The Pliocene is important in the geological evolution of the high northern latitudes. It marks the transition from restricted local- to extensive regional-scale glaciations on the circum-Arctic continents between 3.6 and 2.4 Ma. Since the Arctic Ocean is an almost land-locked basin, tectonic activity and sea-level fluctuations controlled the geometry of ocean gateways and continental drainage systems, and exerted a major influence on the formation of continental ice sheets, the distribution of river run-off, and the circulation and water mass characteristics in the Arctic Ocean. The effect of a water mass exchange restricted to the Bering and Fram Straits on the oceanography is unknown, but modelling experiments suggest that this must have influenced the Atlantic meridional overturning circulation. Cold conditions associated with perennial sea-ice cover might have prevailed in the central Arctic Ocean throughout the Pliocene, whereas colder periods alternated with warmer seasonally ice-free periods in the marginal areas. The most pronounced oceanographic change occurred in the Mid-Pliocene when the circulation through the Bering Strait reversed and low-salinity waters increasingly flowed from the North Pacific into the Arctic Ocean. The excess freshwater supply might have facilitated sea-ice formation and contributed to a decrease in the Atlantic overturning circulation.

Keywords: Arctic Ocean; Pliocene; palaeoceanography; glaciation history; tectonic activity; climate optima

1. Introduction

The Pliocene in the high northern latitudes was a time of considerable change in climate and environment. Superimposed on the long-term cooling since the Mid-Miocene (e.g. White et al. 1997, 1999; Bezrukova et al. 1999; Demske et al. 2000), distinct excursions to warmer climates structured this relatively

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short geological epoch in the Arctic region (Matthews & Ovenden 1990; Brigham-Grette & Carter 1992; Cronin et al. 1993; Funder et al. 2001; Elias et al. 2006). A relatively warm Early Pliocene with small-scale glaciations was succeeded by a progressive intensification in the Mid-Pliocene, leading to the build-up of large ice sheets in North America and Eurasia (e.g. Fronval & Jansen 1996; Mudelsee & Raymo 2005). Since the Arctic Ocean was in the Pliocene, as it is today, an almost land-locked basin with mostly small and shallow conduits to the World Ocean except for the deep Fram Strait that connects the eastern Arctic Ocean and the North Atlantic Ocean via the Nordic Seas (figure 1).

Figure 1. Bathymetry of the Arctic Ocean (Jakobsson et al. 2000) and major surface currents (black: warm currents; white: cold currents). The summer (thin white line) and winter (thick white line) extent of sea-ice cover is shown (AB, Arctic Basin; BB, Baffin Bay; BS, Barents Sea; BES, Bering Strait; BMB, Beaufort–Mackenzie Basin; CAA, Canadian Arctic Archipelago; ESS, East Siberian Sea; FS, Fram Strait; GSR, Greenland–Scotland Ridge; KS, Kara Sea; LS, Labrador Sea; LAS, Laptev Sea; NS, Nordic Seas; NC, Norwegian Current; EGC, East Greenland Current). The locations of shallow-marine sediments that partly record warmer water conditions and ODP holes are shown (1, North Alaska Coastal Plain, Brigham-Grette & Carter (1992) and Cronin et al. (1993); 2, Beaufort–Mackenzie Basin, McNeil et al. (2001); 3, Meighen Island, McNeil (1990); 4, Hvitland Beds, Fyles et al. (1998); 5, Kap København, Funder et al. (2001) and Cronin et al. (1993); 6, ODP Hole 911A; 7, ODP Hole 909C; 8, Île de France, Bennike et al. (2002); 9, Lodin Elv, Feyling-Hanssen et al. (1983) and Funder (1989); 10, Baffin Island (Clyde Foreland, Qivituq Peninsula), Feyling-Hanssen (1985) and Cronin et al. (1993)).
changes in the geometry of the Arctic gateways by sea-level fluctuations and
tectonic activity must have exerted a strong influence on the palaeoenviron-
mental evolution.

Furthermore, the hydrological cycle of the Arctic region is characterized by a
large input of freshwater to the Arctic Ocean by river run-off and import through
the Bering Strait, which, together with net precipitation, lead to a low-salinity
surface water layer (Serreze et al. 2006). Heat loss at the sea surface results in the
formation and maintenance of the sea-ice cover. The formation of a halocline
below the mixed layer insulates the saline and warmer intermediate waters from
the atmosphere and thus supports the sea-ice cover (e.g. Rudels et al. 1996).
Freshwater is exported through the gateways and may influence the Atlantic
meridional overturning circulation (AMOC) by altering deep convection in the
North Atlantic Ocean (Dickson et al. 2007).

Despite the undisputed role of the Arctic Ocean and subarctic seas in the
modern and Pliocene climate system (Driscoll & Haug 1998; Haug et al. 2001,
2005), the Arctic has attracted much attention only in the past few years since
the public has become aware of the ongoing fundamental change in the Arctic
cryosphere as a possible response to global warming, which will have a major
impact on ecosystems and human society (e.g. IPCC 2007). A number of Ocean
Drilling Program (ODP) holes have been drilled in the marginal Arctic Ocean
and subarctic seas in the past 25 years, recently complemented by the first
scientific drill holes in the central Arctic Ocean (IODP Expedition 302, ACEX—
‘Arctic Coring Expedition’, Backman et al. 2006), but these datasets have rarely
been summarized to comprehensively reconstruct the Pliocene palaeoclimate and
palaeoceanographic evolution of the high northern latitudes (Mudie et al. 1990).
Therefore, we attempt to give an overview on the current knowledge of Pliocene
palaeoceanography of the Arctic Ocean and subarctic seas as well as the major
external forcing factors (tectonic activity, freshwater supply) that may have
influenced water mass characteristics, circulation and sea-ice cover.

2. Pliocene chronostratigraphy

The chronostratigraphy of Pliocene sediments in the high northern latitudes is
principally based on magnetostratigraphy supported by age datums of various
microfossil groups. Stable isotope records on benthic and planktic foraminifers
play only a subordinate role owing to strongly discontinuous records (Fronval &
Jansen 1996; Haug et al. 2005). In the Atlantic sector of the high northern
latitudes, calcareous nanofossils and planktonic foraminifers and to a small
extent diatoms and radiolarians provide the basic age control, whereas in the
North Pacific Ocean biosiliceous microfossils are more important. Stratigraphic
resolution of calcareous and biosiliceous microfossils decreases strongly with
increasing latitude, whereas palynomorphs have a larger although not fully
explored stratigraphic potential in the Arctic Ocean (e.g. Eldholm et al. 1989;
Srivastava et al. 1989; Rea et al. 1995; Thiede et al. 1996; Raymo et al. 1999;
Backman et al. 2006).

