The Late Mesozoic-Cenozoic Arctic Ocean Climate and Sea Ice History: A Challenge for Past and Future Scientific Ocean Drilling

Ruediger Stein1,2

1Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research, Bremerhaven, Germany, 2MARUM and Faculty of Geosciences, University of Bremen, Bremen, Germany

Abstract

Over the past 3–4 decades, coincident with global warming and atmospheric CO₂ increase, Arctic sea ice has significantly decreased in its extent as well as in thickness. When extrapolating this alarming trend, the central Arctic Ocean might become ice-free during summers within about the next 2–5 decades. Paleoclimate records allow us to better understand the processes controlling modern climate change and distinguish between natural and anthropogenic forcing. In this context, detailed studies of the earlier Earth history characterized by a much warmer global climate with elevated atmospheric CO₂ concentrations are important. The main focus of this review paper is the long-term late Mesozoic-Cenozoic Arctic Ocean climate history from Greenhouse to Icehouse conditions, with special emphasis on Arctic sea ice history. Starting with some information on the Cretaceous Arctic Ocean climate, this paper will concentrate on selected results from Integrated Ocean Drilling Program (IODP) Expedition 302 (Arctic Ocean Coring Expedition (ACEX)), the first scientific drilling in the permanently ice-covered Arctic Ocean, dealing with the Cenozoic climate history. While these results from ACEX were unprecedented, key questions related to the Cenozoic Arctic climate history remain unanswered, largely due to the major mid-Cenozoic hiatus (if existing) and partly to the poor recovery of the ACEX record. Following-up ACEX and its cutting-edge science, a second scientific drilling on Lomonosov Ridge with a focus on the reconstruction of the continuous and complete Cenozoic Arctic Ocean climate history has currently been proposed and scheduled as IODP Expedition 377 for 2021.

1. Introduction

During the last few decades, the remote Arctic Ocean, characterized by strong seasonal forcing and variability in runoff, sea ice formation, and sunlight, became a major focus within the ongoing debate of climate change as the high latitudes mirror the global warming trend strongly amplified (AMAP, 2017; Park et al., 2018; Serreze & Barry, 2011; Stocker et al., 2013). Due to complex feedback processes (collectively known as "polar amplification"), the Arctic is both a contributor of climate change and a region that will be most affected by global warming (ACIA, 2004, 2005; Serreze & Barry, 2011; Stocker et al., 2013). That means, the Arctic Ocean and surrounding areas are (in real time) and were (over historic and geologic timescales) subject to rapid and dramatic change. Although the Arctic Ocean was and is a key player in the past, present, and future global climate/Earth system (AMAP, 2017; Stocker et al., 2013), the marine geoscience research in the Arctic lags behind other world oceans, especially from a paleoclimatic perspective (e.g., Jakobsson et al., 2014; Stein, 2008). That means, detailed records of the short- and long-term geological and paleoclimatic history of the Arctic Ocean are still sparse. This lack in knowledge is mainly due to the major technological/logistical problems in operating within the permanently ice-covered Arctic region. During the last ~20 years, several international expeditions have greatly advanced our knowledge on central Arctic Ocean paleoenvironment and its variability through Quaternary times (Stein et al., 2015). Prior to 2004, however, central Arctic Ocean piston and gravity coring was mainly restricted to obtaining near-surface sediments; that is, only the upper 15 m could be sampled. Thus, almost all studies were restricted to the late Pliocene/Quaternary time interval, with only a few exceptions based on short sediment cores representing some isolated sections of late Mesozoic/early Cenozoic climate history. A breakthrough in Arctic Ocean paleoclimate research was the first scientific drilling in the permanently ice-covered central Arctic Ocean that was carried out within the Integrated Ocean Drilling Program (IODP) in 2004, that is,
IODP Expedition 302—the Arctic Ocean Coring Expedition (ACEX), penetrating 428 m of Upper Cretaceous to Quaternary sediments on the crest of Lomonosov Ridge between 87° and 88°N (Backman et al., 2006; Moran et al., 2006). The main objectives of the ACEX drilling campaign focused on the reconstruction of Cenozoic Arctic ice, temperatures, and climates from early (Paleogene) Greenhouse to modern (Neogene) Icehouse conditions. Some of the key questions to be answered from ACEX were framed around the evolution of sea ice and ice sheets, the past physical oceanographic structure, Arctic gateways, links between Arctic land and ocean climate, and major changes in depositional environments (Backman et al., 2006, Backman et al., 2008; Backman & Moran, 2009; for reviews, see O’Regan et al., 2011; Stein, Blackman, et al., 2014; Stein, 2017). By studying the unique ACEX sedimentary sequence, >100 papers are now published in highly ranked international peer-reviewed scientific journals (see Stein, Weller, et al., 2014, Stein et al., 2015, for references). Some of the ACEX paleoceanographic themes and highlights with reference to the original literature are summarized in some more detail (see section 5.2). To date, the ACEX site remains the one and only scientific drilling location in the central Arctic Ocean (Figure 1).

