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Past Climate Variability and Change in the Arctic and at High Latitudes

Chapter 7 — History of Sea Ice in the Arctic

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ABSTRACT

The volume and areal extent of Arctic sea ice is rapidly declining, and to put that decline into perspective we need to know the history of Arctic sea ice in the geologic past. Sedimentary proxy records from the Arctic Ocean floor and from the surrounding coasts can provide clues. Although incomplete, existing data outline the development of Arctic sea ice during the last several million years. Some data indicate that sea ice consistently covered at least part of the Arctic Ocean for no less than 13–14 million years, and that ice was most widespread during the last approximately 2 million years in relationship with Earth's overall cooler climate. Nevertheless, episodes of considerably reduced ice cover or even a seasonally ice-free Arctic Ocean probably punctuated even this latter period. Ice diminished episodically during warmer climate events associated with changes in Earth's orbit on the time scale of tens of thousands of years. Ice cover in the Arctic began to diminish in the late 19th century and this shrinkage has accelerated during the last several decades. Shrinkages that were both similarly large and rapid have not been documented over at least the last few thousand years, although the paleoclimatic record is sufficiently sparse that similar events might have been missed. Orbital changes have made ice melting less likely than during the previous millennia since the end of the last ice age, making the recent changes especially anomalous. Improved reconstructions of sea-ice history would help clarify just how anomalous these recent changes are.

7.1 Introduction

The most defining feature of the surface of the Arctic Ocean and adjacent seas is its sea ice cover, which waxes and wanes with the seasons, and which also changes in extent and thickness on interannual and longer time scales. These changes in ice cover are related to climate, notably temperature changes (e.g., Smith et al., 2003), and themselves affect atmospheric and hydrographic conditions in high latitudes (Kinnard et al., 2008; Steele et al., 2008). Observations during the past several decades document substantial retreat and thinning of the Arctic sea ice cover: retreat is accelerating, and it is expected to continue. The Arctic Ocean may become seasonally ice free as early as 2040 (Holland et al., 2006a; Comiso et al., 2008; Stroeve et al., 2008). A reduction in sea ice will promote Arctic warming through a feedback mechanism between ice and its reflectivity (the ice-albedo feedback mechanism), and this reduction will thus influence weather systems in the northern high and perhaps middle latitudes. Changes in ice cover and freshwater flux out of the Arctic Ocean will also affect oceanic circulation of the North Atlantic, which has profound influence on climate in Europe and North America (Seager et al., 2002; Holland et al., 2006b). Furthermore, continued retreat of sea ice will accelerate coastal erosion owing to increased wave action. Ice loss will modify the Arctic Ocean food web and its large predators, such as polar bears and seals, that depend on the ice cover. These changes, in turn, will affect indigenous human populations that harvest such species. All of these possibilities make it important to know how fast Arctic ice will diminish and the consequences of that reduction, a task that requires thorough understanding of the natural variability of ice cover in the recent and longer term past.

7.2 Background on Arctic Sea-Ice Cover

7.2.1 Ice Extent, Thickness, Drift and Duration

Arctic sea ice cover grows to its maximum extent by the end of winter and shrinks to a minimum in September. For the period of reliable satellite observations (1979–2007), extremes in Northern Hemisphere ice extent are 16.44×10^6 square kilometers (km^2) for March 1979 and 4.28×10^6 km^2 for September 2007 (http://nsidc.org/data/seaice_index/; Stroeve et al., 2008). Ice extent is defined as the region of the ocean of which at least 15% is covered by ice. The ice cover can be broadly divided into a perennial ice zone where ice is present throughout the year and a seasonal ice zone where ice is present only seasonally. A considerable fraction of Arctic sea ice is perennial, which differs strongly from Antarctic sea ice which is nearly all seasonal. Ice concentrations in the perennial ice zone typically exceed 97% in winter but fall to 85–95% in summer. Sea ice concentrations in the seasonal ice zone are highly variable, and in general (but not always) they decrease toward the southern sea ice margin.

The thickness of sea ice, which varies markedly in both space and time, can be described by a probability distribution. For the Arctic Ocean as a whole, the peak of this distribution (as thick as the ice ever gets) is typically cited at about 3 meters (m) (Serreze et al., 2007b), but growing evidence (discussed below) suggests that during recent decades not only is the area of sea ice shrinking, but that it is also thinning substantially. Although many different types of sea ice can be defined, the two basic categories are first-year ice, which represents a single year's growth, and multiyear ice, which has survived one or more melt seasons. Undeformed first-year ice typically is as much as 2 m thick. Although in general multiyear ice is thicker (greater than 2 m), first-year ice that is locally pushed into ridges can be as thick as 20–30 m.

Under the influence of winds and ocean currents, the Arctic sea ice cover is in nearly constant motion. The large-scale circulation principally consists of the Beaufort Gyre, a mean annual clockwise motion in the western Arctic Ocean with a drift speed of 1–3 centimeters per second, and the Transpolar Drift, the movement of ice from the coast of *Siberia* east across the pole and into the North Atlantic by way of *Fram Strait*, which lies between northern *Greenland* and *Svalbard*. Ice velocities in the Transpolar Drift increase toward Fram Strait, where the mean drift speed is 5–20 centimeters per second (Figure 7.1) (Thorndike, 1986; Gow and Tucker, 1987). About 20% of the total ice area of the Arctic Ocean is discharged each year through *Fram Strait*, the majority of which is multiyear ice. This ice subsequently melts in the northern North Atlantic, and since the ice is relatively fresh compared with sea water, this melting adds freshwater to the ocean in those regions.

FIGURE 7.1 NEAR HERE

7.2.2 Influences on the Climate System

Seasonal changes in the amount of heat at the surface (net surface heat flux) associated with sea ice modulate the exchange and transport of energy in the atmosphere. Ice, as sheets or as sea ice, reflects a certain percentage of incoming solar radiation back into the atmosphere. The albedo (reflectivity) of ice cover ranges from 80% when it is freshly snow covered to around 50% during the summer melt season (but lower in areas of ponded ice). This high reflectivity contrasts with the dark ocean surface, which has an albedo of less than 10%. Ice's high albedo and its large surface area, coupled with the solar energy used to melt ice and to increase the sensible heat content of the ocean, keep the Arctic atmosphere cool during summer. This cooler polar atmosphere helps to maintain a steady poleward transport of atmospheric energy (heat) from lower latitudes into the Arctic. During autumn and winter, energy derived from incoming solar radiation is small or nonexistent in Polar areas. However, heat loss from the surface adds heat to the atmosphere, and it reduces the requirements for atmospheric heat to be transported poleward into the Arctic (Serreze et al., 2007a).

Model experiments have addressed potential changes in the regional and large-scale aspects of atmospheric circulation that are associated with loss of sea ice. The models commonly use ice conditions that have been projected through the 21st century (see following section). Magnusdottir et al. (2004) found that a reduced area of winter sea ice in the North Atlantic modified the modeled circulation in the same way as the North Atlantic Oscillation; declining ice promotes a negative North Atlantic Oscillation response: storm tracks are weaker and shifted to the south. Many observations show that sea ice in this region affects the development of mid- and high-latitude cyclones because of the strong horizontal temperature gradients along the ice margin (e.g., Tsukernik et al., 2007). Singarayer et al. (2006) forced a model by combining the area of sea ice in 1980–2000 and projected reductions in sea ice until 2100. In one simulation, mid-latitude storm tracks were intensified and they increased winter precipitation throughout western and southern Europe. Sewall and Sloan (2004) found that reduced ice cover led to less rainfall in the American west. In summary, although these and other simulations point to the importance of sea ice on climate outside of the Arctic, different models may produce very different results. Coordinated experiments that use a suite of models is needed to help to reduce uncertainty.

Climate models also indicate that changes in the melting of and export of sea ice to the North Atlantic can modify large-scale ocean circulation (e.g., Delworth et al., 1997; Mauritzen and Hakkinen, 1997; Holland et al., 2001). In particular, exporting more freshwater from the Arctic increases the stability of the upper ocean in the northern North Atlantic. This may suppresses convection, leading to reduced formation of North Atlantic Deepwater and weakening of the Atlantic meridional overturning cell (MOC). This suppression may have far-reaching climate consequences. The considerable freshening of the North Atlantic since the 1960s has an Arctic source (Peterson et al., 2006). Total Arctic freshwater output to the North Atlantic is projected to increase through the 21st century, and decreases in the export of sea ice will be more than balanced by the export of liquid freshwater (derived from the melting of Arctic ice and increased net precipitation). However, less ice may melt in the *Greenland-Iceland-Norwegian (GIN) seas* because less ice is moved through *Fram Strait* into those seas. These changes may increase vertical instability in the ocean regions where deep water forms and counteract the tendency of a warmer climate to increase ocean stability (Holland et al., 2006b). However, this possible instability may be mitigated somewhat if less sea ice accumulates in the *Greenland-Iceland-Norwegian seas*. Additionally, as discussed by Levermann et al. (2007), the reduction in sea ice may help to stabilize the Atlantic meridional overturning circulation by removing the insulating ice cover which, perhaps counterintuitively, limits the amount of heat lost by the ocean to the atmosphere. Thus, sea ice may help to maintain the formation of deep water in the *Greenland-Iceland-Norwegian seas*. Overall, a smaller area of sea ice influences the Atlantic meridional overturning circulation in sometimes competing ways. How they will ultimately affect future climate is not yet certain.

7.2.3 Recent Changes and Projections for the Future

On the basis of satellite records, the extent of sea ice has diminished in every month and most obviously in September, for which the trend for the period 1979–2007 is 10% per decade (Figure 7.2). (Satellite records originated in the National Snow and Ice Data Center (http:/nsidc.org/data/seaice_index/) and combine information from the Nimbus-7 Scanning Multichannel Microwave Radiometer (October 1978–1987) and the Defense Meteorological Satellite Program Special Sensor Microwave/Imager (1987–-present.) Conditions in 2007 serve as an exclamation point on this ice loss (Comiso et al., 2008; Stroeve et al., 2008). The average September ice extent in 2007 of 4.28 million km^2 was not only the least ever recorded but also 23% lower than the previous September record low of 5.56 million km^2 set in 2005. The difference in areas corresponds with an area roughly the size of Texas and California combined. On the basis of an extended sea ice record, it appears that area of ice in September 2007 is only half of its area in 1950–70 (estimated by use of the Hadley Centre sea ice and sea surface temperature data set (HadlSST) (Rayner et al., 2003)..

FIGURE 7.2 NEAR HERE

Many factors may have contributed to this ice loss (as reviewed by Serreze et al., 2007b), such as general Arctic warming (Rothrock and Zhang, 2005), extended summer melt (Stroeve et al., 2006), effects of the changing phase of the Northern Annular Mode and the North Atlantic Oscillation. These and other atmospheric patterns have flushed some older, thicker ice out of the Arctic and left thinner ice that is more easily melted out in summer (e.g., Rigor and Wallace, 2004; Rothrock and Zhang, 2005; Maslanik et al., 2007a), changed ocean heat transport (Polyakov et al., 2005; Shimada et al., 2006), and increased recent spring cloud cover that augments the longwave radiation flux to the surface (Francis and Hunter, 2006). Strong evidence for a thinning ice cover comes from an ice-tracking algorithm applied to satellite and buoy data, which suggests that the area of the Arctic Ocean covered by predominantly older (and hence generally thicker) ice (ice 5 years old or older) decreased by 56% between 1982 and 2007. Within the central Arctic Ocean, the coverage of old ice has declined by 88%, and ice that is at least 9 years old (ice that tends to be sequestered in the Beaufort Gyre) has essentially disappeared. Examination of the distribution of ice of various thickness suggests that this loss of older ice translates to a decrease in mean thickness for the Arctic from 2.6 m in March 1987 to 2.0 m in 2007 (Maslanik et al., 2007b).

The role of greenhouse gas forcing on the observed ice loss finds strong support from the study of Zhang and Walsh (2006). These authors show that for the period 1979–1999, the multi-model mean trend projected by models discussed in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC-AR4) is downward, as are trends from most individual models. However, Stroeve et al. (2007) find that few or none (depending on the time period of analysis) of the September trends from the IPCC-AR4 runs are as large as observed. If the multi-model mean trend is assumed to be a reasonable representation of change forced by increased concentrations of greenhouse gases, then 33–38% of the observed September trend from 1953 to 2006 is externally forced and that percentage increases to 47–-57% from 1979 to 2006, when both the model mean and observed trend are larger. Although this trend argues that natural variability has strongly contributed to the observed trend, Stroeve et al. (2006) concluded that, as a group, the models underestimate the sensitivity of sea ice cover to forcing by greenhouse gases. Overly thick ice assumed by many of the models appears to provide at least a partial explanation.

The Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC-AR4) models driven with the SRES A1B emissions scenario (in which CO_2 reaches 720 parts per million (ppm), in comparison to the current value of 380 ppm, by the year 2100), point to complete or nearly complete loss (less than 1×10^6 km^2) of September sea ice anywhere from year 2040 to well beyond the year 2100, depending on the model and particular run (ensemble member) for that model**.** Even by the late 21st century, most models project a thin ice cover in March (Serreze et al., 2007b). However, given the findings just discussed, the models as a group may be too conservative—predict a later rather than earlier date—when the Arctic Ocean will be ice-free in summer.

Abrupt change in future Arctic ice conditions is difficult to model. For instance, the extent of end-of-summer ice is sensitive to ice thickness in spring (simulations based on the Community Climate System Model, version 3 (Holland et al., 2006a)). If the ice is already thin in the spring, then a "kick" associated with natural climate variability might make it melt rapidly in the summer owing to ice-albedo feedback. In the Community Climate System Model, version 3 events, anomalous ocean heat transport acts as this trigger. In one ensemble member, the area of September ice decreases from about 6×10^6 km^2 to 2×10^6 km^2 in 10 years, resulting in an essentially ice-free September by 2040. This result is not just an artifact of Community Climate System Model, version 3: a number of other climate models show similar rapid ice loss.

These recent reductions in the extent and thickness of ice cover and the projections for its further shrinkage necessitate a comprehensive investigation of the longer term history of Arctic sea ice. To interpret present changes we need to understand the Arctic's natural variability. A special emphasis should be placed on the times of change such as the initiation of seasonal and then perennial ice and the periods of its later reductions.

7.3 Types of Paleoclimate Archives and Proxies for the Sea-Ice Record

The past distribution of sea ice is recorded in sediments preserved on the sea floor and in deposits along many Arctic coasts. Indirect information on sea-ice extent can be derived from cores drilled in glaciers and ice sheets such as the *Greenland Ice Sheet*. Ice cores record atmospheric precipitation, which is linked with air-sea exchanges in surrounding oceanic areas. Such paleoclimate information provides a context within which the patterns and effects of the current and future ice-reduced state of the Arctic can be evaluated.

7.3.1 Marine Sedimentary Records

The most complete and spatially extensive records of past sea ice are provided by sea-floor sediments from areas that are or have been covered by floating ice. Sea ice affects deposition of such sediments directly or indirectly through physical, chemical, and biological processes. These processes and, thus, ice characteristics can be reconstructed from a number of sediment proxies outlined below.

Sediment cores that represent the long-term history of sea ice embracing several million years are most likely to be found in the deep, central part of the Arctic Ocean where the sea floor was not eroded during periods of lower sea-level (and larger ice sheets). On the other hand, rates of sediment deposition in the central Arctic Ocean are generally low, on the order of centimeters or even millimeters per thousand years (Backman et al., 2004; Darby et al., 2006), so that sedimentary records from these areas may not capture short-term variations in paleoenvironments. In contrast, cores from Arctic continental margins usually represent a much shorter time interval, less than 20 thousand years (k.y.) since the last glacial maximum, but they sometimes provide high-resolution records that capture events on century or even decadal time scales. Therefore, investigators need sediment cores from both the central basin and continental margins of the Arctic Ocean to fully characterize sea-ice history and its relation to climate change.

Until recently, and for logistical reasons, most cores relevant to the history of sea ice cover were collected from low-Arctic marginal seas, such as the *Barents Sea* and the *Norwegian-Greenland Sea*. There, modern ice conditions allow for easier ship operation, whereas sampling in the central Arctic Ocean requires the use of heavy icebreakers. Recent advances in drilling the floor of the Arctic Ocean—notably the first deep-sea drilling in the central Arctic Ocean (ACEX: Backman et al., 2006) and the 2005 Trans-Arctic Expedition (HOTRAX: Darby et al., 2005)—provide new, high-quality material from the Arctic Ocean proper with which to characterize variations in ice cover during the late Cenozoic (the last few million years).

A number of sediment proxies have been used to predict the presence or absence of sea ice in down-core studies. The most direct proxies are derived from sediment that melts out or drops from ice owing to the following sequence of processes: (1) sediment is entrained in sea ice, (2) this ice is transported by wind and surface currents to the sites of interest, and (3) sediment is released and deposited. The size of sediment grains is commonly analyzed to identify ice-rafted debris. The entrainment of sediments in sea ice mostly occurs along the shallow continental margins during periods of ice freeze-up and is largely restricted to silt and clay-size sediments and rarely contains grains larger than 0.1 millimeters (mm) (Lisitzin, 2002; Darby, 2003). Coarser ice-rafted debris is mostly transported by floating icebergs rather than by regular sea ice (Dowdeswell et al., 1994; Andrews, 2000). A small volume of coarse grains are shed from steep coastal cliffs onto land-fast ice. To link sediment with sea ice may require investigations other than measurement of grain size: for example, examination of shapes and surface textures of quartz grains will help distinguish sea-ice-rafted and iceberg-rafted material (Helland and

Holmes, 1997; Dunhill et al., 1998). Detailed grain-size distributions say something about ice conditions. For example, massive accumulation of silt-size grains (mostly larger than 0.01 mm) may indicate the position of an ice margin where melting ice is the source of most sediment (Hebbeln, 2000).

Some indicators (sediment provenance indicators) help to establish the source of sediment and thus help to track ice drift. Especially telling is sediment carrying some diagnostic peculiarity that is foreign to the site of deposition and that can be explained only by ice transport—such as the particular composition of iron-oxide sand grains, which can be matched with an extensive data base of source areas around the Arctic Ocean (Darby, 2003). Bulk sediment analyzed by quantitative methods such as X-ray diffraction can also be used in those instances where minerals that are “exotic” relative to the composition of the nearest terrestrial sources are deposited. Quartz in *Iceland* marine cores (Moros et al., 2006; Andrews and Eberl, 2007) and dolomite (limestone rich in magnesium), in sediments deposited along eastern *Baffin Island* and Labrador are two examples (Andrews et al., 2006).

Sediment cores commonly contain skeletons of microscopic organisms (for example foraminifers, diatoms, and dinocysts). These findings are widely used for deciphering the past environments in which these organisms lived. Some marine planktonic organisms live in or on sea ice or are otherwise associated with ice. Their skeletons in bottom sediments indicate the condition of ice cover above the study site. Other organisms that live in open water can be used to identify intervals of diminished ice. Remnants of ice-related algae such as diatoms and dinocysts have been used to infer changes in the length of the ice-cover season (Koç and Jansen, 1994; de Vernal and Hillaire-Marcel, 2000; Mudie et al., 2006; Solignac et al., 2006). To quantify the relationship between these organisms and paleoenvironment, three major research steps are required. The first is to develop a database of the percent compositions in a certain group of organisms from water-column or surficial sea-floor samples that span a wide environmental range. Second, various statistical methods must be used to express the relationship (usually called "transfer functions") between these compositions and key environmental parameters, such as sea-ice duration and summer surface temperatures. Finally, after sediment cores are analyzed and transfer functions are developed on the modern data sets, they are then applied to the temporal (i.e., down-core) data. The usefulness of the transfer functions, however, depends upon the accuracy of the environmental data, which is commonly quite limited in Arctic areas.

Bottom dwelling (benthic) organisms in polar seas are also affected by ice cover because it controls what food can reach the sea floor. The particular suite of benthic organisms preserved in sediments can help to identify ice-covered sites. For instance, environments within the pack ice produce very little organic matter, whereas environments on the margin of the ice produce a great deal. Accordingly, species of bottom-dwelling organisms that prefer relatively high fluxes of fresh organic matter can indicate, for the Arctic shelves, the location of the ice margin (Polyak et al., 2002; Jennings et al., 2004). In the central Arctic Ocean, benthic foraminifers and ostracodes also offer a good potential for identifying ice conditions (Cronin et al., 1995; Wollenburg and Kuhnt, 2000; Polyak et al., 2004).

The composition of organic matter in sediment, including specific organic compounds (biomarkers), can also be used to reconstruct the environment in which it formed. For instance, a specific biomarker, IP25, can be associated with diatoms living in sea ice (Belt et al., 2007). The method has been tested by the analysis of sea-floor samples from the *Canadian Arctic* and is being further applied to down-core samples for characterization of past ice conditions.

It is important to understand that although all of the above proxies have a potential for identifying the former presence of or the seasonal duration of sea-ice cover, each of them has limitations that complicate interpretations based on a single proxy. For instance, by use of a dinocyst transfer function from East Greenland, it was estimated that the sea-ice duration is about 2–3 months (Solignac et al., 2006) when in reality it is closer to 9 months (Hastings, 1960). Agreement among many proxies is required for a confident inference about variations in sea-ice conditions. A thorough understanding of sea-ice history depends on the refining of sea-ice proxies in sediment taken from strategically selected sites in the Arctic Ocean and along its continental margins.

7.3.2 Coastal Records

In many places along the Arctic and subarctic coasts, evidence of the extent of past sea ice is recorded in coastal-plain sediments, marine terraces, ancient barrier island sequences, and beaches. Deposits in all of these formerly marine environments are now above water owing to relative changes in sea level caused by eustatic, glacioisostatic, or tectonic factors. Although these coastal deposits represent a limited time span and geographic distribution, they provide critical information that can be compared with marine sediment records. The primary difference between coastal and sea-floor records is in the type of fossils recovered. Notably, the spacious coastal exposures (as compared with sediment cores) enable large paleontological material such as plant remains, driftwood, whalebone, and relatively large mollusks to be recovered. These items contribute valuable information about past sea-surface and air temperatures, the northward expansions of subarctic and more temperate species, and the seasonality of past sea-ice cover. For example, fossils preserved in these sequences document the dispersals of coastal marine biota between the Pacific, Arctic, and North Atlantic regions, and they commonly carry telling evidence of ice conditions. Plant remains in their turn provide a much-needed link to documented information about past vegetation on land throughout Arctic and subarctic regions. The location of the northern tree line that is presently controlled by the July 7°C mean isotherm is a critical paleobotanic indicator for understanding ice conditions in the Arctic. Nowhere in the Arctic do trees exist near shores lined with perennial sea ice; they thrive only in southerly reaches of regions of seasonal ice. The combination of spatial relationships between marine and terrestrial data allows a comprehensive reconstruction of past climate.

7.3.3 Coastal Plains and Raised Marine Sequences

A number of coastal plains around the Arctic are blanketed by marine sediment sequences laid down during high sea levels. Although these sequences lie inland of coastlines that today are bordered by perennial or by seasonal sea ice, they commonly contain packages of fossil-rich sediments that provide an exceptional record of earlier warm periods. The most well-documented sections are those preserved along the eastern and northern coasts of *Greenland* (Funder et al., 1985, 2001), the eastern *Canadian Arctic* (Miller et al., 1985), *Ellesmere Island* (Fyles et al., 1998), *Meighen Island* (Matthews, 1987; Matthews and Overden, 1990; Fyles et al., 1991), *Banks Island* (Vincent, 1990; Fyles et al., 1994), the *North Slope of Alaska* (Carter et al., 1986; Brigham-Grette and Carter, 1992); the *Bering Strait* (Kaufman and Brigham-Grette, 1993; Brigham-Grette and Hopkins, 1995), and in the western *Eurasian Arctic* (Funder et al., 2002) (Figure 7.3). In nearly all cases the primary evidence used to estimate the extent of past sea ice is *in situ* molluscan and microfossil assemblages. These assemblages, from many sites, coupled with evidence for the northward expansion of tree line during interglacial intervals (e.g., Funder et al., 1985; Repenning et al., 1987; Bennike and Böcher, 1990; CAPE, 2006), provide an essential view of past sea-ice conditions with direct implications for sea surface temperatures, sea ice extent, and seasonality.

FIGURE 7.3 NEAR HERE

7.3.4 Driftwood

The presence or absence of sea ice may be inferred from the distribution of tree logs, mostly spruce and larch found in raised beaches along the coasts of *Arctic Canada* (Dyke et al., 1997), *Greenland* (Bennike, 2004), *Svalbard* (Haggblom, 1982), and *Iceland* (Eggertsson, 1993). Coasts with the highest numbers of driftwood probably were once near a sea-ice margin, whereas coasts hosting more modest amounts were near either too much ice or too open water—neither of which deliver much driftwood. Most of the logs found are attributed to a northern Russian source, although some can be traced to northwest Canada and Alaska. Logs can drift only about 1 year before they become waterlogged and sink (Haggblom, 1982). The logs are probably derived from rivers flooded by spring snowmelt, which bring sediment and trees onto **landfast ice** around the margin of the Arctic Basin. In areas other than Iceland, the glacial isostatic uplift of the land has led to a staircase of raised beaches hosting various numbers of logs with time. An extensive database catalogs these variations in the beaching of logs during the present interglacial (Holocene). These variations have been associated with the growth and disappearance of landfast sea ice (which restricts the beaching of driftwood) and changes in atmospheric circulation with resulting changes in ocean surface circulation (Dyke et al., 1997).