The chronostratigraphy of central Arctic Ocean (CAO) sediments has been
intensively debated for more than 30 years (see Backman et al. (2004) for a
review). Pliocene sediments could not unequivocally be identified until the IODP
Expedition 302 drilled on the Lomonosov Ridge in the CAO (Backman et al. 2006, 2008). This new chronostratigraphy confirms the conclusion by Backman et al. (2004) that average sedimentation rates in the CAO were an order of magnitude higher (cm instead of mm ka\(^{-1}\)) than previously assumed in many studies. Therefore, the Pliocene age of sediments that has been inferred from millimetre-scale sedimentation rates in sediment cores obtained prior to IODP Exp. 302 must be revised, requiring also a re-evaluation of palaeoceanographic interpretations (e.g. Aksu & Mudie 1985; Herman et al. 1989; Scott et al. 1989; Mudie et al. 1990; Clark 1996; Spielhagen et al. 1997; Grantz et al. 2001). Therefore, the reconstruction of the Pliocene palaeoceanographic history of the CAO is based on a single site that has been dated by \(^{10}\)Be stratigraphy and a few biostratigraphic data (Backman et al. 2008; Frank et al. 2008).

Shallow-marine sediments of Pliocene age that are exposed on the circum-Arctic continents provide a wealth of palaeoenvironmental information, but absolute age assignments are as critical as in the CAO (e.g. Brigham-Grette & Carter 1992; Cronin et al. 1993; Harrison et al. 1999; Funder et al. 2001; McNeil et al. 2001; Polyakova 2001; Bennike et al. 2002; figure 1). These sequences comprise rather short periods, and radiometric and/or isotopic ages that are required to place these records into an absolute chronology are rare. Biostratigraphic datums are rarely well calibrated as, for example, illustrated by the stratigraphic range of the benthic foraminifer *Cibicides grossus* that is widespread during sea-level highstands in the Arctic region (Fyles et al. 1998; McNeil et al. 2001). This species has been used as a stratigraphic marker for the Mid- to early Late Pliocene (2.4–3.5 Ma) in a number of shallow-water sequences (McNeil et al. 2001), but it ranges into the Early Pleistocene in deeper-water sediments in the North Sea (King 1989) and on the Yermak Plateau (Osterman 1996). A younger last occurrence in deeper-water sediments is supported by new \(^{87}\)Sr/\(^{86}\)Sr data from the Beaufort–Mackenzie Basin, giving a stratigraphic range of 2–1 Ma (McNeil et al. 2001). Despite chronological uncertainties, some sequences may have been formed at approximately the same time (Matthews & Ovenden 1990; McNeil 1990; Brigham-Grette & Carter 1992; Cronin et al. 1993; Fyles et al. 1998; Funder et al. 2001; Elias et al. 2006; figure 2).

In this review, the stratigraphic concept of Gradstein et al. (2004) is used but a proposed revision of the stratigraphic status of the Quaternary may lead to a much shorter Pliocene epoch (http://www.stratigraphy.org).

3. Palaeoenvironmental boundary conditions in the Arctic region

The exchange of water masses between the Arctic Ocean and the World Ocean is strongly controlled by the geometry of the conduits, and, in particular, tectonic activity and fluctuating sea level may have altered transport rates of shallow- and deep-water masses between the different basins. Furthermore, tectonic uplift may have influenced the geometry of drainage basins and thus the distribution of river run-off and the build-up of large continental ice sheets that supplied sediments to the continental margins. Since these boundary conditions played a more important role for the palaeoclimate and palaeoceanography in the Arctic region than for any other ocean basin in the Pliocene, they are described in the following sections.
The topography of the circum-Arctic continents and the geometry of the Arctic Ocean and its gateways to the North Pacific Ocean, the Nordic Seas, the Labrador Sea and Baffin Bay have been changed frequently by the complex interaction between tectonic activity and sea-level fluctuations in the Pliocene. The Arctic Basin is particularly sensitive to sea-level fluctuations owing to the extensive shelf seas (mostly shallower than 200 m) that make up more than 50 per cent of the area.

(a) The Arctic Basin in the Pliocene

The topography of the circum-Arctic continents and the geometry of the Arctic Ocean and its gateways to the North Pacific Ocean, the Nordic Seas, the Labrador Sea and Baffin Bay have been changed frequently by the complex interaction between tectonic activity and sea-level fluctuations in the Pliocene. The Arctic Basin is particularly sensitive to sea-level fluctuations owing to the extensive shelf seas (mostly shallower than 200 m) that make up more than 50 per cent of the area.
modern Arctic Ocean (Jakobsson et al. 2003). A falling sea level will expose huge shelf areas in the Siberian Arctic and will close the shallow gateways through the Canadian Arctic Archipelago and the Bering Strait. Eustatic sea level varied on average by $\pm 20$ m around the modern level until 3.3 Ma, when a trend to consistently lower maxima and minima commenced (figure 2; Miller et al. 2005).

The deep Arctic Ocean and adjacent seas have experienced little change in their geometry since the Mid-Miocene except for a minor expansion and deepening of the Fram Strait and the continuous sea-floor spreading in the Eurasian Basin and the Nordic Seas (Kristoffersen 1990; Lawver et al. 1990; Jakobsson et al. 2007). The history of the shallow-water connections to the Atlantic and Pacific Oceans and the development of their geometry are less known (figure 2). The Bering Strait possibly opened at the Miocene–Pliocene transition ca 5.5–5.4 Ma (Gladenkov et al. 2002; Gladenkov 2006), although there is evidence from diatom analysis for intermittently open connections since the Early Miocene (Polyakova 2001). Flooding of the Bering Strait was probably facilitated by large-scale tectonic events and/or eustatic sea-level rise (Marincovich 2000). The Barents shelf was possibly subaerially exposed in the Pliocene (Torsvik et al. 2002) and major parts probably subsided below sea level ca 1 Ma (Butt et al. 2002). The sill depth of the Greenland–Scotland Ridge has changed with time due to a variable activity of the Iceland plume (Poore et al. 2006). Shallow conduits through the Canadian Arctic Archipelago might not have existed in the Pliocene. Harrison et al. (1999) assumed that Mid- to Upper Pliocene deposits in the Canadian Arctic form a single depositional sequence. A drainage system that cut down through the regionally extensive Pliocene sediment cover might have developed in the Early Pleistocene. Fluvio-deltaic to deltaic–shallow-marine depositional environments prevailed along the Canadian Arctic continental margin (Dixon et al. 1992; Torsvik et al. 2002). Therefore, the Bering and Fram Straits might have been the only Pliocene gateways to the Arctic Ocean.

