



## History of sea ice in the Arctic

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### ABSTRACT

Arctic sea-ice extent and volume are declining rapidly. Several studies project that the Arctic Ocean may become seasonally ice-free by the year 2040 or even earlier. Putting this into perspective requires information on the history of Arctic sea-ice conditions through the geologic past. This information can be provided by proxy records from the Arctic Ocean floor and from the surrounding coasts. Although existing records are far from complete, they indicate that sea ice became a feature of the Arctic by 47 Ma, following a pronounced decline in atmospheric  $p\text{CO}_2$  after the Paleocene–Eocene Thermal Optimum, and consistently covered at least part of the Arctic Ocean for no less than the last 13–14 million years. Ice was apparently most widespread during the last 2–3 million years, in accordance with Earth's overall cooler climate. Nevertheless, episodes of considerably reduced sea ice or even seasonally ice-free conditions occurred during warmer periods linked to orbital variations. The last low-ice event related to orbital forcing (high insolation) was in the early Holocene, after which the northern high latitudes cooled overall, with some superimposed shorter-term (multidecadal to millennial-scale) and lower-magnitude variability. The current reduction in Arctic ice cover started in the late 19th century, consistent with the rapidly warming climate, and became very pronounced over the last three decades. This ice loss appears to be unmatched over at least the last few thousand years and unexplainable by any of the known natural variabilities.

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### 1. Introduction

The most defining feature of the surface of the Arctic Ocean and adjacent seas is its ice cover (Fig. 1), which waxes and wanes with the seasons, and changes in extent and thickness on inter-annual and longer time scales. These changes, while driven by climate, themselves affect atmospheric and hydrographic

conditions in high latitudes on various time scales (e.g., Smith et al., 2003; Kinnard et al., 2008; Steele et al., 2008; Miller et al., 2010). Observations during the past several decades document substantial, accelerating retreat and thinning of the Arctic sea-ice cover (e.g., Comiso et al., 2008; Stroeve et al., 2008). Based on climate simulations, the Arctic Ocean may become seasonally ice-free as early as around 2040 (Holland et al., 2006a; Wang and Overland, 2009).

A reduction in sea-ice extent will promote strong Arctic warming through the ice-albedo feedback mechanism and may influence weather systems beyond the Arctic (e.g., Francis et al., 2009). Changes

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**Fig. 1.** Overview map of the Arctic and adjacent regions showing the extent of sea ice: magenta and blue lines – March and September medians for 1979–2000, respectively; white field – September 2007 extent (courtesy National Snow and Ice Data Center, Boulder, Colorado). Major circulation systems are schematically shown by green arrows. BG – Beaufort Gyre, TPD – Transpolar Drift, BS – Bering Strait, FS – Fram Strait.

in ice cover and freshwater flux out of the Arctic Ocean may also affect circulation in the North Atlantic, which has profound influence on climate in Europe and North America (Seager et al., 2002; Holland et al., 2006b). Continued ice retreat will accelerate coastal erosion owing to increased wave action (Jones et al., 2009) and will have cascading effects on the Arctic Ocean food web including top predators, such as polar bears and seals (e.g., Derocher et al., 2004; Durner et al., 2009). This will affect indigenous human populations that harvest ice-dependent species. Recent years have already witnessed an intrusion of exotic planktonic biota into the high Arctic (Hegseth and Sundfjord, 2008), while some Pacific plankters penetrated via the Arctic into the North Atlantic for the first time since ca 800 ka (Reid et al., 2007). Another consequence of reduced ice is enhanced marine access to the Arctic Ocean. While providing opportunities for commercial shipping and natural-resource exploitation, this raises a broad set of environmental concerns such as contamination and infringement on natural habitats (e.g., ACIA, 2005).

Interpreting observed recent changes and modeling future changes in the sea-ice cover require a longer-term perspective, from which we can better understand the Arctic's natural variability and response to external forcing over time. Although the targeted investigation of paleo-sea-ice conditions is a fairly new direction of paleoclimate research, a considerable amount of

relevant data appeared recently, and some of the paleo-records generated earlier can be re-evaluated from the sea-ice perspective. This paper provides an overview of the Arctic sea-ice history spanning multiple climate regimes, from the early Cenozoic to present.

## 2. Background on Arctic sea-ice cover

### 2.1. Ice extent, thickness, drift, and duration

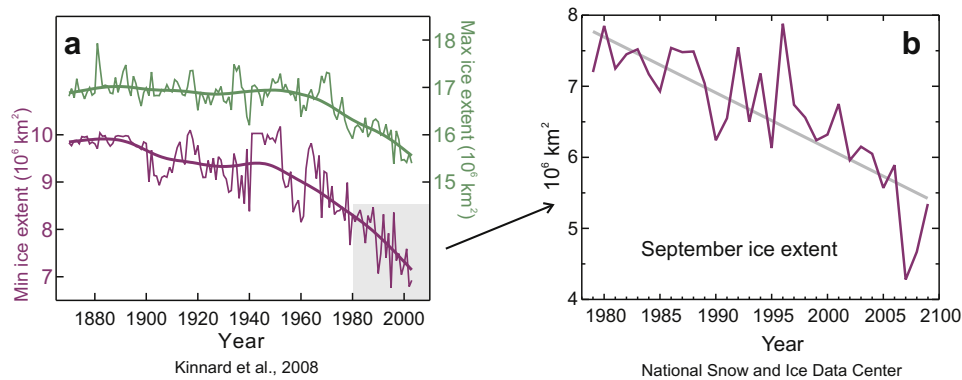
Arctic sea-ice cover attains its maximum seasonal extent in March and shrinks through spring and summer to a minimum extent in September. For the period of reliable satellite observations (1979 onwards), extremes in Northern Hemisphere ice extent, defined as the ocean region with at least 15% ice cover, are  $16.44 \times 10^6 \text{ km}^2$  for March 1979 and  $4.28 \times 10^6 \text{ km}^2$  for September 2007 (Fetterer and Knowles, 2002, updated; Stroeve et al., 2008). 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 (Fig. 1; Weeks and Ackley, 1986; Wadhams, 2000). 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) 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 has been typically cited at about 3 m (Williams et al., 1975; Wadhams, 1980), but there is growing evidence (discussed below) that shrinking ice extent over recent decades has been attended by substantial thinning. Although many different types of sea ice can be defined, the two basic categories are: (1) first-year ice, which represents a single year's growth, and (2) multi-year ice, which has survived one or more melt seasons (Weeks and Ackley, 1986). New ice forms during autumn in seasonally open water, mostly over continental shelves, and is then transported into the central Arctic basin, and can thicken through bottom growth. Undeformed first-year ice can reach as much as 1.5–2 m in thickness. Although multi-year ice is generally thicker, first-year ice that undergoes convergence and/or shear can produce ridges 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  $1\text{--}3 \text{ cm s}^{-1}$ , and the Transpolar Drift, the movement of ice from the coast of Siberia eastward across the pole and into the North Atlantic by way of Fram Strait (Fig. 1). Ice velocities in the Transpolar Drift increase toward Fram Strait, where the mean drift speed is  $5\text{--}20 \text{ cm s}^{-1}$  (Thorndike, 1986; Gow and Tucker, 1987). About 20% of the total ice area of the Arctic Ocean and nearly all of the annual ice export is discharged each year through Fram Strait, the majority of which is multi-year 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.

### 2.2. Recent changes and projections for the future

The composite historical record of Arctic ice margins shows a general retreat of seasonal ice since about 1900, and accelerated retreat of both seasonal and annual ice during the last five decades (Fig. 2a) (Kinnard et al., 2008). The most reliable observations are



**Fig. 2.** (a) Maximal (winter) and minimal (summer) Arctic sea-ice extent time series, 1870–2003 (from Kinnard et al., 2008). Smooth 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. Shaded area highlights the period of summer ice extent observations shown in 2b. (b) Extent of Arctic sea ice (15% concentration) in September, 1979–2009 (National Snow and Ice Data Center, Boulder, Colorado). The 30-yr linear trend shows a decline of 11% per decade. Note different methods of ice area estimate in (a) and (b) resulting in slightly higher values in (a) (details in Kinnard et al., 2008).

from 1979 onwards, corresponding to the modern satellite era. Patterns of ice-margin retreat may differ between different periods and regions of the Arctic, but the overall retreat trend is clearly larger than decadal-scale variability, consistent with observations and modeling of the 20th-century ice concentrations and water temperatures (Polyakov et al., 2005; Kauker et al., 2008; Steele et al., 2008). The severity of present ice loss can be highlighted by the breakup of ice shelves at the northern coast of Ellesmere Island (Mueller et al., 2008), which have been stable until recently for at least several thousand years based on geological data (England et al., 2008). On the basis of satellite records, negative trends in sea-ice extent encompass all months, with the strongest trend in September. As assessed by the U.S. National Snow and Ice Data Center, the September trend over the period 1979–2009 is 11% per decade (Fig. 2b; [http://nsidc.org/data/seaice\\_index](http://nsidc.org/data/seaice_index)). 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<sup>2</sup> was the lowest in the satellite record and 23% lower than the previous September 2005 record low of 5.56 million km<sup>2</sup>. On the basis of an extended sea-ice record, it appears that the September 2007 ice extent is only half of that estimated for the period 1950–1970 based on the Hadley Center sea ice and sea-surface temperature data set (HadISST) (Rayner et al., 2003). While the ice extent rebounded slightly in September 2008 and 2009, these months rank second and third lowest in the satellite record, respectively.

Many factors may have contributed to this ice loss (Serreze et al., 2007a), such as general Arctic warming (Rothrock and Zhang, 2005), extended summer melt season (Stroeve et al., 2006), and effects of the changing phase of large-scale atmospheric patterns such as the Northern Annular Mode and the Dipole Anomaly (Wang et al., 2009). These atmospheric forcings have flushed some thicker multi-year ice out of the Arctic and left thinner first-year 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 long-wave 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 amount of the oldest and thickest ice within the multi-year pack has declined significantly (Maslanik et al., 2007b). The area of the Arctic Ocean covered by predominantly older ice (5 or more

years old) 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 thicknesses suggests that this loss of older ice translates to a decrease in mean thickness for the Arctic from 2.6 m in March 1987–2.0 m in 2007 (Maslanik et al., 2007b).

The role of greenhouse gas forcing on the observed sea-ice area trends finds strong support from the study of Zhang and Walsh (2006). These authors showed that for the period 1979–1999, the multi-model mean trend projected by the Coupled Model Inter-comparison Project, version 3 (CMIP3; IPCC, 2007) is downward, as are trends from most individual simulations. However, Stroeve et al. (2007) found that few or none (depending on the time period of analysis) of the September trends from the CMIP3 runs are as large as that observed in satellite data. If the multi-model mean sea-ice 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 analysis argues that natural variability has strongly contributed to the observed trend, Stroeve et al. (2007) concluded that, as a group, the models underestimate the sensitivity of sea-ice cover to forcing by greenhouse gases. Overly thick ice simulated by several of the models appears to provide at least a partial explanation for the disparity between model results and observed trends.

About half of the CMIP3 models driven with the middle-range SRES A1B emissions scenario (which reaches 720 ppm CO<sub>2</sub> levels by 2100) simulate complete or nearly complete loss (less than 1 × 10<sup>6</sup> km<sup>2</sup>) of September sea ice, with some of them obtaining near ice-free September conditions as early as 2040 (Arzel et al., 2006). These simulations retain winter sea ice throughout the 21st century, although this thins considerably compared to late 20th century conditions. As discussed by Holland et al. (2009), the intermodel scatter in changing sea-ice mass budgets over the 21st century is considerable, with changes in downwelling longwave and net shortwave fluxes in the future Arctic climate being uncertain. This in turn suggests that uncertainties in evolving cloud and surface albedo conditions are important players in the range of future sea-ice loss simulated by current climate models. Nevertheless, as discussed above, these models as a group may be too

conservative, and predict a later rather than earlier date, for when the Arctic Ocean will be ice-free in summer.