Thus, thinking about scientific ocean drilling campaigns that have been carried out under the umbrella of the Deep-Sea Drilling Project (DSDP), the Ocean Drilling Program (ODP), and/or the Integrated Ocean Drilling Program/International Ocean Discovery Program (IODP) in the other world oceans during the last five decades (Stein, Blackman, et al., 2014; Koppers et al., 2019), the Arctic Ocean is still a "mare incognitum" (Figure 1).

This review paper is structured as follows: After an overview about the main characteristics, significance, and recent climate change of the Arctic Ocean (sections 2 and 3) and some basic information about proxies used for the paleoenvironmental reconstructions (section 4), the main focus is on the long-term late
Mesozoic-Cenozoic climate history from Greenhouse to Icehouse conditions, with special emphasis on the Arctic sea ice history during its course from ice-free to seasonal to permanent ice conditions (section 5). Starting with some information on the Cretaceous climate based on short gravity cores, this paper will concentrate on selected results from IODP Expedition 302 (ACEX) dealing with the Cenozoic climate history. Finally, an outlook on the key goals of the coming IODP Expedition 377 scheduled for 2021 will be presented (section 6).

2. The Arctic Ocean: Characteristics and Significance for the Climate System

What are the main characteristics that make the Arctic Ocean so unique in comparison to the other world oceans? Why is the Arctic Ocean so important in the global climate/Earth system? Several aspects can be listed:

2.1. Physiography and Water Mass Exchange

The Arctic Ocean is surrounded by continents and the world's largest shelf seas, with limited connections to the Pacific and Atlantic Oceans, making the Arctic Ocean a "Mediterranean Sea" (Figure 1). With an area of about 9.5 × 10^6 km^2, the entire Arctic Ocean comprises 2.6% of the total area of the world's oceans, but less than 1% of the volume (about 13 × 10^6 km^3; Jakobsson, 2002; Jakobsson et al., 2012). The relatively small total volume is due to the large shelf areas, which make up as much as 52.7% of the total area of the Arctic Ocean, resulting in a shallow mean depth of the entire Arctic ocean of about 1,360 m.

Fram Strait between northeastern Greenland and Svalbard is the only deep-water connection with the world ocean via the Atlantic Ocean (Figures 1 and 2). Furthermore, two major currents, with southward flowing waters on the west and northward flowing waters on the east, exchange surface-water water between the Arctic and the North Atlantic through Fram Strait (Figure 3; Aagaard et al., 1985; Rudels et al., 1994; Schauer et al., 2002): The cold, ice-transporting East Greenland Current is the main current out of the Arctic Ocean. In contrast, the eastern West Spitsbergen Current, an extension of the North Atlantic-Norwegian Current, carries warm, relatively saline Atlantic Water into the Arctic Ocean. This Atlantic Water branch cools down and, together with Atlantic water that is flowing across the Barents Sea and through the St. Anna Trough, follows the continental slope and ridges of the deep basins at intermediate water depths. Toward the Pacific Ocean, only a shallow-water connection exists. From there, relative low-saline, nutrient-rich Pacific Water enters the Arctic Ocean through the Bering Strait (Figure 2), primarily driven by a pressure-head difference between the Pacific and the Arctic Oceans (e.g., Coachman et al., 1975; Hunt et al., 2012; Woodgate et al., 2005).
2.2. Surface Water Characteristics: Temperature and Salinity

Whereas the central Arctic Ocean mean sea-surface temperatures (SST) are $<-1^\circ$C throughout the year (resulting in a permanent sea-ice cover), a distinct seasonal variability is obvious in the marginal seas (Environmental Working Group, 1998; Singh et al., 2013). In the latter, mean SST reaches values of 0 to 5 $^\circ$C during summer at times of increased river discharge and Atlantic Water inflow. During winter times, low mean SST of $<-1^\circ$C similar to those of the central Arctic Ocean were also measured in the marginal seas. An exception from this general picture is the Barents Sea, strongly influenced by Atlantic Water inflow. Here mean summer SST may increase to 8 $^\circ$C and even during winter time; values of up to 4 $^\circ$C were determined in the southwestern part.