Both tectonic activity and glacial erosion shaped the topography of the circum-Arctic continents and influenced the atmospheric circulation in the Pliocene (e.g. White et al. 1997; Zhisheng et al. 2001; Duk-Rodkin et al. 2004; Praeg et al. 2005; Stoker et al. 2005). Tectonic and glacial activity, sea-level change and bottom currents caused substantial hiatuses at the continental margins (e.g. Bohrmann et al. 1990; Wolf & Thiede 1991; Brigham-Grette & Carter 1992; Solheim et al. 1998; McNeil et al. 2001; Geissler & Jokat 2004; Dahlgren et al. 2005; Praeg et al. 2005; Stoker et al. 2005). McNeil et al. (2001) described a regional unconformity in the Beaufort–Mackenzie area of Arctic Canada encompassing the period between ca 5.2 and 3.5 Ma that ‘is probably recognizable on a circum-Arctic scale and beyond’. The unconformity extends from the continental rise to the hinterland of the Beaufort–Mackenzie Delta and was initiated by shelf exposure during the latest Late Miocene to earliest Pliocene eustatic sea-level lowstand, further accentuated by tectonic uplift of the basin margin. In the Siberian Arctic, the base Pliocene unconformity in the Laptev Sea was probably caused by global sea-level fall at the Miocene–Pliocene transition. The transgression towards the end of the Pliocene reached the present shoreline and led to deposition of shallow-marine sediments (Drachev et al. 1998; Kos’ko & Trufanov 2002). In the Late Pliocene to Early Pleistocene, large coastal regions of the Eurasian Arctic were flooded due to tectonic subsidence (Polyakova 2001).
Japsen & Chalmers (2000) compiled evidence for tectonic uplift along the continental margins of West and East Greenland and Norway in the Neogene but the timing of these events is partly controversial. Along the continental margins of the Nordic Seas, Upper Pliocene sediments usually rest unconformably on Upper Miocene or on a thin layer of Lower Pliocene sediments probably due to erosion of Lower Pliocene sequences during an extensive global sea-level fall (e.g. Eidvin et al. 2000; Ryseth et al. 2003; Hjelstuen et al. 2004; Tsikalas et al. 2005). The Intra-Pliocene unconformity forms the base of large prograding sedimentary wedges between Rockall Trough and the Voring Plateau (e.g. Dahlgren et al. 2005; Praeg et al. 2005; Stoker et al. 2005). Tilting from the Early Pliocene (<4 Ma ± 0.5 Ma) to recent resulted in a basinward progradation of shelf-slope wedges, from uplifts along the inner margin and from offshore highs. The extensive development of the continental slope wedges in the Late Pliocene was controlled by tectonic and/or climate evolution (Dahlgren et al. 2005; Praeg et al. 2005; Stoker et al. 2005).

The completeness of the stratigraphic record in the CAO cannot be assessed with the available data. Since IODP Exp. 302 drilled on the relatively shallow Lomonosov Ridge in water depths of less than 1300 m and sedimentation rates were relatively low throughout the Neogene sequence (Backman et al. 2006, 2008), Moore et al. (2006) expected to find other hiatuses than the large Eocene–Miocene hiatus. Subsequently, Frank et al. (2008) described a short-term hiatus in the Late Miocene. Numerous reflectors have been identified in the Pliocene section of a synthetic seismic model, but they could not be related to hiatuses due to the absence of distinct changes in the lithology and the present low stratigraphic resolution (Backman et al. 2008), suggesting a rather continuous sedimentation in the Pliocene. However, local sea-floor erosion by grounding icebergs has removed Pliocene and Pleistocene sediments on the Lomonosov Ridge (Frank et al. 2008).

(b) Glaciation history of the circum-Arctic continents

Large-scale glaciations probably had an asymmetric geographical distribution in the circum-Arctic regions in the Pliocene similar to the Late Pleistocene because Eastern Siberia has only been covered by locally restricted mountain glaciations rather than by large ice sheets (e.g. Prokopenko et al. 2001; Glushkova & Smirnov 2007).

The Miocene–Pliocene transition is marked by the onset of small-scale glaciations in the Northern Hemisphere at ca 7.2–6.0 Ma that persisted until ca 3 Ma (Fronval & Jansen 1996). St. John & Krissek (2002) assumed that glaciations on Greenland started ca 7.3 Ma and glaciers were large enough to reach sea level in southeastern and western Greenland (see also Solheim et al. 1998). In Baffin Bay and the Labrador Sea, sporadic ice rafting has been dated at 8–7 Ma (Korstgård & Nielsen 1989; Wolf & Thiede 1991). In ODP Hole 909C from the Fram Strait an increase in the accumulation (figure 3) and a change in the composition of the coarse fraction between 7.5 and 6.9 Ma is interpreted as indicating a strengthening of the Northern Hemisphere glaciations (NHGs; Winkler et al. 2002).

Glaciations were probably not widespread on the circum-Arctic continents in the Early Pliocene. Geirsdóttir (2004) stated that locally restricted glacial deposits are as old as 5 Ma but local ice caps may have already existed in the

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Miocene. Debris flow deposits and glacial diamictons in ODP Holes 918 and 987 off east Greenland suggest a larger-scale glacial advance across the Greenland shelf in the Early Pliocene (Solheim et al. 1998). In southeastern Alaska, tidewater glaciers were probably present since the Late Miocene but definitely occurred since 4.2–4.3 Ma (Krissek 1995; Rea & Snoeckx 1995; Lagoe & Zellers 1996; Duk-Rodkin et al. 2004). A Cordilleran ice sheet covered southeastern Alaska and small local ice caps developed in northwestern Canada (Barendregt & Duk-Rodkin 2004). In the North Pacific Ocean, ice-rafted debris (IRD) records suggest minor glaciations in both Alaska and Kurile/Kamchatka during the Early and Mid-Pliocene (Krissek 1995).

Both terrestrial and marine records provide consistent evidence for the intensification of continental glaciations in the Northern Hemisphere since the Mid-Pliocene. This transition was probably not abrupt but rather gradual between 3.6 and 2.4 Ma (Krissek 1995; Rea & Snoeckx 1995; Lagoe & Zellers 1996; Duk-Rodkin et al. 2004). A Cordilleran ice sheet covered southeastern Alaska and small local ice caps developed in northwestern Canada (Barendregt & Duk-Rodkin 2004). In the North Pacific Ocean, ice-rafted debris (IRD) records suggest minor glaciations in both Alaska and Kurile/Kamchatka during the Early and Mid-Pliocene (Krissek 1995).

Figure 3. Ratio of the clay mineral groups smectite versus illite and chlorite (lower curve) and accumulation rates (acc. rate) of coarse fraction (upper curve) in Hole 909C from the Fram Strait in the Late Miocene and the Pliocene (Winkler et al. 2002). The shaded areas indicate the NHG and the intensification of small-scale glaciations in the Late Miocene (LMG).

Nevertheless, the available data suggest a somewhat coeval development of major ice sheets around the Arctic Basin since the late Mid-Pliocene. A compilation of North Atlantic Ocean and Nordic Seas IRD records by
Kleiven et al. (2002) shows temporal and spatial differences in initial supply from the circum-North Atlantic ice sheets since 3.6 Ma, but fluctuations of all major ice sheets were synchronous since 2.75–2.72 Ma. This coincides with the onset of large-scale IRD deposition in the North Pacific Ocean at 2.73 Ma (Haug et al. 2005). Geirsdóttir (2004) compiled evidence for the initial development of a major ice sheet on Iceland at ca 2.9 Ma, and subsequent intensifications at ca 2.7 Ma and from 2.4 to 2.5 Ma. The Lodin Elv and Kap København formations contain the earliest albeit circumstantial evidence for glaciations in east Greenland at ca 2.4–2.5 Ma (Funder 1989; Funder et al. 2001). Marine IRD records suggest an intensification of glaciations on eastern and southern Greenland slightly earlier at ca 3 Ma (Wolf & Thiede 1991; Kleiven et al. 2002; St. John & Krissek 2002) that may have resulted in an ice-sheet advance onto the East Greenland Shelf (Vanneste et al. 1995). The size of the Greenland ice sheet might have fluctuated considerably, being reduced to small mountainous ice caps in southeast Greenland during interglacials (Funder et al. 2001).

