over the Alpha-Mendeleyev Ridge in Canada Basin separates the region of Pacific dominance from an “Atlantic domain” in the eastern hemisphere. A dramatic shift of this front from the Lomonosov Ridge in the early 1990s flooded a wide area of former Pacific dominance with warmer and less stratified Atlantic water (Carmack et al., 1995).

The interplay of Atlantic and Pacific influence in the Arctic Ocean, the inflows of freshwater, and the seasonal cycle of freezing and melting create a layered structure in the Arctic Ocean (Treshnikov, 1959). These layers, from top to bottom, include snow; sea ice; surface sea-

water strongly diluted by precipitation, river discharge, and ice melt; warm summer intrusions from ice-free seas (principally the Bering Sea); cold winter intrusions from freezing seas; cool winter intrusions from ice-free seas (principally the Barents Sea); warm intrusions of the Fram Strait Branch; cool intrusions of the Barents Sea Branch; recently-formed deep waters; and relict deep waters. The presence and properties of each layer vary with location across the Arctic Ocean.

The cold and cool winter intrusions form the arctic cold halocline, an approximately isothermal zone wherein salinity increases with depth. The halocline isolates sea ice from warm deeper water because its density gradient inhibits mixing, and its weak temperature gradient minimizes the upward flux of heat. The cold halocline is a determining factor in the existence of year-round sea ice in the present climate. Areas of seasonal sea ice either lack a cold halocline (e.g., Baffin Bay, Labrador Shelf, Hudson Bay) or experience an intrusion of warm water in summer that overrides it (e.g., Chukchi Sea, coastal Beaufort Sea, eastern Canadian Archipelago). The stability of the cold halocline is determined by freshwater dynamics in the Arctic and its low temperature is maintained by cooling and ice formation in recurrent coastal polynyas (Cavaliere and Martin, 1994; Melling, 1993; Melling and Lewis, 1982; Rudels et al., 1996). Polynyas are regions within heavy winter sea ice where the ice is thinner because the oceanic heat flux is locally intense or because existing ice is carried away by wind or currents. The locations and effectiveness of these “ice factories” are functions of present-day wind patterns (Winsor and Björk, 2000).

2.3.4. Sea ice

Sea ice is the defining characteristic of the marine Arctic. It is the primary method through which the Arctic exerts leverage on global climate, by mediating the exchange of radiation, sensible heat, and momentum between the atmosphere and the ocean (see section 2.5). Changes to sea ice as a unique biological habitat are in the forefront of climate change impacts in the marine Arctic.

The two primary forms of sea ice are seasonal (or first-year) ice and perennial (or multi-year) ice. Seasonal or first-year ice is in its first winter of growth or first summer of melt. Its thickness in level floes ranges from a few tenths of a meter near the southern margin of the marine cryosphere to 2.5 m in the high Arctic at the end of winter. Some first-year ice survives the summer and becomes multi-year ice. This ice develops its distinctive hummocky appearance through thermal weathering, becoming harder and almost salt-free over several years. In the present climate, old multi-year ice floes without ridges are about 3 m thick at the end of winter.

The area of sea ice decreases from roughly 15 million km² in March to 7 million km² in September, as much of the first-year ice melts during the summer (Cavaliere et al., 1997). The area of multi-year sea ice, mostly over the

Arctic Ocean basins, the East Siberian Sea, and the Canadian polar shelf, is about 5 million km² (Johannessen et al., 1999). A transpolar drift carries sea ice from the Siberian shelves to the Barents Sea and Fram Strait. It merges on its eastern side with clockwise circulation of sea ice within Canada Basin. On average, 10% of arctic sea ice exits through Fram Strait each year. Section 6.3 provides a full discussion of sea ice in the Arctic Ocean.

Sea ice also leaves the Arctic via the Canadian Archipelago. Joined by seasonal sea ice in Baffin Bay, it drifts south along the Labrador coast to reach Newfoundland in March. An ice edge is established in this location where the supply of sea ice from the north balances the loss by melt in warm ocean waters. Sea-ice production in the source region in winter is enhanced within a polynya (the North Water) formed by the persistent southward drift of ice. Similar “conveyor belt” sea-ice regimes also exist in the Barents and Bering Seas, where northern regions of growth export ice to temperate waters.

First-year floes fracture easily under the forces generated by storm winds. Leads form where ice floes separate under tension, exposing new ocean surface to rapid freezing. Where the pack is compressed, the floes buckle and break into blocks that pile into ridges up to 30 m thick. Near open water, notably in the Labrador, Greenland, and Barents Seas, waves are an additional cause of ridging. Because of ridging and rafting, the average thickness of first-year sea ice is typically twice that achievable by freezing processes alone (Melling and Riedel, 1996). Heavily deformed multi-year floes near the Canadian Archipelago can average more than 10 m thick.

Information on the thickness of northern sea ice is scarce. Weekly records of land-fast ice thickness obtained from drilling are available for coastal locations around the Arctic (Canada and Russia) for the 1940s through the present (Melling, 2002; Polyakov et al., 2003a). Within the Arctic Ocean, there have been occasional surveys of sea ice since 1958, measured with sonar on nuclear submarines (Rothrock et al., 1999; Wadhams, 1997; Winsor, 2001). In Fram Strait and the Beaufort Sea, data have been acquired continuously since 1990 from sonar operated from moorings (Melling, 1993; Melling and Moore, 1995; Melling and Riedel, 1996; Vinje et al., 1998). The average thickness of sea ice in the Arctic Ocean is about 3 m, and the thickest ice (about 6 m) is found along the shores of northern Canada and Greenland (Bourke and Garrett, 1987). There is little information about the thickness of the seasonal sea ice that covers more than half the marine Arctic.

Land-fast ice (or fast ice) is immobilized for up to 10 months each year by coastal geometry or by grounded ice ridges (stamukhi). There are a few hundred meters of land-fast ice along all arctic coastlines in winter. In the present climate, ice ridges ground to form stamukhi in

depths of up to 30 m, as the pack ice is repeatedly crushed against the fast ice by storm winds. In many areas, *stamukhi* stabilize sea ice for tens of kilometers from shore. Within the Canadian Archipelago in late winter, land-fast ice bridges channels up to 200 km wide and covers an area of 1 million km². Some of this ice is trapped for decades as multi-year land-fast ice (Reimnitz et al., 1995). The remobilization of land-fast ice in summer is poorly understood. Deterioration through melting, flooding by runoff at the coast, winds, and tides are contributing factors.

Many potential impacts of climate change will be mediated through land-fast ice. It protects unstable coastlines and coastal communities from wave damage, flooding by surges, and ice ride-up. It offers safe, fast routes for travel and hunting. It creates unique and necessary habitat for northern species (e.g., ringed seal (*Phoca hispida*) birth lairs) and brackish under-ice migration corridors for fish. It blocks channels, facilitating the formation of polynyas important to northern ecosystems in some areas, and impeding navigation in others (e.g., the Northwest Passage).

2.4. Terrestrial water balance

The terrestrial water balance and hydrologic processes in the Arctic have received increasing attention, as it has been realized that changes in these processes will have implications for global climate. There are large uncertainties concerning the water balance of tundra owing to a combination of:

- the sparse network of *in situ* measurements of precipitation and the virtual absence of measurements of evapotranspiration in the Arctic;
- the difficulty of obtaining accurate measurements of solid precipitation in cold windy environments, even at manned weather stations;
- the compounding effects of elevation on precipitation and evapotranspiration in topographically complex regions of the Arctic, where the distribution of observing stations is biased toward low elevations and coastal regions; and
- slow progress in exploiting remote sensing techniques for measuring high-latitude precipitation and evapotranspiration.

Uncertainties concerning the present-day distributions of precipitation and evapotranspiration are sufficiently large that evaluations of recent variations and trends are problematic. The water budgets of arctic watersheds reflect the extreme environment. Summer precipitation plays a minor role in the water balance compared to winter snow, since in summer heavy rains cannot be absorbed by soils that are near saturation. In arctic watersheds, precipitation exceeds evapotranspiration, and snowmelt is the dominant hydrologic event despite its short duration. In the boreal forest, water balance dynamics are dominated by spring snowmelt; water is stored in wetlands, and evapotranspiration is also a

major component in the water balance (Metcalf and Buttle, 1999). Xu and Halldin (1997) suggested that the effects of climate variability and change on streamflow will depend on the ratio of annual runoff to annual precipitation, with the greatest sensitivity in watersheds with the lowest ratios.

2.4.1. Permanent storage of water on land

The great ice caps and ice sheets of the world hold 75% of the global supply of freshwater; of these, the Greenland Ice Sheet contains 2.85 million km³ of freshwater (IPCC, 2001c). The northern portions of mid-latitude cyclones carry most of the water that reaches arctic ice caps, with the result that precipitation generally decreases from south to north. Runoff often exceeds precipitation when ice caps retreat. The behavior of glaciers depends upon climate (see section 6.5).

Temperature and precipitation variations influence the arctic ice caps; for example, temperature increases coupled with decreased precipitation move the equilibrium line (boundary between accumulation and ablation) higher, but with increased precipitation, the line moves lower (Woo and Ohmura, 1997). Small shifts in precipitation could offset or enhance the effect of increasing temperatures (Rouse et al., 1997). Water is also stored in permanent snowfields and firn (compact, granular snow that is over one year old) fields, perched lakes (lakes that are raised above the local water table by permafrost), and as permafrost itself. Whitfield and Cannon (2000) implicated shifts between these types of storage as the source of increases in arctic streamflow during recent warmer periods. The IPCC (2001b) stated: "Satellite data show that there are very likely to have been decreases of about 10% in the extent of snow cover since the late 1960s, and ground-based observations show that there is very likely to have been a reduction of about two weeks in the annual duration of lake and river ice cover in the mid- and high latitudes of the Northern Hemisphere, over the 20th century".

2.4.2. Hydrology of freshwater in the Arctic

The Arctic has four hydrologic periods: snowmelt; outflow breakup period (several days in length but accounting for 75% of total annual flow); a summer period with no ice cover and high evaporation; and a winter period where ice cover thicker than 2 m exists on lakes. Four types of arctic rivers show different sensitivity to climatic variations:

- Arctic–nival: continuous permafrost where deep infiltration is impeded by perennially frozen strata, base flow and winter flow are low, and snowmelt is the major hydrologic event.
- Subarctic–nival: dominated by spring snowmelt events, with peak water levels often the product of backwater from ice jams. Groundwater contributions are larger than those in arctic–nival systems. In some areas, complete winter freezing occurs.

- Proglacial: snowmelt produces a spring peak, but flows continue throughout the summer as areas at progressively higher elevations melt. Ice-dammed lakes are possible.
- Muskeg: large areas of low relief characterized by poor drainage. Runoff attenuation is high because of large water-holding capacity and flow resistance.

Fens (peatlands) are wetlands that depend upon annual snowmelt to restore their water table, and summer precipitation is the most important single factor in the water balance (Rouse, 1998). Actual evapotranspiration is a linear function of rainfall. If summer rainfall decreases, there would be an increase in the severity and length of the water deficit. Water balance has a significant effect on the carbon budget and peat accumulation; under drier conditions, peatlands would lose biomass, and streamflows would decrease. Krasovskaia and Saelthun (1997) found that monthly flow regimes in Scandinavia have stable average patterns that are similar from year to year. They demonstrated that most rivers are very sensitive to temperature rises on the order of 1 to 3 °C, and that nival (snow-dominated) rivers become less stable while pluvial (rain-dominated) rivers become more stable. Land storage of snow is important in the formation of the hydrograph in that the distributed nature of the snow across the land “converts” the daily melt into a single peak. Kuchment et al. (2000) modeled snowmelt and rainfall runoff generation for permafrost areas, taking into account the influence of the depth of thawed ground on water input, water storage, and redistribution.

Where they exist, perennial snow banks are the major source of runoff, and as little as 5% of watershed area occupied by such snow banks will enhance runoff compared to watersheds without them. The resulting stream discharge is termed “proglacial”, and stored water contributes about 50% of the annual runoff. During winter, when biological processes are dormant, the active layer freezes and thaws. Spring snowmelt guarantees water availability about the same time each year, at a time when rainfall is minimal but solar radiation is near its maximum. Summer hydrology varies from year to year and depends upon summer precipitation patterns and magnitudes. Surface organic soils, which remain saturated throughout the year (although the phase changes), are more important hydrologically than deeper mineral soils. During dry periods, runoff is minimal or ceases. During five years of observations at Imnavait Creek, Alaska, an average of 50 to 66% of the snowpack moisture became runoff, 20 to 34% evaporated, and 10 to 19% added to soil moisture storage (Kane et al., 1989). All biological activity takes place in the active layer above the permafrost. Hydraulic conductivity of the organic soils is 10 to 1000 times greater than silt. Unlike the organic layer, the mineral layer remains saturated and does not respond to precipitation events. Soil properties vary dramatically over short vertical distances. The snowmelt period is brief, lasting on the order of 10 days, and peak flow happens within 36 hours of the onset of flow.

Evapotranspiration is similar in magnitude to runoff as a principal mechanism of water loss from a watershed underlain by permafrost. Water balance studies indicate that cumulative potential evaporation is greater than cumulative summer precipitation.

Snowmelt on south-facing slopes occurred one month earlier than on north-facing slopes in subarctic watersheds (Carey and Woo, 1999). On south-facing slopes, the meltwater infiltrated and recharged the soil moisture but there was neither subsurface flow nor actual runoff. The north-facing slopes had infiltration barriers, thus meltwater was impounded in the organic layer and produced surface and subsurface flows. Permafrost slopes and organic horizons are the principal controls on streamflow generation in subarctic catchments. Seppälä (1997) showed that permafrost is confining but not impermeable. Quinton et al. (2000) found that in tundra, subsurface flow occurs predominantly through the saturated zone within the layer of peat that mantles hill slopes, and that water flow through peat is laminar.

Beltaos (2002) showed that temperature increases over the past 80 years have increased the frequency of mild winter days, which has augmented flows to the extent that they can affect breakup processes. There are several implications of this change, including increases in the frequency of mid-winter breakup events; increased flooding and ice-jam damages; delayed freeze-up dates; and advanced breakup dates. Prowse and Beltaos (2002) suggested that climate change may alter the frequency and severity of extreme ice jams, floods, and low flows. These climate-driven changes are projected to have secondary effects on fluvial geomorphology; river modifying processes; aquatic ecology; ice-induced flooding that supplies water and nutrients to wetlands; biological templates; dissolved oxygen depletion patterns; transportation and hydroelectric generation; and ice-jam damage.

The hydrology and the climate of the Arctic are intricately linked. Changes in temperature and precipitation directly and indirectly affect all forms of water on and in the landscape. If the storage and flux of surface water changes, a variety of feedback mechanisms will be affected, but the end result is difficult to project. Snow, ice, and rivers are considered further in Chapters 6 and 8.

2.5. Influence of the Arctic on global climate

2.5.1. Marine connections

Although the marine Arctic covers a small fraction of the globe, positive feedback between the Arctic Ocean and the climate system has the potential to cause global effects. The thermohaline circulation is the global-scale overturning in the ocean that transports significant heat via a poleward flow of warm surface water and an equatorward return of cold, less saline water at depth. The overturning crucial to this transport in the Northern Hemisphere occurs in the Greenland, Irminger, and

Labrador Seas (Broecker et al., 1990). The occurrence and intensity of overturning is sensitive to the density of water at the surface in these convective gyres, which is in turn sensitive to the outflow of low-salinity water from the Arctic. An increase in arctic outflow is very likely to reduce the overturning and therefore the oceanic flux of heat to northern high latitudes. The overturning also moderates anthropogenic impacts on climate because it removes atmospheric CO₂ to the deep ocean. A comprehensive description of the dynamics and consequences of the marine connections is given in section 9.2.3.

2.5.1.1. Ice-albedo feedback to warming and cooling

Sea ice is an influential feature of the marine Arctic. It reflects a large fraction of incoming solar radiation and insulates the ocean waters against loss of heat and moisture during winter. Sea ice also inhibits the movement and mixing of the upper ocean in response to wind. By stabilizing the upper ocean through melting, it may control the global heat sink at high northern latitudes (Manabe et al., 1991; Rind et al., 1995).