The distribution of the sea-surface salinity (SSS) is strongly controlled by the river discharge and the Atlantic Water inflow (Figure 2; Environmental Working Group, 1998). The high river discharge results in a very low salinity in the Arctic Ocean proper as well as the marginal seas, except for the Barents Sea, influenced by more saline Atlantic Water inflow. The Beaufort, East Siberian, Laptev, and Kara seas are characterized by mean SSS of <29 during the summer, strongly decreasing toward the river mouths; in most part of the central Arctic, mean SSS is below 32. In the Barents Sea and adjacent continental slope, mean SSS...
increases to $>34$. Due to seasonal variability in river discharge, the SSS also displays a strong variability, especially in the marginal seas.

### 2.3. Sea-Ice Cover

A key phenomenon of the Arctic Ocean is the perennial sea-ice cover in its central part and a strong seasonal variation of sea ice in the surrounding marginal (shelf) areas (Figure 3; Johannessen et al., 2004; Thomas & Diekmann, 2010; https://nsidc.org/data/seaice_index/). In winter these areas are usually completely covered with ice, whereas in summer they are mostly ice-free. An exception is the Barents Sea and the area around Svalbard, strongly influenced by the warm Norwegian/West Spitsbergen Current system. There, large parts are ice-free throughout the year.

In general, there are two primary forms of sea ice: seasonal (or first-year) ice and perennial (or multi-year) ice. The thickness of first-year ice ranges from a few tenths of a meter near the southern margin of the marine cryosphere to 2.5 m in the high Arctic at the end of winter. Some first-year ice survives the summer and becomes multiyear ice. In the present climate, old multiyear ice floes without ridges are about 3 m thick at the end of winter. In addition, land-fast ice (or fast ice) occurs in the coastal circum-Arctic area, which is immobilized for up to 10 months each year by coastal geometry or by grounded ice ridges (ACIA, 2004, 2005).

Sea-ice concentration (percent ice area per unit area) and derived parameters such as ice extent (the area within the ice- ocean margin defined as 15% ice concentration) can be reliably retrieved through satellite-borne passive microwave sensors, which are available continuously since 1978 and operating independently of cloud cover and light conditions (Gloersen et al., 1992; Johannessen et al., 2004). Satellite data have shown that the area of sea ice decreases from roughly $14-15 \times 10^6 \text{km}^2$ in March to $<5-7 \times 10^6 \text{km}^2$ in September, as much of the first-year ice melts during the summer (Figure 3; Gloersen et al., 1992; Cavalieri et al., 1997; Johannessen et al., 2004; Stroeve et al., 2012; https://nsidc.org/). The area of multiyear sea ice, mostly over the Arctic Ocean basins, the East Siberian Sea, and the Canadian polar shelf, is about $4-5 \times 10^6 \text{km}^2$ having decreased significantly during the last decades (e.g., AMAP, 2017; Johannessen et al., 1999; Nghiem et al., 2007).

The mean sea-ice drift pattern, a direct response to surface atmospheric pressure gradients and resulting wind patterns, is quite well known from drift buoy and geostrophic wind field data ( Colony & Thorndike, 1985; Hibler, 1989; Pfirman et al., 1997; Thorndike & Colony, 1982). North of the Alaskan coast, the ice circulates in a big anticyclonic structure, the Beaufort Gyre (Figure 3), corresponding to the high-pressure system over the Beaufort Sea. Speed varies from almost 0 in the center to 3 cm/s in the marginal zones. Toward the Siberian Arctic, this gyre opens up to a linear structure, the Transpolar Drift (Figure 3). Its origin is in the area of the Chukchi-East Siberian Sea, and it leads to the West. Ice drifts with mean velocities of about 2 cm/s past the North Pole and Franz Josef Land, accelerating to mean values of up to 10 cm/s through Fram Strait between Svalbard and Greenland.

Sea ice is a very critical component of the Arctic system that responds sensitively to changes in atmospheric circulation, incoming radiation, atmospheric and oceanic heat fluxes, and the hydrological cycle (Figure 4). Ice significantly reduces the heat flux between ocean and atmosphere; through its high albedo, it has a strong influence on the radiation budget of the entire Arctic. The albedo of open water is as low as 0.10, whereas the ice albedo ranges between 0.6 and 0.8 (Figure 4; Barry, 1996; Serreze & Barry, 2011). Thus, over ice surfaces as compared with open water, up to 8 times as much of the incoming shortwave radiation is reflected, resulting in lower surface temperatures. Furthermore, the sea-ice cover strongly affects the biological productivity (e.g., Sakshaug, 2004), as a more closed sea-ice cover restricts primary production due to low light influx into the surface waters (Figure 4). Sea ice is also an important agent for sediment transport from the shelves into, across, and out of the Arctic Ocean. Increased export of sea ice via the East Greenland Current will reduce the surface salinity in the Greenland Sea, which subsequently diminishes the global thermohaline circulation by slowing the production of North Atlantic Deep Water (Aagaard & Carmack, 1989; Broecker, 1997; Raymo et al., 1990).