Recent modeling studies have discussed the possibility of rapid change in future Arctic summer ice conditions. Simulations based on the Community Climate System Model, version 3 (CCSM3) (Holland et al., 2006a) indicate that the end-of-summer ice extent is sensitive to ice thickness in spring. If the ice thins to a more vulnerable state, a “kick” associated with natural climate variability can result in rapid summer ice loss enhanced by the ice-albedo feedback. In the CCSM3 events, anomalous ocean heat transport acts as this trigger. In one ensemble member, the area of September ice decreases from about  $6 \times 10^6 \text{ km}^2$  to  $2 \times 10^6 \text{ km}^2$  in 10 years, resulting in a nearly ice-free September by 2040. This result is not just an artifact of CCSM3, as a number of other climate models show similar rapid ice loss. Additionally, the broad similarities of these model results to the large ice loss event of 2007 provide further support that instances of rapid summer ice loss may well occur in a near-future Arctic system.

### 2.3. Influences of sea-ice loss on the climate system

Seasonal changes in the net surface heat flux associated with sea-ice processes modulate atmospheric energy transports and exchange. The albedo of sea-ice ranges from over 80% when it is freshly snow-covered to around 50% during the summer melt season, and can be lower in areas of water ponded on ice (e.g., Perovich et al., 2002). This high reflectivity contrasts with the dark ocean surface, which has an albedo of less than 10%. The high albedo and large surface area of Arctic sea ice, 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 atmospheric poleward heat transport 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, reducing the requirements for atmospheric heat to be transported poleward into the Arctic (Serreze et al., 2007b).

Model experiments have addressed potential changes in the regional and large-scale aspects of atmospheric circulation that may result from a loss of sea ice. Magnusdottir et al. (2004) found that a reduced area of winter sea ice in the North Atlantic promoted a negative phase of the North Atlantic Oscillation in the model, with storm tracks weaker and shifted southward. 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 an atmospheric 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 with 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 produce different results. Coordinated experiments that use a suite of models are needed to help to reduce uncertainty.

Models consistently simulate amplification of Arctic surface warming in response to rising  $\text{CO}_2$  levels (e.g., Manabe and Stouffer, 1980; Holland and Bitz, 2003). While a number of processes cause this Arctic amplification, much of the signal can be attributed to the loss of sea ice and surface albedo change. Retreat of the ice margin allows the dark, low-albedo ocean to readily absorb solar energy, increasing the summer heat content in the ocean mixed layer. Ice formation in autumn and winter is delayed. This allows for a large

upward heat transfers from the ocean to the atmosphere (e.g., Hall, 2004; Winton, 2006; Graverson and Wang, 2009; Serreze et al., 2009). However, the magnitude of Arctic amplification does vary considerably across different climate models. Models with relatively thin initial sea-ice tend to exhibit higher polar amplification (Rind et al., 1995; Holland and Bitz, 2003). Recent observations (Serreze et al., 2009) suggest that an amplified warming signal is now emerging in the Arctic, much like that projected by climate models.

Climate models also indicate that changes in the melting 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 suppress 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 (Manabe and Stouffer, 1999; Clark et al., 2002). 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 glacial ice, increased net precipitation, and increased river input). However, the net effect of these various changes in freshwater buoyancy forcing for the net MOC response remains uncertain.

## 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 seafloor and in deposits along many Arctic coasts. Indirect information on sea-ice extent can also be derived from terrestrial paleoclimate archives such as the coastal vegetation and ice cores. Such paleoclimate information provides a context, within which the patterns and effects of current and projected future ice conditions can be evaluated.

### 3.1. Marine sedimentary records

The most complete and spatially extensive records of past sea-ice conditions are provided by seafloor 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 the 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 millions of years are most likely to be found in the deep, central part of the Arctic Ocean, where the seafloor was not eroded during periods of lower sea-level and the passage of large 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; Polyak et al., 2009), so that sedimentary records from these areas may not capture short-term (submillennial or even millennial-scale) variations in paleoenvironments. By contrast, cores from Arctic continental margins sometimes provide high-resolution records that capture events on century or even decadal time scales, but these cores usually cover a relatively short time interval since the Last Glacial Maximum (LGM) and the corresponding low sea-level stand (<ca 20 kyr). Furthermore, paleoenvironmental signals in these cores are often overwhelmed by local freshwater fluxes and redeposition of terrestrial material during the post-LGM flooding of the shallow shelves fed by large rivers, a problem that is especially



common for the Siberian continental margin (e.g., Bauch et al., 2001; Stein et al., 2004). A combination of sediment cores from the central basin and from continental margins of the Arctic Ocean is needed to fully characterize sea-ice history and its relation to climate change.

Until recently, most cores relevant to the history of sea-ice cover were collected from low-Arctic marginal seas, such as the Barents Sea and the Nordic Seas. There, restricted ice conditions allow for easier ship operation, whereas sampling in the central Arctic Ocean requires the use of heavy icebreakers. Recent advances – notably several icebreaker coring cruises such as the 2005 Trans-Arctic Expedition (HOTRAX: Darby et al., 2005) and the first deep-sea drilling in the central Arctic Ocean (ACEX: Backman et al., 2006) (Fig. 3) — provide new, high-quality material from the Arctic Ocean proper, with which to characterize past variations in ice cover.

### 3.1.1. Ice-rafted sediment

The most direct proxies for the presence of floating ice are derived from sediment that melts out or drops from ice transported by wind and surface currents from the sites of sediment entrainment. Ice-rafted debris (IRD), usually defined as the coarse sediment fraction ( $>63\ \mu\text{m}$  or sometimes even coarser fractions) is commonly used as an indicator of deposition from ice, which can include both iceberg and sea-ice rafting (Lisitzin, 2002, and references therein). Distinguishing between these two ice-transport processes is important, yet challenging, for a correct interpretation of glaciomarine paleoenvironments.

Icebergs can carry sediment grains of all sizes, from clay to boulders, as inherited from sediment material entrained in glaciers. Although in some cases this material can be relatively fine grained, typically iceberg-rafted sediment has a high content of coarse IRD, exceeding 10–20%  $>63\ \mu\text{m}$  (e.g., Clark and Hanson, 1983; Dowdeswell et al., 1994; Andrews, 2000). During glacial periods sedimentation in the Arctic is interpreted to be predominantly iceberg-derived due to low sea levels, and thus exposure of shallow continental shelves, and huge ice-sheet fronts surrounding the Arctic Ocean. This inference is consistent with greater IRD abundances in glacial and deglacial intervals in sediment cores (e.g., Phillips and Grantz, 2001; Adler et al., 2009; Polyak et al., 2009). During interglacials, exemplified by the modern conditions, contribution of icebergs to sediment transport and deposition in the Arctic is limited and plays a significant role only in areas proximal to calving glaciers such as around Greenland and the eastern Canadian Arctic (e.g., Andrews et al., 1997; Lisitzin, 2002). In contrast, the role of sea ice during interglacials is greatly enhanced in the Arctic by the availability of broad and shallow continental shelves, the major sites of sea ice formation.

Sediment entrainment in sea ice occurs mostly during periods of ice-freeze-up (frazil-ice formation) on the shallow shelves and is largely restricted to fine silt and clay-size sediments available in the suspension (e.g., Kempema et al., 1989; Nürnberg et al., 1994; Lisitzin, 2002; Darby, 2003), although the relative contribution of these grain-size fractions may vary geographically (Hebbeln, 2000; Lisitzin, 2002). While sea-ice transported sediment is dominantly fine-grained, there are conditions in which sea ice can contain coarser grains. Coarse sediment shed from coastal cliffs can be transported by landfast ice, but its distribution is mostly limited to near-coastal areas. Anchor ice, which forms on the seafloor in super-cooled conditions, can entrain any sediment available on the shallow shelves (Reimnitz et al., 1987); however, the relative contribution of anchor ice to the overall Arctic ice cover is not fully understood. Studies of sediment in modern sea-ice samples show the amount of  $>63\ \mu\text{m}$  grains as less than 5–10% (e.g., Clark and Hanson, 1983; Pfirman et al., 1990; Nürnberg et al., 1994; Darby, 2003; Darby et al., 2009), and these numbers are probably biased towards higher values due to preferential sampling of easily

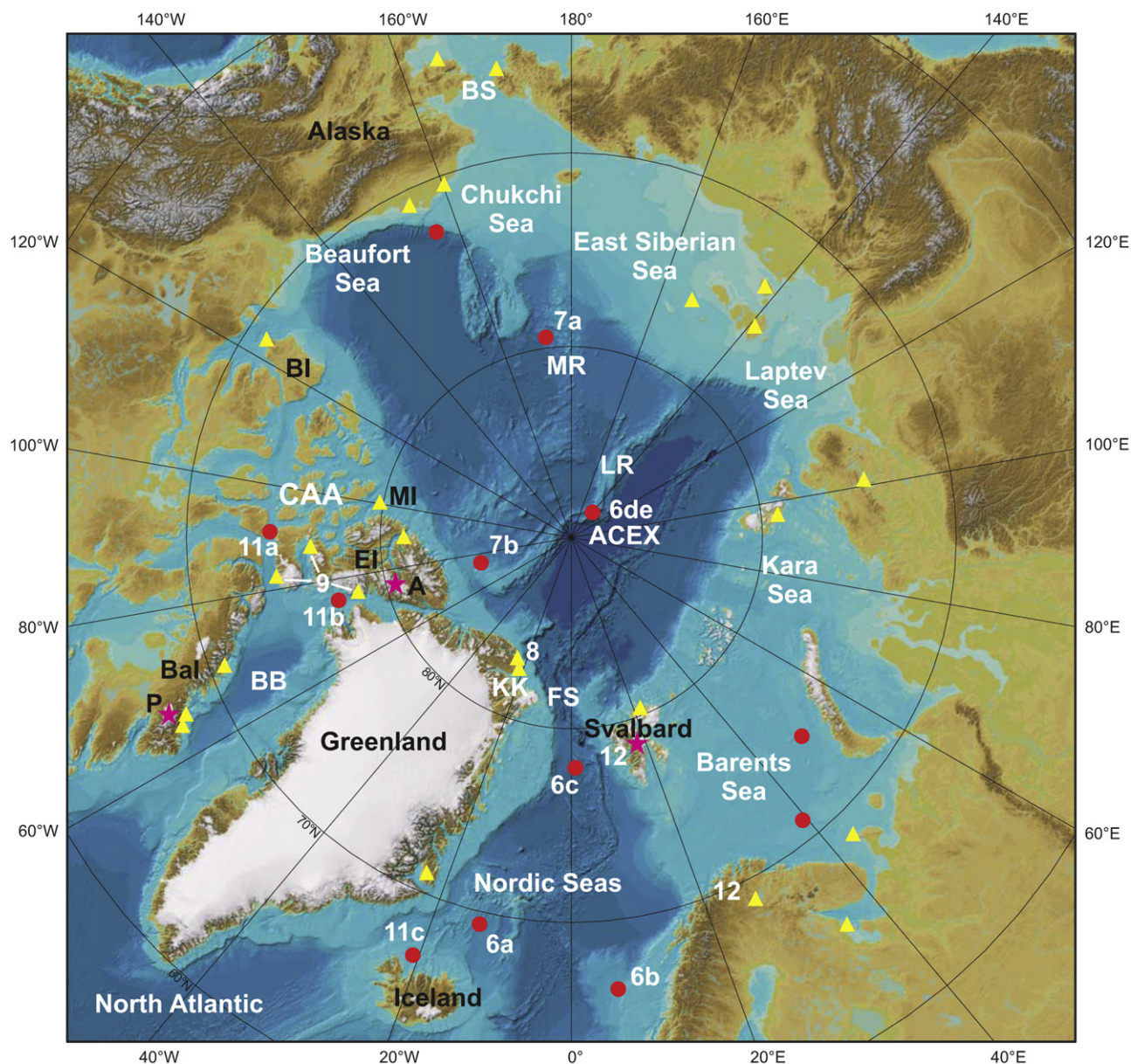
identifiable patches of sandy ice. Overall it can be concluded that sediment in the Arctic Ocean with IRD abundances exceeding the above numbers is indicative of iceberg deposition, whereas low-IRD numbers can be produced by both iceberg and sea ice affected environment. We note that in some subarctic seas, such as in the northwest Pacific, sea-ice sediment may have higher IRD content than in the Arctic due to different sedimentary environments in the coastal zone (Lisitzin, 2002).