The global impact of ice-albedo feedback is predicated on the existence of a strong relationship between atmospheric temperature increases and sea-ice extent. The seasonal analogue of climate change effects on the marine cryosphere is the dramatic expansion of sea-ice extent in winter and its retreat in summer, in tune with (at a lag of several months) the seasonal variation in air temperature. Another relevant analogue is the seasonal progression from frequently clear skies over the marine cryosphere in winter to dominance by fog and stratiform cloud in summer. The increased moisture supply at the melting surface of the ice pack promotes the formation of low clouds that reflect most of the incoming solar radiation in summer, replacing the weakened reflecting capability of melting sea ice. Thus, cloud cover is an important partner to sea ice in the albedo feedback mechanism.

2.5.1.2. Freshwater feedback to poleward transport of heat and freshwater

Deep convection in the northwest Atlantic Ocean is a crucial part of the global thermohaline circulation. Water freshened by arctic outflow is cooled, causing it to sink deep into the ocean, from where it flows either south to the North Atlantic or north into the Arctic Basin (Aagaard and Greisman, 1975; Nikiforov and Shpaikher, 1980). Deep convection has considerable interannual variability controlled by atmospheric circulation. It operates to link the stochastic effects of atmospheric variability to slow oscillations in the ocean–atmosphere system via the oceanic transports of heat and “freshwater” in the global thermohaline circulation (Broecker, 1997, 2000).

The Greenland Sea is one region where new deep water forms (Swift and Aagaard, 1981). Here, warm and saline water of Atlantic origin meets cold arctic water

of lower salinity. Extremely low temperatures cause rapid cooling of the sea surface, which may trigger either deep convective mixing or intensified ice formation, depending on the density of waters at the sea surface. Convection can reach depths of about 2000 m (Visbeck et al., 1995) and the temperature change in the water at that depth is an indicator of the intensity of deep-water formation, with warmer temperatures indicating less deep-water formation. Observations show periods of deep-water temperature increases in the Greenland Sea in the late 1950s and between 1980 and 1990, and temperature decreases in the early 1950s and in the late 1960s. A large increase (0.25 °C) in deep-water temperature occurred in the 1990s (Alekseev et al., 2001). The decrease in deep-water formation implied by increasing deep-water temperatures has weakened the thermohaline circulation, leading to a decreased overflow of deep water through the Faroe–Shetland channel (Hansen et al., 2001).

A reduction in the vertical flux of salt and reduced deep-water formation is likely to trigger a prolonged weakening of the global thermohaline circulation. With less bottom-water formation, there is likely to be a reduction in upwelling at temperate and subtropical latitudes. Paleoclimatic shifts in the thermohaline circulation have caused large and sometimes abrupt changes in regional climate (section 2.7). Dickson et al. (2002) demonstrated that the flows of dense cold water over sills in the Faroe–Shetland Channel and in Denmark Strait are the principal means of ventilating the deep waters of the North Atlantic. Both the flux and density structure of “freshwater” outflow to the North Atlantic are critical to the arctic influence on global climate via the thermohaline circulation (Aagaard and Carmack, 1989).

2.5.2. Sea level

Global average sea level rose between 0.1 and 0.2 m during the 20th century (IPCC, 2001b), primarily because of thermal expansion of warming ocean waters. Although the thermal expansion coefficient for seawater is small, integrated over the 6000 m depth of the ocean the resulting change in sea level can generate changes of significance to ecosystems and communities near coastlines. The warming of arctic seawater will have a negligible impact on local sea level because cold (<0 °C) seawater expands very little with an increase in temperature. However, arctic sea level will respond to changes in the levels of the Atlantic and Pacific Oceans via dynamic links through Bering Strait, Fram Strait, and the Canadian Archipelago. In many areas of the Arctic, sea level is also changing very rapidly as a result of postglacial rebound of the earth’s crust. For example, the land at Churchill, Canada (on the western shore of Hudson Bay), rose one meter during the 20th century. In many parts of the Arctic, changes in the elevation of the shoreline due to crustal rebound are likely to exceed the rise in sea level resulting from oceanic warming.

The Arctic Ocean stores a large volume of “freshwater”. Arctic sea level is sensitive to “freshwater” storage and will rise if this inventory increases, or fall if “freshwater” storage declines. Changes in northern hydrology are therefore likely to have an important effect on arctic sea level by changing “freshwater” storage in the Arctic.

On a timescale of centuries, and with a sufficient increase in temperature, accumulation or ablation of terrestrial ice caps in Greenland and Antarctica are very likely to be the dominant causes of global changes in sea level. There is an interesting aspect to sea-level change in the vicinity of these ice caps: a sea-level increase caused by the ice cap melting, distributed globally, may be offset by changes in the local gravitational anomaly of the ice, which pulls the sea level up towards it. As a result, it is possible that sea level could actually drop at locations within a few hundred kilometers of Greenland, despite an average increase in sea level worldwide. Sections 6.5 and 6.9 provide further details related to ice caps, glaciers, and sea-level rise.

2.5.3. Greenhouse gases

Arctic ecosystems are characterized by low levels of primary productivity, low element inputs, and slow element cycling due to inhibition of these processes by very cold climatic conditions. However, arctic ecosystems still tend to accumulate organic matter, carbon (C), and other elements because decomposition and mineralization processes are equally inhibited by the cold, wet soil environment (Jonasson et al., 2001). Owing to this slow decomposition, the total C and element stocks of wet and moist arctic tundra frequently equal or exceed stocks of the same elements in much more productive ecosystems in temperate and tropical latitudes. Methane (CH₄) production, for example, is related to the position of the water table in the active layer, which will be affected by changes in active-layer depth and/or permafrost degradation. Natural gas hydrates are also found in the terrestrial Arctic, although only at depths of several hundred meters. Currently, arctic and alpine tundra is estimated to contain 96×10^{12} kg of C in its soil and permafrost. This is roughly 5% of the world’s soil C pool (IPCC, 2001c). An additional 5.7×10^{12} kg of C is stored in arctic wetland, boreal, and tundra vegetation, for a total of 102×10^{12} kg of terrestrial C (Jonasson et al., 2001). This is fully discussed in section 7.4.2.1. Thawing of permafrost has the potential to release large stores of CO₂ and CH₄ that are presently contained in frozen arctic soils, both as a direct consequence of thawing and as an indirect consequence of changes in soil wetness (Anisimov et al., 1997; Fukuda, 1994). Although it is not clear whether the Arctic will be a net source or sink of C in the future, the large amounts of C that could be taken up or released make improved understanding of arctic processes important.

The Arctic Ocean was not initially believed to be a significant sink of C because its sizeable ice cover prevents atmosphere–ocean exchange and biological production

in the central ocean was believed to be small. Under warmer climate conditions, however, the amount of C that the Arctic Ocean can sequester is likely to increase significantly. In the northern seas, hydrated CH₄ is trapped in solid form at shallow depths in cold sediments. Gas hydrates are likely to decompose and release CH₄ to the atmosphere if the temperature of water at the seabed rises by a few degrees (Kennett et al., 2000) over centuries to a millennium. This is discussed further in section 9.5.5.

2.6. Arctic climate variability in the twentieth century

2.6.1. Observing systems and data sources

All arctic countries maintain programs of synoptic observations to support their economic activity and the sustainability of communities in the Arctic. Due to the harsh environment and the sparseness of the observation network, the need for meteorological observations is often a major (or even the only) reason for the existence of many arctic settlements. Systematic *in situ* arctic meteorological observations started in the late 18th century in the Atlantic sector (Tabony, 1981). In Fennoscandia, the oldest systematic climatic observations north of 65° N were made in Tornio, Finland, between 1737 and 1749, and regular weather stations were established around 1850. At Svalbard, the first permanent weather station was established in 1911. The first station in the Russian north was established at Arkhangelsk in 1813. Most of the meteorological network in central and northern Alaska was established in the 1920s, with the first station, Kotzebue, opening in 1897. The first meteorological observations in southern Alaska (Sitka at 57° N) were made in 1828. In northern Canada, systematic meteorological observations started in the 1940s.

Meteorological observations in the Arctic Ocean began with the first research voyage of Fridtjof Nansen onboard *Fram* (1894–1896). Additional observations were made during the 1920s and 1930s by ships trapped in the pack ice. A new phase of Arctic Ocean observations began in the mid-1930s with the establishment of North Pole ice stations (Arctic Climatology Project, 2000).

Economic issues led to a significant reduction in the existing meteorological network in northern Russia and Canada in the 1990s. Thus, during the past decade, the number of arctic meteorological stations has noticeably decreased, and the number of the stations conducting atmospheric measurements using balloons has decreased sharply.

The national meteorological services of the Nordic countries, Canada, Russia, and the United States maintain extensive archives of *in situ* observations from their national networks. The station density in these networks varies substantially, from 2 per 1000 km² in Fennoscandia to 1 per 100 000 km² in Canada north of

60° N, northern Alaska, and (since the 1990s) in northern regions of Siberia.

In seeking to assemble a high-quality record, different levels of quality assurance, data infilling, and homogenization adjustments are required. The Global Historical Climatology Network (GHCN) dataset includes selected quality controlled long-term stations suitable for climate change studies, while the Global Daily Climatology Network dataset goes through a more limited screening. The Integrated Surface Hourly Dataset incorporates all synoptic observations distributed through the Global Telecommunication System during the past 30 years. All rawinsonde observations in the Arctic (300 stations north of 50° N and 135 stations north of 60° N) are currently collected in the Comprehensive Aerological Reference Data Set.

The sea-ice boundaries in the Atlantic sector of the Arctic Ocean have been documented since the beginning of the 20th century. Since the late 1950s, sea-ice observations have been conducted throughout the year. The development of shipping along the coast of Siberia in the mid-1920s led to sea-ice monitoring in Siberian Arctic waters. By the late 1930s, aviation had become the main observation tool; since the 1970s, satellite remote sensing has been used. The notes of seamen and the logs of fishing, whaling, and sealer vessels operating in arctic waters serve as an important source of information about changes in the state of arctic sea ice throughout the 19th and 20th centuries (Vinje, 2001). A significant amount of historical data on sea ice near the shores of Iceland was preserved and generalized in many studies (Ogilvie and Jónsdóttir, 2000). Information on sea-ice thickness is scarce; observations of ice draft using upward-looking sonar from submarines and stationary systems are the primary source of this information (see section 2.3.4).

Sea-ice data are concentrated at two World Data Centers. Datasets of satellite observations of sea-ice and snow-cover extent (Ramsay, 1998), snow water equivalent from the Special Sensor Microwave Imager (Armstrong and Brodzik, 2001; Grody and Basist, 1996), cloudiness (Rossow and Schiffer, 1999), and the radiation budget (Wielicki et al., 1995) are available from the National Aeronautics and Space Administration (Goddard Institute for Space Studies, Langley Atmospheric Sciences Data Center) and the National Oceanic and Atmospheric Administration (National Climatic Data Center, National Snow and Ice Data Center – NSIDC). A suite of arctic-related datasets is available from the NSIDC (<http://nsidc.org/index.html>).

Oceanographic measurements in the central Arctic were initiated in 1894 by Nansen (1906). They were restarted in the 1930s, interrupted during the Second World War, and resumed in the late 1940s with the help of aviation and drifting stations. Since 1987, icebreakers with conductivity, temperature, and depth sondes have been used to make observations.

The hydrologic network in the Arctic is probably the weakest of the arctic observation networks, which makes information about the arctic water budget quite uncertain (Vörösmarty et al., 2001). A circumpolar river discharge dataset is available online (R-ArcticNET, 2003).

2.6.2. Atmospheric changes

2.6.2.1. Land-surface air temperature

Although several analyses (e.g., Comiso, 2003; Polyakov et al., 2003b) have examined large-scale temperature variations in the Arctic, no study has evaluated the spatial and temporal variations in temperature over all land areas in the zone from 60° to 90° N for the entire 20th century. Consequently, temperature trends are illustrated for this latitude band using the Climatic Research Unit (CRU) database (Jones and Moberg, 2003) and the GHCN database (updated from Peterson and Vose, 1997). Both databases were used in the IPCC Third Assessment Report (IPCC, 2001c) to summarize the patterns of temperature change over global land areas since the late 19th century. While the impact of urbanization on large-scale temperature trends in the Arctic has not been assessed, the results of Jones et al. (1990), Easterling et al. (1997), Peterson et al. (1999), and Peterson (2003) indicate that urbanization effects at the global, hemispheric, and even regional scale are small (<0.05 °C over the period 1900 to 1990).

Figure 2.6 depicts annual land-surface air temperature variations in the Arctic (north of 60° N) from 1900 to 2003 using the GHCN dataset. The CRU time series is virtually identical to the GHCN series, and both document a statistically significant warming trend of 0.09 °C/decade during the period shown (Table 2.1). The arctic trend is greater than the overall Northern Hemisphere trend of 0.06 °C/decade over the same period (IPCC, 2001b). In general, temperature increased from 1900 to the mid-1940s, then decreased

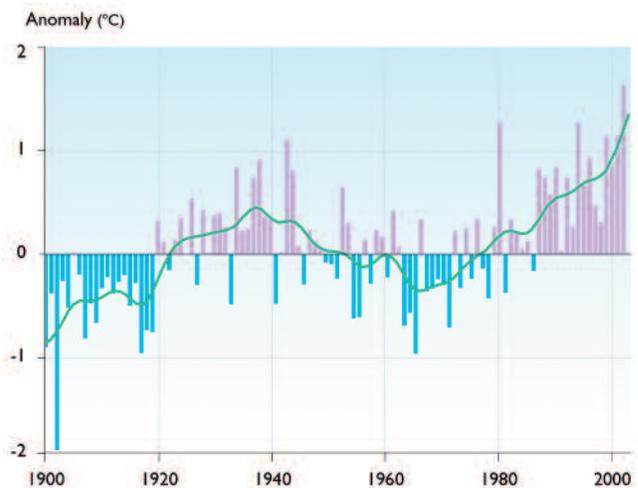


Fig. 2.6. Annual anomalies of land-surface air temperature in the Arctic (60° to 90° N) for the period 1900 to 2003 using the GHCN dataset (updated from Peterson and Vose, 1997). Anomalies are calculated relative to the 1961–1990 average. The smoothed curve was created using a 21-point binomial filter, which approximates a 10-year running mean.

Table 2.1. Least-squares linear trends in annual anomalies of arctic (60° to 90° N) land-surface air temperature (°C/decade) from the GHCN (updated from Peterson and Vose, 1997) and CRU (Jones and Moberg, 2003) datasets. Anomalies are calculated relative to the 1961–1990 average.

	1900–2003	1900–1945	1946–1965	1966–2003
GHCN dataset	0.09 ^a	0.29 ^a	-0.14	0.40 ^a
CRU dataset	0.09 ^a	0.31 ^a	-0.20	0.40 ^a

^avalue significant at the 1% level of confidence or better

until about the mid-1960s, and increased again thereafter. The general features of the arctic time series are similar to those of the global time series, but decadal trends and interannual variability are greater in the Arctic. Comparisons of the arctic and global time series are discussed later in this section.

Figure 2.7 illustrates the patterns of land-surface air temperature change in the Arctic between 1900 and 2003 and for three periods therein. Trends were calculated from annually averaged gridded anomalies using the method of Peterson et al. (1999) with the requirement that annual anomalies include a minimum of ten months of data. For the period 1900 to 2003, trends were calculated only for those 5° x 5° grid boxes with annual anomalies in at least 70 of the 104 years. The minimum number of years required for the shorter time periods (1900–1945, 1946–1965, and 1966–2003) was 31, 14, and 26, respectively. The three periods selected for this figure correspond approximately to the warming–cooling–warming trends seen in Fig. 2.6. The spatial coverage of the region north of 60° N is quite varied. During the first period (1900–1945), there are only three stations meeting the requirement of 31 years of data in Region 4 and only four in Region 3 (see section 18.3 for map and description of the ACIA regions). The highest density of stations is found in Region 1. The coverage for the second and third periods is more uniform.