### 2.4. River Discharge

One of the main characteristics of the Arctic Ocean is its huge river discharge (e.g., Aagaard & Carmack, 1989; Holmes et al., 2002, 2018). Today, the annual direct freshwater inflow by major rivers into the Arctic Ocean reaches a total of about $3,300 \text{km}^3$ (Holmes et al., 2002, 2018; Rachold et al., 2004). For
comparison, the amount of water discharged by the world rivers to the present-day ocean is estimated to be between 32 and 37 × 10³ km³/year (Milliman & Meade, 1983; Chakrapani, 2005). That means, the Arctic Ocean discharge is equivalent to >10% of the global runoff. Major contributors are the Yenisei (620 km³/year), the Ob (429 km³/year), the Lena (525 km³/year), and the MacKenzie (249 km³/year; Figure 1; Holmes et al., 2002, 2018; Rachold et al., 2004; McClelland et al., 2012).

The overall importance of Arctic river discharge for the Arctic Ocean and its significance for the global ocean system can be summarized as follows:

- The fluvial fresh-water supply is essential for the maintenance of the approx. 200-m-thick low-salinity layer of the central Arctic Ocean and, thus, contributes significantly to the strong stratification of the near-surface water masses (Figure 2), encouraging sea-ice formation. Thus, changes in the fresh-water balance would influence the extent of sea-ice cover.
- The fresh-water exported from the Arctic Ocean through Fram Strait influences global thermohaline circulation (Figure 2). Today, the annual liquid fresh-water export of the East Greenland Current is about 1,160 km³. Estimates based on sea-ice export even reach values of 1,680 (Aagaard & Carmack, 1989) to 2,850 km³/year (Eicken, 2004). Because factors such as the global thermohaline circulation as well as sea-ice cover and earth albedo have a strong influence on the Earth’s climate system, the fresh-water input to the Arctic and its change through time may have triggered global climate changes in the past (e.g., the initiation of the Northern Hemisphere Glaciation during mid-Pliocene times; Driscoll & Haug, 1998; for further discussion, see section 6).
- The Arctic rivers transport large amounts of nutrients, dissolved and particulate material (i.e., chemical elements, siliciclastic suspension, terrestrial organic matter, etc.; Martin et al., 1993; Gordeev et al., 1996; Stein & Macdonald, 2004; Bring et al., 2016), and anthropogenic pollutants (i.e., radioactive elements, heavy metals, etc.; Nies et al., 1998; Crane et al., 2001) onto the shelves (see above) where it is accumulated or further transported by different mechanisms (sea ice, icebergs, turbidity currents, etc.) toward the open ocean (e.g., Stein, 2008 for review). Thus, river-derived materials contribute in major proportions to the entire Arctic Ocean sedimentary and chemical budgets.
2.5. Permafrost

Large areas around the Arctic Ocean reaching about one quarter of the landmass of the Northern Hemisphere (Brown et al., 1997) are occupied by permafrost, that is, perennially frozen ground (Romanovskii et al., 2004; Vonk et al., 2015). The thickness of permafrost ranges from very thin layers of only a few centimeters thick to several hundreds of meters; in unglaciated areas of Siberia, even thicknesses of about 1,500 m may be reached (Washburn, 1980; Nelson et al., 1999). Furthermore, thermal modelling and geophysical data (e.g., Rekant et al., 2005; Romanovskii et al., 2005) indicate that large areas of the Arctic shelves, as a result of their exposure during the Last Glacial Maximum, are thought to be almost entirely underlain by subsea permafrost from the coastline down to a water depth of about 100 m (Rachold et al., 2007).

Many of the potential environmental and socioeconomic impacts of global warming in the Arctic are associated with permafrost. The effects of climatic warming on permafrost and the overlying "active layer" can severely disrupt ecosystems and human infrastructure and may intensify global warming (e.g., Frey et al., 2007; Jorgenson et al., 2001; Nelson et al., 1999). Huge amounts of soil organic carbon are currently stored frozen in permafrost soils (e.g., Tarnocai et al., 2009; Zimov et al., 2009), and vast amounts of methane, in a solid form as gas hydrates, are trapped in permafrost and at shallow depths in cold ocean sediments (e.g., Romanovskii et al., 2005). With increasing temperature of the permafrost (Romanovsky et al., 2010) and the water at the seafloor, as a result of increased surface warming in Arctic regions (Stockert et al., 2013), a widespread increase in the thickness of the thawed layer and the decomposition of hydrates could lead to the release of large quantities of CH₄ (and CO₂) to the atmosphere (ACIA, 2004, 2005; Anisimov et al., 1997; Goulden et al., 1998; Michaelson et al., 1996) as well as an enhanced mobilization and export of old, previously frozen soil-derived organic carbon (e.g., Bröder et al., 2018; Bröder et al., 2019; Schuur et al., 2008; Tesi et al., 2016; Vonk et al., 2012; Winterfeld et al., 2015, 2018). The release of these greenhouse gases in turn would create a positive feedback mechanism that can amplify regional and global warming.