Detailed grain-size analyses show that both sea ice and seafloor Holocene deposits in the Arctic Ocean contain sediments with modes in clay or fine silt fractions, but also with a bimodal distribution, where the second mode is in coarser silt to sand fractions (Darby et al., 2009). The fine-grained types are likely related to the frazil-ice sediment entrainment in the water column, while the bimodal type has been interpreted as anchor-ice sediment. This latter type can merge with grain-size distributions resulting from iceberg deposition, which demonstrates that granulometry may not always be sufficient for a conclusive differentiation of sea-ice vs iceberg deposits. In-depth studies of IRD, such as examination of shapes and surface textures of quartz grains, can provide additional insight (Helland and Holmes, 1997; Dunhill, 1998; Stickley et al., 2009). For example, the abundance of quartz grains displaying mechanical abrasion surface textures was found to broadly correspond to increases in total IRD abundance in a study by Stickley et al. (2009); these changes were interpreted as indicating greater transport of sediment by icebergs. Interpretation of ice-transport mode based on grain surface textures is complicated however by the potentially complex sediment history of mineral grains prior to ice transport and deposition in the marine environment.

Sediment provenance indicators can help to establish the source of sediment and track ice drift. If the ice-transported sediment can be traced to a shelf that has not been glaciated, then this can be used to identify sea-ice-rafting as opposed to iceberg rafting. Especially telling is sediment carrying some diagnostic peculiarity that is foreign to the site of deposition. Chemical fingerprinting has been developed for iron-oxide sand grains, which can be matched to specific source areas with an extensive data base of circum-Arctic shelf and coastal sediment samples analyzed for the same elements by electron microprobe (Darby, 2003). Fe-oxide grains accept relatively large amounts of element substitution into the crystal structure during igneous or metamorphic crystallization such that unique compositions are frequently created (Darby, 2003, 2008). Quantitative X-ray diffraction analysis of the clay fraction (Vogt and Knies, 2008) or bulk sediment  $<2\ \text{mm}$  (Andrews et al., 2009) can also be used in those instances where minerals are “exotic” relative to the composition of the nearest terrestrial sources. For example, the abundance of quartz in Holocene marine cores from the Nordic seas is used as a measure of Arctic ice drift dominated by sea ice (Moros et al., 2006; Andrews et al., 2009). Similarly the presence of detrital carbonates (mostly dolomite) is characteristic of iceberg pulses from the Laurentide Ice Sheet in sediment cores both in the North Atlantic (e.g., Andrews, 2000; Parnell et al., 2007) and the Arctic Ocean (Polyak et al., 2009; Stein et al., in press).

### 3.1.2. Other marine proxies

Skeletons of microscopic organisms in bottom sediment such as foraminifers, diatoms, and dinocysts may indicate the condition of ice cover above the study site. Some planktonic organisms live in or on sea ice, such as some diatoms or the epipelagic ostracode *Acetabulostoma* (Cronin et al., 1995), or are otherwise associated with ice. Organisms that require 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 the former presence of sea ice and even to estimate the length of the ice-cover season



**Fig. 3.** Index map of the Arctic showing seafloor core sites (circles), ice cores (stars), and key coastal/terrestrial sites (triangles) mentioned in the paper. Figure numbers are shown next to respective sites. LR – Lomonosov Ridge, MR – Mendeleev Ridge, FS – Fram Strait, BS – Bering Strait, CAA – Canadian Arctic Archipelago, BB – Baffin Bay, Bal – Baffin Island, EI – Ellesmere Island, MI – Meighen Island, BI – Banks Island, KK – Kap Kobenhavn. P and A – Penny and Agassiz Ice Caps (stars). Base map is the International Bathymetric Chart of the Arctic Ocean (IBCAO-2; Jakobsson et al., 2008a).

based on transfer functions (e.g., Koç and Jansen, 1994; de Vernal and Hillaire-Marcel, 2000; de Vernal et al., 2005, 2008; Stickley et al., 2009; Caissie et al., 2010). The reliability of the applied transfer functions, however, depends upon the spatial distribution of calibration data sets and the accuracy of the environmental data, which is commonly quite limited in the Arctic.

Benthic organisms in polar seas are also affected by ice cover because it controls primary production, and thus food reaching the seafloor. Benthic species that prefer relatively high fluxes of fresh organic matter can indicate the location of the ice margin on the Arctic shelves (Polyak et al., 2002; Jennings et al., 2004). In the central Arctic Ocean, benthic foraminifers and ostracodes also offer a potential for identifying ice conditions through an understanding of Pelagic–Benthic coupling (Cronin et al., 1995; Wollenburg et al., 2001; Polyak et al., 2004). We note that many of the changes in benthic assemblages that used to be interpreted in terms of

temperature and salinity are now seen as related primarily to trophic changes (e.g., Wollenburg and Kuhnt, 2000).

The composition of organic matter in sediment, including specific organic compounds (biomarkers), can also be used to characterize the environment in which it formed. A new, powerful tool for sea-ice reconstruction is offered by a specific biomarker, IP<sub>25</sub>, that is associated with diatoms living in sea ice, and thus reflects the occurrence of ice in spring (Belt et al., 2007; Massé et al., 2008; Vare et al., 2009). The method has been tested by the analysis of seafloor samples from the Canadian Arctic and has been further applied to downcore samples for characterization of past ice conditions. This method may provide a new level of understanding of past ice conditions; however, its applicability to sediments in the central Arctic Ocean, where both biological production and sedimentation rates are very low, is yet to be tested.



Reconstructing the history of meltwater and brines provides another approach for evaluating paleo-sea-ice distribution and seasonal conditions. Because oxygen isotopes fractionate during sea-ice formation and melting,  $\delta^{18}\text{O}$  variations in fossil calcite may record the intensity of these processes. For example,  $\delta^{18}\text{O}$  depletions in planktonic foraminifers at halocline levels have been interpreted to indicate enhanced brine production and, thus, seasonal sea-ice formation in the early Holocene Arctic Ocean and during Heinrich events in the North Atlantic (de Vernal et al., 2008; Hillaire-Marcel and de Vernal, 2008). Noble gases in sediment pore water also offer a potential for tracking the history of brine production due to their differential solubility in ice (Hood et al., 1998; Postlethwaite, 2002).

A potentially novel tool for evaluating paleo-sea-ice variation in the Arctic using mercury isotopes is under development (Gleason et al., 2009), exploiting the photochemically sensitive, mass-independent fractionation effect in Hg isotopes (Bergquist and Blum, 2007). Initial studies from Alaskan waters show that Hg-isotope composition changes with latitude, possibly under the influence of sea-ice concentration (Point et al., 2008).

It is important to understand that although all of the above proxies have a potential for identifying the former presence or the seasonal duration of sea ice, each has limitations that complicate interpretations based on a single proxy. Identification of the sea-ice signal can be obscured by other hydrographic controls such as temperature and salinity, as well as nutrient-related changes in biological productivity. 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). A multi-proxy approach is desirable for making confident inferences about sea-ice variations. A thorough understanding of sea-ice history depends on refining sea-ice proxies in sediment taken from strategically selected sites in the Arctic Ocean and along its continental margins.

### 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 sediments and landforms such as coastal plains, marine terraces, and beaches. Deposits from each of these formerly marine environments can now be found 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 valuable information that can be compared with seafloor sediment records. While in sediment cores usually only microfossils provide quantitative paleobiological material, the spacious coastal exposures enable abundant recovery of larger fossils such as plant remains, driftwood, whalebone, and relatively large mollusks. These items contribute information about past sea-surface and air temperatures, circulation, sea-ice conditions, and the expansions of extralimital species. For example, marine 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 shallow-water environments including temperature and ice conditions. Pollen and plant macrofossils, on the other hand, are especially useful for recording vegetation and environmental change on nearby coasts. In addition to the content of coastal deposits, associated landforms such as beach ridges can also be indicative of landfast ice conditions. As beach ridges are formed by wave action, raised beach ridges on coasts bordered by permanent landfast ice must indicate past periods with seasonally open water.

#### 3.2.1. 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 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; Bennike et al., 2002), the eastern Canadian Arctic (Miller, 1985), Ellesmere Island (Fyles et al., 1998), Meighen Island (Matthews, 1987; Matthews and Ovenden, 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 area (Kaufman and Brigham-Grette, 1993; Brigham-Grette and Hopkins, 1995), and in the western Eurasian Arctic (Funder et al., 2002) (triangles in Fig. 3). In nearly all cases the primary evidence used to estimate the extent of past sea ice is derived from *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; Bennike et al., 2002; CAPE, 2006), provide an essential view of sea-surface temperatures, sea-ice extent, and seasonality.

For example, fossils found in deposits of early-Pleistocene marine transgressions (Bigbendian and Fishcreekian) on coastal plains along the northern and western shores of Alaska suggest that the limit of seasonal sea ice in Alaskan waters was at least 1600 km farther north than at present (Carter et al., 1986; Brigham-Grette and Carter, 1992; Kaufman and Brigham-Grette, 1993). This interpretation is based on extralimital fauna including the gastropod *Littorina squalida* and bivalve *Clinocardium californiense*, whose modern northern limit is in the Bering Sea (Carter et al., 1986). The Fishcreekian deposits also include *Natica* (*Tectonatica*) *janthostoma*, which today is limited to waters adjoining northern Japan and the Kamchatka Peninsula. Both units include the fossil remains of sea otters, while modern sea otters do not tolerate severe seasonal sea-ice conditions. Pollen assemblages in the Bigbendian deposits suggest that the nearby territories probably supported an open spruce-birch woodland or even parkland like in modern southern Alaska, possibly with scattered pine (Brigham-Grette and Carter, 1992). Pollen from Fishcreekian deposits suggest more severe conditions of herbaceous tundra with open larch forest; these environments still indicate a considerable northward migration of vegetation zones in comparison with modern conditions. Based on both marine faunal and paleobotanic evidence, perennial sea ice must have been severely restricted or absent, and winters were warmer than at present during these two sea-level highstands. Another prominent example of a faunal range extension is provided by the findings of a northern Pacific bivalve *Cyrtodaria kurriana* at the western coasts of the Canadian Arctic Archipelago (CAA) during the last deglaciation (England and Furze, 2008). Unlike climate-induced range extensions discussed above, this event was probably enabled by the clearance of summer sea ice by voluminous meltwater inputs from the retreating Laurentide ice sheet.

#### 3.2.2. Driftwood

Sea-ice conditions, especially the presence or absence of landfast ice, may be inferred from the distribution of driftwood logs, mostly spruce and larch, found in raised beaches along the coasts of Arctic Canada (Blake, 1975; Dyke et al., 1997; England et al., 2008), Greenland (Bennike, 2004; Funder et al., 2009), Svalbard (Hagblom, 1982), and Iceland (Eggertsson, 1993). Driftwood is delivered primarily to landfast-free coasts near sea-ice margins, because ice is essential for long-distance transport of the wood, which otherwise becomes water logged and sinks after about a year

(Hagblom, 1982). The Transpolar Drift Stream appears to have been the main agent of wood transport in the Arctic Ocean (Tremblay et al., 1997). Most of the larch logs found are attributed to a northern Siberian source, whereas most of the spruce comes from the Mackenzie and Yukon Rivers. In areas other than Iceland, the glacial isostatic uplift of the land has led to a staircase of raised beaches hosting various numbers and compositions of logs (e.g., the ratio of spruce to larch) with time. An extensive radiocarbon database catalogs these variations, which have been associated with the growth and disappearance of landfast sea ice (Dyke et al., 1997; England et al., 2008) and changes in the trajectory of the Transpolar Drift under different dominant wind fields (Dyke et al., 1997; Tremblay et al., 1997).