Although it is difficult to assess trends in some areas, air temperature appears to have increased throughout the Arctic during most of the 20th century. The only period characterized by widespread cooling was 1946 to 1965, and even then large areas (e.g., southern Canada and southern Eurasia) experienced significant increases in temperature. Temperatures in virtually all parts of the Arctic increased between 1966 and 2003, with trends exceeding 1 to 2 °C/decade in northern Eurasia and northwestern North America. The average trend between 1966 and 2003 over the Arctic was 0.4 °C/decade, approximately four times greater than the average for the century. While most pronounced in winter and spring, all seasons experienced an increase in temperature over the past several decades (Fig. 2.8). The trends shown in Fig. 2.8 were calculated from seasonally-averaged gridded anomalies using the method of Peterson et al. (1999) with the requirement that the calculation of seasonal anomalies should include all three months. Trends were calculated only for those 5° x 5° grid boxes containing seasonal anomalies in at least 26 of the 38 years. The updated analysis by Chapman and Walsh (2003) showed generally similar

patterns of change over the period 1954 to 2003, but with areas of cooling around southern Greenland.

The instrumental record of land-surface air temperature is qualitatively consistent with other climate records in the Arctic (Serreze et al., 2000). For instance, temperatures in the marine Arctic (as measured by coastal land stations, drifting ice stations, and Russian North Pole stations) increased at the rate of 0.05 °C/decade during the 20th century (Polyakov et al., 2003b). As with the land-only record, the increases were greatest in winter and spring, and there were two relative maxima during the century (the late 1930s and the 1990s). For periods since 1950, Polyakov et al. (2003b) found the rate of temperature increase in the marine Arctic to be similar to that noted for the GHCN dataset. Because of the scarcity of data prior to 1945, it is very difficult to say whether the Arctic as a whole was as warm in the 1930s and 1940s as it was during the 1990s. In the Polyakov et al. (2003b) analysis, only coastal stations were chosen and most of the stations contributing to the high average temperatures in the 1930s were in Scandinavia. Interior stations, especially those between 60° N (the southern limit for the analysis in this section) and 62° N (the southern limit for the Polyakov et al. (2003b) study), have warmed more than coastal stations over the past few decades. As discussed in section 2.6.3, arctic sea-ice extent contracted from 1918 to 1938 and then expanded between 1938 and 1968 (Zakharov, 2003). The expansion after 1938 implies that the Arctic was cooling during that period.

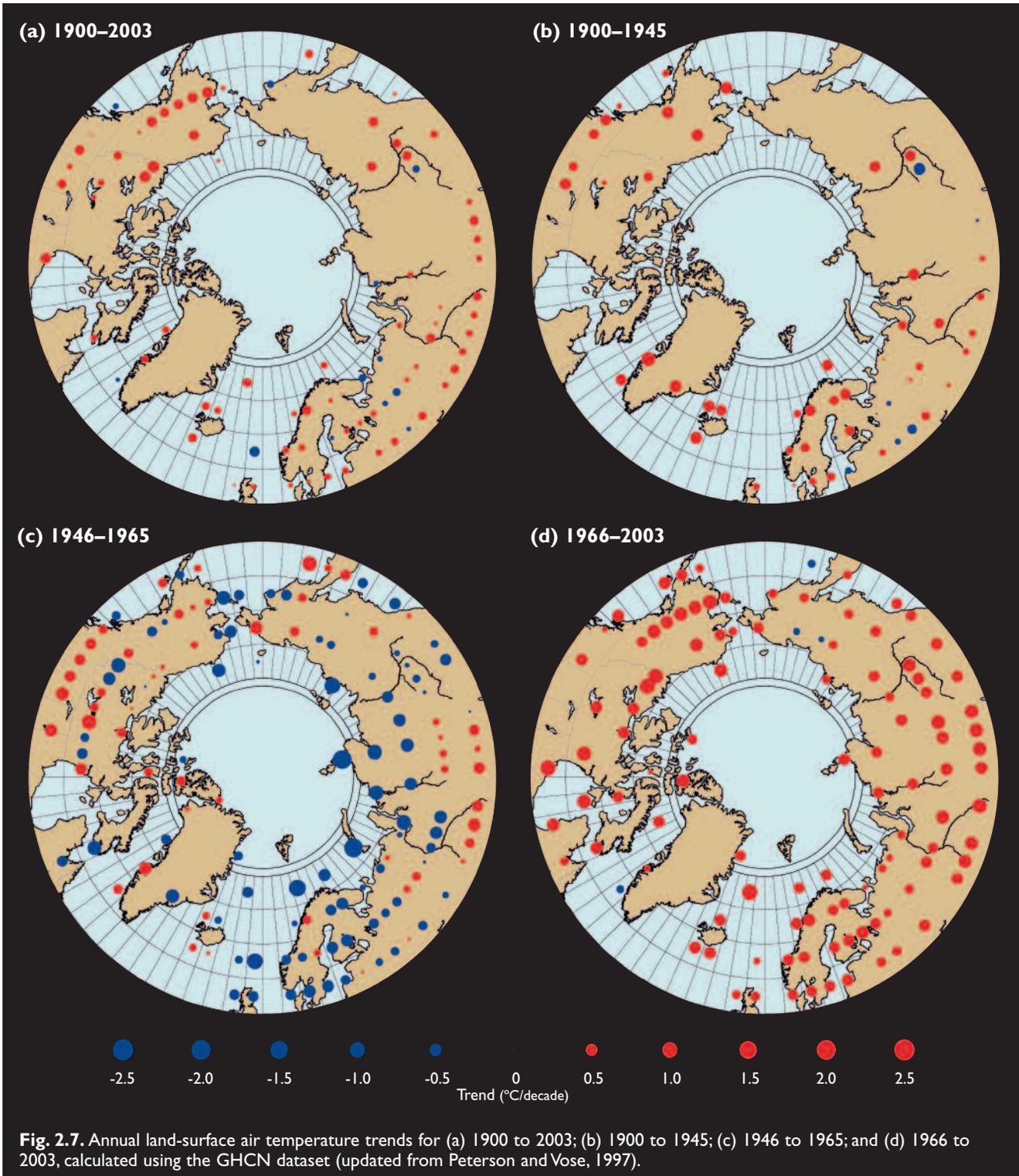
Polyakov et al. (2002) presented the temperature trends (°C/yr) for their dataset (as described above) and for the Jones et al. (1999) Northern Hemisphere dataset. For comparison purposes, the temperature trends (°C/yr) for the land-surface temperatures (GHCN dataset) were computed for the latitude bands 60° to 90° N and 0° to 60° N (Fig. 2.9). The trend shown for any given year before present is the average trend from that year through 2003; for example, the value corresponding to 60 years before present is the average trend for the period 1944 to 2003. In both latitudinal bands, the trend over any period from 120 years ago to the present is positive (i.e., the Arctic is warming). Although the trends for both bands have been increasing over the past 60 years, the trend in the 60° to 90° N band is larger. The rate of temperature increase in the Arctic (as defined here) exceeds that of lower latitudes. Due to natural variability and sparse data in the Arctic, the arctic trend shows more variability and the confidence limits are wider. Over the past 40 years, the arctic warming

trend was about $0.04\text{ }^{\circ}\text{C}/\text{yr}$ ($0.4\text{ }^{\circ}\text{C}/\text{decade}$) compared to a trend of $0.025\text{ }^{\circ}\text{C}/\text{yr}$ for the lower latitudes.

Likewise, surface temperatures derived from satellite thermal infrared measurements, which provide circum-polar coverage from 1981 to 2001, exhibited statistically significant warming trends in all areas between 60° to 90° N except Greenland (Comiso, 2003). The warming trends were $0.33\text{ }^{\circ}\text{C}/\text{decade}$ over the sea ice, $0.50\text{ }^{\circ}\text{C}/\text{decade}$ over Eurasia, and $1.06\text{ }^{\circ}\text{C}/\text{decade}$ over North America. In addition, the recent reduction in sea-ice thickness (Rothrock et al., 1999), the retreat of sea-ice cover (Parkinson et al., 1999), and the decline in peren-

nial sea-ice cover (Comiso, 2002) are consistent with large-scale warming in the Arctic. In view of this evidence, it is very probable that the Arctic has warmed over the past few decades.

Global climate model simulations generally indicate that increasing atmospheric GHG concentrations will result in greater temperature increases in the Arctic than in other parts of the world. As stated by the IPCC (2001c), model experiments show “a maximum warming in the high latitudes of the Northern Hemisphere”. In reference to warming at the global scale, the IPCC (2001b) also concluded, “There is new and stronger evidence that

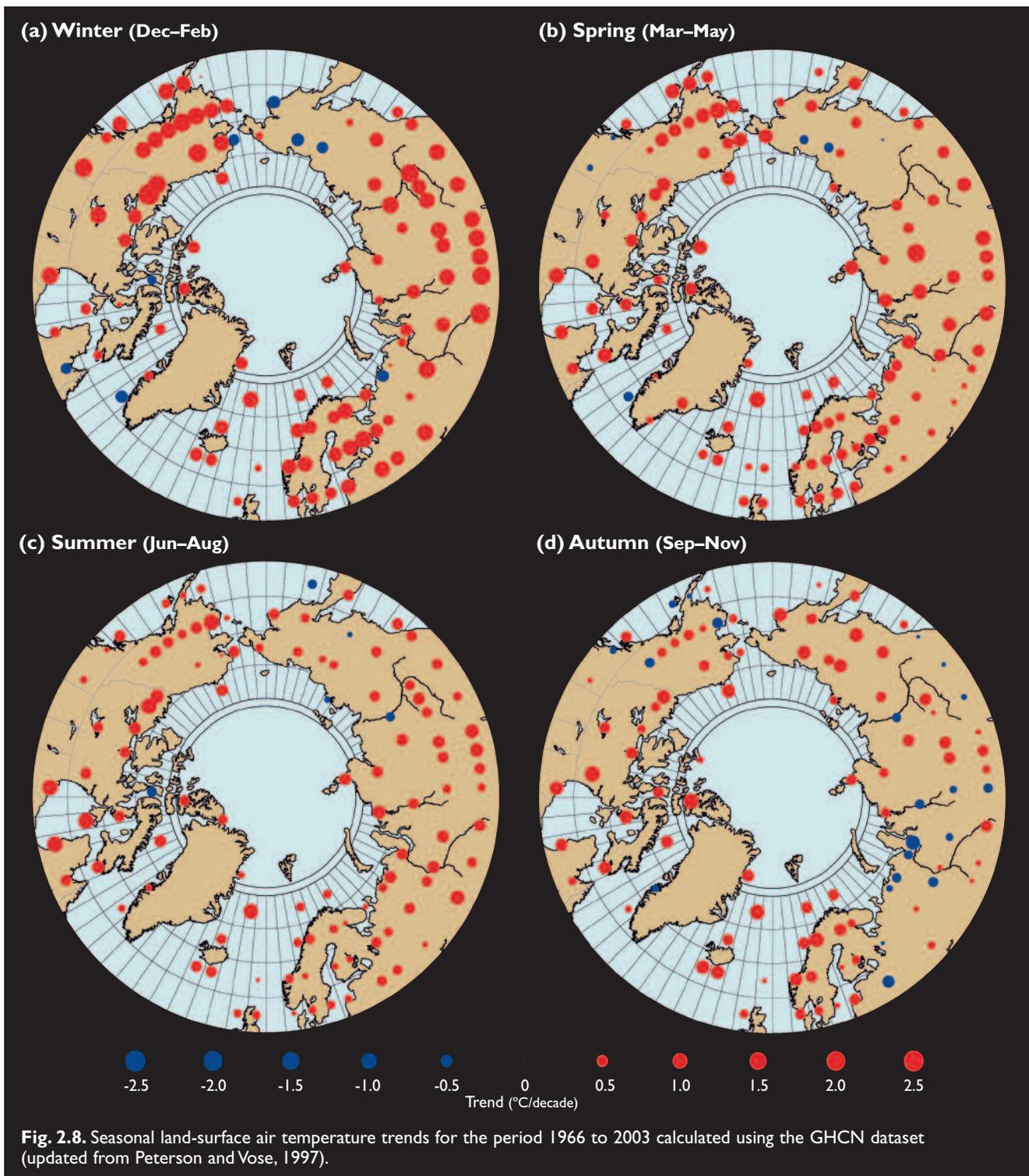


most of the warming observed over the past 50 years is attributable to human activities”.

The question is whether there is definitive evidence of an anthropogenic signal in the Arctic. This would require a direct attribution study of the Arctic, which has not yet been done. There are studies showing that an anthropogenic warming signal has been detected at the regional scale. For example, Karoly et al. (2003) concluded that temperature variations in North America during the second half of the 20th century were probably not due to natural variability alone. Zwiers and Zhang (2003) were able to detect the combined effect of changes in

GHGs and sulfate aerosols over both Eurasia and North America for this period, as did Stott et al. (2003) for northern Asia (50°–70° N) and northern North America (50°–85° N). In any regional attribution study, the importance of variability must be recognized. In climate model simulations, the arctic signal resulting from GHG-induced warming is large but the variability (noise) is also large. Hence, the signal-to-noise ratio may be lower in the Arctic than at lower latitudes. In the Arctic, data scarcity is another important issue.

A related question is whether the warming in the Arctic is enhanced relative to that of the globe (i.e., polar



amplification). For example, Polyakov et al. (2002) concluded that observed trends in the Arctic over the entire 20th century did not show polar amplification. A number of studies (e.g., Comiso, 2003; Thompson and Wallace, 1998) suggested that much of the recent warming resulted from changes in atmospheric circulation; specifically, during the past two decades when the AO was in a phase that brought relatively warm air to the same areas that experienced the greatest increases in temperature (i.e., northern Eurasia and northwestern North America). Serreze et al. (2000), however, noted that the changes in circulation were “not inconsistent” with anthropogenic forcing.

It is clear that trends in temperature records, as evidence of polar amplification, depend on the timescale chosen. Over the past 100 years, it is possible that there has been polar amplification but the evidence does not allow a firm conclusion. As noted previously, the analysis presented here of land stations north of 60° N does show more warming over the past 100 years (0.09 °C/decade for the Arctic compared to 0.06 °C/decade for the

globe). However, an analysis of coastal stations north of 62° N (Polyakov et al., 2002) found arctic warming of 0.05 °C/decade over the past 100 years. Johannessen et al. (2004) found, with a more extensive dataset, that the “early warming trend in the Arctic was nearly as large as the warming trend for the last 20 years” but “spatial comparison of these periods reveals key differences in their patterns”. The pattern of temperature increases over the past few decades is different and more extensive than the pattern of temperature increases during the 1930s and 1940s, when there was weak (compared to the present) lower-latitude warming.

In conclusion, for the past 20 to 40 years there have been marked temperature increases in the Arctic. The rates of increase have been large, and greater than the global average. Two modeling studies have shown the importance of anthropogenic forcing over the past half century for modeling the arctic climate. Johannessen et al. (2004) used a coupled atmosphere–ocean general circulation model (AOGCM) to study the past 100 years and noted, “It is suggested strongly that whereas the earlier warming was natural internal climate-system variability, the recent SAT (surface air temperature) changes are a response to anthropogenic forcing”. Goosse and Renssen (2003) simulated the past 1000 years of arctic climate with a coarser resolution AOGCM and were able to replicate the cooling and warming until the mid-20th century. Without anthropogenic forcing, the model simulates cooling after a temperature maximum in the 1950s. There is still need for further study before it can be firmly concluded that the increase in arctic temperatures over the past century and/or past few decades is due to anthropogenic forcing.