3. Recent Arctic Climate Change, Future Predictions, and Paleorecords

During the 1990s, it became widely recognized that the Arctic was undergoing a dramatic change (Dickson et al., 2000; Macdonald, 1996; Morison et al., 2000; Moritz et al., 2002; Serreze et al., 2000). Over the past decades, a significant increase in Siberian river discharge, associated with a warmer climate and enhanced precipitation in the river basins, has been observed (Semiletov et al., 2000; Serreze et al., 2000; Peterson et al., 2002). At the same time, an increase in the amount and temperature of Atlantic Water inflow into the Arctic, a reduced sea-ice cover, a thawing of permafrost, and a retreat of small Arctic glaciers are observed (Dickson et al., 2000; Serreze et al., 2000).

The most obvious indicator for the ongoing rapid changes in the climate state of our planet, however, is certainly the observed loss in Arctic sea ice (Serreze et al., 2007; Notz & Stroeve, 2018; Stroeve & Notz, 2018). Over the past three to four decades, coincident with global warming and atmospheric CO₂ increase, Arctic sea ice has significantly decreased in its extent (Figure 5) as well as in thickness (Cavalieri & Parkinson, 2012; Kwok & Cunningham, 2015; Stocker et al., 2013; Stroeve et al., 2007, 2012). The loss of sea ice results in a distinct decrease in albedo, causing further warming of ocean surface waters. When extrapolating this trend, the central Arctic Ocean might become ice-free during summers within about the next three to five decades or even sooner, depending on the different IPCC scenarios used for prediction (Figure 5; Wang & Overland, 2012; Stocker et al., 2013; Niederdrenk & Notz, 2018; Notz & Stroeve, 2018; Stroeve & Notz, 2018; Screen & Deser, 2019). Furthermore, the recent decrease in sea ice seems to be more rapid than predicted by climate models (Figure 5; Stroeve et al., 2007, 2012), suggesting that the processes causing these recent rapid climate changes are still not fully understood and subject of intense scientific and societal debate. Based on the relationship between the September sea-ice extent and cumulative atmospheric CO₂ emissions, however, Notz and Stroeve (2016) have demonstrated quite clearly that the observed decrease in sea-ice cover is mainly driven by anthropogenic warming. Based on their records, these authors emphasized the importance of limiting future CO₂ emissions to values compatible with a global warming well below 2 °C in order to keep the ice cover during summer (cf. Sigmond et al., 2018; Stroeve & Notz, 2018). In any case, a key aspect within the debate about present climate change is to distinguish and quantify...
more precisely natural and anthropogenic forcing of global climate change and related sea ice decrease (AMAP, 2017; Stocker et al., 2013).

In order to better understand the processes controlling modern climate change and distinguish between natural and anthropogenic forcing, it is certainly very useful and important to study paleoclimate records. In general, such paleoclimate records document the natural climate, rates of change, and variability prior to anthropogenic influence. The instrumental records of temperature, salinity, precipitation, and other environmental observations span only a very short interval (<150 years) of the Earth’s climate history and provide an inadequate perspective of natural climate variability, as they are biased by an unknown amplitude of anthropogenic forcing. Paleoclimate reconstructions, however, can be used to assess the sensitivity of the Earth’s climate system to changes of different forcing parameters (e.g., CO₂) and to test the reliability of climate models by evaluating their simulations for conditions very different from the modern climate. In this context, not only high-resolution studies of the most recent (Holocene) climate history are of importance but also detailed studies of the earlier Earth history characterized by a much warmer (Greenhouse-type) global climate with elevated atmospheric CO₂ concentrations, such as the Late Cretaceous and the Paleogene (Hong & Lee, 2012; Kent & Muttoni, 2013; O’Brien et al., 2017; Zachos et al., 2008). A precise knowledge of rates and scales of past climate change are the only mode to separate natural and anthropogenic forcings and will enable us to further increase the reliability of climate change predictions. Thus, understanding the mechanisms of natural climate change is one of the major challenges for mankind in the coming years. In this context, the polar regions certainly play a key role, and detailed climate records from the Arctic Ocean spanning time intervals from the Cretaceous-Paleogene Greenhouse world to the Neogene-Quaternary Icehouse world will give new insight into the functioning of the Arctic Ocean within the global climate system.