7.3.5 Whalebone

Reconstructions of sea-ice conditions in the Canadian Arctic Archipelago have to date been derived mainly from the distribution in space and time of marine mammal bones in raised marine deposits (Dyke et al., 1996, 1999; Fisher et al., 2006). Several large marine mammals have strong affinities for sea ice: polar bear, several species of seal, walrus, narwhal, beluga (white) whale, and bowhead (Greenland right) whale. Of these, the bowhead has left the most abundant, hence most useful, fossil record, followed by the walrus and the narwhal. Radiocarbon dating of these remains has yielded a large set of results, largely available through Harington (2003) and Kaufman et al. (2004).

Former sea-ice conditions can be reconstructed from bowhead whale remains because seasonal migrations of the whale are dictated by the oscillations of the sea-ice pack. The species is thought to have had a strong preference for ice-edge environments since the Pliocene (2.6–5.3 million years ago (Ma)), perhaps because that environment allows it to escape from its only natural predator, the killer whale. The Pacific population of bowheads spends winter and early spring along the ice edge in the *Bering Sea* and advances northward in the summer ice into the Canadian *Beaufort Sea* region along the western edge of the *Canadian Arctic Archipelago*. The Atlantic population spends winter and early spring in the northern *Labrador Sea* between southwest Greenland and northern Labrador and advances northward in summer into the eastern channels of the *Canadian Arctic Archipelago*. In normal summers, the Pacific and Atlantic bowheads are prevented from meeting by a large, persistent, plug of sea-ice that occupies the central region of the *Canadian Arctic Archipelago*; i.e., the central part of the Northwest Passage (Figure 7.4). Both populations retreat southward upon autumn freeze-up.

FIGURE 7.4 NEAR HERE

However, the ice-edge environment is hazardous, especially during freeze-up, and individuals or pods may become entrapped (as has been observed today). Detailed measurements of fossil bowhead skulls (a proxy of age) now found in raised marine deposits allow a reconstruction of their lengths (Dyke et al., 1996; Savelle et al., 2000). The distribution of lengths compares very closely with the length distribution of the modern *Beaufort Sea* bowhead population (Figure 7.5), indicating that the cause of death of many bowheads in the past was a catastrophic process that affected all ages indiscriminately. This process can be best interpreted as ice entrapment.

FIGURE 7.5 NEAR HERE

7.3.6 Ice Cores

Among paleoenvironmental archives, ice cores from glaciers and ice sheets have a particular strength as a direct recorder of atmospheric composition, especially in the polar regions, at a fine time resolution. The main issue is whether ice cores contain any information about the past extent of sea ice. Such information may be inferred indirectly: for example, one can imagine that higher temperatures recorded in an ice core are associated with reduced sea ice. However, the real goal is to find a chemical indicator whose concentration is mainly controlled by past sea-ice extent (or by a combination of ice extent and other climate characteristics that can be deduced independently). Any such indicator must be transported for relatively long distances, as by wind, from the sea ice or the ocean beyond. Such an indicator frozen into ice cores would then allow ice cores to give an integrated view throughout a region for some time average, but the disadvantage is that atmospheric transport can then determine what is delivered to the ice.

The ice-core proxy that has most commonly been considered as a possible sea ice indicator is sea salt, usually estimated by measuring a major ion in sea salt, sodium (Na). In most of the world's oceans, salt in sea water becomes an aerosol in the atmosphere by means of a bubble bursting at the ocean surface, and formation of the aerosol is related to wind speed at the ocean surface (Guelle et al., 2001). Expanding sea ice moves the source region (open ocean) further from ice core sites, so that a first assumption is that a more extensive sea ice cover should lead to less sea salt in an ice core.

A statistically significant inverse relationship between annual average sea salt in the *Penny Ice Cap* ice core (*Baffin Island*) and the spring sea ice coverage in *Baffin Bay* (Grumet et al., 2001) was found for the 20th century, and it has been suggested that the extended record could be used to assess the extent of past sea ice in this region. However, the correlation coefficient in this study was low, indicating that only about 7% of the variability in the abundance of sea salt was directly linked to variability in position of sea ice. The inverse relationship between sea salt and sea-ice cover in *Baffin Bay* was also reported for a short core from *Devon Island* (Kinnard et al., 2006). However, more geographically extensive work is needed to show whether these records can reliably reconstruct past sea ice extent.

For *Greenland*, the use of sea salt in this way seems even more problematic. Sea salt in aerosol and snow throughout the Greenland plateau tends to peak in concentration in the winter months (Mosher et al., 1993; Whitlow et al., 1992), when sea ice extent is largest, which already suggests that other factors are more important than the proximity of open ocean. Most authors carrying out statistical analyses on sea salt in Greenland ice cores in recent years have found relationships with aspects of atmospheric circulation patterns rather than with sea ice extent (Fischer, 2001; Fischer and Mieding, 2005; Hutterli et al., 2007). Sea-salt records from Greenland ice cores have therefore been used as general indicators of storminess (inducing production of sea salt aerosol) and transport strength (Mayewski et al., 1994; O'Brien et al., 1995), rather than as sea ice proxies.

An alternative interpretation has arisen from study of Antarctic aerosol and ice cores, where the sea ice surface itself can be a source of large amounts of sea-salt aerosol in coastal Antarctica (Rankin et al., 2002). It has then been argued that, although sea salt concentrations and fluxes may be dominated by transport effects on a year-to-year basis, they could be used as an indicator of regional sea ice extent for Antarctica over longer time periods (Fischer *et al.*, 2007a; Wolff *et al.*, 2003). An Antarctic sea ice record covering 740 ka has been presented on this basis, showing extended sea ice at times of low temperature (Wolff *et al.*, 2006). The obvious question arises as to whether this inverted model of the relationship between sea salt and sea ice might also be applicable in the Arctic (Rankin *et al.*, 2005). Current ideas about the source of sea-ice relate it to the production of new, thin ice. In the regions around *Greenland* and the nearby islands, much of the sea ice is old ice that has been advected, rather than new ice. It therefore seems unlikely that the method can easily be applied under present conditions (Fischer et al., 2007). The complicated geometry of the oceans around *Greenland* compared with the radial symmetry of Antarctica also poses problems in any interpretation. It is possible that under the colder conditions of the last glacial period, new ice produced around *Greenland* may have led to a more dominant sea-ice source, opening up the possibility that there may be a sea ice record available within this period. However, there is no published basis on which to rely at the moment (2008), and the balance of importance between salt production and salt transport in the Arctic needs further investigation.

One other chemical (methanesulfonic acid, MSA) has been used as a sea-ice proxy in the Antarctic (e.g., Curran et al., 2003). However, studies of MSA in the Arctic do not yet support any simple statistical relationship with sea ice there (Isaksson et al., 2005).

In summary, sea salt in ice cores has the potential to add a well-resolved and regionally integrated picture of the past extent of sea ice extent. At one site weak statistical evidence supports a relationship between sea ice extent and sea salt. However, the complexities of aerosol production and transport mean that no firm basis yet exists for using sea salt in ice cores to estimate past sea-ice extent in the Arctic. Further investigation is warranted to establish whether such proxies might be usable: investigators need a better understanding of the sources of proxies in the Arctic region, further statistical study of the modern controls on their distribution, and modeling studies to assess proxies' sensitivity to major changes in sea-ice extent.

7.3.7 Historical Records

Historical records may describe recent paleoclimatic processes such as weather and ice conditions. The longest historical records of ice cover exist from ice-marginal areas that are more accessible for shipping, as exemplified by a compilation for the *Barents Sea* covering four centuries in variable detail (Vinje, 1999, 2001). Systematic records of the position of sea-ice margin around the Arctic Ocean have been compiled for the period since 1870 (Walsh, 1978; Walsh and Chapman, 2001). These sources vary in quality and availability with time. More reliable observational data on ice concentrations for the entire Arctic are available since 1953, and the most accurate data from satellite imagery is available since 1972 (Cavalieri et al., 2003).

Seas around *Iceland provide* a rare opportunity to investigate the ice record in a more distant past because Iceland has for1200 years recorded observations of drift ice (i.e., sea ice and icebergs) following the settlement of the island in approximately 870 CE (Koch, 1945; Bergthorsson, 1969; Ogilvie, 1984; Ogilvie et al., 2000). This long record has facilitated efforts to quantify the changes in the extent and duration of drift ice around the Iceland coasts during the last 1200 years (Koch, 1945; Bergthorsson, 1969). During times of extreme drift-ice incursions, ice wraps around *Iceland* in a clockwise motion. Ice commonly develops off the northwest and north coasts and only occasionally extends into southwest Iceland waters (Ogilvie, 1996). Historical sources have been used to construct a sea-ice index that compares well with springtime temperatures at a climate station in northwest Iceland (Figure 7.6).

FIGURE 7.6 NEAR HERE

7.4 History of Arctic Sea-Ice Extent and Circulation Patterns

7.4.1 Pre-Quaternary History (Prior to ~2.6 Ma ago)

The shrinkage of the perennial ice cover in the Arctic and predictions that it may completely disappear within the next 50 years or even sooner (Holland et al., 2006a; Stroeve et al., 2008) are especially disturbing in light of recent discoveries that sea ice in the Arctic has persisted for the past 2 million years and may have originated several million years earlier (Darby, 2008; Krylov et al., 2008). Until recently, evidence of long-term (million-year scale) climatic history of the north polar areas was limited to fragmentary records from the Arctic periphery. The *ACEX deep-sea drilling borehole* in the central Arctic Ocean (Backman et al., 2006) provides new information about its Cenozoic history for comparison with circum-Arctic records. Drilling results confirmed that about 50 Ma, during the Eocene Optimum (Figure 3.8 in Chapter 3), the Arctic Ocean was considerably warmer than it is today, as much as 24°C at least in the summers, and fresh-water subtropical aquatic ferns grew in abundance (Moran et al., 2006). This environment is consistent with forests of enormous *Metasequoia* that stood at the same time on shores of the Arctic Ocean—such as on *Ellesmere Island* across lowlying delta floodplains riddled with lakes and swamps (Francis, 1988; McKenna, 1980) Coarse grains occurring in ACEX sediment as old as about 46 Ma indicate the possible onset of drifting ice and perhaps even some glaciers in the Arctic during the cooling that followed the thermal optimum (Moran et al., 2006; St. John, 2008). This cooling matches the timing of a large-scale reorganization of the continents, notably the oceanic separation of Antarctica and of a sharp decrease in atmospheric CO_2 concentration of more than1,000 parts per million (ppm) (Pearson and Palmer, 2000; Lowenstein and Demicco, 2006; also see Figure 4.24). However, in the Eocene the *ACEX site* was at the margin of rather than in the center of the Arctic Ocean (O'Regan et al., 2008) and therefore coarse grains may have been delivered to this site by rivers rather than by drifting ice. The circum-Arctic coasts at this time were still occupied by rich, high-biomass forests of redwood and by wetlands characteristic of temperate conditions (LePage et al., 2005; Williams et al., 2003). Continued cooling, punctuated by an abrupt temperature decrease at the Eocene-Oligocene boundary about 34 Ma, triggered massive Antarctic glaciation. It may have also led to the increase in winter ice in the Arctic. This inference cannot yet be verified in the central Arctic Ocean because the ACEX record contains no sediment deposited between about 44 to 18 Ma. Mean annual temperatures at the Eocene-Oligocene transition (about 33.9 Ma) dropped from nearly 11°C to 4°C in southern Alaska (Wolfe, 1980, 1997) at this time, whereas fossil assemblages and isotopic data in marine sediments along the coasts of the *Beaufort Sea* suggest waters with a seasonal range between 1°C and 9°C (Oleinik et al., 2007). The first glaciers may have developed in *Greenland* about the same time, on the basis of coarse grains interpreted as iceberg-rafted debris in the North Atlantic (Eldrett et al., 2007). Sustained, relatively warm conditions lingered during the early Miocene (about 23–16 Ma) when cool-temperate *Metasequoia* dominated the forests of northeast *Alaska* and the *Yukon* (White and Ager, 1994; White et al., 1997), and the central Canadian Arctic Islands were covered in mixed conifer-hardwood forests similar to those of southern Maritime Canada and New England today. Such forests and associated wildlife would have easily tolerated seasonal sea ice, but they would not have survived the harshness of perennial ice cover on the adjacent ocean (Whitlock and Dawson, 1990).