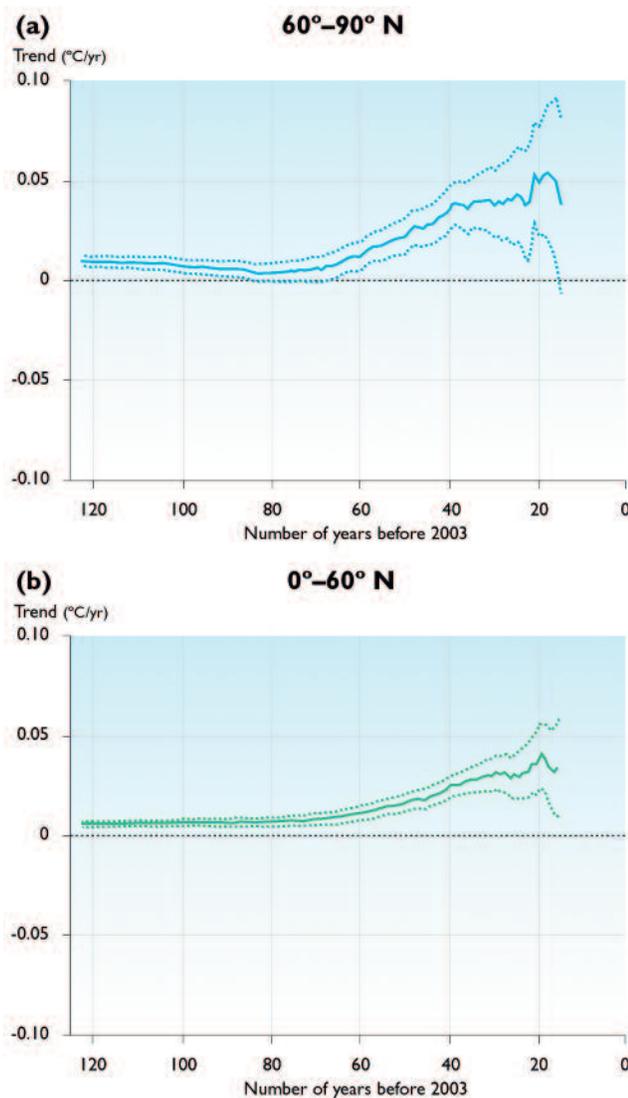


Fig. 2.9. Trends in annual mean land-surface air temperatures (solid lines) and their 95% significance levels (dashed lines) over the past 120 years for (a) 60° to 90° N and (b) 0° to 60° N (data from the GHCN dataset, updated from Peterson and Vose, 1997).

2.6.2.2. Precipitation

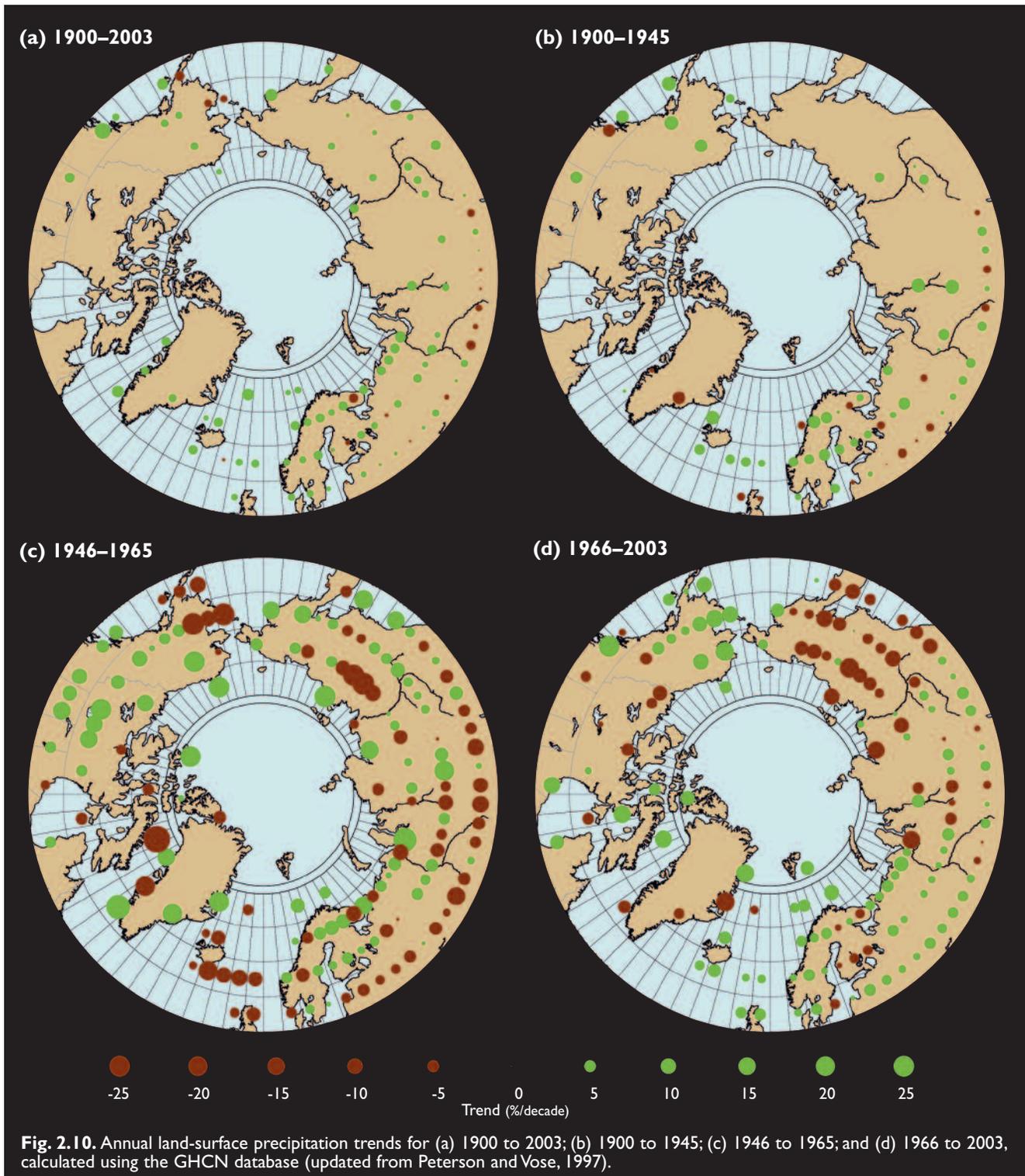
Precipitation assessments in the Arctic are limited by serious problems with the measurement of rainfall and snowfall in cold environments (Goodison et al., 1998). Methods for correcting measured precipitation depend upon observing practices, which differ between countries; in addition, metadata (information about the data) needed to perform corrections are often inadequate. The IPCC (2001b) concluded, “It is very likely that precipitation has increased by 0.5 to 1% per decade in the 20th century over most mid- and high latitudes of the Northern Hemisphere continents”.

An updated assessment of precipitation changes north of 60° N is presented in Figs. 2.10 and 2.11. These figures provide estimates of linear trends in annual and seasonal precipitation based on available data in the GHCN database (updated from Peterson and Vose, 1997). Trends were calculated from annually-averaged gridded anomalies (baseline 1961–1990) using the method of Peterson et al. (1999) with the requirement that annual anomalies include a minimum of ten months of data (Fig. 2.10) and that seasonal anomalies must contain all three months (Fig. 2.11). In Fig. 2.10, for the period 1900 to 2003 trends were calculated only for those 5° x 5° grid boxes containing annual

anomalies in at least 70 of the 104 years; the minimum number of years required for the shorter time periods (1900–1945, 1946–1965, and 1966–2003) was 31, 14, and 26 respectively. In Fig. 2.11, trends were calculated only for those $5^\circ \times 5^\circ$ grid boxes containing seasonal anomalies in at least 26 of the 38 years.

For the entire period from 1900 to 2003, Table 2.2 indicates a significant positive trend of 1.4% per decade. New et al. (2001) used uncorrected records and found that terrestrial precipitation averaged over the 60° to 80° N band exhibited an increase of 0.8% per decade over the period from 1900 to 1998. In gen-

eral, the greatest increases were observed in autumn and winter (Serreze et al., 2000). Fig. 2.10 shows that most high-latitude regions have experienced an increase in annual precipitation. During the arctic warming in the first half of the 20th century (1900–1945), precipitation increased by about 2% per decade, with significant positive trends in Alaska and the Nordic region. During the two decades of arctic cooling (1946–1965), the high-latitude precipitation increase was roughly 1% per decade. However, there were large regional contrasts with strongly decreasing values in western Alaska, the North Atlantic region, and parts of Russia. Since 1966, annual precipitation



has increased at about the same rate as during the first half of the 20th century. In eastern Russia, annual precipitation decreased, mostly because of a substantial decrease during winter (Fig. 2.11). Average Alaskan precipitation has increased in all seasons.

These trends are in general agreement with the results of a number of regional studies. For example, there have been positive trends in annual precipitation (up to a 20% increase) during the past 40 years over Alaska (Karl et al., 1993), Canada north of 55° N (Mekis and Hogg, 1999), and the Russian permafrost-

free zone (Groisman and Rankova, 2001). Siberian precipitation exhibited little change during the second half of the 20th century (Groisman et al., 1991, updated). Hanssen-Bauer et al. (1997) documented that annual precipitation in northern Norway increased by 15% during the 20th century. Førland et al. (1997) and Hanssen-Bauer and Førland (1998) found precipitation increases of about 25% over the past 80 years in Svalbard. Lee et al. (2000) described details of specific periods and changes in annual precipitation, which generally increased between 1880 and 1993 in the Arctic.

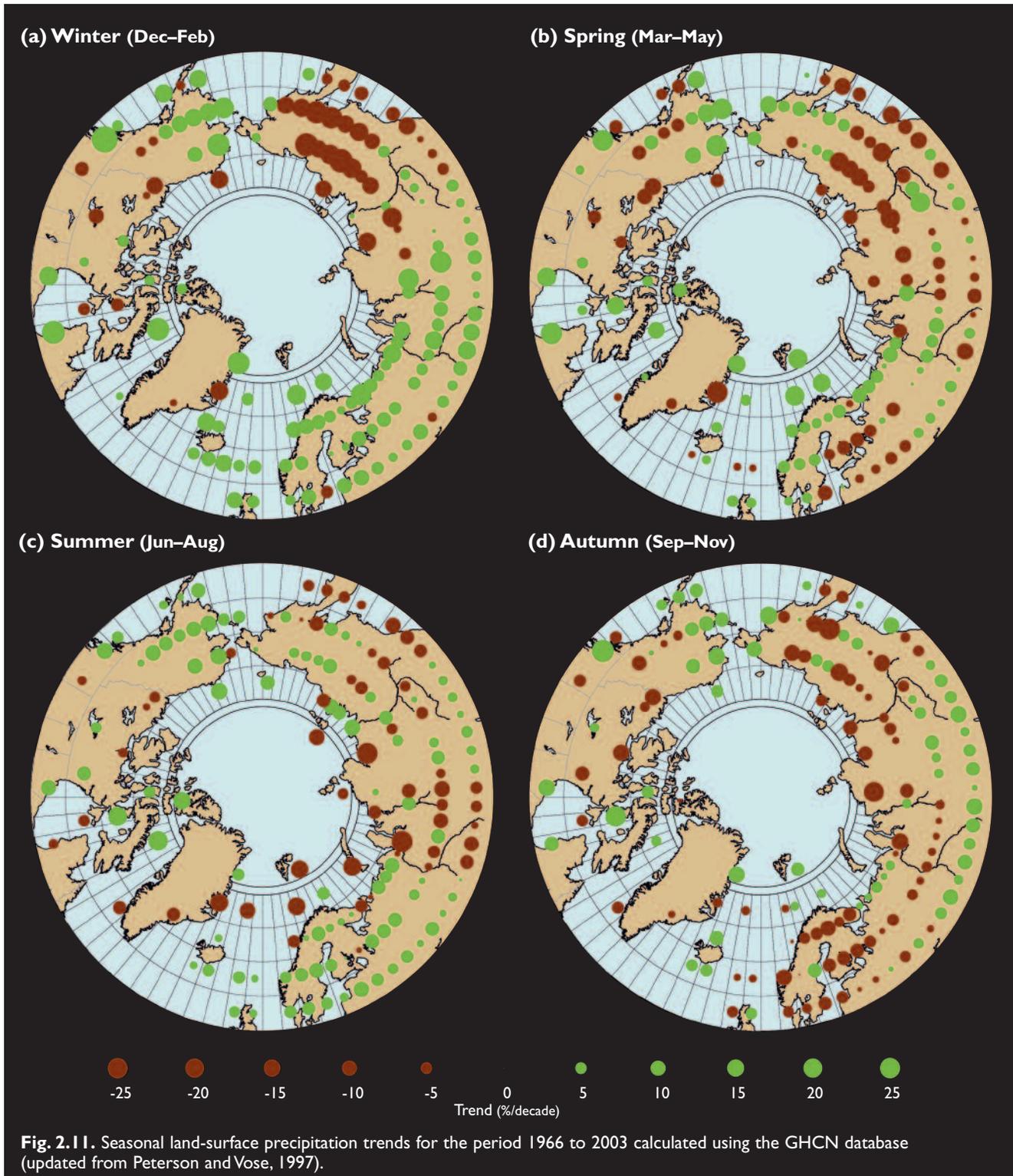


Table 2.2. Least-squares linear trends in annual anomalies of arctic (60° to 90° N) precipitation (% per decade) from the GHCN database (updated from Peterson and Vose, 1997). Anomalies are calculated relative to the 1961–1990 average.

	1900–2003	1900–1945	1946–1965	1966–2003
GHCN	1.4 ^a	2.3 ^a	1.3	2.2
^a value significant at the 1% level of confidence or better				

Overall, it is probable that there was an increase in arctic precipitation over the past century.

A few studies suggest that the fraction of annual precipitation falling as snow has diminished, which is consistent with widespread temperature increases in the Arctic. For the period 1950 to 2001, Groisman et al. (2003) found an increase of 1.2% per decade in annual rainfall north of 50° N, an increase that was partially due to an additional fraction of liquid precipitation during spring and autumn. In addition, Førland and Hanssen-Bauer (2003) found that the fraction of solid precipitation diminished at all stations in the Norwegian Arctic between 1975 and 2001. If the fraction of precipitation falling in solid form has decreased, it is possible that the arctic trend in total precipitation may be overestimated because gauge under-catch is less problematic during liquid precipitation events (Førland and Hanssen-Bauer, 2000).

Reporting changes in precipitation frequency in cold environments is difficult because many events result in small precipitation totals and many gauges have accuracy problems during light precipitation events. To avoid uncertainties in the measurement of small amounts of precipitation, Groisman et al. (2003) assessed the climatology and trends in the annual number of days with precipitation >0.5 mm for the period 1950 to 2001. They found no significant change in the number of wet days over North America and northern Europe (including western Russia) but a significant decrease in wet days over eastern Russia. The latter was first reported by Sun et al. (2001) for eastern Russia south of 60° N and is remarkable because it is accompanied by an increase in heavy precipitation in the same region (Groisman et al., 1999a,b; Sun et al., 2001).