There is only marine evidence for an intensification of glaciations on Scandinavia and in the Barents Sea at 2.8–2.5 Ma, but these were probably of intermediate size until the Mid-Pleistocene (e.g. Mangerud et al. 1996; Solheim et al. 1998). The first ice advances to the coast and onto the shelf occurred locally since ca 2.8–2.7 Ma (Fronval & Jansen 1996; Eidvin et al. 2000; Knies et al. 2002; Geissler & Jokat 2004). The centre of glaciations was possibly located farther north and maximum glacial erosion occurred in the Barents Sea in the Pliocene (Mangerud et al. 1996). The intensification of glaciations on Fennoscandia and the Northern Barents Sea is marked by lithological changes in marine records and increased accumulation of terrigenous sediments (Henrich et al. 1989; Wolf & Thiede 1991; Knies et al. 2002; Hjelstuen et al. 2004).

Thus, relatively high sand contents in Hole 911A reflect the advance of the northern Barents Sea ice sheet onto the outer shelf since 2.7 Ma (figure 4), coeval with a substantial increase in large IRD (Knies et al. 2002). The provenance of sediments abruptly shifted from a distal to a proximal source at 2.4 Ma because high magnetic susceptibility and clay mineral group ratio S/(I+C) values are associated with a particular source region in the hinterland of the Kara Sea (Stein et al. 2004). The clay mineral group ratio indicates that sediment composition changed from a distal Westsiberian (smectite) to a proximal Spitsbergen provenance (illite, chlorite) at Site 911 on Yermak Plateau. A comparable trend at Site 909 in the Fram Strait with a decrease of the clay mineral group ratio and an increase of coarse fraction accumulation rates during the intensification of NHG has been attributed to increased glacial weathering and erosion on the adjacent continents (figure 3; Winkler et al. 2002). This implies that sediment supply from the adjacent ice sheets on northeastern Greenland and in the northwestern Barents Sea almost simultaneously increased since 2.3–2.4 Ma.

Duk-Rodkin et al. (2004) reconstructed regional-scale Cordilleran and local ice caps in northern Alaska and the Beaufort–Mackenzie area between 2.6 and 2.9 Ma. IRD deposition started in the North Pacific at 2.73 Ma (e.g. Haug et al. 2005) and the provenance of IRD points to SE Alaska as one major source area (McKelvey et al. 1995). The Atlanta till in the mid-USA (2.41±0.14 Ma) is possibly the oldest direct evidence of continental glaciations in the Northern Hemisphere and was nearly the most southerly advance of the Laurentide Ice Sheet (Balco et al. 2005).
The intensification of glaciations coincided with tectonic uplift since ca. 4 Ma that, in conjunction with glacial erosion, ultimately led to the transfer of huge amounts of sediments to the continental margins and the formation of sedimentary wedges and deep-sea fans (e.g. Arthur et al. 1989; Krissek 1995; Solheim et al. 1998; McNeil et al. 2001; Winkler et al. 2002; Duk-Rodkin et al. 2004; Geissler & Jokat 2004; Dahlgren et al. 2005; Praeg et al. 2005, Stoker et al. 2005). The uplifted land areas along the European continental margins might have been the topographic precondition for nucleation and growth of glaciations (e.g. Eyles 1996; Dahlgren et al. 2005; Tsikalas et al. 2005). The progradation and

Figure 4. The Mid-Pliocene to Lower Pleistocene section of ODP Hole 911A. The shaded areas mark the local Pliocene climate optimum and the change in provenance of sediments (PS) in the Late Pliocene. The age model is based on magnetostratigraphy and calcareous nanofossils (Thiede et al. 1996; Sato & Kameo 1996; Sato et al. 2004). (Sources of data: lithological units IA and IB, Thiede et al. 1996; seismic units YP3 and YP2, Geissler & Jokat (2004); magnetic susceptibility, Thiede et al. 1996; dinoflagellate cysts, Matthissen & Brenner (1996) and Knies et al. (2002); TOC, Stein & Stax (1996) and Knies et al. (2002); clay mineral group ratio smectite/(illite + chlorite), sm/(il+chl), and sand contents, Knies et al. (2002)). Circles, *Brigantedinium* spp.; squares, *F. filifera* s.l.; crosses, *O. centrocarpum*. 

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aggradation of shelves along Norway, the Barents Sea, Greenland, the Beaufort Sea and the North Pacific Ocean since the Early Pliocene was partly related to large-scale glaciations in these areas.

The onset of sedimentary wedge formation was not necessarily synchronous along the high-latitude continental margins such as those off northwest Europe, but a glacial influence on sedimentation is generally recorded since 2.74–1.6 Ma (Dahlgren et al. 2005). The upper part of prograding wedges in the Norwegian Sea younger than 1.5 Ma (Sejrup et al. 2005) record advances of ice sheets across the shelves. Shelf progradation probably occurred mainly during glacial periods in the vicinity of cross-shelf troughs where fast-flowing ice streams supplied sediments to submarine fans. Ice streams were important transport pathways along the Greenland, Norwegian and Arctic continental margins during the Last Glacial Maximum, but there are few indications of earlier activity in the Pliocene (e.g. Stokes & Clark 2001 and references therein; Geissler & Jokat 2004; Ottesen et al. 2005, 2007; Wilken & Mienert 2006).

Furthermore, the timing and interaction of tectonic uplift and erosion of topographic barriers influenced the availability and distribution of moisture from different source regions that might have controlled ice-sheet growth in coastal Alaska (Duk-Rodkin et al. 2004). Difference in timing and extent of glaciations in northwest Canada and east-central Alaska was probably related to proximity and size of moisture source (Duk-Rodkin et al. 2004). The renewed uplift in coastal southeast Alaska since 4 Ma resulted in a progressive barrier to Pacific moisture, resulting in increased continentality, and if the Arctic Ocean was ice free, it might have been a significant moisture source for continental glaciations (White et al. 1999; Duk-Rodkin et al. 2004).

The continuous tectonic uplift of mountain ranges in the Asian hinterland of the Arctic Ocean, such as the Tibetan Plateau, might have influenced Arctic climate in various ways. Zhisheng et al. (2001) compiled evidence to show that the atmospheric circulation system was strongly changed by uplift and extension of the Tibetan plateau in the Pliocene. Uplift of portions of the Tibetan Plateau and ice build-up were probably the driving forces behind long-term desertification in China with distinct increases at 3.6 and 2.6 Ma (Guo et al. 2004). The Late Pliocene uplift at Lake Baikal coincides with early glaciations in southeastern Siberia (Prokopenko et al. 2001).