A large unconformity (a surface in a sequence of sediments that represents missing deposits, and thus missing time) in the ACEX record prevents us from characterizing sea-ice conditions between about 44–18 Ma (Backman et al., 2008). Sediments overlying the unconformity contain little ice-rafted debris, and they indicate a smaller volume of sea ice in the Arctic Ocean at that time (St. John, 2008). Marked changes in Arctic climate in the middle Miocene were concurrent with global cooling and the onset of Antarctic reglaciation (Figure 3.8 in Chapter 3). These changes may have been promoted by the opening of the *Fram Strait* between the Eurasian and Greenland margins about 17 Ma, which allowed the modern circulation system in the Arctic Ocean to develop (Jakobsson et al., 2007). Resultant cooling led to a change from pine-redwood-dominated to larch-spruce-dominated floodplains and swamps at the Arctic periphery at about16 Ma as recorded, for example, on *Banks Island* by extensive peats with stumps in growth position (Fyles et al., 1994; Williams, 2006). A combination of cooling and increased moisture from the North Atlantic caused ice masses on and around *Svalbard* to grow and icebergs to discharge into the eastern Arctic Ocean and the *Greenland Sea* at about 15 Ma (Knies and Gaina, 2008). The source of sediment in the central Arctic Ocean changed between 13–14 Ma and indicates the likelihood that sea ice was now perennial (Krylov et al., 2008), although the ice's geographic distribution and persistence is not yet understood. Evidence of perennial ice can be found in even older sediments, starting from at least 14 Ma (Darby, 2008). Several pulses of more-abundant-than-normal ice-rafted debris in the late Miocene ACEX record indicate further growth of sea ice (St. John, 2008). This interpretation is consistent a cooling climate indicated by the spread of pine-dominated forests in northern *Alaska* (White et al., 1997). On the other hand, paleobotanical evidence also suggests that throughout the late Miocene and most of the Pliocene in at least some intervals perennial ice was severely restricted or absent. Thus, extensive braided-river deposits of the Beaufort Formation (early to middle Pliocene, about 5.3–3 Ma) that blanket much of the western Canadian Arctic Islands enclose abundant logs and other woody detritus representing more than 100 vascular plants such as pine (2 and 5 needles) and birch, and dominated at some locations by spruce and larch (Fyles, 1990; Devaney, 1991). Although these floral remains indicate overall boreal conditions cooler than in the Miocene, extensive perennial sea ice is not likely to have existed in the adjacent *Beaufort Sea* during this time. This inference is consistent with the presence of the bivalve Icelandic Cyprine (*Arctica islandica)* in marine sediments capping the Beaufort Formation on *Meighen Island* at 80°N and dated to the peak of Pliocene warming, about 3.2 Ma (Fyles et al., 1991). Foraminifers in Pliocene deposits in the Beaufort-Mackenzie area are also characteristic of boreal but not yet high-Arctic waters (McNeil, 1990), whereas the only known pre-Quaternary foraminiferal evidence from the central Arctic Ocean indicates seasonally ice-free conditions in the early Pliocene about 700 km north of the Alaskan coast (Mullen and McNeil, 1995).

Cooling in the late Pliocene profoundly reorganized the Arctic system: tree line retreated from the Arctic coasts (White et al., 1997; Matthews and Telka, 1997), permafrost formed (Sher et al., 1979; Brigham-Grette and Carter, 1992), and continental ice masses grew around the Arctic Ocean—for example, the *Svalbard* ice sheet advanced onto the outer shelf (Knies et al., 2002) and between 2.9–2.6 Ma ice sheets began to grow in North America (Duk-Rodkin et al., 2004). The ACEX cores record especially large volumes of high ice-rafted debris in the Arctic Ocean around 2 Ma (St. John, 2008). Despite the overall cooling, extensive warm intervals during the late Pliocene and the initial stages of the Quaternary (about 2.4–3 Ma) are repeatedly documented at the Arctic periphery from northwest Alaska to northeastern Greenland (Feyling-Hanssen et al., 1983; Funder et al., 1985, 2001; Carter et al., 1986; Bennike and Böcher, 1990; Kaufman, 1991; Brigham-Grette and Carter, 1992). For example, beetle and plant macrofossils in the nearshore high-energy sediments of the upper *Kap København* Formation on northeast Greenland, dated about 2.4 Ma, mimic paleoenvironmental conditions similar to those of southern Labrador today (Funder et al., 1985; 2001; Bennike and Böcher, 1990). At the same time, marine conditions were distinctly Arctic but, analogous with present-day faunas along the Russian coast, open water must have existed for 2 or 3 months in the summer. These results imply that summer sea ice in the entire Arctic Ocean was probably much reduced.

A more complete history of perennial versus seasonal sea ice and ice-free intervals during the past several million years requires additional sedimentary records distributed throughout the Arctic Ocean and a synthesis of sediment and paleobiological evidence from both land and sea. This history will provide new clues about the stability of the Arctic sea ice and about the sensitivity of the Arctic Ocean to changing temperatures and other climatic features such as snow and vegetation cover.

7.4.2 Quaternary Variations (the past 2.6 Ma)

The Quaternary period of Earth's history during the past 2.6 million years (m.y.) or so is characterized by overall low temperatures and especially large swings in climate regime (Figure 3.9 in Chapter 3). These swings are related to changes in insolation (incoming solar radiation) modulated by Earth's orbital parameters with periodicities of tens to hundreds of thousand years (see Chapter 3 for more detail). During cold periods when large ice masses are formed, such as during the Quaternary, these variations are amplified by powerful feedbacks due to changes in the albedo (reflectivity) of Earth's surface and concentration of greenhouse gases in the atmosphere. Quaternary climate history is composed of cold intervals (glacials) when very large ice sheets formed in northern Eurasia and North America and of interspersed warm intervals (interglacials), such as the present one, referred to as the Holocene (which began about 11.5 thousand years ago (ka). Temperatures at Earth's surface during some interglacials were similar to or even somewhat warmer than those of today; therefore, climatic conditions during those times can be used as approximate analogs for the conditions predicted by climate models for the 21st century (Otto-Bliesner et al., 2006; Goosse et al., 2007). One of the biggest questions in this respect is to what degree sea-ice cover was reduced in the Arctic during those warm intervals. This issue is insufficiently understood because interglacial deposits at the Arctic margins are exposed only in fragments (CAPE, 2006) and because sedimentary records from the Arctic Ocean generally have only low resolution. Even the age assigned to sediments that appear to be interglacial is commonly problematic because of the poor preservation of fossils and various stratigraphic complications (e.g., Backman et al., 2004). A better understanding has begun to emerge from recent collections of sediment cores from strategic sites drilled in the Arctic Ocean such as ACEX (Backman et al., 2006) and HOTRAX (Darby et al., 2005). The severity of ice conditions (widespread, thick, perennial ice) during glacial stages is indicated by of the extreme rarity of biological remains in cool-climate sediment layers and possible non-deposition intervals due to especially solid ice (Polyak et al., 2004; Darby et al., 2006; Cronin et al., 2008). In contrast, interglacials are characterized by higher marine productivity that indicates reduced ice cover. In particular, planktonic foraminifers typical of subpolar, seasonally open water lived in the area north of Greenland during the last interglacial (marine isotope stage 5e), 120–130 ka (Figure 7.7, Nørgaard-Pedersen et al., 2007a,b). Given that this area is presently characterized by especially thick and widespread ice, most of the Arctic Ocean may have been free of summer ice cover in the interval between 120–130 ka. Investigators need to carefully examine correlative sediments throughout the Arctic Ocean to determine how widespread were these low-ice or possibly ice-free conditions. Some intervals in sediment cores from various sites in the central Arctic have been reported to contain subpolar microfauna (e.g., Herman, 1974; Clark et al., 1990), but their age was not well constrained. New sediment core studies are needed to place these intervals in the coherent stratigraphic context and to reconstruct corresponding ancient ice conditions. This task is especially important as only those records from the central Arctic Ocean can provide direct evidence for ocean-wide ice-free water.

FIGURE 7.7 NEAR HERE

Some coastal exposures of interglacial deposits such as marine isotope stage 11 (about 400 ka) and 5e (about 120–130 ka) also indicate water temperatures warmer than present and, thus, reduced ice. For example, deposits of the last interglacial on the Alaskan coast of the

Chukchi Sea (the so-called Pelukian transgression) contain some fossils of species that are limited today to the northwest Pacific, whereas inter-tidal snails found near *Nome*, just slightly south of the *Bering Strait*, suggest that the coast here may have been annually ice free (Brigham-Grette and Hopkins, 1995; Brigham-Grette et al., 2001). On the Russian side of the *Bering Strait*, formaninifer assemblages suggest that coastal waters were fairly warm, like those in the Sea of Okhotsk and Sea of Japan (Brigham-Grette et al., 2001). Deposits of the same age along the northern Arctic coastal plain show that at least eight mollusk species extended their distribution ranges well into the *Beaufort Sea* (Brigham-Grette and Hopkins, 1995). Deposits near *Barrow* include at least one mollusk and several ostracode species known now only from the North Atlantic. Taken together, these findings suggest that during the peak of the last interglacial, about 120–130 ka, the winter limit of sea ice did not extend south of the Bering Strait and was probably located at least 800 km north of historical limits (such as on Figure 7.1), whereas summer sea-surface temperatures were warmer than present through the *Bering Strait* and into the *Beaufort Sea*.

7.4.3 The Holocene (the most recent 11.5 ka)

The present interglacial that has lasted approximately 11.5 k.y. is characterized by much more paleoceanographic data than earlier warm periods, because Holocene deposits are ubiquitous on continental shelves and along many coastlines. Owing to relatively high sedimentation rates at continental margins, ice drift patterns can be constructed on sub-millennial scales from some sedimentary records. Thus, the periodic influx of large numbers of iron oxide grains from specific sources, as into the Siberian margin-to-sea-floor area north of Alaska, has been linked to a certain mode of the atmospheric circulation pattern (Darby and Bischof, 2004).

If this link is proven, it will signify the existence of longer term atmospheric cycles in the Arctic than the decadal Arctic Oscillation observed during the last century (Thompson and Wallace, 1998).

Many proxy records indicate that early Holocene temperatures were warmer than today and that the Arctic contained less ice. This climate is consistent with a higher intensity of insolation that peaked about 11 ka owing to Earth's orbital variations. Evidence of warmer temperatures appears in many paleoclimatic records from the high Arctic—*Svalbard* and northern *Greenland*, northwestern North America, and eastern *Siberia* (Kaufman et al., 2004; Blake, 2006; Fisher et al., 2006; Funder and Kjær, 2007). Decreased sea-ice cover in the western Arctic during the early Holocene has also been inferred from high sodium concentrations in the *Penny Ice Cap* of *Baffin Island* (Fisher et al., 1998) and the *Greenland Ice Sheet* (Mayewski et al., 1994), although the implications of salt concentration is yet to be defined. Areas that were affected by the extended melting of the *Laurentide Ice Sheet*, especially the northeastern sites in North America and the adjacent North Atlantic, show more complex patterns of temperature and ice distribution (Kaufman et al., 2004).

An extensive record has been compiled from bowhead whale findings along the coasts of the *Canadian Arctic Archipelago* straits (Dyke et al., 1996, 1999; Fisher et al., 2006). Understanding the dynamics of ice conditions in this region is especially important for modern-day considerations because ice-free, navigable straits through the *Canadian Arctic Archipelago* will provide new opportunities for shipping lanes. The current set of radiocarbon dates on bowheads from the *Canadian Arctic Archipelago* coasts is grouped into three regions: western, central, and eastern (Figure 7.8). The central region today is the area of normally persistent summer sea ice; the western region is within the summer range of the Pacific bowhead; the eastern region is within the summer range of the Atlantic bowhead. These three graphs allow us to draw the following conclusions:

1. Bowhead bones have been most commonly found in all three regions in early Holocene (10–8 ka) deposits. At that time Pacific and Atlantic bowheads were able to intermingle freely along the length of the Northwest Passage indicating at least periodically ice-free summers.
2. Following an interval (8–5 ka) containing fewer bones, abundant bowhead bones have been found in deposits in the eastern channels during the middle Holocene (5–3 ka). At times, the Atlantic bowheads penetrated the central region, particularly 4.5–4.2 ka. The Pacific bowhead apparently did not extend its range at this time.
3. A final peak of bowhead bones has been found about 1.5–0.75 ka in all three regions, suggesting an open Northwest Passage during at least some summers. During this interval the bowhead-hunting Thule Inuit (Eskimo) expanded eastward out of the *Bering Sea* region and ultimately spread to *Greenland* and Labrador.
4. The decline of bowhead abundances during the last few centuries is evident in all three graphs. Thule bowhead hunters abandoned the high Arctic of *Canada* and *Greenland* during the Little Ice Age cooling (around 13th to 19th centuries) and *Thule* living in more southern Arctic regions increasingly focused on alternate resources.