### 3.2.3. Whalebone

Reconstructions of sea-ice conditions in the CAA have to date been derived mainly from the distribution in space and time of marine mammal bones in raised marine deposits, where abundances have been documented in systematic surveys (e.g., Dyke et al., 1996, 1999; Fisher et al., 2006; Atkinson, 2009). Understanding the dynamics of ice conditions in this region is especially important for modern-day considerations because an ice-free, navigable North-west Passage will provide new opportunities for shipping. 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, mainly limited to the Holocene, has yielded a large set of results (e.g., Harington, 2003; 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, 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 summer into the Beaufort Sea along the western edge of the CAA. The Atlantic population advances in summer from the northern Labrador Sea into the eastern channels of the CAA. 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 CAA; i.e., the central part of the North-west Passage. Both populations retreat southward upon autumn freeze-up.

However, the ice-edge environment is hazardous, especially during freeze-up, and individuals or pods may become entrapped (as has been directly observed). Detailed measurements of fossil bowhead skulls and mandibles (a proxy of age) now found in raised marine deposits allow a reconstruction of their lengths (Dyke et al., 1996; Saville et al., 2000). The distribution of lengths compares very closely with the length distribution of the modern Beaufort Sea bowhead population (Fig. 4), indicating that the cause of death of many bowheads in the past was a catastrophic process that affected all ages indiscriminately. One such process is ice entrapment. For example, the northernmost Holocene bowheads found in the CAA are relatively rare finds that appear to represent the remains of doomed strays that failed to retreat early enough from coastal leads extending north from northernmost polynyas accessible to them (Dyke and England, 2003).

The rich bowhead record of Svalbard may eventually be equally informative, and it shows some similarities to that from the CAA (Dyke et al., 1996). However, systematic surveys of Svalbard bone abundance through time are lacking.

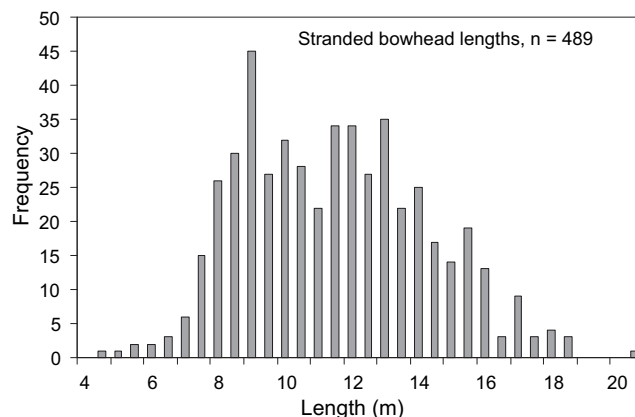


Fig. 4. The reconstructed lengths of Holocene bowhead whales on the basis of skull measurements (485 animals) and mandible measurements (an additional 4 animals) (Saville et al., 2000). This distribution is very similar to the distribution of lengths in living Pacific bowhead whales; thus, past whale strandings affected all age classes. [Reproduced by permission of Arctic Institute of North America.].

## 3.3. Terrestrial records

### 3.3.1. Vegetation

Plant remains and pollen provide a much-needed link to information about the past climate throughout Arctic and subarctic regions. While the evidence for spatial patterns of paleo-vegetation is not even across various parts of the Arctic, and is especially meager beyond the range of the Holocene, there exists a number of key stratigraphic sections, which allow for an evaluation of the past plant ecosystems and related climatic environments (e.g., Francis, 1988; Bennike and Böcher, 1990; Fyles et al., 1994; White et al., 1997; Andreev et al., 2003, 2009). Notably, the location of the northern tree line, which is presently controlled by the July 7 °C mean isotherm, is a critical paleobotanic indicator for understanding Arctic paleoclimate including sea-ice conditions. Nowhere today do trees exist near shores lined with perennial sea ice; forests thrive only in southerly reaches of regions with seasonal ice. In addition to the composition of vegetation, the width of tree rings in maritime regions can also be used as an indirect proxy for sea-ice conditions via regional changes in precipitation and temperature. Macias-Fauria et al. (2009) reconstructed the millennial history of sea-ice conditions in the Nordic Seas based on the combined tree-ring and ice-core data from northern Scandinavia and Svalbard. The validity of this reconstruction, corroborated by a comparison with historical sea-ice observations for the last two centuries, is based on spatially coherent patterns in sea ice, SST, and atmospheric circulation in this region due to a strong influence of the ice extent on the North Atlantic overturning and related atmospheric conditions.

### 3.3.2. 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 continuous, often sub-annual resolution. The question to address 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 by means of such proxies as melt layers or oxygen isotopes are associated with reduced sea ice (Isaksson et al., 2005a; Macias-Fauria et al., 2009). Atmospheric patterns inferred from ice-core data can also help in reconstructing former sea-ice conditions (e.g., Alt et al., 1985). However, the real goal for paleo-sea-ice studies is to find a chemical indicator that has a concentration mainly controlled by 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. In most of the world's oceans, salt in sea water becomes an aerosol in the atmosphere by means of bubble-bursting at the ocean surface, and formation of the aerosol is related to surface wind speed (Guelle et al., 2001). Expanding sea ice moves the source region (open ocean) farther 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 low correlation coefficient meant that only about 7% of the variability in the abundance of sea salt was directly linked to variability in position of sea ice. An 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 (Whitlow et al., 1992; Mosher et al., 1993), 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 Antarctic studies of aerosols and ice cores, where the sea ice surface itself can be a source of large amounts of sea salt aerosol in coastal regions (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 (Wolff et al., 2003; Fischer et al., 2007). Wolff et al. (2003) identified negative non-sea-salt sulfate as providing additional evidence of a sea ice source of sea salt, owing to precipitation of mirabilite before aerosolization from “frost flowers” growing on new sea ice. An Antarctic sea ice record covering 740 ka has been presented on the basis of sea salt aerosol, and shows extended sea ice at times of low temperature (Wolff et al., 2006). The obvious question is 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 salt in ice cores relate it to the production of new, thin sea ice. In the regions around Greenland and the nearby islands, much of the sea ice is old ice that has been transported into the region. It therefore seems unlikely that the method can easily be applied to long northern ice cores 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.

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., 2005b).

In summary, ice cores have the potential to add a well-resolved and regionally integrated picture of the past sea ice extent, but this potential has not yet been realized in the Arctic, and ice core data will not feature strongly in the paleo-reconstructions discussed in this paper.

### 3.4. Historical records

Historical records may describe recent paleoclimatic features such as weather and ice conditions. The longest historical records of ice cover exist from coastal seas accessible to shipping. This is exemplified by the Barents Sea, where a record of varying detail has been compiled covering four centuries (Vinje, 1999, 2001; Divine and Dick, 2006). Systematic records of the position of the 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 content with time. Direct observations on ice concentrations spanning the Arctic are available since 1953. Coverage from early era satellites began in 1972, with the modern satellite record starting in 1979 (Cavalieri et al., 2003).

The most remarkable case of historical sea-ice records is provided by Iceland. For 1200 years, the peoples of Iceland have recorded observations of drift ice (i.e., sea ice and icebergs), following the settlement of the island in approximately 870 AD (Koch, 1945; Bergthorsson, 1969; Ogilvie, 1984; Ogilvie et al., 2000). These historical sources have been used to construct a sea ice index that compares well with springtime temperatures at a climate station in northwest Iceland (Fig. 5). Ice commonly develops off the northwest and north coasts and only occasionally extends into southwest Iceland waters (Ogilvie, 1996).

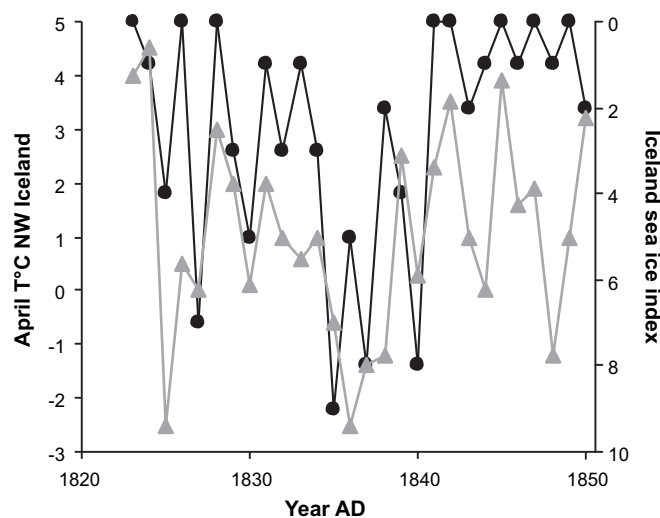


Fig. 5. The sea ice index on the Iceland shelf (black line) plotted against springtime air temperatures in northwest Iceland that are affected by the distribution of ice in this region (from Ogilvie, 1996). Higher ice index indicates more ice.

#### 4. History of Arctic sea-ice extent and circulation patterns based on proxy records

##### 4.1. Pre-Quaternary history

Until recently, information on the long-term (million-year scale) climatic history of the polar areas of the Northern Hemisphere was limited to fragmentary records from the Arctic periphery. The ACEX deep-sea drilling borehole on the Lomonosov Ridge 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 Paleocene–Eocene Thermal Maximum (PETM), the Arctic Ocean was considerably warmer than it is today, with summer temperatures estimated as high as 24 °C and freshwater subtropical aquatic ferns growing in abundance (Moran et al., 2006). This warm 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 low-lying delta floodplains riddled with lakes and swamps (McKenna, 1980; Francis, 1988). Later, an abundance of the sea ice diatom *Synedropsis* spp., concurrent with the first occurrence of sand-sized debris at ~47 Ma and a doubling of its flux at ~46 Ma, indicate the possible onset of drifting sea ice and perhaps even some glaciers in the Arctic during the cooling that followed the thermal optimum (Moran et al., 2006; St. John, 2008; Stickley et al., 2009). This was the time of a large-scale reorganization of the continents, notably the oceanic separation of Antarctica, and of a decrease in atmospheric CO<sub>2</sub> concentration that may have been more than 1000 ppm (Table 1) (Pearson and Palmer, 2000; Lowenstein and Demicco, 2006). It is important to note, however, that during the Eocene the ACEX site was at the margin of the Arctic Ocean and may not be fully representative of environments in its central part.

The middle to late Eocene was the time of an overall continued cooling, but the high-Arctic coasts at this time were still occupied by rich, high-biomass forests of redwood and by wetlands characteristic of temperate conditions (Williams et al., 2003; LePage et al., 2005). Cooling culminated in an abrupt decrease in atmospheric pCO<sub>2</sub> and temperature at the Eocene–Oligocene boundary about 34 Ma, triggering massive Antarctic glaciation (Table 1) (e.g., Zachos et al., 2008; Pearson et al., 2009). The first Greenland glaciers may have developed about the same time, on the basis of coarse grains interpreted as IRD in the North Atlantic sediments deposited between 38 and 30 Ma (Eldrett et al., 2007). In Alaska this climatic deterioration caused a dramatic floral turnover with generic extinction in woody plants estimated as high as 40% (Wolfe, 1980, 1997). Central Arctic conditions during this time cannot yet be evaluated as the ACEX record has a large hiatus between ca 44 and 18 Ma (Backman et al., 2008), whereas fossil assemblages and isotopic data in Late Oligocene marine sediments along the coasts

of the Beaufort Sea suggest frigid waters with a seasonal range between 1 °C and 9 °C (Oleinik et al., 2007).

Sustained climate conditions lingered through 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 CAA was covered in mixed conifer-hardwood forests similar to those of southern Maritime Canada and New England today, the southerly reaches of seasonal ice.

ACEX sediments overlying the 44–18 Ma unconformity contain relatively low numbers of IRD from 17.5 to ~16 Ma, but abundances ramp up after that (St. John, 2008). This may have been caused by the growth of ice masses on and around Svalbard and iceberg discharge into the eastern Arctic Ocean and the Greenland Sea at about 15 Ma (Knies and Gaina, 2008), which was possibly related to the establishment of the modern circulation system in the Arctic Ocean after the opening of Fram Strait at about 17 Ma (Jakobsson et al., 2007).