Figure 5. September sea ice extent in the Arctic Ocean (1890–2090) based on historical data, direct observations/measurements, and projected by different climate models and different IPCC scenarios toward 2090. RCP = Representative Concentration Pathways of greenhouse gases in relation to the probable radiative forcing in 2100. The shaded area represents the standard deviation (Stocker et al., 2013; Stroeve et al., 2007; Stroeve et al., 2012). Map shows September sea-ice extent in 1980, 2007, and 2012 (Source: National Snow and Ice Data Center; https://nsidc.org/).
4. Proxies for Reconstructing Past Arctic Ocean Paleoenvironmental Conditions

What type of proxies are available for reconstruction of Arctic paleoclimate and, especially, paleo-sea ice conditions? A large number of studies are based on sedimentological, mineralogical, geochemical, and geophysical data (e.g., Darby, 2003, 2008, 2014; Jakobsson et al., 2014; Knies et al., 2000; Nørgaard-Pedersen et al., 2003; Polyak et al., 2010; Spielhagen et al., 2004), as well as microfossils such as diatoms, dinoflagellates, ostracods, coccoliths, and foraminifers (e.g., Koç et al., 1993; Matthiessen et al., 2001, Matthiessen, Brinkhuis, et al., 2009; Sarnthein et al., 2003; Polyak et al., 2010; de Vernal & Rochon, 2011; Cronin et al., 2013; De Vernal, Gersonde, et al., 2013; De Vernal, Rochon, et al., 2013; Seidenkrantz, 2013; Limoges et al., 2018). These proxies may give information about water-mass characteristics (e.g., temperature, salinity, oxygenation, and ocean currents), primary productivity, sea-ice cover, iceberg-related processes and discharge, sources and pathways of terrigenous sediment fractions, etc. (for overviews, see Hillaire-Marcel & de Vernal, 2007; Stein, 2008). As the reconstruction of Arctic sea ice history is a major focus of this review, sea-ice indicative proxies are discussed in some more detail in the following paragraphs.

**Figure 6.** Proxies for reconstruction of Arctic (paleo) sea ice conditions and sea-surface temperature (SST). (a) Ice-rafted debris, (b) examples of microfossils (from left to right: modern sea-ice algae/Brown et al., 2014; Eocene sea-ice algae/ Stickley et al., 2009; dinocyst Islandinium minutum/de Vernal & Rochon, 2011), and (c) specific biomarkers indicative for sea ice cover (IP$_{25}$), phytoplankton open-water productivity (specific sterols, highly branched isoprenoids (HBIs), and long-chain alkenones) and SST. By combining IP$_{25}$ and open-water phytoplankton biomarkers (“PIP$_{25}$”; cf. Müller et al., 2011) or specific HBIs (“HBI-T$_{25}$”; cf. Belt et al., 2019), even information about temporal and spatial changes in sea-ice coverage can be obtained (for further details, see text).
Sea ice often contains large amount of terrigenous material that has been entrained into the ice in the marginal seas (sediment-laden or “dirty” sea ice; Figure 6a; e.g., Pfirman et al., 1989; Nürnberg et al., 1994; Eicken et al., 1997, Eicken et al., 2005; Dethleff et al., 2000). In areas of sea-ice melting, sediment particles are released and deposited at the sea floor, contributing significantly to the present and past Arctic Ocean sedimentary budget. This sea-ice sediment—or “ice-rafted debris (IRD)” —mainly consists of terrigenous material with clay minerals, quartz, feldspars, heavy minerals, and specific Fe oxide grains as the main components. The mineralogy of sea-ice sediments may be very variable in time and space, depending on the geology of the source areas in the hinterland. Thus, studies of the mineralogical composition may allow the identification of source areas of the IRD particles and, based on these data, the reconstruction of past and present transport pathways (e.g., Nürnberg et al., 1994; Wahsner et al., 1999; Darby, 2003, 2014; see Stein, 2008 for review). IRD, however, can also be transported by icebergs (Figure 6a). If IRD is related to iceberg transport and not sea ice, the IRD proxy records the existence of extended continental ice sheets reaching the shelf-break. Here a first-order proxy to discriminate between sea-ice and iceberg-rafted sediments is the grain-size distribution. It is generally accepted that very coarse-grained material >250 μm up to pebble size (Figure 6a) are mainly restricted to iceberg transport, whereas the dominance of finer-grained (silt and clay) sediments are more typical for sea-ice transport (e.g., Clark & Hanson, 1983; Nørgaard-Pedersen et al., 1998; Spielhagen et al., 2004; Dethleff, 2005).