A number of Late Pliocene continental glaciations have been recorded from northern Alaska, northern Canada, Iceland and Siberia, but extensive glaciations might only have occurred in these regions at the Pliocene–Pleistocene boundary (Vanneste et al. 1995; Prokopenko et al. 2001; Duk-Rodkin et al. 2004; Geirsdóttir 2004; Wilken & Mienert 2006). Since the formation of reflector R7 at the Barents Sea continental margin, which marks the intensification of NHG at 2.3–2.5 Ma, only moderate glaciations were recorded along the Norwegian continental margin until the Early Pleistocene (Henrich et al. 1989; Jansen & Sjøholm 1991; Fronval & Jansen 1996; Solheim et al. 1998; Hjelstuen et al. 2007). Between ca 3 and 1 Ma a high-frequency and low-amplitude variability of IRD deposition occurred on obliquity time scales, suggesting reduced glacial activity (stable ice margins) around the Nordic Seas (Bohrmann et al. 1990; Jansen et al. 2000). Along the European continental margin from Ireland to Svalbard ice sheets reached the shelf edge earlier in the north in the latest Pliocene and advanced to the continental slopes first since

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the Mid-Pleistocene while glacial advances were more frequent only in the past 0.5 Ma (Solheim et al. 1998; Knies et al. 2002, 2007; Geissler & Jokat 2004; Sejrup et al. 2005).

(c) Freshwater supply to the Arctic Ocean

In contrast to other ocean basins, freshwater supply by rivers and by inflow through the Bering Strait together with net precipitation plays a major role in the hydrological cycle (Serreze et al. 2006). Approximately 3200 km$^3$ are annually discharged by rivers into the Arctic Ocean, mainly from the Mackenzie, Lena, Yenisey and Ob watersheds (Serreze et al. 2006). Climate change and tectonic activity must have had a considerable impact on the geometry of catchment areas and on the volumes of freshwater and sediments discharged into the Arctic Ocean. Thus, Driscoll & Haug (1998) proposed that the intensification of the North Atlantic Drift since the earliest Pliocene, due to the closure of the Central American Isthmus, might have enhanced moisture delivery to Eurasia. This freshwater is supplied by the Siberian rivers to the Arctic Ocean and is further transported into the Nordic Seas and North Atlantic Ocean where it may have a major impact on deep-water formation.

Catchment areas and river systems differed from the modern configuration but the impact on distribution and rates of discharge cannot currently be evaluated. Discharge rates cannot be estimated from fluvial sediment supply because there need not be a linear relationship between fluvial supply and freshwater charge. Today, the major Arctic rivers differ strongly in their sediment load. Although the Mackenzie River transports the largest sediment load of all Arctic rivers its run-off is much smaller than that of Lena, Ob and Yenisei (Gordeev et al. 1996).

In the Barents Sea, fluvial systems might have existed during the early phase of the sedimentary wedge growth in the Early Pliocene (Solheim et al. 1998; Dahlgren et al. 2005). Siberia was probably drained to the Arctic Ocean in the Pliocene (Driscoll & Haug 1998 and references therein). Franke et al. (2001) stated that Late Miocene and Pliocene tectonic deformation associated with the Indo–Asian collision modified the drainage pattern across the Siberian craton, resulting in large freshwater inflow and influx of terrigenous sediment since the early Late Miocene. Prior to the uplift of the Primorsky Range north of Lake Baikal in the Pleistocene, the Lena drained a much larger area including Lake Baikal whereas today it is drained by the Yenisei (Mats et al. 2000).

In the Canadian Arctic, the modern Mackenzie River system is a relatively young feature existing since the Late Pleistocene (Duk-Rodkin & Hughes 1994). During the Pliocene, the area drained to the Arctic Ocean was less than 10 per cent of the Mackenzie drainage area today, suggesting that river run-off was probably much smaller in the Pliocene. In contrast to the modern drainage systems, the northern cordillera was the headwater of drainage systems that reached three oceans and a much larger area of Arctic Canada was drained to the east, in particular into the Labrador Sea (Duk-Rodkin & Hughes 1994). Greenland might have been an additional significant freshwater source for the Labrador Sea during interglacials when the ice sheet was absent and the island was drained to the west (Funder et al. 2001).

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The predominant siliciclastic sediments that have low biogenic carbonate, opal and marine organic carbon contents limit our ability to reconstruct changes in water mass characteristics, surface-water productivity and sea-ice cover of the Arctic Ocean (e.g. Thiede et al. 1996; Stein & Stax 1996; Backman et al. 2006). The interpretation of the records is further complicated by diagenetic processes leading, for example, to the formation of authigenic carbonates, manganese and iron oxides, and iron sulphides in the sediments (e.g. Thiede et al. 1996; Chow et al. 1996; Henrich et al. 2002; Expedition 302 Scientists 2006; Haley et al. 2008a; Krylov et al. 2008; St. John 2008). Sedimentological, petrographical and mineralogical proxies have not yet been fully explored, but recent studies demonstrate that they might be useful for palaeoceanographic reconstructions (e.g. Darby 2008; Krylov et al. 2008; St. John 2008).

Pliocene marine deposits have been recovered from the marginal Arctic Ocean and from isolated outcrops in the circum-Arctic hinterland. The shallow-marine sequences cover only short periods in the Pliocene (e.g. Yakhimovich et al. 1990; Brigham-Grette & Carter 1992; Mullen & McNeil 1995; Thiede et al. 1996; Fyles et al. 1998; Harrison et al. 1999; Funder et al. 2001; McNeil et al. 2001; Polyakova 2001; Bennike et al. 2002) and therefore sites drilled during ODP Legs 104, 105, 145, 151 and 162 in the subarctic seas still provide the majority of palaeoenvironmental information. Palaeoceanographic reconstructions in the high northern latitudes are furthermore biased to the period since the intensification of NHG in the Mid-Pliocene (e.g. Henrich et al. 2002; Haug et al. 2005). The Early Pliocene did not receive much attention probably owing to more equally warm climate conditions.

(a) Large-scale circulation in the high northern latitudes

The large-scale circulation in the Pliocene was similar to that of today. The modern Arctic estuarine circulation with a surface-water outflow and deep inflow through the Fram Strait was probably established in the Early Miocene (Jakobsson et al. 2007). Radioisotope and mineralogical studies suggest that the Transpolar Drift and possibly also the Beaufort Gyre have been persistent features since the Mid-Miocene (Darby 2008; Haley et al. 2008b; Krylov et al. 2008; St. John 2008). Interpretations differ with respect to the contribution of the different current systems to the circulation in the CAO, being dominated either by the Transpolar Drift (Krylov et al. 2008; Haley et al. 2008b) or by the Beaufort Gyre circulation (Darby 2008). In the Nordic Seas, relatively warm waters flowed northwards along the Scandinavian continental margin whereas colder waters were exported along the Greenland continental margin from the Arctic Ocean to the North Atlantic Ocean (Henrich et al. 1989, 2002; Bohrmann et al. 1990; Wolf & Thiede 1991; Fronval & Jansen 1996). Similarly, the North Atlantic Drift penetrated continuously into the Labrador Sea in the Pliocene (de Vernal & Mudie 1989).

The most significant change in the surface circulation occurred in the Mid-Pliocene. The circulation through the Bering Strait reversed from a southward flow between the opening at 5.5–5.4 and 3.6 Ma to a northward flow after 3.6 Ma as indicated by the migration of molluscs between the North Atlantic–Arctic Oceans and the North Pacific Ocean (Marincovich & Gladenkov 2001;
Gladenkov et al. 2002; Gladenkov 2006). The reversal in flow direction might have been an effect of the oceanographic changes associated with the closure of the Central American Seaway until 2.7 Ma, which may have further induced an intensification of the North Atlantic Drift and the formation of North Atlantic deep water (Haug et al. 2001). The reversal in flow direction is associated with the onset of a prominent sea-level fall that might have led to the short-term closure of the Bering Strait during marine isotope stages M2–MG2 at 3.3 Ma (figure 2).