The mineralogical change in the ACEX record between 14 and 13 Ma indicates the likelihood that sea ice was now perennial, based on a shift from relatively proximal (Kara Sea) to more distal sources requiring more than 1 year of ice drift under present circulation rates (Krylov et al., 2008), although this transition yet needs to be examined in records from other Arctic sites to estimate the geographic distribution and persistence of the ice. The presence of perennial ice at the ACEX site at this time is also demonstrated by the provenance of Fe-oxide grains in IRD from the eastern Siberian shelves that bear no apparent evidence of glaciation and would require more than a year to reach the ACEX site (Darby, 2008), and by low fluxes of cosmogenic <sup>10</sup>Be to the seafloor (Frank et al., 2008). Several pulses of relatively abundant IRD in the late Miocene ACEX record indicate further expansion of sea ice or the growth of glaciers shedding icebergs into the Arctic Ocean (St. John, 2008). This interpretation is consistent with a cooling climate indicated by the spread of pine-dominated forests in northern Alaska around 12–13 Ma (Table 1) (White et al., 1997; Wolfe, 1997) and a reduction in warm-water or high-productivity foraminiferal species in the Beaufort–Mackenzie basin (termination of *Asterigerina staeschei* assemblage; McNeil, 1990). In the global context this climatic transition was characterized by an expansion of the Antarctic glaciation (Shevenell et al., 2004) and a considerable decrease in atmospheric pCO<sub>2</sub> (Kürschner et al., 2008; Tripathi et al., 2009).

After this climatic deterioration, throughout the late Miocene and most of the Pliocene climate was still considerably warmer than in the Pleistocene. For example, extensive braided-river deposits of the Beaufort Formation (early to middle Pliocene, about 5.3–3 Ma) that blanket much of the western CAA 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

**Table 1**  
Major Cenozoic (pre-Quaternary) climatic transitions reflected in the Arctic marine and terrestrial records.

Age of transition	ACEX record (central Arctic Ocean)	Marine records, Arctic margins	Alaska/CAA botanical record	Arctic glaciers	Global background
Early–Mid Eocene (47–46 Ma)	Appearance of sea-ice diatoms and IRD	Insufficient data	Insufficient data	Not known	pCO <sub>2</sub> drop; first ice in Antarctica?
Eocene–Oligocene (34 Ma)	No record (hiatus)	Insufficient data	T drop, spread of cool-temperate (conifer-hardwood) forests	First Greenland glaciers?	pCO <sub>2</sub> drop; massive glaciation in Antarctica
Mid Miocene (14–12 Ma)	Provenance change indicating perennial ice?	Faunal change; cooling/ice spread?	T drop; spread of boreal pine-dominated forests	Svalbard glaciation from 15 Ma	pCO <sub>2</sub> drop; expansion of Antarctic ice-sheet
Pliocene–Pleistocene (3–2.5 Ma)	No evident change (strong dissolution)	Faunal change; cooling/ice spread?	Southward migration of treeline; permafrost	Ice sheets in Eurasia and N. America	pCO <sub>2</sub> drop; onset of regular glacier cycles (“icehouse world”)

boreal conditions, cooler than in the Miocene, they are not compatible with the existence of extensive perennial sea ice in the adjacent Beaufort Sea during this time. Studies of the Mid-Pliocene climatic optimum indicate a strong influx of warm Atlantic waters into the Arctic resulting in SST as high as 18°C on the Yermak Plateau (ODP site 911) (Robinson, 2009). This climatic reconstruction is consistent with the presence of the subarctic bivalve *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 and ostracodes in Pliocene deposits in the Beaufort–Mackenzie area and several other sites around the Arctic margins are also characteristic of cool but not yet high-Arctic waters (Feyling-Hanssen, 1985; McNeil, 1990; Cronin et al., 1993). Deep-sea Arctic microfauna of potentially Pliocene age is only known from one site about 700 km north of the Alaskan coast (Mullen and McNeil, 1995). The abundance of species characteristic of seasonal phytodetritus pulses (e.g., *Epistominella exigua*) and the absence of Quaternary Arctic endemics adapted to very low-productivity environments, such as *Stetsonia arctica*, indicate seasonally ice-free conditions for this record.

Cooling in the late Pliocene, after ca 3 Ma, profoundly reorganized the Arctic system (Table 1). The tree line retreated from the Arctic coasts (Matthews and Telka, 1997; White et al., 1997), permafrost formed (Brigham-Grette and Carter, 1992; Burn, 1994), shallow marine faunas underwent the last major turnover (Feyling-Hanssen, 1985; McNeil, 1990), and continental ice masses grew at several sites at the Arctic margins – for example, the Svalbard ice sheet advanced onto the outer shelf (Knies et al., 2009), and between 2.9 and 2.6 Ma ice sheets began to grow in northern North America (Duk-Rodkin et al., 2004). New proxy measurements of pCO<sub>2</sub> (Foster et al., 2008; Tripathi et al., 2009) corroborate a widespread inference that this cooling is related to the decrease in atmospheric pCO<sub>2</sub> after the warm middle Pliocene. An additional factor that likely affected the Arctic Ocean was freshwater build-up due to inputs from ice sheets and the increasing transport of low-salinity Pacific water, which could have enhanced sea ice formation (Haug et al., 2001; Matthiessen et al., 2009). Many of the marine sites in the North Pacific, North Atlantic and Nordic Seas show a marked increase in the amount of IRD beginning in the late Pliocene (Fig. 6a–c) (e.g., Kriisek, 1995; Wolf-Welling et al., 1996; Jansen et al., 2000). Surprisingly, this same pattern is not found in IRD record from ACEX, where more subtle shifts are recorded between the Miocene and late Pleistocene (Fig. 6d,e; St. John, 2008). Biogenic proxies are also nearly absent in ACEX Neogene sediments, probably due to a pervasive dissolution (Backman et al., 2008; Cronin et al., 2008), which severely restricts our ability to reconstruct changes in Arctic Ocean environments during the Pliocene–Pleistocene transition based on this record alone.

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, 1985; 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 (North 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). Marine fossils from the same sediments are analogous with present-day faunas along the Siberian coast indicating open water for 2 or 3 months during summer as opposed to permanently ice-locked coast at the Kap København site today. Considering that present sea ice has highest concentration and thickness in the area north of Greenland, these

results imply that summer sea ice in the entire Arctic Ocean was considerably reduced during the Kap København event.

A noticeable reduction in the amount of IRD in the ACEX record occurred in the early Pleistocene, between ca 1.5–2 Ma, followed by only slightly varying IRD abundance during both glacial and interglacial intervals until the Late Pleistocene (Fig. 6d,e; St. John, 2008; O'Regan et al., in press). This observation does not appear to be an artifact of changing sedimentation rates. As reduction of ice sheets around the Arctic Ocean at that time is unlikely, this low-IRD record may be an indicator of a rather stable ice pack, which would reduce the transport and melting of debris-laden sea ice and icebergs across the central Arctic. Alternatively, the decrease in IRD deposition may be related to lower inputs of Atlantic-derived intermediate water, which would reduce basal melting of sea ice, or to less icebergs entering the Transpolar Drift from the Eurasian margin. The most dramatic changes in sediment composition, biogenic preservation, and IRD abundance in the ACEX record and adjacent cores occurred during Marine Isotope Stages (MIS) 1–6, that is, during the last 190 kyr (Fig. 6e; O'Regan et al., 2008b). These depositional patterns are linked to advances and retreats of the Barents–Kara ice sheet, which extended to the shelf edge during MIS 6 and succeeding glaciations, and thus directly affected sedimentation in the central Arctic areas affected by the Transpolar Drift (e.g., Spielhagen et al., 2004). The Pleistocene depositional history appears to be different in the western Arctic Ocean, where a pronounced increase in IRD abundances occurred earlier than on the Lomonosov Ridge, at estimated MIS16, about 650 ka (Polyak et al., 2009; Stein et al., in press). The western Arctic is largely controlled by the Beaufort Gyre circulation and sediment fluxes from the North American margin and, thus, reflects primarily the history of the Laurentide ice sheet and possibly different sea-ice conditions than in the eastern Arctic.

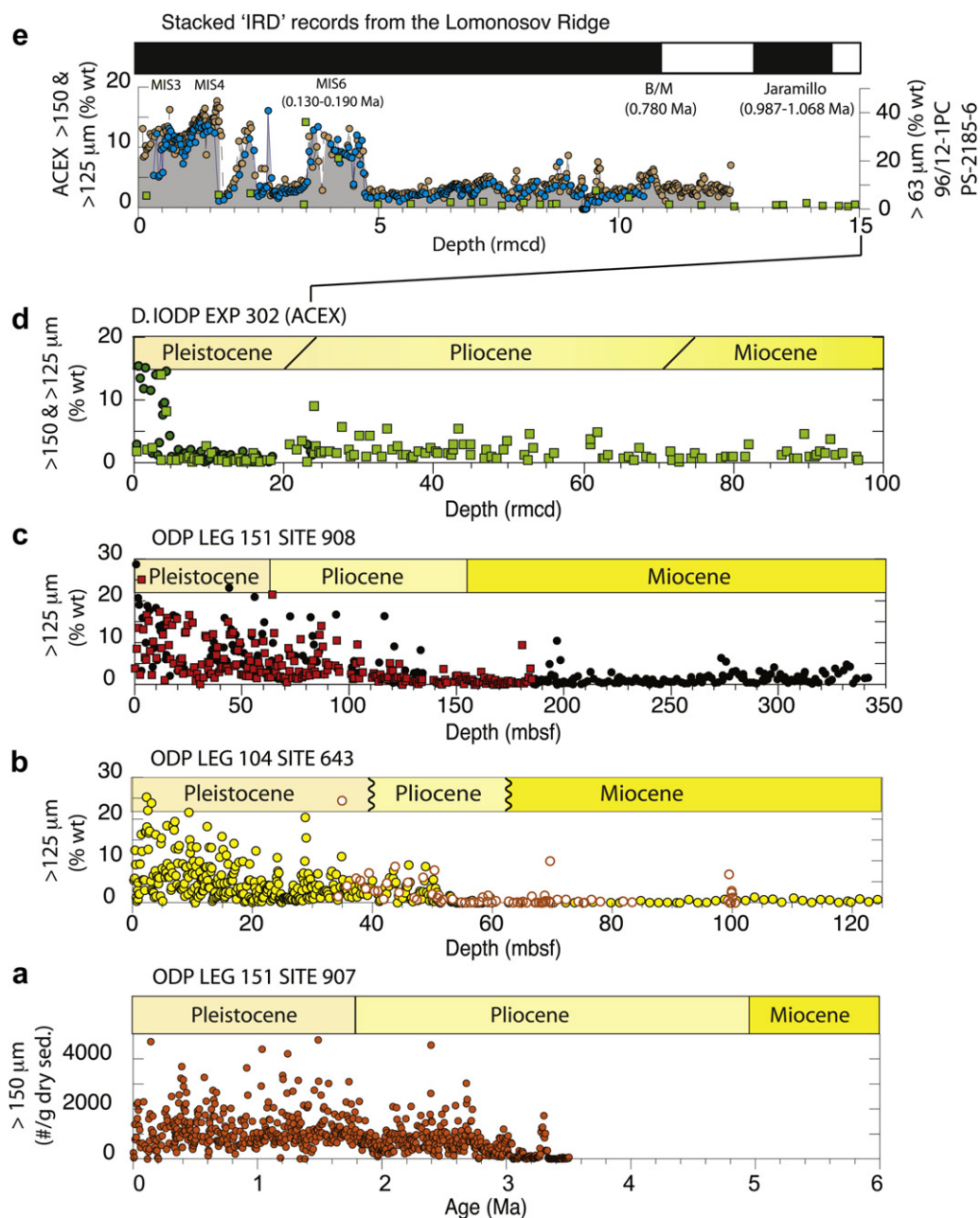
A more complete history of perennial versus seasonal sea ice and ice-free intervals during the past several million years requires additional sedimentary records distributed throughout the Arctic Ocean, and a synthesis of sediment and paleobiological evidence from both land and sea. This history will provide additional information on 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.