Sea-ice associated organisms like pennate ice diatoms are frequently used for reconstructing present and past sea-ice conditions (e.g., Koç et al., 1993; Stickley et al., 2009). However, it has also been shown that the preservation of fragile siliceous diatom frustules can be relatively poor in surface sediments from the Arctic realm, and the same is also true (if not worse) for calcareous-walled microfossils, thus limiting their applicability (Matthiessen et al., 2001; Schlüter et al., 2000; Steinsund & Hald, 1994). De Vernal, Rochon, et al. (2013) present a status report about reconstructing past sea-ice cover of the Northern Hemisphere using dinoflagellate cysts. Based on a study of Arctic Greenland shelf sediments, Limoges et al. (2018) provide new insights into the distribution of a sea ice-associated dinoflagellate species, species Polarella glacialis and its cysts, and propose that this species might have the potential to become a useful proxy for reconstructing past seasonal sea ice.

A multiproxy approach considering sediment texture in combination with lithology and sediment provenance as well as micropaleontological and geochemical data indicative for surface-water characteristics certainly provides a much sounder interpretation of sea-ice versus iceberg-rafted sediments and reconstruction of past sea-ice conditions.

The ability to (semi-)quantitatively reconstruct paleo-sea-ice distributions has been significantly improved by a biomarker approach based on determination of a highly branched isoprenoid (HBI) with 25 carbons (C25 HBI monoene—“IP25”; Figure 6c; Belt et al., 2007). This biomarker is only biosynthesized by specific diatoms living in Arctic sea ice (Figure 6b; Brown et al., 2014); that is, its presence in the sediment is a direct proof of the presence of past Arctic sea ice. Furthermore, IP25 appears to be a specific, sensitive, and stable proxy for Arctic sea ice in sedimentary sections representing late Miocene to Holocene times (Belt, 2018; Belt & Müller, 2013; Knies et al., 2014; Stein et al., 2012, 2016, and references therein). When using this proxy, however, one has to consider that IP25 is more or less absent under a permanent sea-ice cover that limits light penetration and, in consequence, sea-ice algal growth. The same consequence applies to totally ice-free conditions where no ice algae live, that is, IP25 = 0. This difficulty in interpreting IP25 data can be overcome by the additional use of phytoplankton-derived, open-water biomarkers such as brassicasterol and dinosterol (Müller et al., 2009, 2011) or specific tri-unsaturated HBIs (Belt et al., 2015, 2019; Limoges et al., 2018; Ribeiro et al., 2017; Smilk et al., 2016). By combining the environmental information carried by IP25 and the different phytoplankton biomarkers (Figure 6c), even more semiquantitative estimates of present and past sea-ice coverage, seasonal variability, and marginal ice zone situations are possible (Müller et al., 2011; Belt et al., 2015, 2019; Stein et al., 2017; for recent and critical review of this biomarker approach, see Belt, 2018).

Another useful option to distinguish between the two “IP25 = 0” extremes, that is, ice-free versus thick closed ice coverage, is the determination of SST. SST values significantly above 0 °C give important information about surface water characteristics per se, but also clearly point to ice-free conditions if IP25 = 0. For reconstructions of past SST conditions in the central Arctic Ocean during the Cretaceous, Paleocene-Eocene and Miocene, very promising biomarker SST tools such as the alkenone-based U37 index (Brassell et al., 1986;
Prahl & Wakeham, 1987) and the TEX$_{86}$ index (Schouten et al., 2002, 2004) have been used successfully (Jenkyns et al., 2004; Suijs et al., 2006, Suijs et al., 2009; Weller & Stein, 2008; Stein, Weller, et al., 2014; for further discussion, see below).

The alkenone-based U$^{K}_{37}$ approach evolved from the observation that certain microalgae of the class Prymnesiophyceae, notably the marine coccolithophors Emiliania huxleyi and Gephyrocapsa oceanica (e.g., Conte et al., 1992; Marlowe et al., 1984; Volkman et al., 1980), and presumably other living and extinct members of the family Gephyrocapsaceae (Marlowe et al., 1990; Müller et al., 1998), have or had the capability to synthesize alkenones whose extent of unsaturation changes with growth temperature (Brassell et al., 1986; Marlowe et al., 1984; Prahl & Wakeham, 1987). Based on this correlation, paleo-SST can be calculated from the so-called ketone unsaturation index U$^{K}_{37}$ or the simplified version of the index U$^{K}_{37}'$ (Figure 6c; Brassell et al., 1986; Prahl & Wakeham, 1987; Müller et al., 1998). The TEX$_{86}$ approach is based on specific membrane lipids, that is, tetraether lipids (glycerol dialkyl glycerol tetraethers) with 0–4 cyclopentane rings and, in one case, a cyclohexane ring, synthesized by marine Crenarchaeota (Schouten et al., 2002). As shown in a study of marine surface sediments from the world oceans, these organisms adjust the composition of these tetraether lipids in response to SST (Ho et al., 2014; Kim et al., 2008; Schouten et al., 2002, 2004; Weijers et al., 2006; Wuchter et al., 2004). This response is quantified as the so-called TEX$_{86}$ (tetraether index of 86 carbon atoms).