The IPCC (2001c) concluded that it was likely that there had been a widespread increase in heavy and extreme precipitation events in regions where total precipitation had increased (e.g., in the high latitudes of the Northern Hemisphere). Subsequent to the study of Groisman et al. (1998, 1999a), which reported that heavy precipitation events appear to have increased at high latitudes, a series of analyses demonstrated that changes in heavy precipitation over Canada are spatially diverse (Akinremi et al., 1999; Mekis and Hogg, 1999; Stone et al., 2000). Mekis and Hogg (1999) noted that the magnitude of heavy precipitation events making up the top 10% of such events increased in northern Canada but decreased in the south during the 20th-century period of record. Zhang X. et al. (2001) reported a marked decadal

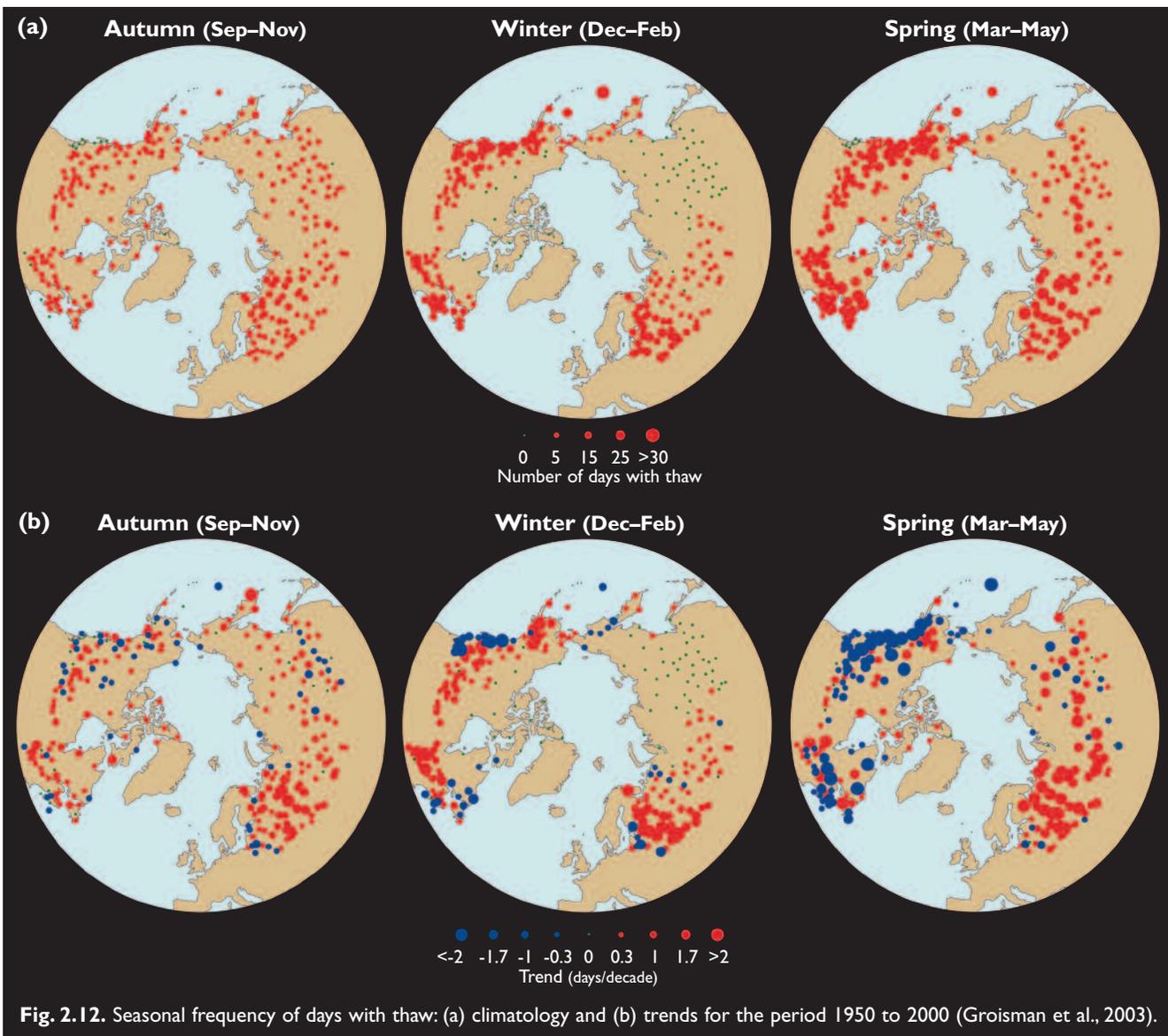
increase in heavy snowfall events in northern Canada. Stone et al. (2000) showed that during the second half of the 20th century, heavy and intermediate daily precipitation events became significantly more frequent in over one-third of all districts of Canada, indicating a shift to more intense precipitation. Groisman et al. (2002) reported an increase in the frequency of “very heavy” (upper 0.3%) rain events over the European part of Russia during the past 65 years. Comparing two multi-decadal periods in the second half of the 20th century, Frich et al. (2002) reported an increase in most indicators of heavy precipitation events over western Russia, eastern Canada, and northern Europe, while Siberia showed a decrease in the frequency and/or duration of heavy precipitation events.

2.6.2.3. Sea-level pressure

Sea-level pressure data from the Arctic Buoy Program (1979–1994) show decreasing pressures in the central arctic that are greater than decreases in sea-level pressure found anywhere else in the Northern Hemisphere (Walsh et al., 1996). These decreases are greatest and statistically significant in autumn and winter, and are compensated by pressure increases over the subpolar oceans. It is very probable that sea-level pressure has been dropping over the Arctic during the past few decades. A useful review of other evidence for these and related changes, and their linkages to the AO, was undertaken by Moritz et al. (2002). A similar pattern of sea-level pressure change is simulated by some climate models when forced with observed variations in GHGs and tropospheric sulfate aerosols (Gillett et al., 2003). However, for unknown reasons, the magnitude of the anthropogenically forced pressure change in climate models is generally much less than the observed pressure change. Longer-term (>100 years) observations, based in part on recently released Russian meteorological observations from coastal stations, show that the decrease in central arctic sea-level pressure during recent decades is part of a longer-term (50 to 80 year) oscillation that peaked in the late 1800s and again in the 1950s and 1960s (Polyakov et al., 2003b). This suggests that the decrease in central arctic sea-level pressure during recent decades may be a result of the combined effects of natural and anthropogenic forcing.

2.6.2.4. Other variables

Groisman et al. (2003) assessed a set of atmospheric variables derived from daily temperatures and precipitation that have applications in agriculture, architecture, power generation, and human health management. This section provides examples of these derived variables, including a temperature derivative (frequency of thaws) and a temperature and precipitation derivative (rain-on-snow-events). A day with thaw (i.e., snowmelt) can be defined as a day with snow on the ground when the daily mean temperature is above -2 °C (Brown R., 2000). During these days, snow deteriorates, changes its physical properties, and (eventually) disappears. In win-



ter and early spring at high latitudes, thaws negatively affect transportation, winter crops, and the natural environment including vegetation and wild animals. In late spring, intensification of thaw conditions leads to earlier snow retreat and spring onset (Brown R., 2000; Cayan et al., 2001; Groisman et al., 1994). Gradual snowmelt during the cold season affects seasonal runoff in northern watersheds, reducing the peak flow of snowmelt origin and increasing the mid-winter low flow (Vörösmarty et al., 2001). Figures 2.12 and 2.13 show the climatology of thaws in North America and Russia and quantitative estimates of change during the past 50 years. The time series shown in Fig. 2.13 demonstrate statistically significant increasing trends for winter and autumn of 1.5 to 2 days per 50 years. This change constitutes a 20% (winter) to 40% (autumn) increase in the thaw frequency during the second half of the 20th century.

Rain falling on snow causes a faster snowmelt and, when the rainfall is intense, may lead to flash floods. Along the western coast of North America, rain-on-snow events are the major cause of severe flash floods (Wade et al., 2001). Groisman et al. (2003) found a significant

increase in the frequency of winter rain-on-snow events in western Russia (50% over the 50 years from 1950 to 2000) and a significant reduction (of similar magnitude) in western Canada. During spring, there has been a

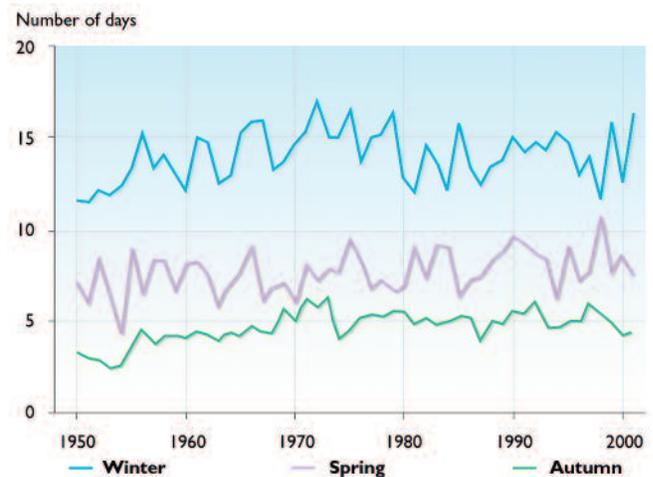


Fig. 2.13. Time series of variations in the frequency of days with thaw, averaged over the arctic land area for winter, spring, and autumn (Groisman et al., 2003).

significant increase in rain-on-snow events in Russia (mostly in western Russia), but there has also been a significant decrease in the frequency of these events in western Canada. The decrease in western Canada is mostly due to reduced snow-cover extent.

2.6.3. Marine Arctic

The 20th-century record of sea-ice extent reveals phases of expansion between 1900 and 1918 and between 1938 and 1968, and of contraction between 1918 and 1938 and from 1968 to the present (Zakharov, 2003). The IPCC (2001b) concluded, "Northern Hemisphere spring and summer sea-ice extent has decreased by about 10 to 15% since the 1950s. It is likely that there has been about a 40% decline in Arctic sea-ice thickness during late summer to early autumn in recent decades and a considerably slower decline in winter sea-ice thickness". Since 1978, the average rate of decrease has been $2.9 \pm 0.4\%$ per decade (Fig. 2.14; Cavalieri et al., 1997). The extent of perennial sea ice has decreased at more than twice this rate (Johannessen et al., 1999). The contraction has been greatest in August and September, while in January and February there has been a small expansion (Serreze et al., 2000). Recent trends vary considerably by region. In eastern Canadian waters, ice cover has actually increased (Parkinson, 2000). Interdecadal variations are large and dominated by anomalies of opposite sign in the Eurasian and American sectors (Deser et al., 2000). These variations are associated with anomalies in atmospheric pressure and temperature that closely resemble those associated with the AO (section 2.2).

The thickness of sea ice has decreased by 42% since the mid-1970s within a band across the central Arctic Ocean between the Chukchi Sea and Fram Strait (Rothrock et al., 1999; Wadhams, 2001). In the 1990s, the average thickness in this zone at the end of summer was only 1.8 m. However, the extent to which this change is a result of greater melt and reduced growth

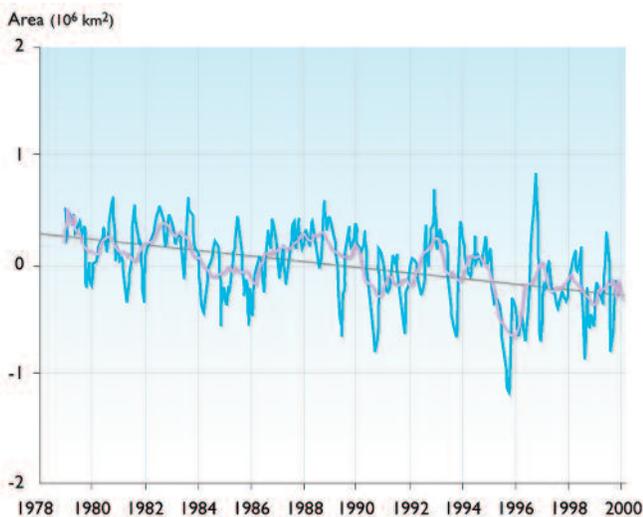


Fig. 2.14. Time series of monthly sea-ice extent anomalies in the Arctic Basin (blue line) with respect to the 1979–2001 average, with 12-month running mean (lavender line) and least-squares linear fit (grey line), based on data from 1979 to 2001 (National Snow and Ice Data Center, Boulder, Colorado).

(namely a warming climate) remains to be resolved. Thick sea ice observed before the 1990s may simply have been driven out of the band of observation by wind changes in the 1990s associated with the AO (Holloway and Sou, 2001; Polyakov and Johnson, 2000; Zhang T. et al., 2000). The impacts of the concurrent inflow of warmer Atlantic water and disappearing cold halocline (see below) are also unclear (Winsor, 2001). There are no systematic trends over the past 50 years in the thickness of coastal land-fast ice in either northern Canada or Siberia (Brown R. and Côté, 1992; Polyakov et al., 2003a). Interannual variability in the thickness of coastal land-fast ice is dominated by changes in winter snow accumulation, not air temperature.

Sea-ice drift in the central Arctic Ocean has been monitored since 1979 under the direction of the International Arctic Buoy Program. The pattern of sea-ice drift changed in the late 1980s coincident with the change in the pattern of surface winds caused by the AO. In the first decade of observations, the transpolar drift of sea ice moved directly from the Laptev Sea to Fram Strait. The Beaufort Gyre extended from the Canadian Archipelago to Wrangel Island. During the 1990s, the source of the transpolar drift moved eastward into the East Siberian Sea and the Beaufort Gyre shrank eastward into the Beaufort Sea (Kwok, 2000; Mysak, 2001; Polyakov and Johnson, 2000). The change in circulation moved the thickest sea ice toward the Canadian Archipelago and may have facilitated increased export of perennial sea ice across the north of Greenland and out through Fram Strait (Maslowski et al., 2000). Section 6.3 provides additional information regarding sea ice.

There are few systematic long-term oceanographic observations in the marine Arctic, except perhaps in the European sector. Russian airborne surveys are the primary source of information for the central Arctic prior to 1990 (Gorshkov, 1980). Because most records are of short duration, confidence in discriminating between trends and variability in the marine Arctic is low.

Most evidence of change in ice-covered waters is recent, emerging from observations made in the 1990s. Temperature and salinity at intermediate depths in the arctic basins were stable between 1950 and 1990. In the early 1990s, the temperature of the Fram Strait Branch north of the Barents Sea increased by 2 °C. The temperature increase has since spread around the perimeter of the basin, reaching the Beaufort Sea north of Alaska in 2001 (Carmack et al., 1997; Ivanov, 2000; Morison et al., 1998; Rudels et al., 1997; Schauer et al., 2002). The spatial pattern has remained consistent with that of past decades, with the highest temperatures in the eastern sector (Alekseev et al., 1998, 1999). However, the ocean front that separates waters of Atlantic and Pacific origin in the Arctic moved 600 km eastward along the Siberian slope in the early 1990s, shifting its alignment from the Lomonosov Ridge to the Alpha-Mendeleev Ridge (Carmack et al., 1995; McLaughlin et al., 1996;

Morison et al., 1998). This shift reflects a 20% increase in the volume of Atlantic water within the Arctic Ocean. The cold halocline separating sea ice from warm Atlantic water in the Eurasian Basin virtually disappeared in the 1990s (Steele and Boyd, 1998). In general, the pattern of spreading and the greater thickness of the Atlantic layer are consistent with increases in the rate of inflow and temperature of Atlantic water entering the Arctic via Fram Strait (Dickson et al., 2000; Grottefendt et al., 1998).

The magnitude and temperature of inflow to the Arctic via the Barents Sea is highly variable, in response to wind changes associated with the AO (Dickson et al., 2000). Both the inflow and its temperature have increased since 1970, mirroring trends observed during the arctic warming of 1910 to 1930 (Loeng et al., 1997).

The properties of the surface layer of the Arctic Ocean vary greatly by season and from year to year. Russian data reveal an average 1.3 m decrease in surface (upper 50 m) “freshwater” storage in the Arctic Ocean in winter between 1974 and 1977 (Alekseev et al., 2000). Surface-layer variability is greatest in the area of seasonal sea ice because varying ice conditions strongly modulate the temperature and salinity of the upper ocean (e.g., Fissel and Melling, 1990). Possible temporal trends in most peripheral seas of the marine Arctic remain obscured by high variability.

In the decade following 1963 there was an obvious freshening of the upper waters of the North Atlantic (the Great Salinity Anomaly: Dickson et al., 1988) which is thought to have originated in the Arctic Ocean (e.g., Hakkinen, 1993; Mysak et al., 1990). Unusually warm summers from 1957 to 1962 in Baffin Bay and the Canadian Archipelago may have promoted melting of land ice and an enhanced freshwater outflow (Alekseev et al., 2000). The Great Salinity Anomaly, first evident north of Iceland (Steffansson, 1969), peaked in the North Atlantic in 1969 (Malmberg and Blindheim, 1994), at the same time that unusually fresh and cold water was observed on the other side of Greenland (Drinkwater, 1994). The simultaneous development of freshening in both areas indicates paths for “freshwater” transport via both Fram Strait and the Canadian Archipelago. A new freshwater pulse may appear in the North Atlantic in response to lengthening of the arctic summer melt period in the 1990s (Belkin et al., 1998). For further discussion of change in the Atlantic sector, refer to section 9.2.4.2.

In the central Greenland Sea, change in the temperature of water at the 2000 m depth is an indicator of varying deep-water formation (Alekseev et al., 2001; Boenisch et al., 1997). The observation of a large temperature increase (0.25 °C) in the central Greenland Sea in the 1990s is consistent with the weakening of the thermohaline circulation implied by decreasing deep-water flow toward the Arctic through the Faroe-Shetland Strait (Hansen et al., 2001).

2.6.4. Terrestrial system

About 24.5% of the exposed land surface in the Northern Hemisphere is underlain by permafrost (Zhang T. et al., 2000), and any increase in temperature is very likely to cause this area to shrink, in turn affecting arctic groundwater supply. Section 6.6.1.2 provides a more detailed description of recent changes in permafrost. Permafrost is very sensitive to climate change, and is dependent upon ambient temperatures for its existence and distinctive properties (Lunardini, 1996). Brown J. et al. (2000) indicated that only the thin active (seasonally thawed) layer of permafrost responds immediately to temperature changes, making it both an indicator and a product of climate change, as its depth is dependent on temperature. Vourlitis and Oechel (1996) reported an annual increase of 10 to 22 cm in the depth of the active layer between June and August in some permafrost areas. Climate change is very likely to accelerate this depth increase, due to both higher temperatures and increased evapotranspiration, which makes the soil drier throughout the active layer and causes a descent of the permafrost table (Vourlitis and Oechel, 1996).