#### 4.2. Quaternary variations

The Quaternary period is characterized by overall low temperatures and especially large swings in climate regime related to changes in insolation modulated by Earth's orbital parameters (e.g., Lisiecki and Raymo, 2005, and references therein). Temperatures at Earth's surface during some interglacials were similar to or even somewhat higher than those of today; therefore, climatic conditions during those times can be used as imperfect analogs for the conditions predicted by climate models for the 21st century (Otto-Bliesner et al., 2006a; Goosse et al., 2007). A question to be addressed is to what extent was the Arctic sea ice reduced during these warm intervals. This issue is insufficiently understood because around the Arctic Ocean margins interglacial sedimentary records are fragmentary (e.g., CAPE, 2006), whereas, ocean sediments generally have low resolution and their dating is commonly problematic due to poor preservation of fossils and additional stratigraphic complications (e.g., Backman et al., 2004; Polyak et al., 2009). A better understanding has begun to emerge from recent collections of sediment cores from strategic sites (Jakobsson et al., 2000; Nørgaard-Pedersen et al., 2007a,b; O'Regan et al., 2008b; Adler et al., 2009).

The severity of ice conditions during glacial stages is indicated by the extreme rarity of biological remains and likely non-deposition intervals due to especially solid ice (Polyak et al.,

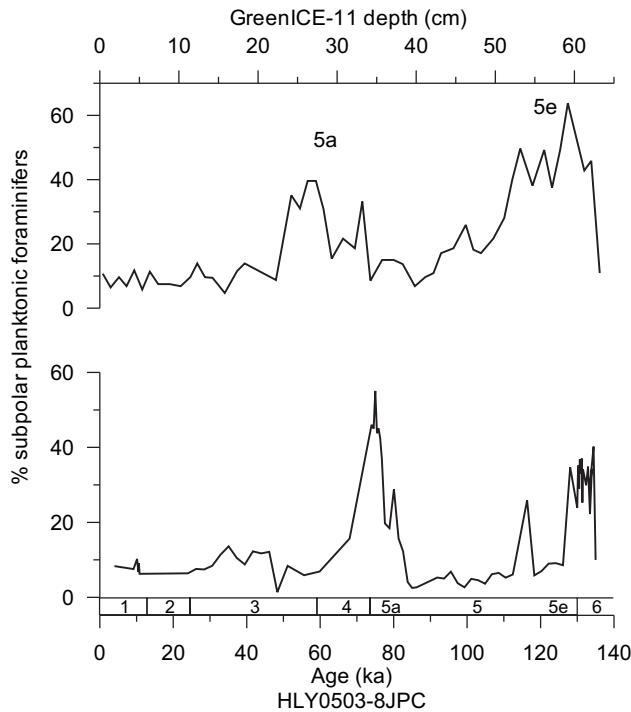




**Fig. 6.** Compilation of Neogene to Quaternary IRD and coarse-fraction records from the Nordic Seas and Arctic Ocean (see Fig. 3 for location). Note that the Pliocene–Quaternary boundary has been recently accepted by the International Commission on Stratigraphy as 2.6 Ma. (a) The counts of >150 µm terrigenous grains from ODP site 907. Data and age model from Jansen et al. (2000). (b) The >125 µm size fraction from ODP Leg 105 Site 643. Data from Wolf (2001) shown in yellow and Bruns and Dullo (2002) in orange. Age model and placement of hiatuses in the recovered record (wavy lines) are from Bleil et al. (1989). (c) The >125 µm size fraction from ODP Leg 151 Site 908. Data from Wolf-Welling et al. (1996) and Winkler (1999) shown in black and red, respectively. (d) The ACEX coarse-fraction record: >150 µm (green squares; St. John, 2008) and >125 µm (green circles; unpublished data courtesy of F. Eynaud). Age model from Backman et al. (2008), with angled lines drawn in to represent uncertainty in the depth of boundaries. (e) Stacked records from ACEX and neighboring cores: PS-2185-6 (blue; Spielhagen et al., 2004; Spielhagen et al., 2005), 96/12-IPC (brown; Jakobsson et al., 2001). Stratigraphic correlations are from O'Regan et al. (2008a) with chron boundaries from O'Regan et al. (2008b) and MIS stages from Spielhagen et al. (2004).

2004; Cronin et al., 2008; Polyak et al., 2009). Ice as thick as several hundred meters entirely or partially covering the Arctic Ocean during some glaciations has been inferred based on theoretical considerations (Bradley and England, 2008) or glaciogenic seafloor morphology on the submarine ridges and plateaus (Polyak et al., 2001; Jakobsson et al., 2008b). In contrast, interglacial and major interstadial intervals are characterized by higher marine productivity suggestive of reduced ice cover. The most prominent evidence is that planktonic foraminifers typical of subpolar, seasonally open

water lived in the central Arctic Ocean during the last interglacial (MIS 5e) as well as during a younger, relatively warm MIS 5a (Fig. 7, Nørgaard-Pedersen et al., 2007a,b; Adler et al., 2009). Given that one of these areas, north of Greenland, is presently characterized by especially thick and widespread ice, most of the Arctic Ocean may have been free of summer ice cover during these intervals. These conditions make sense in the context of high insolation intensity and elevated temperatures in the northern high latitudes during the last interglacial (CAPE, 2006; Axford et al., 2009), but the



**Fig. 7.** Planktonic foraminiferal records from (a) the Mendelev Ridge (HLY0503-8JPC; Adler et al., 2009) and (b) north of Greenland (GreenICE-11; Nørgaard-Pedersen et al., 2007b) (see Fig. 3 for location). 8JPC data is plotted vs age with MIS boundaries indicated; GreenICE data are plotted vs core depth with MIS5a and 5e noted. High numbers of subpolar planktonic foraminifera during MIS 5e (the last interglacial) and MIS 5a indicate warm temperatures and/or reduced-ice conditions. Foraminifera were counted in >63  $\mu\text{m}$  size fraction in GreenICE core and 75–150  $\mu\text{m}$  size in 8JPC.

comparably reduced ice at MIS5a is less expected and requires further investigation. This event exemplifies a complex relationship between sea ice and climate that may involve significant hydrographic and atmospheric controls.

Some coastal exposures of interglacial deposits, such as MIS 11 and 5e, 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 (Pelukian transgression) contain some fossils of species that are limited today to the northwest Pacific, whereas intertidal snails found near Nome, just south of the Bering Strait, suggest that the coast there may have been annually ice-free (Brigham-Grette and Hopkins, 1995; Brigham-Grette et al., 2001). On the Russian side of the Bering Strait, foraminiferal assemblages suggest that coastal waters were similar to those in the Sea of Okhotsk and Sea of Japan (Brigham-Grette et al., 2001). Deposits of the same age along the northern Alaskan Coastal Plain show that at least eight mollusk species extended their 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 the winter limit of sea ice did not extend south of Bering Strait and was probably located at least 800 km north of historical limits, whereas summer sea-surface temperatures were higher than present through Bering Strait and into the Beaufort Sea.

The time-series evaluation of a Late Quaternary sediment record from the Mendelev Ridge using sediment color lightness ( $L^*$ ) showed a combination of precessional ( $\sim 19$ – $20$  kyr) and 100-kyr spectra (Adler et al., 2009). Color lightness in this and correlative Arctic records co-varies with planktonic foraminiferal abundance

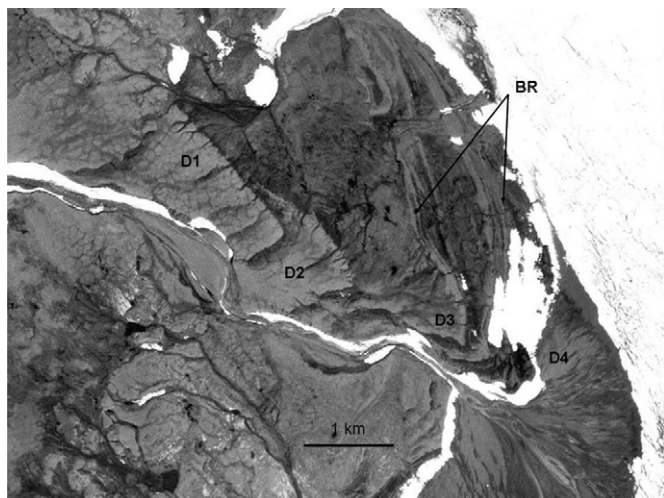
and stable-isotopic ( $\delta^{13}\text{C}$ ) composition and, thus, presumably with biological production and ice conditions, although the linkages need to be investigated further. Spectral results obtained on a longer ACEX record are dominated by the 100-kyr cyclicity after ca 1 Ma, and also show a possibility of a strong precessional control at some intervals (O'Regan et al., 2008b). While the 100-kyr signal is likely related to major Quaternary glaciations (e.g., Lisiecki and Raymo, 2005), a strong precession in the Arctic may be less expected and needs to be examined. The precessional control is commonly associated with low-latitude processes such as monsoons (e.g., Braconnot and Marti, 2003), but the Arctic also has a potential for its amplification through changes in sea ice and snow albedo and the poleward transport of heat and moisture (Jackson and Broccoli, 2003; Khodri et al., 2005). Hence, these precession-driven mechanisms may have affected sea-ice conditions as exemplified by the Holocene history below.

#### 4.3. The Holocene

##### 4.3.1. Long-term changes

The present interglacial that has lasted approximately 11.5 kyr is characterized by much more paleoceanographic data than earlier warm periods, because Holocene deposits are ubiquitous and technically accessible on continental shelves and along many coastlines. Multiple proxy records and climate models indicate that early Holocene temperatures were higher than today and that the Arctic contained less ice, consistent with a high intensity of orbitally-controlled spring and summer insolation that peaked about 11 ka and gradually decreased thereafter (e.g., Crucifix et al., 2002; Goosse et al., 2007). For example, the summer melt record of the Agassiz Ice Cap (north Ellesmere Island) indicates summer temperatures in the early Holocene were about  $3^\circ\text{C}$  above mid-20th century conditions (Koerner and Fisher, 1990; Fisher et al., 2006), although some of the cooling reflected in the mid-to-late Holocene record may have been contributed by glacioisostatic uplift of the site. Evidence of a warm early Holocene appears in many Arctic paleoclimatic records from areas where the cooling influence of the lingering Laurentide Ice Sheet was small (CAPE, 2001; Kaufman et al., 2004; Fisher et al., 2006). Coastal records indicate seasonally ice-free conditions as far north as the northern coasts of Svalbard and Greenland (Blake, 2006; Funder and Kjær, 2007), and temperatures  $4^\circ\text{C}$  warmer than in the 20th century are consistently reconstructed for the northernmost Siberian sites (Makeyev et al., 2003; Andreev et al., 2003, 2008, 2009). Areas that were affected by the prolonged melting of the Laurentide Ice Sheet, especially the northeastern sites in North America and the adjacent North Atlantic, show variously delayed thermal maximum (Kaufman et al., 2004).

Probably the most spectacular evidence of low-ice Arctic conditions in the early Holocene comes from Northeast Greenland (Fig. 8; Funder and Kjær, 2007; Funder et al., 2009). At this northernmost coast in the world, isostatically raised 'staircases' of well-developed wave-generated beach ridges investigated along a total coastline stretch of several hundred kilometers document seasonally open water as far north as  $83^\circ\text{N}$ . Further north, ridges are short and sporadic, restricted to mouths of embayments and valleys, which suggests that permanent sea ice persisted throughout the Holocene at the northernmost stretch of the coast, near  $83.5^\circ\text{N}$ . Overall, the zone of coastal melt was displaced about 500 km north of its present position. Large numbers of striated, apparently iceberg-raftered boulders in and on the marine sediments on this coast also indicate that it was not blocked by landfast ice, as it is now. Presently the entire Northeast Greenland coastline is permanently surrounded by pack ice with numerous pressure ridges and rare locked-in icebergs, with coastal melt occurring maximum to  $76^\circ\text{N}$ . Radiocarbon-dated mollusk shells from the beach ridges show that

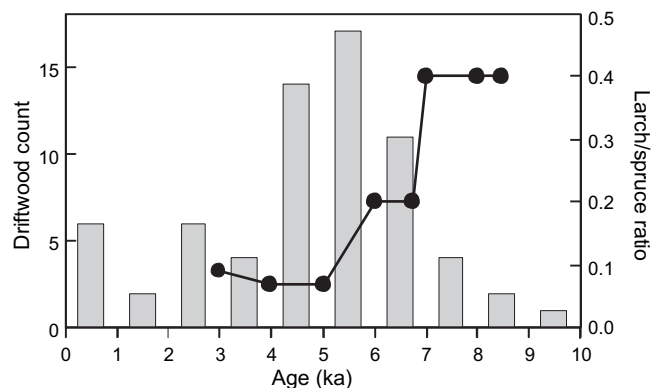


**Fig. 8.** Aerial photo of wave-generated beach ridges (BR) at Kap Ole Chiewitz, north-east Greenland (see Fig. 3 for location). D1–D4 are raised deltas. The oldest, D1, is dated about 10 ka whereas D4 is the modern delta; D3 is associated with wave activity between ca 8.5–6 ka.

they were formed in the early Holocene, within the interval between ca 8.5–6 ka. This interval is progressively shorter from south to north. As glacier margins in North Greenland retreated inland by the onset of the Holocene (Funder, 1989), meltwater discharge could not be a considerable factor in sea-ice displacement here, as it probably was in northern CAA (Dyke et al., 1996; Atkinson, 2009). Therefore, Northeast Greenland beach ridges are likely diagnostic for climatically controlled sea-ice conditions in the adjacent Arctic Ocean.