FIGURE 7.8 NEAR HERE

On the basis of the summer ice melt record of the *Agassiz Ice Cap* (Fisher et al., 2006), summer temperatures that accompanied the early Holocene bowhead maximum are estimated at about 3°C above mid-20th century conditions, when July mean daily temperatures along the central Northwest Passage were about 5°C. Unless other processes, such as a different ocean circulation pattern, were also forcing greater summer sea-ice clearance in the early Holocene, the value of 3°C is an upper bound on the amount of warming necessary to clear the Northwest Passage region of summer sea ice. At times during the middle and late Holocene (especially 4.5–4.2 ka) the threshold condition was approached and, at least briefly, met, as indicated by Atlantic bowhead penetrating the central channels. The threshold condition for clearance of ice from the Northwest Passage was crossed in summer 2007. Whether this will be a regular event and what the consequences might be for Pacific-Atlantic exchanges of biota remains to be seen.

The bowhead record can be compared with the distribution of driftwood. Dated driftwood from raised marine beaches along the Arctic coasts of North America, notably around the margins of *Baffin Bay* (Blake, 1975), has been used to infer changes in the transport of sea ice from the Arctic Basin (Dyke et al., 1997) (Figure 7.9). The ratio of larch (mainly from Russia) to spruce (mainly from northwest Canada) driftwood declines sharply about 7 ka. This abrupt shift might have been caused by the intensity of ice drift from the Arctic Ocean or changes in its trajectories (Tremblay et al., 1997), or it might reflect changes in the composition or extent of forests. The delivery of driftwood, which probably was borne on the East Greenland Current, peaked during the middle Holocene, possibly in conjunction with less ice cover in the Arctic Ocean.

FIGURE 7.9 NEAR HERE

Levac et al. (2001) estimated the duration of sea-ice cover during the Holocene in northern *Baffin Bay* (southern reach of *Nares Strait* between *Ellesmere Island* and northwest *Greenland*) based on transfer functions of dinocyst assemblages. The present-day duration of the ice cover in this area is about 8 months, whereas the predicted duration for the Holocene ranges between 7 and 10–12 months. An interval of minimal sea-ice cover existed until about 4.5 ka, whereas afterwards the sea-ice cover was considerably more extensive (Figure 7.10).

FIGURE 7.10 NEAR HERE

Along the North Greenland coasts, isostatically raised staircases of wave-generated beach ridges (Figure 7.11) document seasonally open water (Funder and Kjær, 2007). Large numbers of striated boulders in and on the marine sediments also indicate that the ocean was open enough for icebergs to drift along the shore and drop their loads. Presently the North Greenland coastline is permanently surrounded by pack ice, and rare icebergs are locked up in sea ice. Radiocarbon-dated mollusk shells from beach ridges show that the beach ridges were formed in the early Holocene, within the interval from about 8.5–6 ka, which is progressively shorter from south to north. These wave-generated shores and abundant iceberg-deposited boulders indicate the possibility that the adjacent Arctic Ocean was free of sea ice in summer at this time.

A somewhat different history of ice extent in the Holocene emerges from the northern North Atlantic and *Nordic seas*, exemplified by the Iceland margin. A 12,000 year record of quartz content in shelf sediment, which is used in this area as a proxy for the presence of drift ice (Eiriksson et al., 2000), has been produced for a core (MD99-2269) from the northern Iceland shelf. The record has a resolution of 30 years per sample (Moros et al., 2006); these results are consistent with data obtained from 16 cores across the northwestern Iceland shelf (Andrews, 2007). These data show a minimum in quartz and, thus, ice cover at the end of deglaciation, whereas the early Holocene area of ice increased and then reached another minimum around 6 ka, after which the content of quartz steadily rose (Figure 7.12). The lagged Holocene optimum in the North Atlantic in comparison with high Arctic records can be explained by the nature of oceanic controls on ice distribution. In particular, the discharge of glacial meltwater from the remains of the Laurentide Ice Sheet slowed the warming in the North Atlantic region in the early Holocene (Kaufman et al., 2004). Additionally, oceanic circulation seesawed between the eastern and western regions of the *Nordic seas* throughout much of the Holocene. For example, in the Norwegian Sea the Holocene ice-rafting peaked in the mid-Holocene, 6.5–3.7 ka (Risebrobakken et al., 2003), and changes in Earth's orbit forced decreasing summer temperatures and decreased seasonality (Moros et al., 2004). By contrast, the middle Holocene is a relatively warm period off East Greenland, and it received a strong subsurface current of Atlantic Water around 6.5–4 ka, while ice-rafted debris was low (Jennings et al., 2002). These patterns are consistent with modern marine and atmospheric temperatures that commonly change in opposite directions on the eastern and western side of the North Atlantic ("seesaw effect" of van Loon and Rogers, 1978).

FIGURE 7.12 NEAR HERE

The Neoglacial cooling of the last few thousand years is considered overall to be related to decreasing summer insolation (Koç and Jansen, 1994). However, high-resolution climate records reveal greater complexity in the system—changes in seasonality and links with conditions in low latitudes and southern high latitudes (e.g., Moros et al., 2004). Variations in the volumes of ice-rafted debris indicate several cooling and warming intervals during Neoglacial time, similar to the so-called “Little Ice Age” and “Medieval Climate Anomaly” cycles of greater and lesser areas of sea ice (Jennings and Weiner, 1996; Jennings et al., 2002; Moros et al., 2006; Bond et al., 1997). Polar Water excursions have been reconstructed as multi-century to decadal-scale variations superimposed on the Neoglacial cooling at several sites in the subarctic North Atlantic (Andersen et al., 2004; Giraudeau et al., 2004; Jennings et al., 2002). In contrast, a decrease in drift ice during the Neoglacial is documented for areas influenced by the North Atlantic Current, possibly indicating a warming in the eastern *Nordic Seas* (Moros et al., 2006). A seesaw climate pattern has been evident between seas adjacent to West Greenland and Europe. For instance, warm periods in Europe around 800–100 BC and 800–1300 AD (Roman and the Medieval Climate Anomalies) were cold periods on West Greenland because little warm Atlantic Water fed into the West Greenland Current. Moreover, a cooling interval in western Europe (during the Dark Ages) correlated with increased meltwater —and thus warming—on West Greenland (Seidenkrantz et al., 2007).

Bond et al. (1997, 2001) suggested that cool periods manifested as past expansions of drift ice and ice-rafted debris (most notably, hematite-stained quartz grains) in the North Atlantic punctuated deglacial and Holocene records at intervals of about 1500 years and that these drift ice events were a result of climates that cycled independently of glacial influence. Bond et al. (2001) concluded that peak volumes of Holocene drift ice resulted from southward expansions of polar waters that correlated with times of reduced solar output. This conclusion suggests that variations in the Sun’s output is linked to centennial- to millennial-scale variations in Holocene climate through effects on production of North Atlantic Deep Water. However, continued investigation of the drift ice signal indicates that although the variations reported by Bond et al.

(2001) may record a solar influence on climate, they likely do not pertain to a simple index of drift ice (Andrews et al., 2006). In addition, those cooling events prior to the Neoglacial interval may stem from deglacial meltwater forcing rather than from southward drift of Arctic ice (Giraudeau et al., 2004; Jennings et al., 2002). In an effort to test the idea of solar forcing of 1500 year cycles in Holocene climate change, Turney et al. (2005) compared Irish tree-ring-derived chronologies and radiocarbon activity, a proxy for solar activity, with the Holocene drift-ice sequence of Bond et al. (2001). They found a dominant 800-year cycle in moisture, reflecting atmospheric circulation changes during the Holocene but no link with solar activity.

Despite many records from the Arctic margins indicating considerably reduced ice covering the early Holocene, no evidence of the decline of perennial ice cover has been found in sediment cores from the central Arctic Ocean. Arctic Ocean sediments contain some ice-rafted debris interpreted to arrive from distant shelves requiring more than 1 year of ice drift (Darby and Bischof, 2004). One explanation is that the true record of low-ice conditions has not yet been found because of low sedimentation rates and stratigraphic uncertainties. Additional investigation of cores by use of many proxies with highest possible resolution is needed to verify the distribution of ice in the Arctic during the warmest phase of the current interglacial.

7.4.4 Historical Period

Arctic paleoclimate records that contain proxies such as lake and marine sediments, trees, and ice cores indicate that from the mid-19th to late 20th century the Arctic warmed to the highest temperatures in at least four centuries (Overpeck et al., 1997). Subglacial material exposed by retreating glaciers in the *Canadian Arctic* indicates that modern temperatures are warmer than any time in at least the past 1600 years (Anderson et al., 2008). Paleoclimatic proxy records of the last two centuries agree well with hemispheric and global data (including instrumental measurements) (Mann et al., 1999; Jones et al., 2001). The composite record of ice conditions for Arctic ice margins since 1870 shows a steady retreat of seasonal ice since 1900; in addition, the retreat of both seasonal and annual ice has accelerated during the last 50 years (Figure 7.13) (Kinnard et al., 2008). The latter observations are the most reliable for the entire data set and are based on satellite imagery since 1972. The rate of ice-margin retreat over the most recent decades is spatially variable, but the overall trend in ice is down. The current decline of the Arctic sea-ice cover is much larger than expected from decadal-scale climatic and hydrographic variations (e.g., Polyakov et al., 2005; Steele et al., 2008). The recent warming and associated ice shrinkage are especially anomalous because orbitally driven insolation has been decreasing steadily since its maximum at 11 ka, and it is now near its minimum in the 21 k.y. precession cycle (e.g., Berger and Loutre, 2004), which should lead to cool summers and extensive sea ice.

FIGURE 7.13 NEAR HERE

7.5 Synopsis

Geological data indicate that the history of Arctic sea ice is closely linked with temperature changes. Sea ice in the Arctic Ocean may have appeared as early as 46 Ma, after the onset of a long-term climatic cooling related to a reorganization of the continents and subsequent formation of large ice sheets in polar areas. Year-round ice in the Arctic possibly developed as early as 13–14 Ma, in relation to a further overall cooling in climate and the establishment of the modern hydrographic circulation in the Arctic Ocean. Nevertheless, extended seasonally ice-free periods were likely until the onset of large-scale Quaternary glaciations in the Northern Hemisphere approximately 2.5 Ma. These glaciations were likely to have been accompanied by a fundamental increase in the extent and duration of sea ice. Ice may have been less prevalent during Quaternary interglacials, and the Arctic Ocean even may have been seasonally ice free during the warmest interglacials (owing to changes in insolation modulated by variations in Earth's orbit that operate on time scales of tens to hundred thousand years). Reduced-ice conditions are inferred, for example, for the previous interglacial and the onset of the current interglacial, about 130 and 10 ka. These low-ice periods can be used as ancient analogs for future conditions expected from the marked ongoing loss of Arctic ice cover. On time scales of thousands and hundreds of years, patterns of ice circulation vary somewhat; this feature is not yet well understood, but large periodic reductions in ice cover at these time scales are unlikely. Recent historical observations suggest that ice cover has consistently shrunk since the late 19th century, and that shrinkage has accelerated during the last several decades. Shrinkage that was both similarly large and rapid has not been documented over at least the last few thousand years, although the paleoclimatic record is sufficiently sparse that similar events might have been missed. The recent ice loss does not seem to be explainable by natural climatic and hydrographic variability on decadal time scales, and is remarkable for occurring when reduction in summer sunshine from orbital changes has caused sea-ice melting to be less likely than in the previous millennia since the end of the last ice age. The recent changes thus appear notably anomalous; improved reconstructions of sea-ice history would help clarify just how anomalous these changes are.