The intermediate- and deep-water circulation was more strongly affected by the variable sill depths of the gateways (e.g. Bohrmann et al. 1990; Henrich et al. 2002; Poore et al. 2006). Neodymium isotopes indicate that Arctic intermediate waters originated from brine formation in the Eurasian shelf regions in the Pliocene rather than being advected from the North Atlantic Ocean as in the past 2 Ma (Haley et al. 2008a). The deep-water overflow from the Nordic Seas to the North Atlantic Ocean was enhanced in the Early and Mid-Pliocene (Henrich et al. 2002; Poore et al. 2006). This phase might have been linked to the southward surface-water flow through the Bering Strait. Since the NHG intensified and a northward flow through the Bering Strait was established, water-mass exchange between the North Atlantic Ocean and the Nordic Seas diminished (Henrich et al. 2002; Poore et al. 2006). Some deep-water formation might have been possible by brine rejection during sea-ice formation.

Apart from the Bering Strait and the Greenland–Scotland Ridge, the possible impact of a change in the cross-sectional area of other gateways on the circulation is not considered in palaeoceanographic reconstructions. Since the Barents Sea and the conduits through the Canadian Arctic Archipelago were probably closed in the Pliocene (Harrison et al. 1999; Butt et al. 2002; Torsvik et al. 2002), the configuration of the Arctic Basin and its gateways differed considerably from the modern situation. The Bering and Fram Straits as well as the Greenland–Scotland Ridge formed the major bottlenecks for the exchange of water masses with the adjacent North Pacific and Atlantic Oceans in the Pliocene. Sea-level change must have further led to a recurrent opening and closing of the shallow (less than 50 m) Bering Strait, and exposure and flooding of the large shallow Siberian shelves. Since the global sea level fluctuated by ±50 m relative to present in the Pliocene (figure 2; Miller et al. 2005), the geometry of the Bering Strait must have ranged from a closed state to a much larger cross-section than today if isostatic movements could be neglected. Short-term closure might have occurred during pronounced sea-level lowstands at 4.9, 4.0, 3.3 and 2.5 Ma (Miller et al. 2005), but sea level was often higher in the Early Pliocene than today, enabling an increased water-mass exchange. Since 3.6 Ma sea level steadily decreased and the Bering Strait might have been closed more frequently.

The results of modelling experiments suggest that the closure of Arctic gateways must have influenced the circulation and distribution of freshwater in the Arctic and North Atlantic Oceans (e.g. Butt et al. 2002; Wadley & Bigg 2002; Hu et al. 2007). On the basis of an Early Pleistocene geometry of the Arctic Basin with an open Canadian Arctic Archipelago but a closed Barents Sea, Butt et al. (2002) calculated an increased input of relatively warm Atlantic waters into the Arctic Ocean. The closed shallow seaways through the Canadian Arctic Archipelago prevent a southward flow of low-salinity Arctic surface waters to
Baffin Bay and the Labrador Sea (Wadley & Bigg 2002). The AMOC will increase due to increased deep-water formation in the Labrador Sea. At the same time, the increased freshwater export from the Arctic Ocean through the Fram Strait may suppress deep-water convection in the Nordic Seas. Closing of the Bering Strait may lead to increased surface salinities in the Arctic Ocean and adjacent Nordic Seas because low-salinity waters are transported northwards through the open Bering Strait, which probably cause a decrease of the strength of the AMOC (Shaffer & Bendtsen 1994; Wadley & Bigg 2002).

The response of the AMOC to freshwater perturbations is today strongly affected by the status of the Bering Strait (Hu et al. 2007). The AMOC recovers faster when the Bering Strait is open because this freshwater is then exported both to the South Atlantic and via the Arctic Ocean to the North Pacific Ocean. Freshwater discharge to the North Atlantic Ocean while the Bering Strait is closed will lead to a recirculation in the Arctic Ocean, an enhanced export of freshwater and sea ice and a prolonged period of recovery of the AMOC (Hu et al. 2007). Since glacial–interglacial cycles are coupled to sea-level fluctuations and thus either an open or closed Bering Strait, the contrasting model scenarios might correspond to long-lasting instabilities in the meridional circulation during glacials in contrast to a rather stable interglacial situations (De Boer & Nof 2004; Hu et al. 2007). Freshwater fluxes need not be continuous to prevent a recovery if the Bering Strait is closed, but an open Bering Strait requires a continuous flux to prevent deep convection from recovery. However, the exact sea-level history of the Bering Strait is unknown so it is not possible to derive the potential temporal variability during glacial–interglacial cycles.

(b) Variability of water masses in the high northern latitudes

The reconstruction of the Pliocene palaeoceanography of the Arctic Ocean and subarctic seas is strongly hampered by a low production and a variable preservation of planktonic and benthic microfossils. Thus, the stable oxygen and stable carbon isotope records of planktonic and benthic foraminifers contain large gaps in the high northern latitudes (e.g. Fronval & Jansen 1996; Haug et al. 2005). The discontinuous benthic isotope record from ODP Sites 642/644 (Vøring Plateau, eastern Nordic Seas) is the exception to the rule and reflects distinct glacial excursions in the past ca 6.5 Ma. Large fluctuations indicating long-lasting changes in either global ice volume or bottom-water temperature or both are recorded between 6.5 and 2.5 Ma (Fronval & Jansen 1996). The gradual cooling since the Late Miocene is also reflected in various planktonic and benthic microfossil records (e.g. Locker & Martini 1989; Mudie et al. 1990; McNeil et al. 2001), whereas in the Labrador Sea and Baffin Bay relatively cold conditions prevailed throughout the Pliocene (Aksu & Kaminski 1989; de Vernal & Mudie 1989). A pronounced expansion of cold surface waters is reflected by the southward migration of warm-adapted calcareous nanofossils in the North Pacific and Atlantic Oceans since 2.74 Ma (Sato & Kameo 1996; Sato et al. 2004).

The absence of calcareous (foraminifers, coccolithophores) and biosiliceous (diatoms, radiolarians, silicoflagellates) microfossils and the low abundances of organic-walled palynomorphs in the CAO prevent the detection of any palaeoceanographic variability in the Pliocene (Backman et al. 2006). In the marginal Arctic Ocean, opal and carbonate contents are generally low, usually
containing negligible amounts of biosiliceous and calcareous microfossils, whereas palynomorphs are rather consistently present (figure 4).

In the subarctic seas, a variable deposition of biogenic opal and carbonate reflects changing hydrographic conditions (Henrich et al. 1989, 2002; Bohrmann et al. 1990; Haug et al. 1995; Rea & Snoeckx 1995). Shifts from opal to carbonate accumulation were attributed to major rearrangements in circulation and water-mass formation in the Nordic Seas since the Mid-Miocene (Bohrmann et al. 1990). Opal deposition and reduced carbonate preservation reflect a diminished exchange of surface and deep waters with the North Atlantic Ocean between 5.4 and 4.8 Ma, whereas increased carbonate accumulation and preservation indicate times of enhanced advection of relatively warm surface waters until ca 3 Ma (Henrich et al. 1989, 2002; Bohrmann et al. 1990; Wolf & Thiede 1991; Fronval & Jansen 1996; figure 2). This stronger inflow is linked to an increased formation of Norwegian Sea deep water and overflow to the North Atlantic Ocean (Bohrmann et al. 1990; Henrich et al. 2002; Poore et al. 2006).