The evidence for more open water is further supported by the distribution of driftwood based on nearly 100 dated samples from Greenland coasts north of 80°N, which are directly affected by outflow from the Arctic Ocean (Funder et al., 2009). A minimum in driftwood fluxes in the early Holocene coincides with the period of beach ridge formation indicating a reduction in the amount of multi-year ice and, thus, limited possibilities for driftwood delivery. Highest numbers of driftwood characterize the period between ca 5.5 and 3 ka, probably marking an increase in multi-year sea ice, while a succeeding minimum between ca 3 and 1 ka may indicate a further barring of the coast by landfast ice, except for short-term maxima occurring in the last millennium. These results are consistent with the northern Ellesmere Island record indicating an abrupt termination in driftwood stranding along most of the north coast (>250 km) after 5.5 ka due to the development of exceptionally thick, multi-year landfast ice (ice shelves) with possibly only short intervals of lesser ice since then (England et al., 2008). The growth of these ice shelves indicates very high ice concentrations in the adjacent Arctic Ocean since ca 5.5 ka, which apparently expanded and ice-locked Northeast Greenland coast by ca 3 ka.

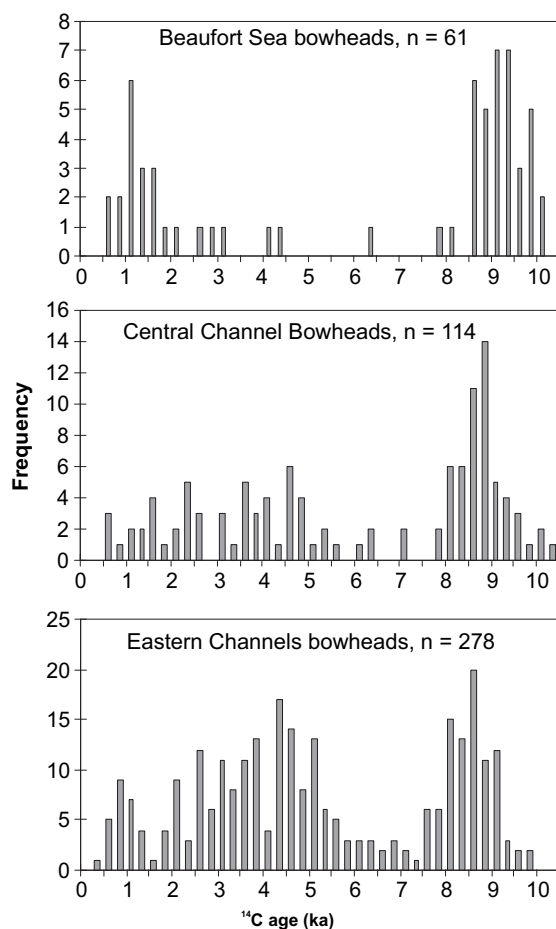
These records can be compared with the distribution of dated driftwood from raised marine beaches along the margins of Baffin Bay, where the maximum influx of driftwood occurred between ca 8 and 4.5 cal. ka (Blake, 1975; Dyke et al., 1997) (Fig. 9). This timing is generally consistent with the northern Greenland and Ellesmere Island records, although the interpretation for this region is more complex and includes both the sea ice and circulation factors. The ratio of larch (mainly from Russia) to spruce (mainly from north-west Canada) in this driftwood declines sharply about 8 cal. ka. This abrupt shift must have been primarily caused by changes in the intensity or trajectories of ice drift from the Arctic Ocean (Tremblay et al., 1997), while changes in the composition or extent of forests were probably less important as the boreal forests never contracted



**Fig. 9.** Distribution of  $^{14}\text{C}$  ages (boxes) and larch to spruce ratio (black line) of Holocene driftwood on the shores of Baffin Bay (data from Dyke et al., 1997). See Fig. 3 for location of major sites sampled. Larch to spruce ratio indicates Russian vs Canadian sources. Ages used in the text have been calibrated. [Reproduced by permission of Arctic Institute of North America].

enough during the Holocene so as to be removed from northward flowing circumpolar drainages.

In the CAA the Holocene record, composed mostly of driftwood and bowhead whale findings (Fig. 10; Dyke et al., 1996; Saville et al., 2000; Dyke and England, 2003; Fisher et al., 2006), is more complex, which may reflect both the uncertainties with proxy



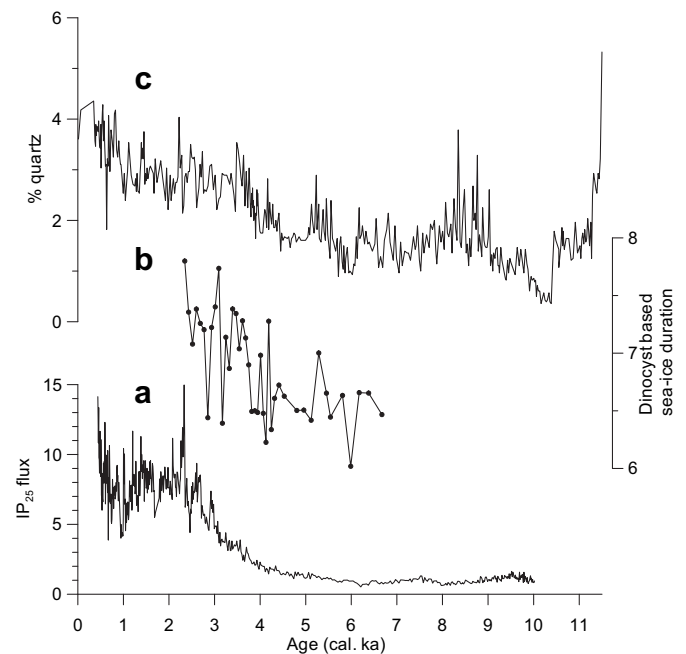
**Fig. 10.** Distribution of  $^{14}\text{C}$  ages of bowhead whales by regions of the Canadian Arctic Archipelago (data from Dyke et al., 1996; Saville et al., 2000). A standard oceanic  $^{14}\text{C}$  reservoir correction of  $-400$  years was applied. Ages used in the text have been calibrated. [Reproduced by permission of Arctic Institute of North America].



interpretations and the complex pattern of circulation and ice conditions in the straits (e.g., Howell et al., 2009). Bowhead bones are most commonly found in all three CAA regions in early Holocene deposits, 11–9 cal. ka. At that time Pacific and Atlantic bowheads were clearly able to intermingle freely along the length of the Northwest Passage, indicating at least periodically ice-free summers. Furthermore, a bowhead skull as old as nearly 12 ka (10.4  $^{14}\text{C}$  ka) was recently discovered in the northernmost part of the archipelago, 700 km north of the previously reported early Holocene range of the species (Atkinson, 2009). One explanation for the bowhead abundance in the early Holocene is that the warming from higher insolation may have been sufficient for the Northwest Passage to be clear of summer sea ice. However, it is also possible that other processes such as a different ocean circulation pattern and, especially, meltwater outflows from melting Laurentide and Innuitian ice sheets were forcing the early Holocene sea-ice clearance (Dyke et al., 1996; Atkinson, 2009). An interval from ca 9 to 5–6 ka contains fewer bones as well as low driftwood fluxes (Dyke et al., 1996, 1997), which appears to indicate higher ice coverage of CAA waters despite a relatively warm climate. A possible explanation is that ice may have been transported from the Arctic Ocean due to low formation of first-year ice in the straits, as has been commonly observed in recent years (Howell et al., 2009). Abundant bowhead bones dated between ca 6 and 3 ka have been found along the eastern channels (Fig. 10). At times during this interval (especially around 5 ka) the Atlantic bowheads penetrated the central region indicating that the passage on the Atlantic side was clear, at least briefly. The Pacific bowhead apparently did not extend its range at this time. The maximum influx of driftwood in the central CAA also occurred in the second half of the Holocene indicating a favorable circulation and considerable mobility of sea ice in the channels and shorelines at least periodically free of landfast ice (Dyke et al., 1997; Tremblay et al., 1997).

A more continuous reconstruction of ice conditions in central CAA is now available based on  $\text{IP}_{25}$ , a biomarker of ice-related diatom spring blooms (Vare et al., 2009). A downcore  $\text{IP}_{25}$  record from the central archipelago demonstrates little ice during the early Holocene, an accelerating increase in ice occurrence between 6 and 3 ka, and high but variable occurrence since then (Fig. 11a). This overall pattern is consistent with the temperature reconstruction from the summer melt record of the Agassiz Ice Cap (Koerner and Fisher, 1990). A similar pattern of ice conditions (duration of ice cover), although for a shorter time interval, was reconstructed from dinocyst assemblages in northern Baffin Bay (Fig. 11b: Levac et al., 2001; Ledu et al., 2008). Decreased sea-ice cover in the 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, as discussed in Section 3.3.2, there are questions regarding interpretation.

A somewhat different history of ice extent in the early Holocene emerges from the northern North Atlantic. The eastern Nordic seas and the adjacent Barents Sea show a well-developed early-Holocene warming (e.g., Moros et al., 2006), which has been slowed in the more western areas as a result of the discharge of glacial meltwater from the remains of the Laurentide Ice Sheet (Kaufman et al., 2004), as well as from potentially intensified export of ice from the Arctic Ocean (de Vernal et al., 2008). For example, a 12,000-yr record of quartz content in 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 (Moros et al., 2006). Results from this record, which has a resolution of 30 years per sample, are consistent with data obtained from 16 cores across the northwestern Iceland shelf (Andrews et al., 2009). These



**Fig. 11.** Comparison of proxy reconstructions of Holocene ice conditions in the Canadian Arctic, northern Baffin Bay, and on the Iceland shelf (see Fig. 3 for location). (a)  $\text{IP}_{25}$  fluxes indicating spring sea ice occurrence in the central CAA straits (Vare et al., 2009). (b) Duration of ice cover (months per year) in northern Baffin Bay based on dinocyst assemblages (Levac et al., 2001). (c) Percentage of quartz, proxy for drift ice on the northern Iceland shelf (Moros et al., 2006).

data show a minimum in ice cover at the end of deglaciation (10–11 ka), after which the area of ice increased and then reached another minimum around 6 ka before the content of quartz started to rise steadily; consistent with sea ice indicators from the Canadian Arctic (Fig. 11c).