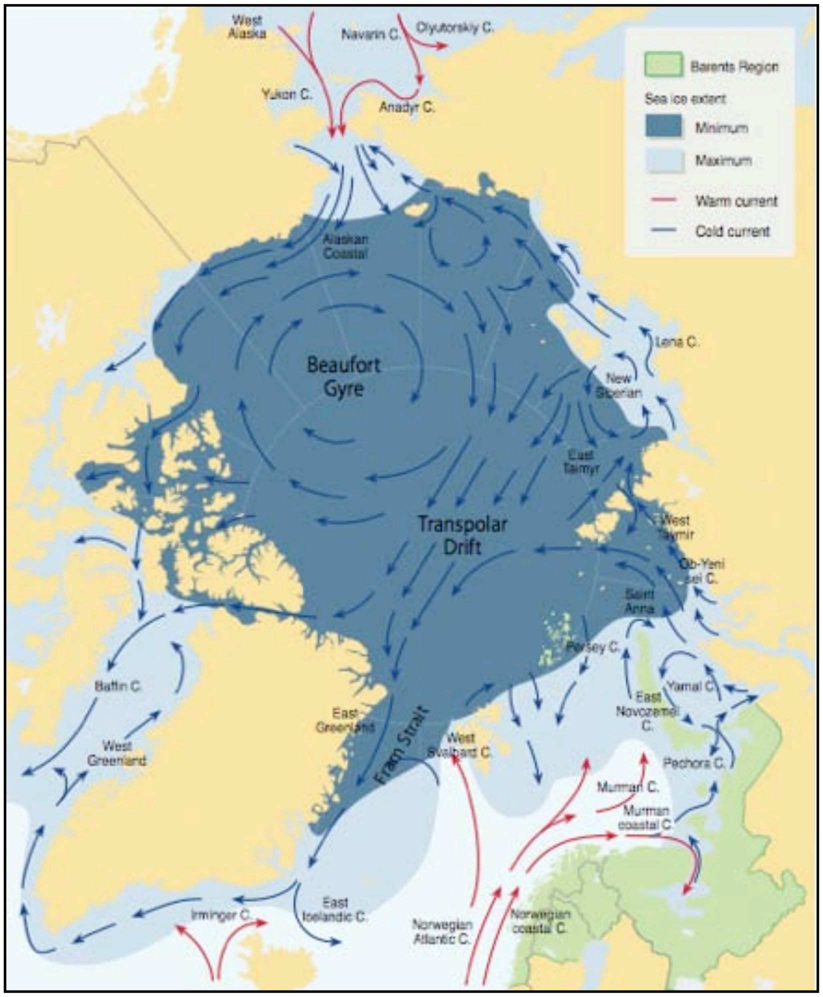


Figure 7.1. Northern ocean currents and extent of sea ice extent. UNEP/GRID-Arendal Maps and Graphics Library. Dec 97. UNEP/GRID-Arendal. 19 Feb 2008. Philippe Rekacewicz, UNEP/GRID-Arendal) http://maps.grida.no/go/graphic/ocean_currents_and_sea_ice_extent.

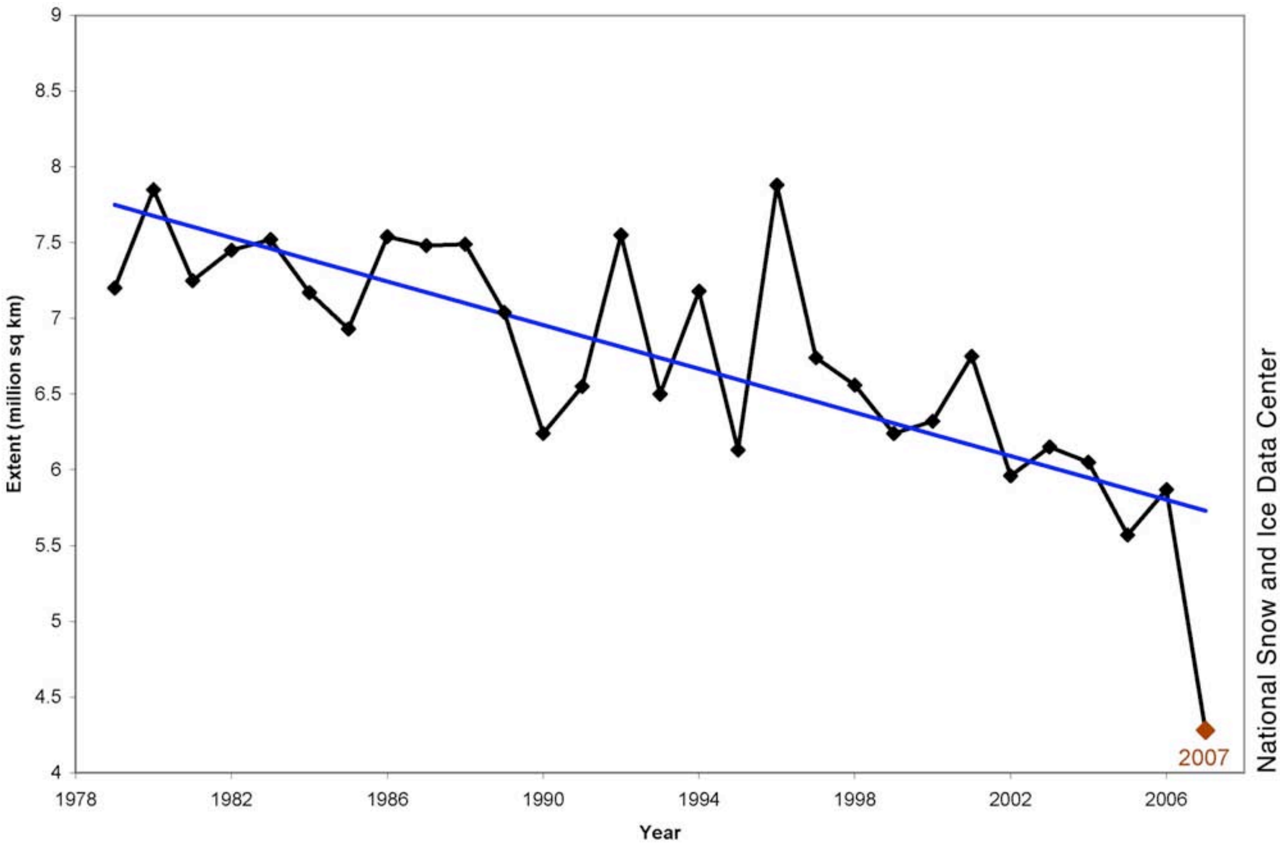


Figure 7.2. Extent of Arctic sea ice in September, 1979–2007. The linear trend (trend line shown in blue) including 2007 shows a decline of 10% per decade (courtesy National Snow and Ice Data Center, Boulder, Colorado).

Figure 7.3. Key marine sedimentary sequences exposed at the coasts of Arctic North America and Greenland.

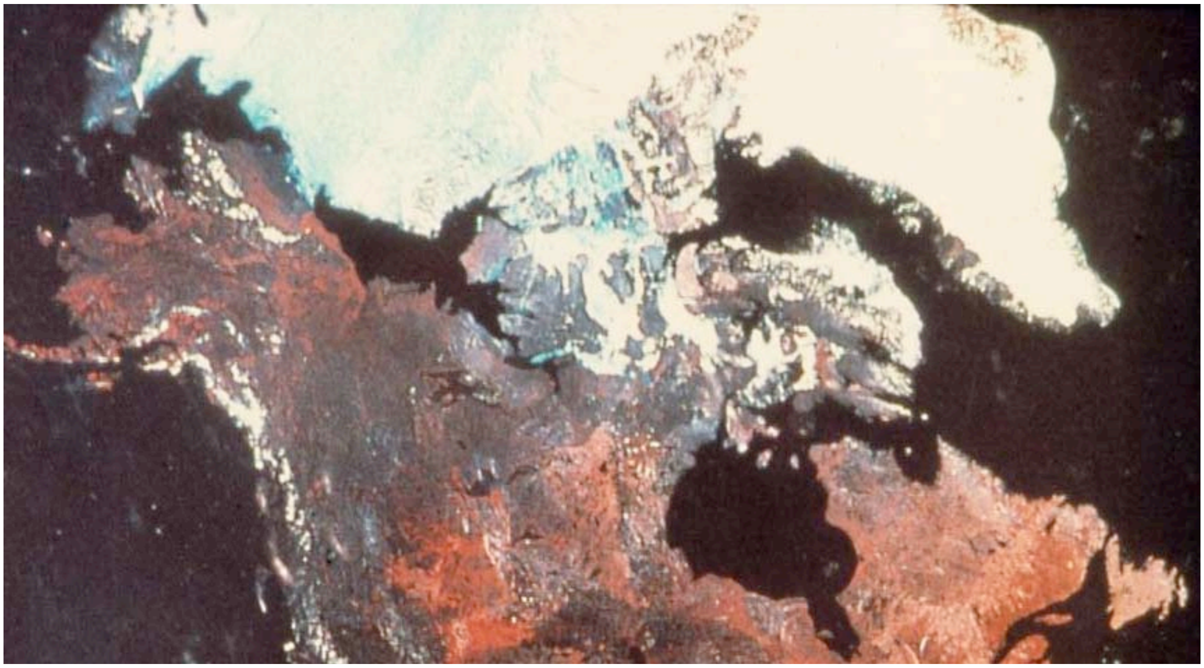


Figure 7.4. Typical late 20th century summer ice conditions in the Canadian Arctic Archipelago. (Dyke et al., 1996)

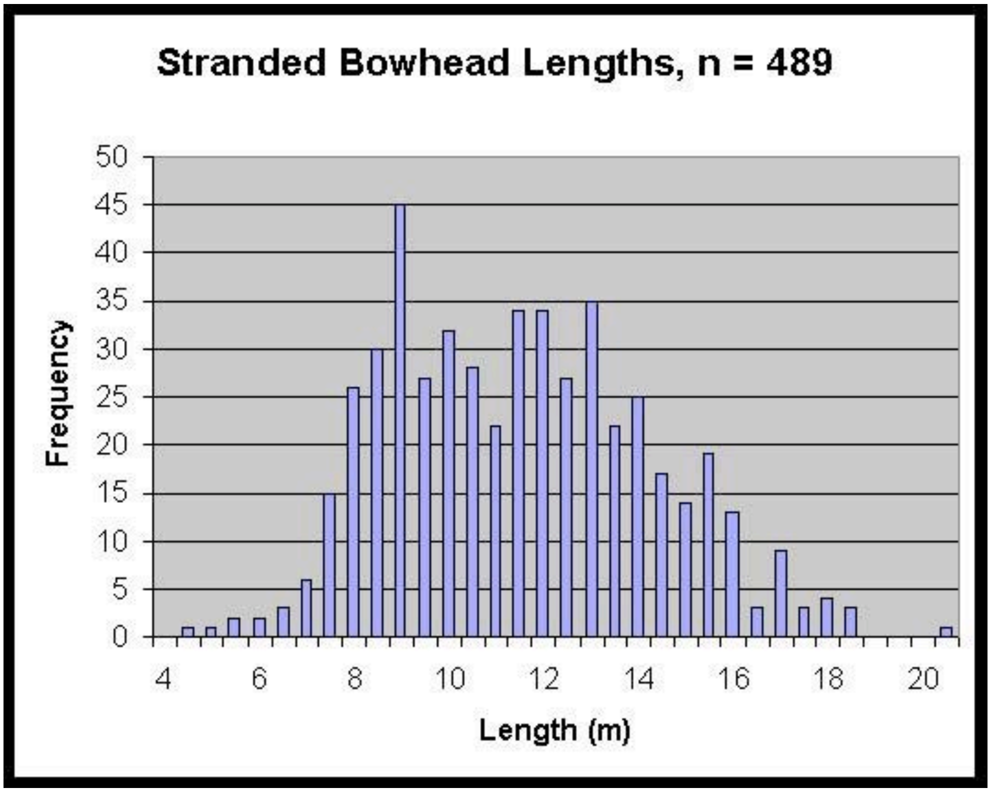


Figure 7.5. The reconstructed lengths of Holocene bowhead whales based on skull measurements (485 animals) and mandible measurements (an additional 4 animals) (Savelle, et al., 2000). This distribution is very similar to the lengths of living Pacific bowheads, indicating that past strandings affected all age classes.