An increased opal production in the eastern Labrador Sea was caused by ice-edge diatom blooms between 4 and 2 Ma, possibly related to the advection of cold and low-salinity sea-ice-covered waters from the Arctic Ocean to the North Atlantic Ocean with the East Greenland Current (Bohrmann et al. 1990). The high opal accumulation in both the Nordic Seas and the North Pacific Ocean in the earliest Pliocene was probably associated with the global biogenic bloom due to increased nutrient supply between 4.5 and 7 Ma (Cortese et al. 2004).

Opal accumulation broke down in the North Pacific Ocean synchronously with increased IRD deposition since 2.74 Ma due to the onset of the permanent stratification that reduced upwelling of nutrient-rich waters, but opal production resumed in the latest Pliocene (Haug et al. 1995, 2005; Rea & Snoeckx 1995). This event is associated with the development of the halocline in the North Pacific Ocean. Seasonal contrasts in sea-surface temperature increased and the warm pool of late summer surface water might have been a potential source of moisture to supply the growing North American ice sheet (Haug et al. 2005; Swann et al. 2006). A significant freshening in surface-water salinities by 2–4 psu (practical salinity units) associated with meltwater supply has helped to maintain the stratification (Swann et al. 2006).

Carbonate preservation in the Nordic Seas was good until 3 Ma succeeded by a time-transgressive change to poor preservation due to a strong reduction in warm-water inflow associated with a reduced deep-water export and the establishment of a strong temperature gradient between the North Atlantic Ocean and the Nordic Seas (Henrich et al. 1989, 2002; Fronval & Jansen 1996). The sediments are virtually carbonate free in the Nordic Seas and Labrador Sea between 2.8 and 1.9 Ma, except in the easternmost Norwegian Sea that had low but variable contents (Henrich et al. 2002). This southeast to northwest gradient in the Nordic Seas in the Late Pliocene might have been related to a strong temperature gradient (Fronval & Jansen 1996). The carbonate preservation was probably poor due to (i) low production of carbonate shells in the surface water, (ii) sluggish renewal of deep waters induced by a rather stable sea-ice cover and/or increased freshwater supply, and/or (iii) production of carbonate corrosive dense brines during sea-ice formation (Henrich et al. 2002). Northern Component Water reconstructions support a steady decline of deep-water overflow to the North Atlantic Ocean between 3 and 2 Ma (Poore et al. 2006). Excess freshwater
that may have helped to suppress deep-water formation in the Nordic Seas might have been increasingly supplied from the North Pacific Ocean since (i) the onset of the modern East Greenland Current at ca 4 Ma (Bohrmann et al. 1990), (ii) the reversal to a northward flow through the Bering Strait at 3.6 Ma (Marincovich & Gladenkov 2001; Gladenkov et al. 2002; Gladenkov 2006), and (iii) the substantial decrease of sea-surface salinities in the North Pacific Ocean at ca 2.7 Ma (Swann et al. 2006). Since the latest Pliocene, the first intrusions of the Proto-Norwegian Current into a narrow corridor in the southeastern Nordic Seas caused a much better carbonate preservation and a resumption of deep-water formation in the Nordic Seas (Henrich et al. 2002).

The Mid- to Late Pliocene cooling in the high latitudes was obviously punctuated by pronounced warming events in the subarctic seas and the marginal Arctic Ocean that were at least intermittently much warmer than today (figure 2). Harrison et al. (1999) suggested a link between global temperature fluctuations and Arctic transgressive–regressive depositional cycles in the past 47 Myr, with sea-level highstands being associated with temperature maxima. This is in agreement with previous studies demonstrating that climates with a warmer fauna and flora are related to transgressions or sea-level highstands often either succeeded or preceded by colder climate conditions (e.g. Funder 1989; Brigham-Grette & Carter 1992; Funder et al. 2001; McNeil et al. 2001). The terrestrialley exposed warmer-water sequences may represent rather short interglacials such as the Kap København Formation (marine isotope stage 95/96(?)) that probably had a duration of less than 20,000 years (Funder et al. 2001). A pronounced climate cyclicity in the Pliocene of the marginal Arctic Ocean is supported by the stable oxygen isotope composition of bivalve shells from the Tjörnes beds in Iceland, which shows a large-scale variability of seawater temperature superimposed on a long-term cooling trend between ca 4.3 and 2.5 Ma (Buchardt & Simonarson 2003).

Uplifted shallow-marine sequences in the marginal Arctic Ocean that indicate warmer temperatures than today with at least seasonally ice-free conditions and establishment of boreal forest to forest tundra have been mainly reported from the Mid- and Late Pliocene (figure 2; Feyling-Hanssen et al. 1983; McNeil 1990; Brigham-Grette & Carter 1992; Cronin et al. 1993; Fyles et al. 1998; Funder et al. 2001; Bennike et al. 2002). The most prominent of these events occurred in the Mid-Pliocene between ca 3.3 and 3.0 Ma when increased meridional heat transport to the high-latitude North Pacific and North Atlantic Oceans led to the inflow of warmer waters into the Arctic Ocean and ice-free conditions in the marginal Arctic Ocean (Cronin et al. 1993; Knies et al. 2002; Dowsett et al. 2005). Advection of warmer waters is reflected by increased carbonate accumulation and abundances of subpolar planktonic foraminifers in the North Pacific Ocean and Nordic Seas probably due to increased production in warmer surface waters and/or reduced dissolution (Henrich et al. 1989; Bohrmann et al. 1990; Haug et al. 1995; Rea & Snelbeck 1995; Baumann et al. 1996). In the Nordic Seas, the carbonate maximum strongly diminishes to the north, but a maximum of the dinoflagellate cyst Operculodinium centrocarpum at the western and northern Barents Sea margin may indicate an increased inflow of Atlantic waters into the eastern Arctic Ocean (e.g. Hole 911A, figure 4; Smelror 1999; Knies et al. 2002; Ryseth et al. 2003). The timing of this event in Hole 911A is different compared with Dowsett et al. (2005), but this may be due to the poor age control in the lower part of Hole 911A.
Dinoflagellate cysts principally record relatively cold, stable and productive sea-surface conditions with a seasonal sea-ice cover (*Brigantedinium* spp.) in the marginal eastern Arctic Ocean in the Mid- to Late Pliocene. Possibly slightly warmer conditions with a reduced sea-ice cover as indicated by abundant *Filisphaera filifera* s.l. occurred between 2.1 and 1.8 Ma (figure 4).