In some areas of the Arctic, notably at the Chukchi Sea margin north of Alaska, Holocene sea-ice reconstructions from sediment cores also show a more complex long-term pattern than expected from insolation changes alone (de Vernal et al., 2005, 2008; McKay et al., 2008; Konno, 2009). These data suggest an increase in ice cover at the northern Chukchi margin in the early Holocene, somewhat similar to the North Atlantic. As inputs of Laurentide meltwater waned in the Chukchi Sea after ca 8.5–9 ka (Lundeen, 2005; McKay et al., 2008; Darby et al., 2009), the inferred sea-ice pattern may be related to the strengthening of the western Arctic cyclonic circulation combined with enhanced ice formation on the Arctic shelves (de Vernal et al., 2008). These factors would have intensified ice transport by the Transpolar Drift and, thus, affected ice conditions both in the western Arctic Ocean and in the North Atlantic. We note that closer to the Alaska Chukchi coast more open-water conditions occurred at the same time with higher ice further north, possibly under the influence of the Alaskan coastal current (Konno, 2009). This geographic inhomogeneity of the early Holocene ice cover emphasizes the complex pattern of Arctic ice response to climatic forcings.

In the central Arctic Ocean, no compelling record of Holocene ice conditions has been generated thus far because of low sedimentation rates and stratigraphic uncertainties. Some indications of low-ice conditions have been found in Holocene benthic assemblages (Cronin et al., 1995), but their age could not be determined accurately. New evidence of a retreated summer ice margin in the western Arctic Ocean potentially stems from elevated early Holocene sedimentation rates identified in some sediment cores (Polyak et al., 2009). More studies targeting higher-resolution cores and using multiple proxies are needed to

verify the distribution of ice in the Arctic during the warmest phase of the current interglacial.

#### 4.3.2. Suborbital variability

Owing to relatively high sedimentation rates at continental margins, paleoceanographic environments and ice drift patterns can be constructed on suborbital (millennial to decadal) scales from some sedimentary records. These high-resolution records reveal a considerable complexity in the system including forcings operating on suborbital scales (such as solar activity, volcanic eruptions, and atmospheric and oceanic circulation), changes in seasonality, and links with lower latitudes. Thus, the periodic, centennial-scale influx of large numbers of iron-oxide grains from the Siberian shelves to the northern Alaska margin has been linked to a reduced Beaufort Gyre and a shift in the Trans-Polar Drift toward North America (Darby and Bischof, 2004). Under present conditions such a shift occurs during a positive phase of the Arctic Oscillation (Rigor et al., 2002; Mysak, 2001). The finding that similar changes may have occurred on century to millennial-scales argues for the existence of longer-term atmospheric variability in the Arctic than the decadal Arctic Oscillation observed during the last century (Thompson and Wallace, 1998).

Variations in the volumes of IRD in the subarctic North Atlantic indicate several cooling and warming intervals during Neoglacial time (late Holocene), similar to the so-called “Little Ice Age” and “Medieval Warm Period” cycles of greater and lesser areas of sea ice known from the last millennium (Jennings and Weiner, 1996; Bond et al., 1997; Jennings et al., 2002; Moros et al., 2006). Southward polar-water excursions have been reconstructed at several sites in this region as multi-century to multidecadal-scale variations superimposed on the longer-term trends (Andersen et al., 2004; Giraudeau et al., 2004; Jennings et al., 2002). Millennial variations in ice conditions have also been suggested for the Barents Sea reflecting changes in atmospheric and oceanic interactions between the North Atlantic and the Arctic (Voronina et al., 2001).

Bond et al. (2001) concluded that peak volumes of Holocene drift ice at intervals of about 1500 years 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 are linked to centennial- to millennial-scale climate variations through effects on the North Atlantic thermohaline circulation, as supported by some numerical simulations (Braun et al., 2005; Dima and Lohmann, 2009). However, continued investigation of the drift-ice signal indicates that the variations reported by Bond et al. (2001) may not pertain to a simple index of drift ice (Andrews et al., 2009), and those cooling events prior to the Neoglacial interval may stem from deglacial meltwater forcing rather than from southward drift of Arctic ice (Jennings et al., 2002; Giraudeau et al., 2004). Furthermore, a comparison of Irish tree-ring chronologies and radiocarbon activity with Bond's drift-ice sequence found a dominant 800-yr cycle in moisture, reflecting atmospheric circulation changes but no clear link with solar activity (Turney et al., 2005).

In the Arctic centennial climate variability during the cooling of the last 2 kyr was more subdued than in the Northern hemisphere as a whole, although major temperature anomalies like the warmings around 1000 and 500 A.D. can be discerned (Kaufman et al., 2009). A final peak of bowhead bones appears to have culminated shortly prior to 1000 A.D. in the Beaufort Sea and somewhat later in the eastern CAA (Fig. 10), suggesting the possibility of temporarily open channels. This inference is consistent with the IP<sub>25</sub> record, which indicates a relatively decreased spring ice occurrence between ca 1.2 and 0.8 ka (800–1200 A.D.) (Fig. 11a; Vare et al., 2009). At the end of this time the bowhead-hunting Thule Inuit (Eskimos) expanded eastward out of

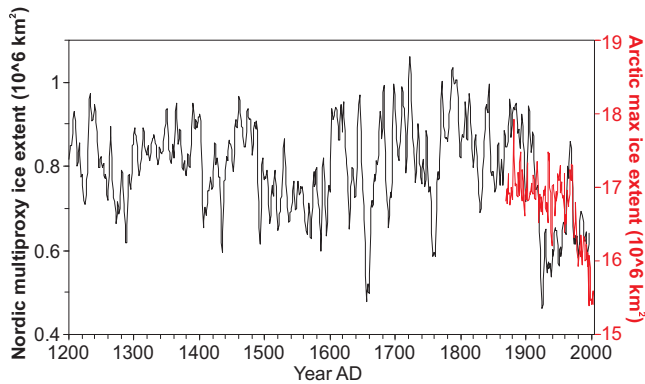
the Bering Sea region and ultimately spread to Greenland and Labrador (McGhee, 1984; Friesen and Arnold, 2008). The subsequent decline of bowhead abundances in the CAA is consistent with the abandonment of the high Arctic of Canada and Greenland by bowhead hunters, while Thule living in more southern Arctic regions increasingly focused on alternate food resources. The warming event around 1500 A.D. is identified by climatic simulations in the Atlantic sector of the Arctic and is explained by the internal variability of atmospheric circulation (Crespin et al., 2009). The subsequent cooling culminated in the “Little Ice Age”, between ca 1600 and 1850 AD, when ice conditions in the high-Arctic remained especially prohibitive for navigation (e.g., Alt et al., 1985).

Historical observations in the Nordic Seas since mid-18th century indicate multidecadal oscillations in ice extent superimposed on an overall trend of retreating ice margin (Divine and Dick, 2006). Similar oscillations, although with somewhat variable frequencies, are inferred for a longer, 800-yr period from an ice-core/tree-ring proxy record (Macias-Fauria et al., 2009). These multidecadal changes are probably related to variability in the North Atlantic thermohaline circulation (Polyakov et al., 2009 and references therein), but their mechanism is not well understood and may involve a combination of internal variability in the circulation with external factors such as solar and aerosol forcings. No record of similar oscillations prior to the 20th century is known from other parts of the Arctic, although most of the paleo-data series existing to date lack sufficient detail.

#### 4.4. Recent warming

Arctic paleoclimate proxies in lake and marine sediments, tree rings, and ice cores indicate that from the mid-19th century the Arctic not only warmed by more than 1 °C average in comparison with the “Little Ice Age” (Overpeck et al., 1997), but also reached the highest temperatures in at least the last two thousand years (Kaufman et al., 2009). This warming sharply reversed the long-term cooling trend that had likely been caused by the orbitally-driven decreasing summer insolation with the positive feedbacks from ice and snow albedo (e.g., Otto-Bliesner et al., 2006b). Subglacial material exposed by retreating glaciers in the Canadian Arctic corroborates that modern temperatures are higher than any time in at least the past 1600 years (Anderson et al., 2008). An even longer perspective for the outstanding magnitude of the modern warming and related ice loss is provided by the history of ice shelves at the northern coast of Ellesmere Island, which are made of super-thickened landfast ice supported by pack ice in the adjacent Arctic Ocean. These ice shelves have been stable for most of the last 5.5 kyr based on driftwood ages (England et al., 2008), but declined by more than 90% during the 20th century and continue to break at a notable rate (Mueller et al., 2008).

An unraveled magnitude and duration of modern sea-ice retreat on a millennial background has been reported for the Nordic Seas based on combined ice core and tree-ring proxy data from Svalbard and Scandinavia (Macias-Fauria et al., 2009). A comparison of this reconstruction with the Arctic-wide compilation of ice extent since the mid-19th century (Kinnard et al., 2008) shows a close match except for an obvious discrepancy in the early 20th century (Fig. 12). This discrepancy reflects the pronounced warming event in the Nordic Seas that was amplified by multidecadal variability of the North Atlantic circulation (Polyakov et al., 2009) and therefore affected primarily the Atlantic sector of the Arctic, somewhat similar to the 15th-century warming anomaly (Crespin et al., 2009). In contrast, a very close match between the Nordic Seas and Arctic-wide records of ice extent during the recent decades emphasizes the pan-Arctic nature of the modern ice loss.



**Fig. 12.** Comparison of a multi-proxy reconstruction of sea-ice extent in the Nordic Seas during 1200–1997 AD (black curve; Macias-Fauria et al., 2009) and maximum Arctic-wide ice extent during 1870–2003 (red curve; Kinnard et al., 2008). The discrepancy between the two records in the early 20th century corresponds to an increase in the Atlantic inflow to the Nordic Seas (e.g., Polyakov et al., 2009).

A climatic simulation by Sedláček and Mysak (*in press*) suggests that after about 1900 AD the slow increase in atmospheric greenhouse gas concentrations was the main driver of sea-ice changes in the Northern Hemisphere, while other forcings such as volcanic activity were mostly responsible for the thermodynamically produced changes in sea ice area and volume during the preceding four centuries. The remarkable modern warming and associated reduction in sea-ice extent are especially anomalous because orbitally-driven summer insolation in the Arctic has been decreasing steadily since its maximum at 11 ka, and is now near its minimum in the precession cycle (Berger and Loutre, 2004).

## 5. Summary

Reviewed geological data indicate that the history of Arctic sea ice is closely linked with climate changes driven primarily by greenhouse and orbital forcings and associated feedbacks. This link is reflected in the persistence of the Arctic amplification, where fast feedbacks are largely controlled by sea-ice conditions (Miller et al., 2010). Based on proxy records, sea ice in the Arctic Ocean appeared as early as 47 Ma, after the onset of a long-term climatic cooling that followed the Paleocene–Eocene Thermal Maximum and led to formation of large ice sheets in polar areas. Year-round ice in some parts of the Arctic Ocean possibly developed as early as at least 13–14 Ma, in relation to a further cooling in climate and, possibly, to the establishment of the near-modern hydrographic circulation in the Arctic (opening of deep-water connection with the Atlantic). Nevertheless, extended periods of reduced ice likely occurred until the development of large-scale Quaternary glaciations in the Northern Hemisphere after approximately 3 Ma. The onset of these glaciations was likely followed by an increase in the extent and duration of sea ice, enabled by the cooling and freshwater build-up.

Ice was probably less prevalent during Quaternary interglacials and major interstadials, and the Arctic Ocean may even have been seasonally ice-free during some of these times. There is some evidence that ice conditions may be precessionally controlled, possibly due to amplification of climate change via albedo feedback and changes in meridional heat and moisture transport. Especially mild ice conditions in the Late Quaternary are inferred for MIS 5e and 5a and the onset of the current interglacial, about 130, 75, and 10 ka, respectively. However, the controls on these ice reductions were clearly more complex than orbital variations alone. This is indicated for example by a considerable geographic inhomogeneity in ice

distribution during the early Holocene, revealed by a better data set than is available for older warmings. Despite many uncertainties remaining, the existing data demonstrate that the Quaternary low-ice periods can provide critical information for understanding seasonally ice-free conditions expected to evolve through the 21st century.

On suborbital time scales, ice distributions varied in the Holocene, but no evidence exists for large, pan-Arctic fluctuations. Historical records indicate that Arctic sea-ice extent has been declining since the late 19th century. Although this decline was accompanied by multidecadal oscillations, the accelerated ice loss during the last several decades lead to conditions not documented in at least the last few thousand years. Taking together the magnitude, wide geographic distribution, and abruptness of this ice loss, it appears to be anomalous in comparison with climatic and hydrographic variability observed on submillennial time scales and longer-term insolation changes.

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