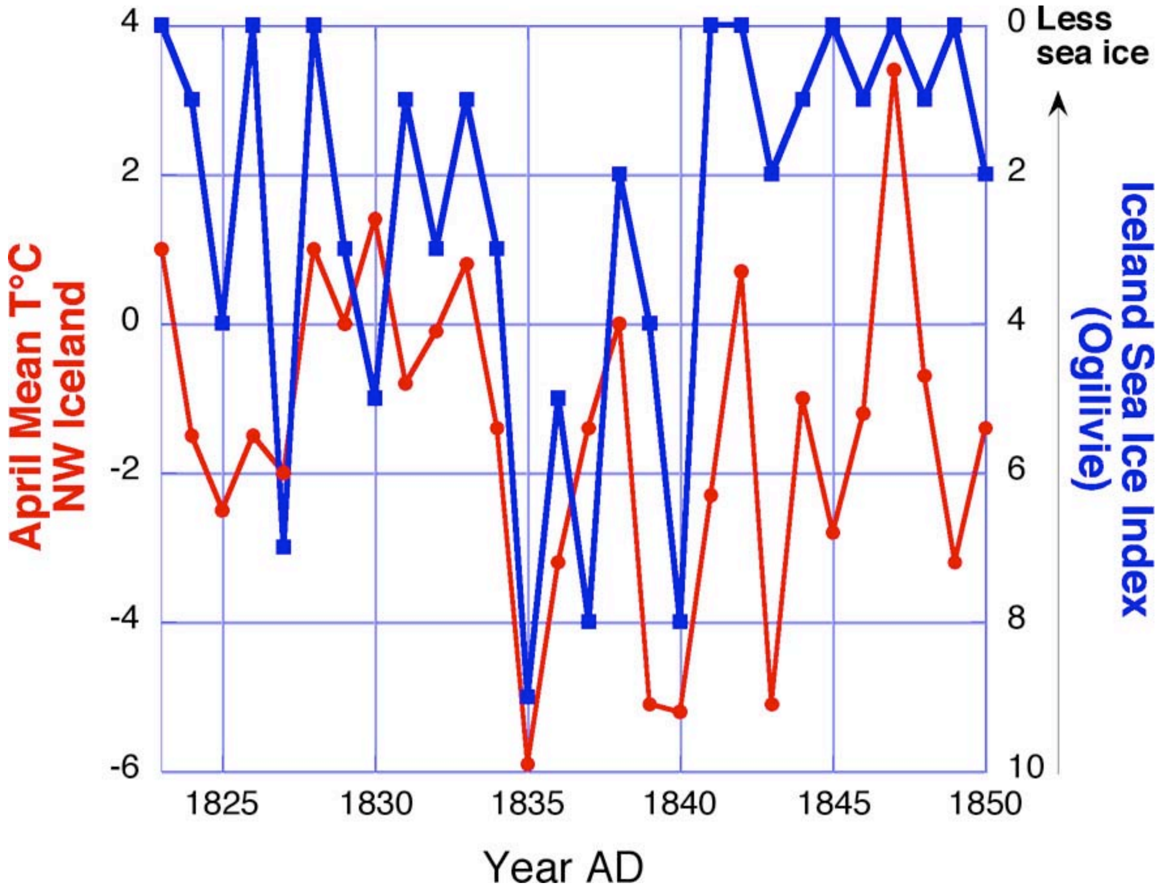


Figure 7.6. The sea-ice index on the Iceland shelf plotted against springtime air temperatures in northwest Iceland that are affected by the distribution of ice in this region (from Ogilvie, 1996). The two correlate well.

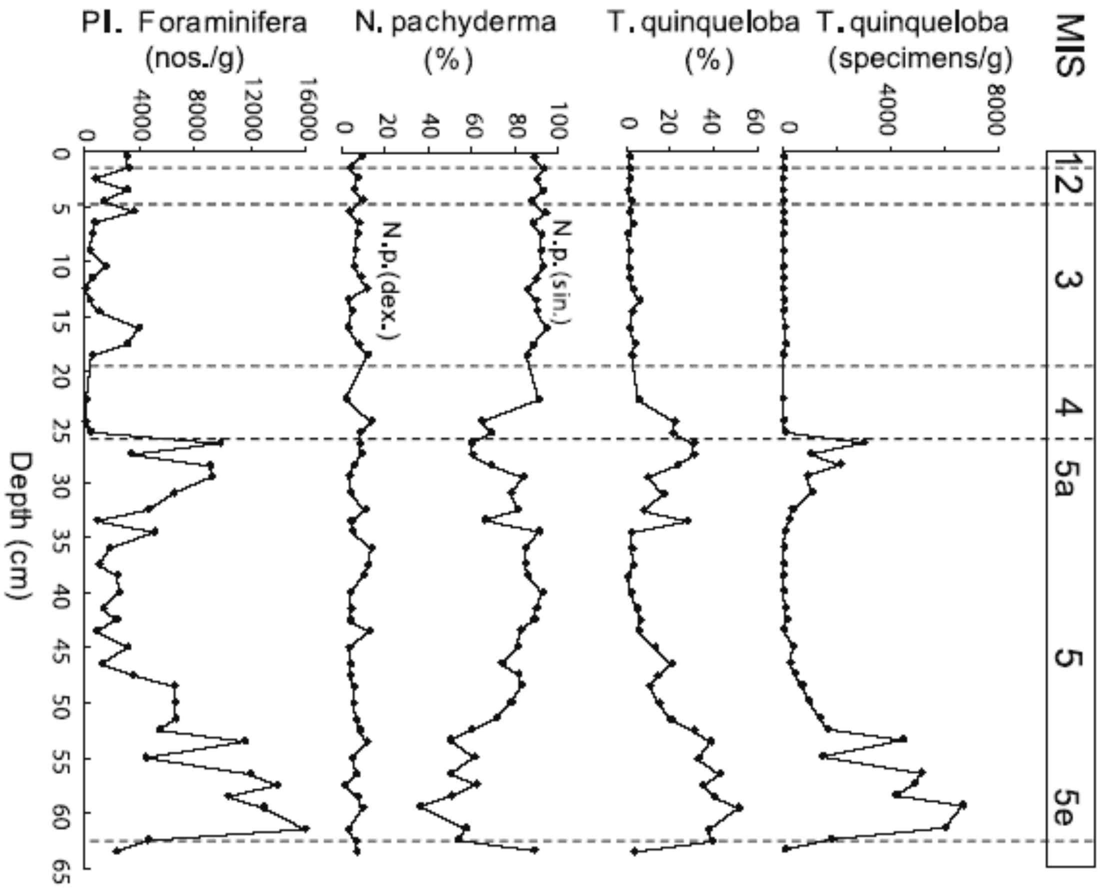


Figure 7.7. Planktonic foraminiferal record, core GreenICE-11, north of Greenland (from Nørgaard-Pedersen et al., 2007b). Note high numbers of a subpolar planktonic foraminifer *T. quinqueloba* during the last interglacial, marine isotopic stage (MIS) 5e; they indicate warm temperatures or reduced-ice conditions (or both) north of Greenland at that time.

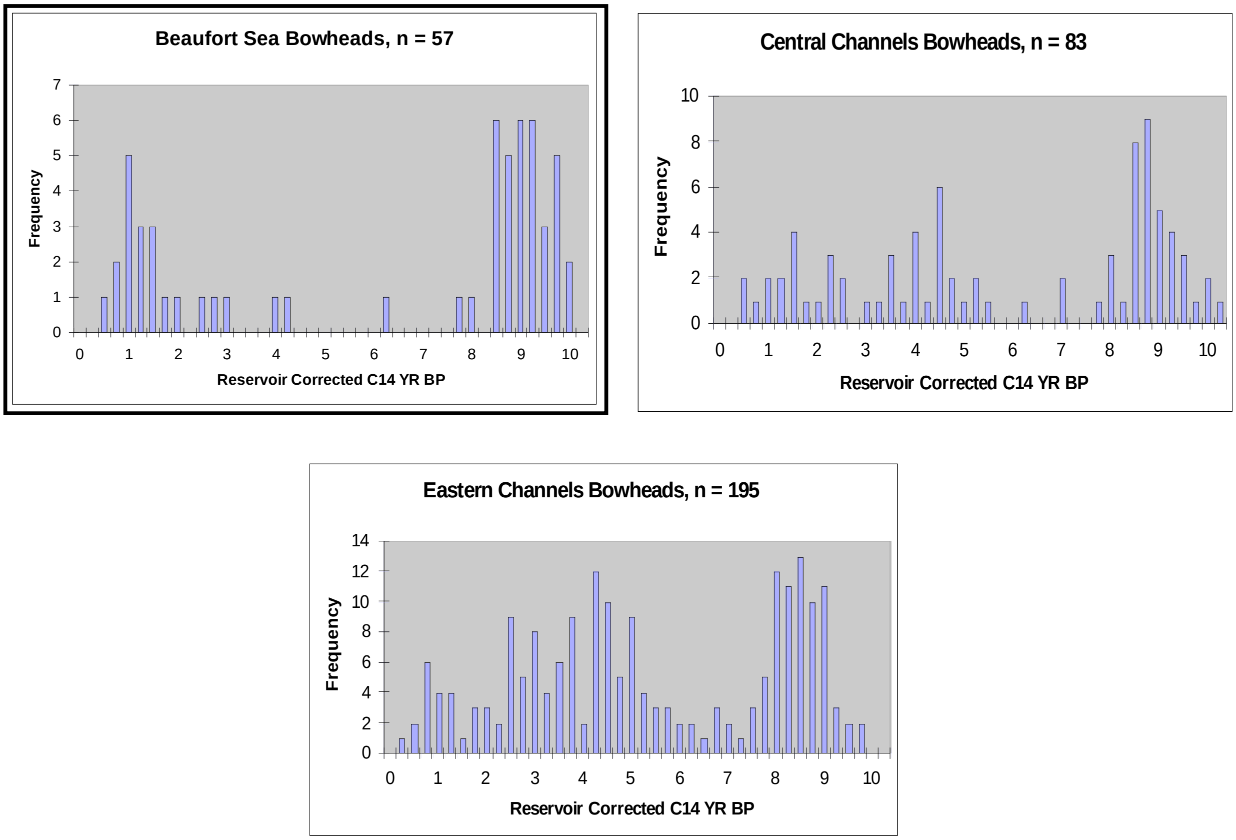
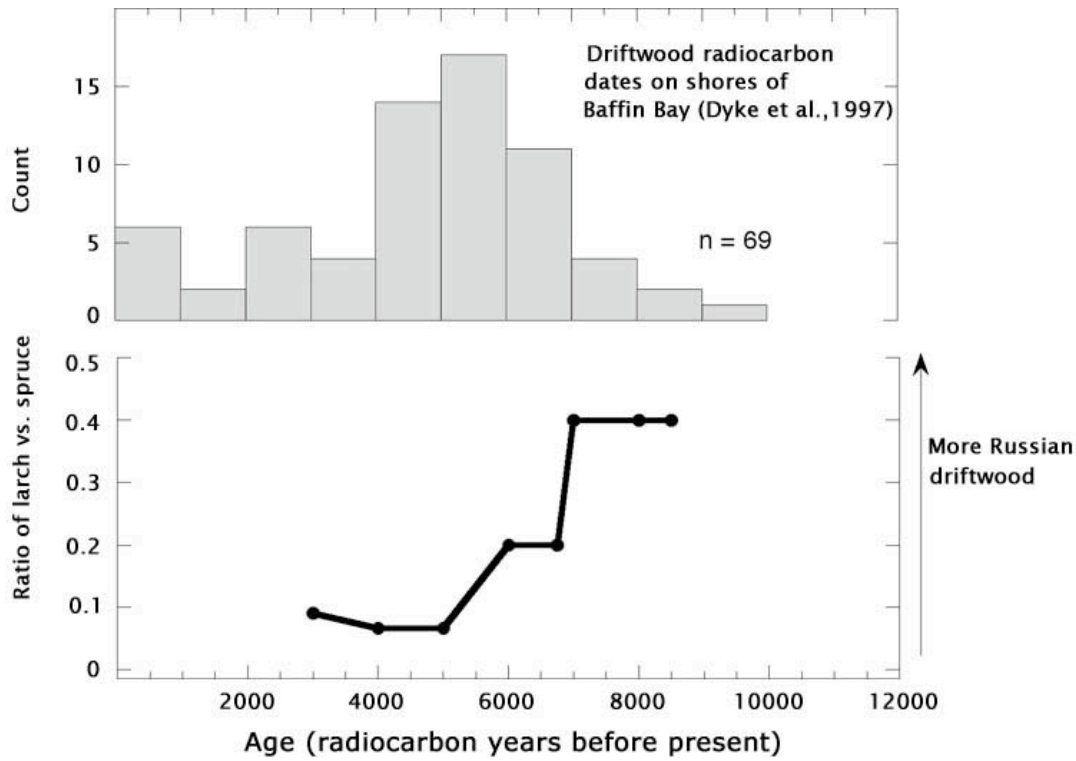


Figure 7.8. Distribution of radiocarbon ages (in thousands of years) of bowhead whales in three regions of the Canadian Arctic Archipelago (data from Dyke et al., 1996; Savelle et al., 2000).