(c) Arctic Ocean sea-ice cover in the Pliocene

The Arctic Ocean was probably covered by sea ice in the Pliocene, but the spatial and seasonal extent is still a controversial issue (Cronin *et al*. 1993, 2008; Polyakova 2001; Dowsett 2007; Darby 2008; Haley *et al*. 2008; St. John 2008). Owing to sparse microfossil records, the distinction between a perennial and seasonal sea-ice cover relies mainly on sedimentological and mineralogical data. The interpretation of these records is further complicated by a possible contribution of IRD from melting icebergs. St. John (2008), who discussed the implications of coarse fraction data (more than 250 and 150–250 μm, respectively) from Lomonosov Ridge (IODP Exp. 302), stated that the low contents throughout the Neogene (generally less than 5%) do not allow one to unequivocally distinguish sea ice from iceberg transport. Darby (2008) suggested that, based on the size spectra of detrital iron oxide grains in the same record, icebergs were of minor importance for the transport of IRD. This may be related to restricted glaciations in the source region of the Lomonosov Ridge sediments. The coarse fraction in CAO sediments substantially increased first in the Mid- to Late Pleistocene (Spielhagen *et al*. 2004), suggesting that since then icebergs more frequently passed the Lomonosov Ridge and the eastern Arctic Ocean. At sites proximal to Pliocene ice sheets such as the Nordic Seas, the Fram Strait and Yermak Plateau, iceberg transport was more important during the intensification of NHG (Henrich *et al*. 1989; Wolf & Thiede 1991; Winkler *et al*. 2002; figures 3 and 4). In general, sea ice was probably the dominant transport mode to distribute IRD in the Pliocene of the CAO, with a minor contribution from iceberg transport.

Sedimentological, mineralogical, radiogenic and cosmogenic isotope data suggest that a permanent sea-ice cover existed in the CAO since the Mid-Miocene (Darby 2008; Frank *et al*. 2008; Haley *et al*. 2008; Krylov *et al*. 2008; St. John 2008). Darby (2008) and Krylov *et al*. (2008) suggested that, based on the provenance of the detrital fraction of sediments and a careful analysis of transport distances and possible drift rates of sea ice, transit times of sea ice from some source areas to the Lomonosov Ridge must have exceeded 1 year, requiring a perennial ice cover as transport agent. The applied drift rates, however, may underestimate the potential pronounced variability as recorded from recent measurements. In 2007, the ice cover drifted at much higher rates in the eastern Arctic Ocean, and the schooner Tara completed her drift from the northeastern Laptev Sea to the Fram Strait in approximately 15 months (Gascard *et al*. 2008). Thus, even calculations using drift rates of 3 cm s$^{-1}$ (Darby 2008) may result in transit times being too long and the CAO might have been easily reached by sea ice from most potential sediment source areas in less than a year. In this case, transit times from Canadian Arctic source regions being still longer than a year may indicate the presence of multi-year sea ice in some regions rather than the development of a year-round sea-ice cover in the whole CAO. Furthermore,
the work by Cronin et al. (2008) disagrees with a perennial ice cover in the CAO since the Mid-Miocene because abundance maxima of agglutinated foraminifers in the Early Pleistocene might have been related to seasonally ice-free conditions. Dowsett (2007) used evidence for a reduced sea-ice cover in the marginal Arctic Ocean and a warmer terrestrial climate on the circum-Arctic continents to suggest a seasonally ice-free Arctic Ocean in the Mid-Pliocene (ca 3.3–3 Ma). Cronin et al. (1993) already proposed that a stronger advection of warmer water to the Arctic Ocean during warming events recorded in circum-Arctic shallow-marine sediments between 3.5 and 2 Ma might have ultimately led to a reduced sea-ice cover in the CAO owing to a weakened or eliminated halocline. These discrepancies between reconstructions may be partly explained by the variable temporal resolution of the records, leading to a possible under-representation of rather short interglacials (cf. Funder et al. 2001) in the studied records.

The circum-Arctic shelf seas were seasonally ice covered in the Pliocene, but the extent of the sea-ice cover is not known. Cronin et al. (1993 and references therein) concluded from ostracod analysis that the North American continental margin was ice free in summer during warmer intervals between 3.5 and 2 Ma, and probably the Arctic Ocean was year-round ice free in some regions at the Canadian Arctic and Alaskan continental margins. Diatom assemblages from the marginal seas of the Eurasian Arctic suggest cold conditions and the presence of a seasonal sea-ice cover since the Late Miocene and probably a sea-ice cover similar to today since the Late Pliocene (Polyakova 2001). Dinoflagellate cysts in Mid- to Upper Pliocene sediments from the Yermak Plateau indicate a variable but still seasonal sea-ice cover (Knies et al. 2002; figure 4). In the Nordic Seas, sea ice played an important role in the hydrological cycle since the latest Miocene (Henrich et al. 1989, 2002; Wolf & Thiede 1991). Polar surface waters with a dense sea-ice cover, at least during wintertime, extended into the Norwegian Sea in the Late Pliocene (Henrich et al. 2002). The occurrence of sea-ice diatoms suggests the presence of sea ice in the marginal northwestern North Pacific Ocean since the Miocene–Pliocene transition (Barron 2003).

The available data suggest strong temporal and spatial variabilities of the sea-ice cover rather than a stable perennial ice cover in the whole Arctic Ocean. At present, micropalaeontological data do not prove or disprove the presence of seasonal or perennial sea-ice cover in the CAO. The ice drift rates, at least in the Early and Mid-Pliocene, might have been quite different from the average modern oceanographic situation owing to a more vigorous exchange of water masses with the North Atlantic Ocean (Henrich et al. 2002; Poore et al. 2006), an increased input of warm Atlantic waters into the Arctic Ocean that might have additionally induced a reduction of the sea-ice cover (Butt et al. 2002) and an export of arctic waters through the Bering Strait until 3.6 Ma (Marincovich 2000). This might have caused a much faster transport of sea ice from the shelves into the CAO and an increased export into the North Pacific Ocean and Nordic Seas. Since 3.6 Ma low-salinity waters might have been increasingly transported from the Pacific Ocean into the Arctic Ocean, causing enhanced sea-ice formation that increased albedo and isolated the high heat capacity of the ocean from the atmosphere (Haug et al. 2001). Evaporative cooling of surface waters during deep-water formation may have provided the necessary moisture for ice-sheet growth to the high northern latitudes.

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5. Conclusions

Although the Arctic Ocean played a crucial role in the global climate development in the Pliocene owing to the build-up of ice sheets on the adjacent continents, relatively few marine data are available to test palaeoceanographic hypotheses based on numerical and conceptual models. In contrast to other ocean basins, tectonic forcing and sea-level fluctuations must have strongly influenced the climate development both on land and in the ocean by changing the geometry of gateways and by altering the topography of the circum-Arctic continents. This might have influenced atmospheric circulation and moisture transport, and subsequently the hydrological cycle of the Arctic Ocean.

A different freshwater supply might have been a decisive factor for triggering changes in water-mass properties, sea-ice formation and deep convection. These boundary conditions make the Pliocene Arctic Ocean much different from the modern basin, but they are rarely considered in palaeoceanographic studies.

Palaeoceanographic reconstructions are generally hampered by low contents of biogenic components in the high-latitude sediments and sparse palaeo records from the Arctic Ocean. Despite these restrictions, there is evidence both for a long-term cooling trend and for pronounced short-term variability. However, there is still a need for high-resolution studies on time scales that may capture palaeoenvironmental variability related, for example, to sea-level change or variable freshwater discharge.

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