Along the land-sea margin of the Arctic, over the past few decades there has been an increase in the magnitude of oscillations of the permafrost bottom (deepest extent of permafrost), and a greater variability in thickness during the warmer part of longer-term cycles, indicating an overall degradation of coastal permafrost (Romanovskii and Hubberten, 2001). Along a north–south transect in Alaska, permafrost temperatures at 15 to 20 m depths have increased between 0.6 and 1.5 °C over the past 20 years, mean annual ground surface temperatures have increased by 2.5 °C since the 1960s, and discontinuous permafrost has begun thawing downward at a rate of about 0.1 m/yr at some locations (Osterkamp, 2003; Osterkamp and Romanovsky, 1999; Romanovsky and Osterkamp, 2000; Romanovsky et al., 2003). The effects of increased permafrost temperatures on groundwater could possibly include the development of large, unfrozen near-surface aquifers with groundwater flow throughout the year; more active recharge and discharge of aquifers and increased groundwater flow would affect the base flow volumes and chemistry of many rivers.

Romanovsky and Osterkamp (2000) took a series of precise permafrost temperature measurements from four observation sites in Alaska, and found that unfrozen water has the greatest effect on the ground thermal regime immediately after freeze-up and during the cooling of the active layer, but is less important during the warming and thawing of the active layer. Jorgenson et al. (2001) showed widespread permafrost degradation in the Tanana Valley due to increases in air and permafrost temperatures. Permafrost temperatures have increased in Alaska, and decreased in eastern Canada (Serreze et al., 2000). Increased permafrost temperatures have also been observed in Siberia (Isaksen et al., 2001). Climate change is likely to drasti-

cally affect wildlife habitat, as permafrost is essential to the existence and functioning of arctic wetlands.

2.7. Arctic climate variability prior to 100 years BP

This section examines the record of past climate change in the Arctic with the objective of providing a context for evaluating evidence of more recent climate change and the possible impacts of future climate change. This review focuses on the past two million years (approximately), and particularly the past 20 000 years. Over geological time periods, the earth's natural climate system has been forced or driven by a relatively small number of external factors. Tectonic processes acting very slowly over millions of years have affected the location and topography of the continents through plate tectonics and ocean spreading. Changes in the orbit of the earth occur over tens to hundreds of thousands of years and alter the amount of solar radiation received at the surface by season and by latitude. These orbital changes drive climate responses, including the growth and decay of ice sheets at high latitudes. Finally, changes in the emissivity of the sun have taken place over billions of years, but there have been shorter-term variations that occurred over decades, centuries, and millennia. As a consequence of changes in these external forces, global climate has experienced variability and change over a variety of timescales, ranging from decades to millions of years.

The sparseness of instrumental climate records prior to the 20th century (especially prior to the mid-19th century) means that estimates of climate variability in the Arctic during past centuries must rely upon indirect "proxy" paleoclimate indicators, which have the potential to provide evidence for prior large-scale climatic changes. Typically, the interpretation of proxy climate records is complicated by the presence of "noise" in which climate information is immersed and by a variety of possible distortions of the underlying climate information (see Bradley, 1999 for a comprehensive review; IPCC, 2001). Careful calibration and cross-validation procedures are necessary to establish a reliable relationship between a proxy indicator and the climatic variable or variables of interest, providing a "transfer" function that allows past climatic conditions to be estimated. Of crucial importance is the ability to date different proxy records accurately in order to determine whether events occurred simultaneously, or whether some events lagged behind others.

Sources of paleoclimatic information in the Arctic are limited to a few, often equivocal types of records, most of which are interpreted as proxies for summer temperature. Little can be said about winter paleoclimate (Bradley, 1990). Only ice cores, tree rings, and lake sediments provide continuous high-resolution records. Coarsely resolved climate trends over several centuries are evident in many parts of the Arctic, including:

- the presence or absence of ice shelves deduced from driftwood frequency and peat growth episodes and pollen content;
- the timing of deglaciation and maximum uplift rates deduced from glacio-isostatic evidence as well as glacial deposits and organic materials over-ridden by glacial advances or exposed by ice recession;
- changes in the range of plant and animal species to locations beyond those of today (Bradley, 1990); and
- past temperature changes deduced from the geothermal information provided by boreholes.

In contrast, large-scale continuous records of decadal, annual, or seasonal climate variations in past centuries must rely upon sources that resolve annual or seasonal climatic variations. Such proxy information includes tree-ring width and density measurements; pollen, diatom, and sediment changes from laminated sediment cores; isotopes, chemistry, melt-layer stratigraphy, acidity, pollen content, and ice accumulation from annually resolved ice cores; and the sparse historical documentary evidence available for the past few centuries. Information from individual paleoclimate proxies is often difficult to interpret and multi-proxy analysis is being used increasingly in climate reconstructions. Taken as a whole, proxy climate data can provide global-scale sampling of climate variations several centuries into the past, and have the potential to resolve both large-scale and regional patterns of climate change prior to the instrumental period.

2.7.1. Pre-Quaternary Period

Over the course of millions of years, the Arctic has experienced climatic conditions that have ranged from one extreme to the other. Based on the fossil record, 120 to 90 million years before present (My BP) during the mid-Cretaceous Period, the Arctic was significantly warmer than at present, such that arctic geography, atmospheric composition, ocean currents, and other factors were quite different from at present. In contrast, as recently as 20 ky BP, the Arctic was in the grip of intense cold and continental-scale glaciation was at its height. This was the latest of a series of major glacial events, which, together with intervening warm periods, have characterized the past two million years of arctic environmental history.

The most recent large-scale development and build-up of ice sheets in the Arctic probably commenced during the Late Tertiary Period (38 to 1.6 My BP). Ice accumulation at that time may have been facilitated initially by plate tectonics (Ruddiman et al., 1989). According to Maslin et al. (1998), the onset of Northern Hemisphere glaciation began with a significant build-up of ice in southern Greenland. However, progressive intensification of glaciation does not seem to have begun until 3.5 to 3 My BP, when the Greenland Ice Sheet expanded to include northern Greenland. Maslin et al. (1998) suggested that the Eurasian Arctic, northeast Asia, and Alaska were glaciated by about 2.7 My BP and northeast

America by 2.5 My BP. Maslin et al. (1998) suggested that tectonic changes, including the deepening of the Bering Strait, were too gradual to be responsible for the speed of Northern Hemisphere glaciation, but suggested that tectonic changes brought the global climate to a critical threshold while relatively rapid variations in the orbital parameters of the earth triggered the glaciation.

2.7.2. Quaternary Period

The Quaternary Period (the last 1.6 My) has been characterized by periodic climatic variations during which the global climate system has switched between interglacial and glacial stages, with further subdivision into stadials (shorter cold periods) and interstadials (shorter mild episodes). Glacial stages are normally defined as cold phases with major glacier and ice sheet expansion. Interglacials are defined as warm periods when temperatures were at least as high as during the present Holocene interglacial.

This interglacial–glacial–interglacial climate oscillation has been recurring with a similar periodicity for most of the Quaternary Period, although each individual cycle appears to have had its own idiosyncrasies in terms of the timing and magnitude of specific events. It has been estimated that there have been between 30 and 50 glacial/interglacial cycles during the Quaternary Period (Ruddiman et al., 1989; Ruddiman and Kutzbach, 1990), primarily driven by changes in the orbit of the earth (Imbrie and Imbrie, 1979).

One of the earliest (and certainly the most well known) hypotheses concerning the effects of orbital configuration on glacial cycles is described by Milankovitch (1941), who presents the argument that a decrease in summer insolation is critical for glacial initiation. Low summer insolation occurs when the tilt of the axis of rotation of the earth is small; the poles are pointing less directly at the sun; the Northern Hemisphere summer solstice is farthest from the sun; and the earth's orbit is highly eccentric. The key orbital parameters involved include changes in the eccentricity of the orbit of the earth with a period of 100 ky; the tilt of the axis of rotation of the earth, which oscillates between 22.2° and 24.5° with a period of 41 ky; and the position of the earth within its elliptical orbit during the Northern Hemisphere summer, which changes over a period of 23 ky. Bradley (1990) pointed out that such orbitally induced radiation anomalies are especially significant at high latitudes in summer when daylight persists for 24 hours. The latitude most sensitive to low insolation values is 65° N, the latitude at which most ice sheets first formed and lasted longest during the last glaciation. The amount of summer insolation reaching the top of the atmosphere at 65° N and nearby latitudes can vary by as much as $\pm 12\%$ around the long-term mean value. Changes in winter insolation also occur with exactly the opposite timing, but they are not considered important to ice-sheet survival. When summer insolation is weak, less radiation is delivered to the surface at high latitudes,

resulting in lower regional temperatures. Lower temperatures reduce the summer ablation of the ice sheets and allow snow to accumulate and ice sheets to grow. Once ice sheets are created, they contribute to their own positive mass balance by growing in elevation, reaching altitudes of several kilometers where prevailing temperatures favor the accumulation of snow and ice.

Other hypotheses concerning the causes of glacial initiation include that of Young and Bradley (1984), who argued that the meridional insolation gradient is a critical factor for the growth and decay of ice sheets through its control over poleward moisture fluxes during summer and resultant snowfall. Snowfall could increase in a cooler high-latitude climate through enhanced storm activity forced by a greater latitudinal temperature gradient. Ruddiman and McIntyre (1981) and Miller and de Vernal (1992) suggested that the North Atlantic Ocean circulation is important for modulating ice volume. There is evidence that the North Atlantic remained warm during periods of ice growth and they proposed that enhanced meridional temperature gradients and evaporation rates during the winter enhanced snow delivery to the nascent ice sheets of northeastern Canada. An active thermohaline circulation in the North Atlantic could increase the moisture supply to sites of glacial initiation through its ability to maintain warmer sea surface temperatures, limit sea-ice growth, and allow for greater evaporation from the ocean surface.

These external forcing mechanisms in turn cause responses and chain reactions in the internal elements of the earth's system (e.g., Bradley, 1985). Changes in one internal element of the system can cause responses in other elements. These can lead to feedback effects that can amplify or attenuate the original signal. Thus, ice sheets, ice caps, and glaciers play an important role in the global climate system. Glacial advance and retreat may therefore be both a consequence and a cause of climate change (Imbrie et al., 1993a,b).

Given the inherent errors in dating techniques, gaps in the stratigraphic record, and the varying rates of response of different biological proxy indicators, there is considerable uncertainty about the timing of specific events and whether climate changes were truly synchronous in different regions. The errors and uncertainties tend to be amplified farther back in the paleoclimatic record, particularly in the Arctic, where much of the paleoclimatic evidence from earlier parts of the Quaternary Period has been removed or obfuscated as a result of later glaciations. Consequently, this review of climate variability in the Arctic during the Quaternary Period focuses first on the more complete and reliable evidence of climate conditions immediately prior to the onset of the most recent glacial–interglacial oscillation (~130 ky BP). This is followed by a brief review of conditions during the Last Glacial Maximum (~20 ky BP) and the subsequent period of deglaciation, and culminates in a review of the evidence of climatic variability during the present interglacial, the Holocene.

2.7.3. Last interglacial and glaciation

2.7.3.1. Last interglacial: The Eemian

Climatic conditions during interglacial periods are generally considered to be broadly comparable to present-day conditions. The most recent interglacial, the Eemian, extended from the end of the penultimate glaciation about 130 ky BP until about 107 ky BP when the last glacial period began (see IPCC, 2001c for further discussion of the Eemian). The Eemian is often regarded as a typical interglacial event with characteristics including relatively high sea level, a retreat to minimum size of global ice sheets, and the establishment of biotic assemblages that closely parallel those at present. According to most proxy data, the last interglacial was slightly warmer everywhere than at present (IPCC, 2001c). Brigham-Grette and Hopkins (1995) reported that during the Eemian the winter sea-ice limit in Bering Strait was at least 800 km farther north than today, and that during some summers the Arctic Ocean may have been ice-free. The northern treeline was more than 600 km farther north, displacing tundra across all of Chukotka (Lozhkin and Anderson, 1995). Western European lake pollen records show deciduous forests (characteristic of warmer conditions) across much of Western Europe that were abruptly replaced by steppic taxa characteristic of colder conditions; this shift is associated with a cold event at 107 ky BP. This relatively prolonged warm period is also detected in northeast Atlantic marine sediments, but is not evident throughout the North Atlantic. Faunal and lithic records from the polar North Atlantic (Fronval and Jansen, 1997) indicate that the first abrupt cooling occurred around 118 to 117 ky BP (Adkins et al., 1997; Keigwin et al., 1994).

Evidence of warmer conditions in the Arctic than exist at present is provided by a re-evaluation of the oxygen isotope ratio ($\delta^{18}\text{O}$) record obtained from Greenland ice core samples (Cuffey and Marshall, 2000). These authors suggest that the Greenland Ice Sheet was considerably smaller and steeper during the Eemian than at present and probably contributed 4 to 5.5 m to global sea level during that period. This implies that the climate of Greenland during the Eemian was more stable and warmer than previously thought, and the consequent melting of the Greenland Ice Sheet more significant.

Some researchers suggest that a paleoclimatic reconstruction of the Eemian provides a means of establishing the mode and tempo of natural climate variability with no anthropogenic influence. However, a general lack of precise absolute timescales and regionally-to-globally synchronous stratigraphic markers makes long-distance correlation between sites problematic, and inferred terrestrial changes are difficult to place within the temporal framework of changes in ice volume and sea level (Tzedakis, 2003).

2.7.3.2. Last glaciation: Wisconsinan/Weichselian

Although the timing of the end of the Eemian interglacial is subject to some uncertainty, high-resolution North Atlantic marine sediment records indicate that the Eemian ended with abrupt changes in deep-water flow occurring over a period of less than 400 years (Adkins et al., 1997; IPCC, 2001c). Evidence in marine sediments of an invasion of cold, low salinity water in the Norwegian Sea at this time has been linked by Cortijo et al. (1994) to a reduction in warm-water transport by the North Atlantic Drift and the thermohaline circulation. It seems likely that this contributed to the onset of widespread arctic glaciation in sensitive areas.

Following the initial cooling event (~107 ky BP), climatic conditions often changed suddenly, followed by several thousand years of relatively stable climate or even a temporary reversal to warmth. Overall, however, there was a decline in global temperatures. The boundaries of the boreal forests retreated southward and fragmented as conditions grew colder. Large ice sheets began to develop on all the continents surrounding the Arctic Ocean. The point at which the global ice extent was at its greatest (~24 to 21 ky BP) is known as the Last Glacial Maximum (LGM – Clark and Mix, 2000).

At its maximum extent, the Laurentide Ice Sheet extended from the Arctic Ocean in the Canadian Archipelago to the midwestern United States in the south, and from the Canadian Cordillera to the eastern edge of the continent. Local ice sheets also covered the Alaska and Brooks Ranges in Alaska. The Eurasian and Laurentide Ice Sheets were responsible for most of the glacio-eustatic decrease in sea level (about 120 m) during the LGM. The pattern of postglacial isostatic rebound suggests that the ice was thickest over Hudson Bay. The different parts of the Laurentide Ice Sheet reached their maximum extent between 24 and 21 ky BP (Dyke et al., 2002). The Innuitian ice buildup appears to have culminated in the east after 20.5 ky BP. Dyke et al. (2002) suggested that the entire ice sheet system east of the Canadian Cordillera responded uniformly to changes in climate. In contrast, the Cordilleran Ice Sheet did not reach its maximum extent until 15.2 to 14.7 ky BP, well after the LGM and the insolation minimum at approximately 21 ky BP (Dyke et al., 2002). This out-of-phase response may be attributable to the effects of growth of the Cordilleran Ice Sheet, which would have intercepted moisture transport to the interior plains at the expense of the Laurentide Ice Sheet. During its maximum extent, the Laurentide Ice Sheet was more than twice the size of the Eurasian Ice Sheet. Changes in climate during the LGM are discussed by the IPCC (2001c).