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935 **Figure 7.9.** Distribution of radiocarbon ages of Holocene driftwood on the shores of Baffin Bay
936 (from Dyke et al., 1997).

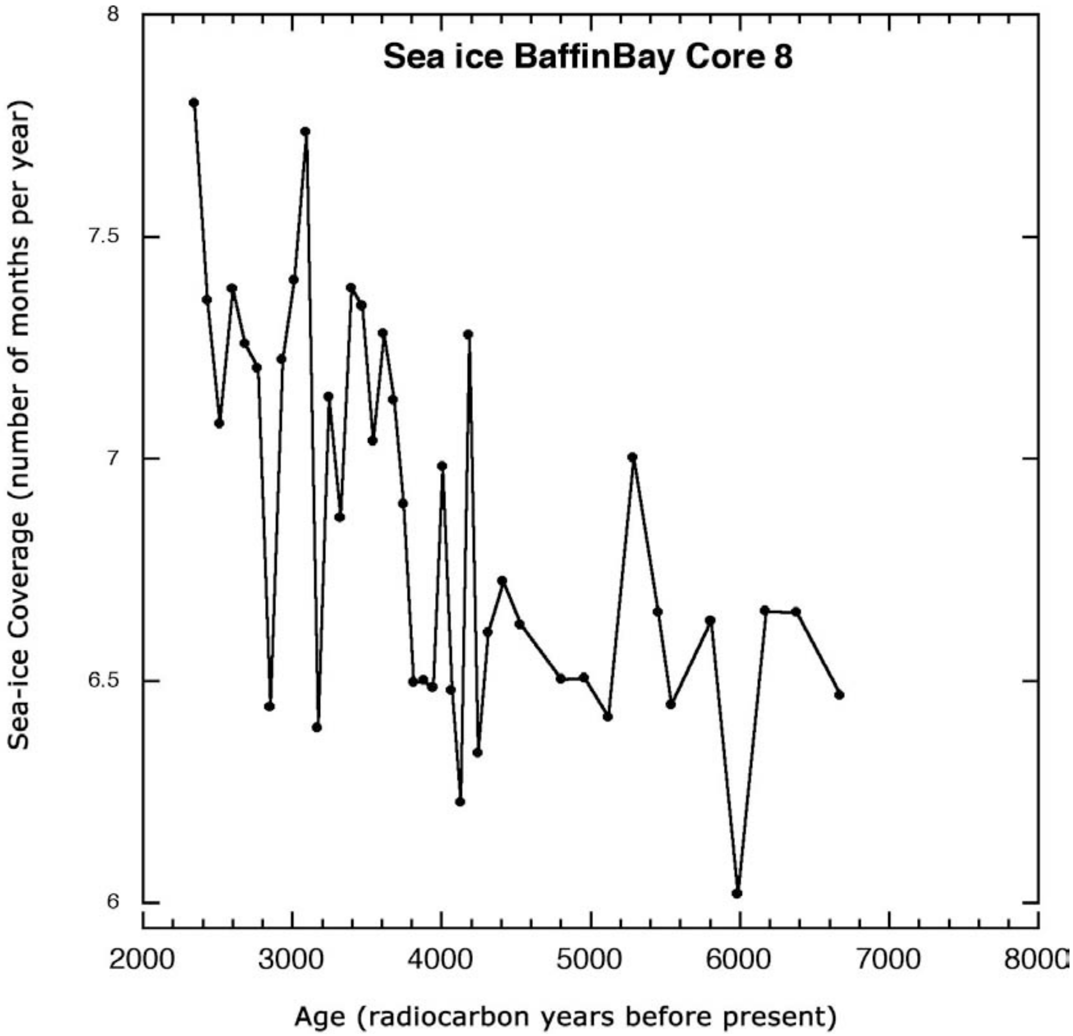

Figure 7.10. Reconstruction of the duration of ice cover (months per year) in northern Baffin Bay during the Holocene based on dinocyst assemblages (modified from Levac et al., 2001).

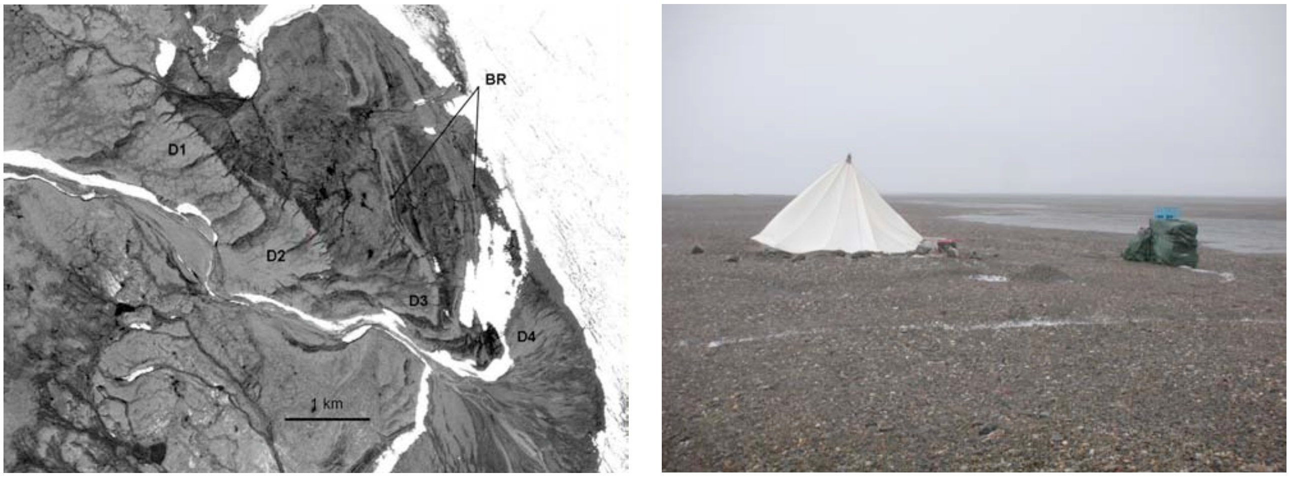

Figure 7.11. Aerial photo (left) of wave-generated beach ridges (BR) at *Kap Ole Chiewitz,* 83°25’N, northeast Greenland. D1-D4 are raised deltas. The oldest, D1, is dated to ~10 ka while D4 is the modern delta. Only D3 is associated with wave activity. The period of beach ridge formation is dated to ca. 8.5–6 ka. The photo on the right shows the upper beach ridge. (Funder, S. and K. Kjær, 2007)

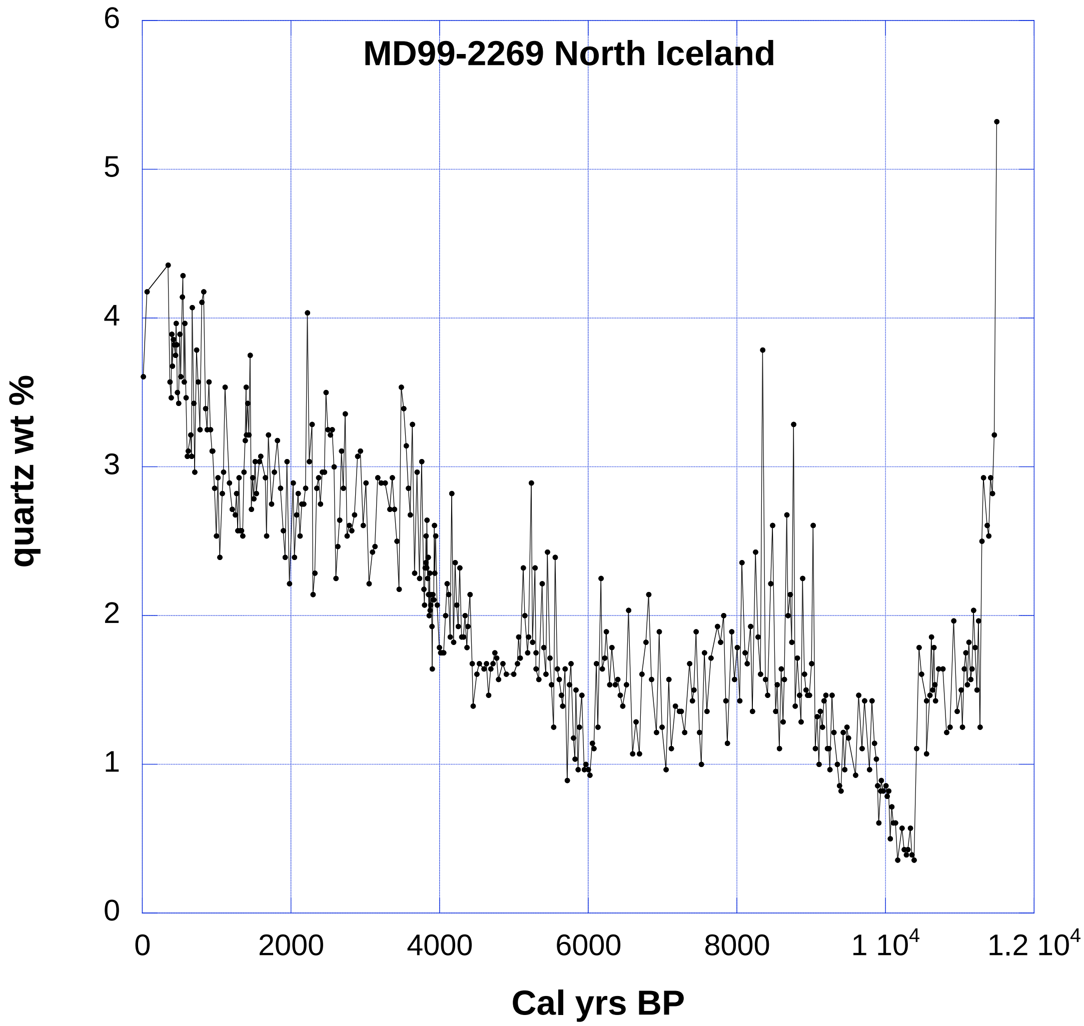

Figure 7.12. Variations in the percentage of quartz (a proxy for drift ice) in Holocene sediments from the northern Iceland shelf (from Moros et al., 2006). BP, before present.

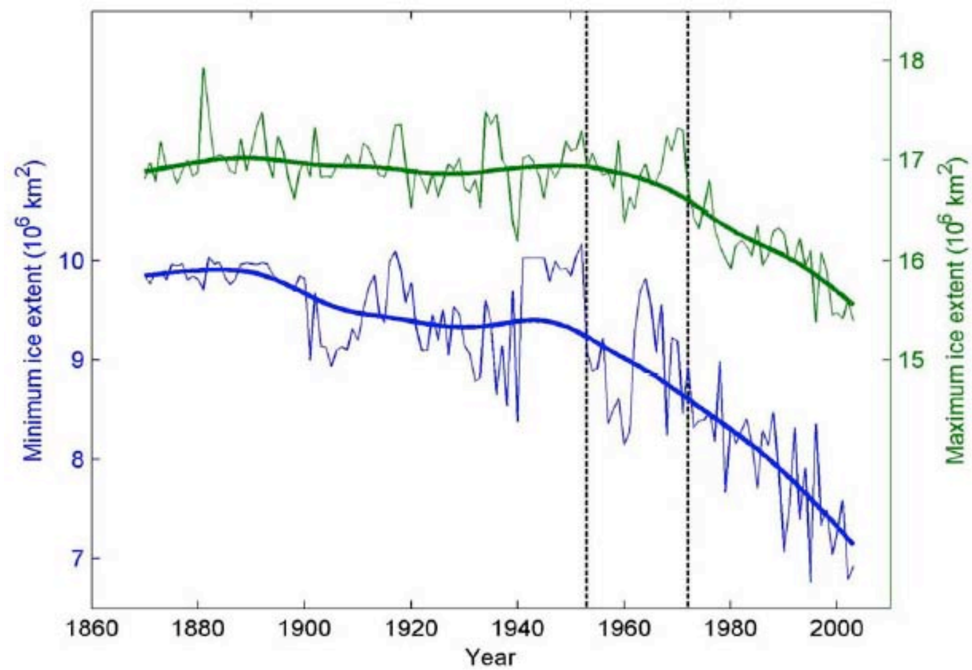

Figure 7.13. Total sea-ice extent time series, 1870–2003 (from Kinnard et al., 2008). Green lines: maximal extent. Blue lines: minimal extent. Thick lines are robust spline functions that highlight low-frequency changes. Vertical dotted lines separate the three periods for which data sources changed fundamentally: earliest, 1870–1952, observations of differing accuracy and availability; intermediate, 1953–1971, generally accurate hemispheric observations; most recent, 1972–2003, satellite period, best accuracy and coverage.

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