The Eurasian Ice Sheet initiation began 28 ky BP as a result of temperature changes that lowered equilibrium line altitudes across the Scandinavian mountains, Svalbard, Franz Josef Land, and Novaya Zemlya. Ice flow

north from Scandinavia and south from Svalbard, in conjunction with eustatic sea-level decreases, caused the margin of the ice sheet to migrate into the Barents Sea. Complete glaciation of the Barents Sea by a grounded ice sheet was achieved by 20 ky BP. The ice sheet at its maximum extent covered Scandinavia and the Barents Shelf and included a marine-based margin along the northern Barents Shelf, the western Barents Sea, western Scandinavia, and northern Great Britain and Ireland. The eastern margin of the ice sheet is generally thought to have been located west of the Taymir Peninsula (Mangerud et al., 2002). It appears that at the LGM, cold arid conditions persisted across much of eastern Siberia and that an East Siberian Sea Ice Sheet did not exist (Brigham-Grette et al., 2003), as some have claimed (Grosswald and Hughes, 2002). Glaciers in the Urals at the LGM appear to have been confined to the mountain valleys, rather than coalescing with the Barents-Kara Ice Sheet, as happened during previous glaciations (Mangerud et al., 2002).

2.7.4. Last glacial/interglacial transition through to mid-Holocene

2.7.4.1. Last glacial/interglacial transition

The most extreme manifestation of climate change in the geological record is the transition from full glacial to full interglacial conditions. Following the LGM (~24 to 21 ky BP), temperatures close to those of today were restored by approximately 10 ky BP (IPCC, 2001c).

The inception of warming appears to have been very rapid (NRC, 2002). The rate of temperature change during the recovery phase from the LGM provides a benchmark against which to assess rates of temperature change in the late 20th century. Available data indicate an average warming rate of about 2 °C per millennium between about 20 and 10 ky BP in Greenland, with lower rates for other regions. On the other hand, very rapid temperature increases at the start of the Bølling-Allerød period (14.5 ky BP; Severinghaus and Brook, 1999) or at the end of the Younger Dryas (~11 ky BP) may have occurred at rates as large as 10 °C per 50 years over substantial areas of the Northern Hemi-

sphere. Almost synchronously, major vegetation changes occurred in Europe and North America and elsewhere (Gasse and van Campo, 1994). There was also a pronounced warming of the North Atlantic and North Pacific (Webb et al., 1998).

Oxygen isotope measurements from Greenland ice cores demonstrate that a series of rapid warm and cold oscillations, called Dansgaard–Oeschger (D-O) events, punctuated the last glaciation, often taking Greenland and northwestern Europe from a full-glacial climate to conditions about as warm as at present (Fig. 2.15). For the period between 115 and 14 ky BP, 24 of these short-lived warm events have been identified in the Greenland ice core data, although many lesser warming events also occurred (Dansgaard et al., 1993). According to Lang et al. (1999), associated temperature changes may have been as great as 16 °C. The D-O oscillations are correlated with sea surface temperature variations derived from several North Atlantic deep-sea cores (Bond et al., 1993). From the speed of the climate changes recorded in the Greenland Ice Sheet (Dansgaard et al., 1989), it is widely thought that the complete change in climate occurred, at least regionally, over only a few decades. These interstadials lasted for varying periods of time, usually a few centuries to about 2000 years, before equally rapid cooling returned conditions to their previous state. The evidence from high-resolution deep sea cores from the North Atlantic (Bond et al., 1997) suggests that during at least the past 30 000 years, interstadials tended to occur at the warmer points of a background North Atlantic temperature cycle that had a periodicity of approximately 1500 years.

Heinrich events appear to be the most extreme of a series of sudden, brief cold events that seem to have occurred very frequently over the past 115 000 years, apparently tending to start at the low point of the same 1500-year temperature cycle. Heinrich events occurred during times of decreasing sea surface temperatures in the form of brief, exceptionally large discharges of icebergs in the North Atlantic from the Laurentide and European Ice Sheets that left conspicuous layers of detrital material in deep-sea sediments. Accompanying the Heinrich events were large decreases in the oxygen isotope ratio of planktonic foraminifera, providing evidence of lowered surface salinity probably caused by melting of drifting ice (Bond et al., 1993). Heinrich events appear at the end of a series of saw-tooth-shaped temperature cycles known as Bond cycles. During the Pleistocene Epoch, each cycle was characterized by the relatively warm interstadials becoming progressively cooler.

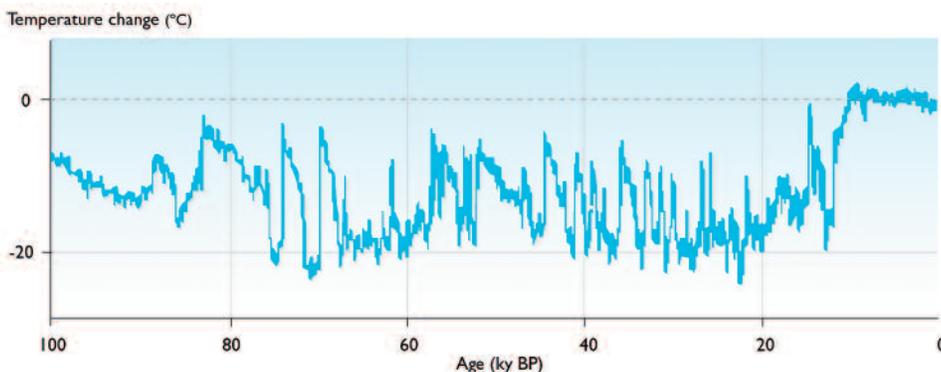


Fig. 2.15. Temperature change over the past 100 ky (departure from present conditions) reconstructed from a Greenland ice core (Ganopolski and Rahmstorf, 2001).

Deep-sea cores also show the presence of ice-rafting cycles in the intervals between Heinrich events (Bond and Lotti, 1995). The duration of these ice-rafting cycles varies between 2000 and 3000 years and they closely coincide with the D-O events. A study of the ice-rafted material suggests that, coincident with the D-O cooling, ice within the Icelandic ice caps and within or near the Gulf of Saint Lawrence underwent nearly synchronous increases in rates of calving. The Heinrich events reflect the slower rhythm of iceberg discharges into the North Atlantic, probably from Hudson Strait.

Air temperature, sea surface temperature, and salinity variations in the North Atlantic are associated with major changes in the thermohaline circulation. A core from the margin of the Faroe-Shetland Channel covering the last glacial period reveals numerous oscillations in benthic and planktonic foraminifera, oxygen isotopes, and ice-rafted detritus (Rasmussen et al., 1996). These oscillations correlate with the D-O cycles, showing a close relationship between the deep-ocean circulation and the abrupt climatic changes during the last glaciation. It is increasingly apparent that large, globally linked environmental feedbacks were involved in the generation of the large, rapid temperature increases that occurred during glacial termination and the onset of D-O events.

During the last glacial–interglacial transition, the movements of the North Atlantic Polar Front have been described as pivoting around locations in the western North Atlantic. Iceland, situated in the middle of the North Atlantic, has glaciers sensitive to changes in oceanic and atmospheric frontal systems (Ingólfsson et al., 1997). The late-glacial (subsequent to the LGM) records from Iceland indicate that relatively warm Atlantic water reached Iceland during the Bølling–Allerød Interstadial, with a short cooling period corresponding with the Older Dryas. Karpuz et al. (1993) suggested that the marine polar front was located close to Iceland during the Bølling–Allerød, and Sarnthein et al. (1995) concluded that sea-surface circulation was mainly in Holocene interglacial mode after 12.8 ky BP.

Warm episodes were also associated with higher sea-surface temperatures and the presence of oceanic convection in the Norwegian Greenland Sea. Cold episodes were associated with low sea surface temperatures, low salinity, and no convection (Rasmussen et al., 1996). Subsequent to the initial phase of deglaciation on Spitsbergen (~13 to 12.5 ky BP), most of the eastern and northern Barents Sea was deglaciated during the Bølling–Allerød Interstadial (Vorren and Laberg, 1996). Vorren and Laberg (1996) speculate on the existence of a close correlation between summer air temperatures and the waxing and waning phases of the northern Fennoscandian and southern Barents Sea Ice Sheets. Most of the ice-sheet decay is attributed by these authors to increases in air temperature, which caused thinning of the ice sheets, making them susceptible to decoupling from the seabed and increased calving.

After a few thousand years of recovery, the Arctic was suddenly plunged back into a new and very short-lived cold event known as the Younger Dryas (see NRC, 2002, for a detailed discussion) leading to a brief advance of the ice sheets. The central Greenland ice core record (from the Greenland Ice Core Project and the Greenland Ice Sheet Project 2) indicates that the return to the cold conditions of the Younger Dryas from the incipient interglacial warming 13 ky BP took place in less than 100 years. The warming phase at the end of the Younger Dryas, which took place about 11 ky BP and lasted about 1300 years, was very abrupt and central Greenland temperatures increased by 7 °C or more in a few decades.

2.7.4.2. Early to mid-Holocene

Following the sudden end of the Younger Dryas, the Arctic entered several thousand years of conditions that were warmer and probably moister than today. Peak high-latitude summer insolation (Milankovitch orbital forcing) occurred during the earliest Holocene, with a maximum radiation anomaly (approximately 8% greater than at present) attained between 10 and 9 ky BP (Berger and Loutre, 1991). Although most of the Arctic experienced summers that were warmer (1–2 °C) than at present during the early to middle Holocene, there were significant spatial differences in the timing of this warm period. This was probably due to the effects of local residual land-ice cover and local sea surface temperatures (Overpeck et al., 1997). In most of the ice cores from high latitudes, the warm period is seen at the beginning of the Holocene (about 11 to 10 ky BP). In contrast, central Greenland (Dahl-Jensen et al., 1998) and regions downwind of the Laurentide Ice Sheet, including Europe (COHMAP Members, 1988) and eastern North America (Webb et al., 1998), did not warm up until after 8 ky BP. The rapid climatic events of the last glacial period and early Holocene are best documented in Greenland and the North Atlantic and may not have occurred throughout the Arctic. This early Holocene warm period appears to have been punctuated by a severe cold and dry phase about 8200 years ago, which lasted for less than a century, as recorded in the central Greenland ice cores (Alley et al., 1997).

Glacio-isostatic evidence indicates deglaciation was underway by the beginning of the Holocene and maximum uplift occurred between 8 and 7 ky BP and even earlier in many areas. By 9 ky BP, Spitsbergen glaciers had retreated to or beyond their present day positions (Sexton et al., 1992) and the marine faunal evidence suggest that this period was as warm if not warmer than at present along the west and north coasts of Svalbard (Salvigsen et al., 1992). The retreat of the largest of the glaciers along the Gulf of Alaska began as early as 16 to 14 ky BP (Mann D. and Hamilton, 1995). Although much of the high and mid-Canadian Arctic remained glaciated, warm summers are clearly registered by enhanced summer melting of the Agassiz Ice Cap (Koerner and Fisher, 1990). Following the large, abrupt change in stable-isotope ratios marking the end of the

last glaciation, $\delta^{18}\text{O}$ profiles from Agassiz Ice Cap cores show a rapid warming trend that reached a maximum between approximately 9 and 8 ky BP.

Deglacial marine sediments in Clements Markham Inlet on the north coast of Ellesmere Island resemble those characteristic of temperate (as opposed to polar) tide-water glaciers, suggesting that climatic conditions in the early Holocene were significantly warmer there than today (Stewart, 1988). Glaciers had retreated past present-day termini in some areas by 7.5 ky BP. Increasing sea surface temperatures in Baffin Bay enhanced precipitation on Baffin Island (Miller and de Vernal, 1992), leading to a widespread early Holocene glacial advance along the east coast. Marine mammals and boreal mollusks were present far north of their present-day range by 7.5 to 6.5 ky BP, as were many species of plants between 9.2 and 6.7 ky BP (Dyke et al., 1996, 1999). Caribou were able to survive in the northernmost valleys of Ellesmere Island and Peary Land by 8.5 ky BP or earlier. Such evidence indicates very warm conditions early in the Holocene (before 8 ky BP).

Early Holocene summer temperatures similar to those at present have been reconstructed in Arctic Fennoscandia by numerous studies using a range of proxies and multi-proxy analyses (Rosén et al., 2001). However, abrupt climatic variations were characteristic of the early Holocene, with distinct cool episodes around 9.2, 8.6, and 8.2 ky BP (Korhola et al., 2002). The most recent of these events might be connected to the widely known “8.2 ky event”, which affected terrestrial and aquatic systems in northern Fennoscandia (Korhola et al., 2002; Nesje et al., 2000; Snowball et al., 1999). Hantemirov and Shiyatov (2002) report that open larch forests were already growing in the Yamal Peninsula of northwestern Siberia 10.5 to 9 ky BP and that the most favorable period for tree growth lasted from 9.2 to 8 ky BP. During the early Holocene, reconstructed mean temperature anomalies for the warmest month, based on pollen data across Northern Europe, show temperatures comparable to those at present (Davis et al., 2003). Temperatures then increased around 6 ky BP, with the onset of this increase delayed to around 9 ky BP in the east.

Boreal forest development across northern Russia (including Siberia) commenced by 10 ky BP (MacDonald G. et al., 2000). Over most of Russia, forests advanced to or near the current arctic coastline between 9 and 7 ky BP, and retreated to their present position by between 4 and 3 ky BP. Forest establishment and retreat were roughly synchronous across most of northern Russia, with the exception of the Kola Peninsula, where both appear to have occurred later. During the period of maximum forest extension, the mean July temperature along the northern coastline of Russia may have been 2.5 to 7.0 °C warmer than present.

The Arctic appears to have been relatively warm during the mid-Holocene, although records differ spatially, temporally, and by how much summer warmth they suggest

(relative to the early Holocene insolation maximum; Bradley, 1990; Hardy and Bradley, 1996).

A review of the Holocene glaciation record in coastal Alaska (Calkin, 1988) suggests that glacier fluctuations in arctic, central interior, and southern maritime Alaska were mostly synchronous. Ager (1983) and Heusser (1995) report that pollen records indicate a dramatic cooling about 3.5 ky BP and suggest an increase in precipitation and storminess in the Gulf of Alaska accompanied by a rejuvenation of glacial activity. In northern Iceland, the Holocene record of glacier fluctuations indicates two glacial advances between 6 and 4.8 ky BP (Stötter, 1991).

In the Canadian Arctic, interior regions of Ellesmere Island appear to have retained extensive Inuitian and/or plateau ice cover until the mid-Holocene (Bell, 1996; Smith, 1999), after which ice margins retreated to positions at or behind those at present. Restricted marine-mammal distributions imply more extensive summer sea ice between 8 and 5 ky BP, and hence cooler conditions (Dyke et al., 1996, 1999). However, marine conditions at 6 ky BP warmer than those at present are suggested by analyses of marine microfossils (Gajewski et al., 2000) performed over a broad area from the high Arctic to the Labrador Sea via Baffin Bay. A multi-proxy summary of marine and terrestrial evidence from the Baffin sector (Williams et al., 1995) suggests that warming began around 8 ky BP, intensified at 6 ky BP, and that conditions had cooled markedly by 3 ky BP.

Dugmore (1989) demonstrated that, in Iceland, the Sólheimajökull glacier extended up to 5 km beyond its present limits between 7 and 4.5 ky BP. Major ice sheet advances also occurred before 3.1 ky BP and between 1.4 and 1.2 ky BP. In the 10th century (1 ky BP), this glacier was also larger than during the period from AD 1600 to 1900, when some other glaciers reached their maximum Holocene extent. Stötter et al. (1999) suggested that major glacier advances in northern Iceland occurred at around 4.7, 4.2, 3.2–3.0, 2.0, 1.5, and 1.0 ky BP.

Evidence for a mid-Holocene thermal maximum in Scandinavia is considerable, and based on a wide range of proxies (Davis et al., 2003). Treelines reached their maximum altitude (up to 300 m higher than at present; Barnekow and Sandgren, 2001), and glaciers were greatly reduced or absent (Seierstad et al., 2002). Pollen and macrofossil records from the Torneträsk area in northern Swedish Lapland indicate optimal conditions for Scots pine (*Pinus sylvestris*) from 6.3 to 4.5 ky BP (Barnekow and Sandgren, 2001) and records of treeline change in northern Sweden show high-elevation treelines around 6 ky BP (Karlén and Kuylénstierna, 1996). These data indicate an extended period in the early to mid-Holocene when Scandinavian summer temperatures were 1.5 to 2 °C higher than at present.

Tree-ring data from the Torneträsk area indicate particularly severe climatic conditions between 2.6 and 2 ky BP

(600–1 BC). This period includes the greatest range in ring-width variability of the past 7400 years in this area, indicating a highly variable but generally cold climate (Grudd et al., 2002). This period is contemporary with a major glacial expansion in Scandinavia when many glaciers advanced to their Holocene maximum position (Karlén, 1988) with major effects on human societies (Burenhult, 1999; van Geel et al., 1996).

Especially severe conditions in northern Swedish Lapland occurred 2.3 ky BP (330 BC), with tree-ring data indicating a short-term decrease in mean summer temperature of about 3 to 4 °C. A catastrophic drop in pine growth at that time is also reported by Eronen et al. (2002), who state that this was the most unfavorable year for the growth of treeline pines in Finnish Lapland in the past 7500 years. Reconstructed Holocene summer temperature changes in Finnish Lapland, based on proxy climate indicators in sediments from Lake Tsuolbmajavri, show an unstable early Holocene between 10 and 8 ky BP in which inferred July air temperatures were about the same as at present most of the time, but with three successive cold periods at approximately 9.2, 8.6, and 8.2 ky BP, and a “thermal maximum” between approximately 8 and 5.8 ky BP, followed by an abrupt cooling (Korhola et al., 2002). Dated subfossils (partially fossilized organisms) show that the pine treeline in northwestern Finnish Lapland retreated a distance of at most 70 km during this cooling, but that the shift was less pronounced in more easterly parts of Lapland (Eronen et al., 2002).

Hantemirov and Shiyatov (2002) reported that the most favorable period for tree growth in the Yamal Peninsula of northwestern Siberia lasted from 9.2 to 8 ky BP. At that time, the treeline was located at 70° N. Then, until 7.6 ky BP, temperatures decreased but this did not result in any significant shift in the treeline. The treeline then moved south until, by 7.4 ky BP, it was located at approximately 69° N. It remained here until 3.7 ky BP when it rapidly retreated (~20 km) to within 2 to 3 km north of its present position south of the Yamal Peninsula. This retreat in the space of only 50 years coincides with an abrupt and large cooling as indicated in the tree-ring data. This cooling event may have been associated with the eruption of the Thera (Santorini) volcano in the southern Aegean around 3.6 ky BP.

Tree-ring data from the Kheta-Khatanga plain region and the Moyero-Kotui plateau in the eastern part of the Taymir Peninsula indicate climatic conditions more favorable for tree growth around 6 ky BP, as confirmed by increased concentrations of the stable carbon isotope ¹³C in the annual tree rings (Naurzbaev et al., 2002). The growth of larch trees at that time was 1.5 to 1.6 times greater than the average radial growth of trees during the last 2000 years, and the northern tree-line is thought to have been situated at least 150 km farther north than at present, as indicated by the presence of subfossil wood of that age in alluvial deposits of the Balakhnya River. During the past 6000 years, the

eastern Taymir tree-ring chronologies show a significant and progressive decrease in tree growth and thus temperature.

2.7.5. Last millennium

Over the last millennium, variations in climate across the Arctic and globally have continued. The term “Medieval Warm Period”, corresponding roughly to the 9th to the mid-15th centuries, is frequently used but evidence suggests that the timing and magnitude of this warm period varies considerably worldwide (Bradley and Jones, 1993; Crowley and Lowery, 2000; IPCC 2001c). Current evidence does not support a globally synchronous period of anomalous warmth during that time frame, and the conventional term of “Medieval Warm Period” appears to have limited utility in describing trends in hemispheric or global mean temperature changes.

The Northern Hemisphere mean temperature estimates of Mann M. et al. (1999), and Crowley and Lowery (2000), show that temperatures during the 11th to the 14th centuries were about 0.2 °C higher than those during the 15th to the 19th centuries, but somewhat below the temperatures of the mid-20th century. The long-term hemispheric trend is best described as a modest and irregular cooling from AD 1000 to around 1850 to 1900, followed by an abrupt 20th-century warming.

Regional evidence is, however, quite variable. Crowley and Lowery (2000) show that western Greenland exhibited local anomalous warmth only around AD 1000 (and to a lesser extent, around AD 1400), and experienced quite cold conditions during the latter part of the 11th century. In general, the few proxy temperature records spanning the last millennium suggest that the Arctic was not anomalously warm throughout the 9th to 14th centuries (Hughes and Diaz, 1994).

In northern Swedish Lapland, Scots pine tree-ring data indicate a warm period around AD 1000 that ended about AD 1100 when a shift to a colder climate occurred (Grudd et al., 2002). In Finnish Lapland, based on a 7500-year Scots pine tree-ring record, Helama et al. (2002) reported that the warmest non-overlapping 100-year period in the record is AD 1501 to 1600, but AD 1601 was unusually cold. Other locations in Fennoscandia and Siberia were also cold in AD 1601, and Briffa et al. (1992, 1995) linked the cold conditions to the AD 1600 eruption of the Huaynaputina volcano in Peru. In northern Siberia, and particularly east of Taymir where the most northerly larch forests occur, long-term temperature trends derived from tree rings indicate the occurrence of cool periods during the 13th, 16th to 17th, and early 19th centuries. The warmest periods over the last millennium in this region were between AD 950 and 1049, AD 1058 and 1157, and AD 1870 and 1979. A long period of cooling began in the 15th century and conditions

remained cool until the middle of the 18th century (Naurzbaev et al., 2002).

For the most part, “medieval warmth” appears to have been restricted to areas in and around the North Atlantic, suggesting that variability in ocean circulation may have played a role. Keigwin and Pickart (1999) suggested that the temperature contrasts between the North Atlantic and other areas were associated with changes in ocean currents in the North Atlantic and may to a large extent reflect century-scale changes in the NAO.

By the middle of the 19th century, the climate of the globe and the Arctic was cooling. Overall, the period from 1550 to 1900 may have been the coldest period in the entire Holocene (Bradley, 1990). This period is usually called the “Little Ice Age” (LIA), during which glaciers advanced on all continents. The LIA appears to have been most clearly expressed in the North Atlantic region as altered patterns of atmospheric circulation (O’Brien et al., 1995). Unusually cold, dry winters in central Europe (e.g., 1 to 2 °C below normal during the late 17th century) were very probably associated with more frequent flows of continental air from the north-east (Pfister, 1999; Wanner et al., 1995). Such conditions are consistent with the negative or enhanced easterly wind phase of the NAO, which implies both warm and cold anomalies over different regions of the North Atlantic sector. Although the term LIA is used for this period, there was considerable temporal and spatial variability across the Arctic during this period.

Ice shelves in northwestern Ellesmere Island probably reached their greatest extent in the Holocene during this interval. On the Devon Island Ice Cap, 1550 to 1620 is considered to have been a period of net summer accumulation, with very extensive summer sea ice in the region. There is widespread evidence of glaciers reaching their maximum post-Wisconsinan positions during the LIA, and the lowest $\delta^{18}\text{O}$ values and melt percentages for at least 1000 years are recorded in ice cores for this interval. Mann M. et al. (1999) and Jones et al. (1998) supported the theory that the 15th to 19th centuries were the coldest of the millennium for the Northern Hemisphere overall. However, averaged over the Northern Hemisphere, the temperature decrease during the LIA was less than 1 °C relative to late 20th-century levels (Crowley and Lowery, 2000; Jones et al., 1998; Mann M. et al., 1999). Cold conditions appear, however, to have been considerably more pronounced in particular regions during the LIA. Such regional variability may in part reflect accompanying changes in atmospheric circulation. Overpeck et al. (1997) summarized arctic climate change over the past 400 years.

There is an abundance of evidence from the Arctic that summer temperatures have decreased over approximately the past 3500 years. In the Canadian Arctic, the melt record from the Agassiz ice core indicates a decline in summer temperatures since approximately

5.5 ky BP, especially after 2 ky BP. In Alaska, widespread glacier advances were initiated at approximately 700 ky BP and continued through the 19th century (Calkin et al., 2001). During this interval, the majority of Alaskan glaciers reached their Holocene maximum extensions. The pattern of LIA glacier advances along the Gulf of Alaska is similar on decadal timescales to that of the well-dated glacier fluctuations throughout the rest of Alaska.

There is a general consensus that throughout the Canadian Archipelago, the late Holocene has been an interval of progressive cooling (the “Neoglacial”, culminating in the LIA), followed by pronounced warming starting about 1840 (Overpeck et al., 1997). According to Bourgeois et al. (2000), the coldest temperatures of the entire Holocene were reached approximately 100 to 300 years ago in this region. Others, working with different indicators, have suggested that Neoglacial cooling was even greater in areas to the south of the Canadian Archipelago (Johnsen et al., 2001). Therefore, even if the broad pattern of Holocene climatic evolution is assumed to be coherent across the Canadian Archipelago, the available data suggest regional variation in the amplitude of temperature shifts.

The most extensive data on the behavior of Greenland glaciers apart from the Greenland Ice Sheet come from Maniitsoq (Sukkertoppen) and Disko Island. Similar to the inland ice-sheet lobes, the majority of the local glaciers reached their maximum Neoglacial extent during the 18th century, possibly as early as 1750. Glaciers started to retreat around 1850, but between 1880 and 1890 there were glacier advances. In the early 20th century, glacier recession continued, with interruptions by some periods of advance. The most rapid glacial retreat took place between the 1920s and 1940s.

In Iceland, historical records indicate that Fjallsjökull and Breidamerkurjökull reached their maximum Holocene extent during the latter half of the 19th century (Kugelmann, 1991). Between 1690 and 1710, the Vatnajökull outlet glaciers advanced rapidly and then were stationary or fluctuated slightly. Around 1750 to 1760 a significant re-advance occurred, and most of the glaciers are considered to have reached their maximum LIA extent at that time (e.g., Grove, 1988). During the 20th century, glaciers retreated rapidly. During the LIA, Myrdalsjökul and Eyjafjallsjökull formed one ice cap, which separated in the middle of the 20th century into two ice caps (Grove, 1988). Drangajökull, a small ice cap in northwest Iceland, advanced across farmland by the end of the 17th century, and during the mid-18th century the outlet glaciers were the most extensive known since settlement of the surrounding valleys. After the mid-19th century advance, glaciers retreated significantly. On the island of Jan Mayen, some glaciers reached their maximum extent around 1850. The glaciers subsequently experienced an oscillating retreat, but with a significant expansion around 1960 (Anda et al., 1985).

In northeastern Eurasia, long-term temperature trends derived from tree rings close to the northern treeline in east Taymir and northeast Yakutia indicate decreasing temperatures during the LIA (Vaganov et al., 2000).

Variations in arctic climate over the past 1000 years may have been the result of several forcing mechanisms. Bond et al. (2001) suggested variations in solar insolation. Changes in the thermohaline circulation or modes of atmospheric variability, such as the AO, may also have been primary forcing mechanisms of century- or millennial-scale changes in the Holocene climate of the North Atlantic. It is possible that solar forcing may excite modes of atmospheric variability that, in turn, may amplify climate changes. The Arctic, through its linkage with the Nordic Seas, may be a key region where solar-induced atmospheric changes are amplified and transmitted globally through their effect on the thermohaline circulation. The resulting reduction in northward heat transport may have further altered latitudinal temperature and moisture gradients.

2.7.6. Concluding remarks

Natural climate variability in the Arctic over the past two million years has been large. In particular, the past 20000-year period is now known to have been highly unstable and prone to rapid changes, especially temperature increases that occurred rapidly (within a few decades or less). These temperature increases occurred during glacial terminations and at the onset of D-O interstadials. This instability implies rapid, closely linked changes within the earth's environmental system, including the hydrosphere, atmosphere, cryosphere, and biosphere. Not only has the climate of the Arctic changed significantly over the past two million years, there have also been pronounced regional variations associated with each change.

The Arctic is not homogeneous and neither is its climate, and past climate changes have not been uniform in their characteristics or their effects. Many of these changes have not been synchronous nor have they had equal magnitudes and rates of change. Climate changes in one part of the Arctic may trigger a delayed response elsewhere, adding to the complexity. The paleoenvironmental evidence for the Arctic suggests that at certain times, critical thresholds have been passed and unpredictable responses have followed. The role that anthropogenic changes to the climate system might play in exceeding such thresholds and the subsequent response remains unclear.

It is clear that between 400 and 100 years BP, the climate in the Arctic was exceptionally cold. There is widespread evidence of glaciers reaching their maximum post-Wisconsinan positions during this period, and the lowest $\delta^{18}\text{O}$ values and melt percentages for at least 1000 years are recorded in ice cores for this interval. The observed warming in the Arctic in the latter half of the 20th century appears to be without

precedent since the early Holocene (Mann M. and Jones, 2003).

2.8. Summary and key findings

This chapter has described the arctic climate system; the region's impact on the global climate system; recent climatic change depicted by the instrumental record; and the historical/paleoclimatic perspective on arctic climatic variability. Features of the arctic climate system that are unique to the region include the cryosphere, the extremes of solar radiation, and the role of salinity in ocean dynamics.

The climate of the Arctic is changing. Trends in instrumental records over the past 50 years indicate a reasonably coherent picture of recent environmental change in northern high latitudes. The average surface temperature in the Arctic increased by approximately 0.09 °C per decade over the past century, and the pattern of change is similar to the global trend (i.e., an increase up to the mid-1940s, a decrease from then until the mid-1960s, and a steep increase thereafter with a warming rate of 0.4 °C per decade). It is very probable that the Arctic has warmed over the past century, at a rate greater than the average over the Northern Hemisphere. It is probable that polar amplification has occurred over the past 50 years. Because of the scarcity of observations across the Arctic before about 1950, it is not possible to be certain of the variation in mean land-station temperature over the first half of the 20th century. However, it is probable that the past decade was warmer than any other in the period of the instrumental record. The observed warming in the Arctic appears to be without precedent since the early Holocene.

It is very probable that atmospheric pressure over the Arctic Basin has been dropping and it is probable that there has been an increase in total precipitation at the rate of about 1% per decade over the past century. It is very probable that snow-cover extent around the periphery of the Arctic, sea-ice extent averaged over the Arctic (during at least the past 40 years), and multi-year sea-ice extent in the central Arctic have all decreased.

These climate changes are consistent with projections of climate change by global climate models forced with increasing atmospheric GHG concentrations, but definitive attribution is not yet possible.

Natural climate variability in the Arctic over the past two million years has been substantial. In particular, the past 20000-year period is now known to have been highly unstable and prone to large rapid changes, especially temperature increases that occurred quickly (within a few decades or less). It is clear that between 400 and 100 years BP the climate in the Arctic was exceptionally cold, and there is widespread evidence of glaciers reaching their maximum post-Wisconsinan positions during this period.

Changes in the Arctic are very likely to have significant impacts on the global climate system. For example, a reduction in snow-cover extent and a shrinking of the marine cryosphere would increase heating of the surface, which is very likely to accelerate warming of the Arctic and reduce the equator-to-pole temperature gradient. Freshening of the Arctic Ocean by increased precipitation and runoff is likely to reduce the formation of cold deep water, thereby slowing the global thermohaline circulation. It is likely that a slowdown of the thermohaline circulation would lead to a more rapid rate of rise of global sea level, reduce upwelling of nutrients, and exert a chilling influence on the North Atlantic region as Gulf Stream heat transport is reduced. It would also decrease the rate at which CO₂ is transported to the deep ocean. Finally, temperature increases over permafrost areas could possibly lead to the release of additional CH₄ into the atmosphere; if seabed temperatures rise by a few degrees, hydrated CH₄ trapped in solid form could also escape into the atmosphere.

Although it is possible to draw many conclusions about past arctic climate change, it is evident that further research is still needed. The complex processes of the atmosphere, sea-ice, ocean, and terrestrial systems should be further explored in order to improve projections of future climate and to assist in interpreting past climate. Reconstructions of the past have been limited by available information, both proxy and instrumental records. The Arctic is a region of large natural variability and regional differences and it is important that more uniform coverage be obtained to clarify past changes. In order for the quantitative detection of change to be more specific in the future, it is essential that steps be taken now to fill in observational gaps across the Arctic, including the oceans, land, ice, and atmosphere.

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