5. Long-Term Climate Change From Greenhouse to Icehouse Conditions

The Mesozoic to Cenozoic time interval is a period when the climate on Earth changed from one extreme (Cretaceous-Paleogene Greenhouse lacking major ice sheets) to another (Neogene Icehouse with bipolar glaciation; Figure 7; Pearson & Palmer, 2000; Norris et al., 2002; Zachos et al., 2008; Friedrich et al., 2012; Kent & Muttoni, 2013; O’Brien et al., 2017; O’Connor et al., 2019). This long-term climatic evolution, including prominent warm and cold phases, is best reflected in marine deep-water and surface-water temperatures determined from benthic and planktic oxygen isotope, Mg/Ca thermometry, and biomarker (TEX$_{86}$ and U$^{K}_{37}$) records, respectively, from low to middle latitudes (e.g., Alsenz et al., 2013; Bijl et al., 2009; Friedrich et al., 2012; Lear et al., 2000; Littler et al., 2011; Liu et al., 2009; Miller et al., 1987; Mutterlose et al., 2010; O’Brien et al., 2017; Zachos et al., 2001; Zachos et al., 2008). Peak warm intervals occurred around the Cenomanian/Turonian boundary, during the Paleocene-Eocene Thermal Maximum (PETM), the Early Eocene Climate Optimum (EECO), and the Middle Eocene Climate Optimum (Figure 7), most probably related to higher atmospheric CO$_2$ concentrations. The long-term, post-PETM high-latitude cooling started near 50 Ma and is punctuated by four major steps in the early mid-Eocene (about 48–45 Ma), at the Eocene/Oligocene boundary (near 34 Ma), in the middle-Miocene (at about 14 Ma), and in the middle-Pliocene (at about 3.5–2.6 Ma; Figure 7; Miller et al., 1987; Lear et al., 2000; Pearson & Palmer, 2000; Zachos et al., 2001, Zachos et al., 2008; Westerhold & Röhl, 2009). Furthermore, long-term climatic changes on global and regional (e.g., Arctic) scales are also controlled by plate tectonic processes such as opening and closure of oceanic gateways, isolating single ocean basins from the World Ocean, configuration of continents, and uplift of mountain ranges and plateaus (Hay et al., 1999; Scotese, 2001; Müller et al., 2008; references with focus on the Arctic Ocean: Kristoffersen, 1990; Grantz et al., 2011; Shephard et al., 2013; Kanao et al., 2015; Petrov et al., 2016).

The long-term paleoclimatic and paleoceanographic history of the Arctic Ocean through this time interval, however, is still poorly known in comparison to other world ocean areas. Major information on the paleoenvironment of the early Arctic is derived from circum-Arctic terrestrial data sets (for references, see next sections) and—concerning the marine records—from petroleum exploration drill holes from the Arctic marginal seas (e.g., Leith et al., 1992; Spencer et al., 2011) and Deep-Sea Drilling Project and ODP drill cores from subarctic regions (Leg 38: Talwani et al., 1976; Leg 104: Eldholm et al., 1989; Leg 105: Srivastava et al., 1989; Leg 151: Thiede et al., 1996; Leg 152: Larsen et al., 1994; Leg 162: Raymo et al., 1999; see Figure 1 for locations). Direct information from sediment cores derived from the central Arctic Ocean is restricted to a very few short sections—at least prior to the IODP-ACEX drilling campaign in 2004 (Figure 7; Thiede et al., 1990; Backman et al., 2008; see below). From Figure 7 it is obvious that for almost the entire Miocene, the Oligocene, Eocene, and Paleocene, no sediment-core data are available at all. For the Cretaceous, data from
central Arctic Ocean cores are limited to three short sections representing isolated, discontinuous fragments of the late Campanian and Maastrichtian climate history (see next section).

5.1. The Cretaceous Arctic Ocean: Warm, Euxinic, and Productive, but Occasionally Ice

Up to the 1980s, the general global climate of the Cretaceous was viewed as a period of great warmth over the entire globe (Frakes, 1979; Hallam, 1985). According to these authors, tropical-subtropical conditions prevailed to at least 45°N and possibly to 70°S, and warm to cool-temperate climates extended to the ice-free polar regions. Such a greenhouse world was characterized by significantly higher temperatures and reduced latitudinal temperature gradients (O’Brien et al., 2017; Robinson et al., 2019; Sinninghe Damsté et al., 2010). The overall warm climatic conditions most likely resulted from an enhanced greenhouse effect due to higher atmospheric CO$_2$ concentrations as inferred from various proxies (Figure 7; Crowley & Berner, 2001; Royer, 2010; Kent & Muttoni, 2013; Foster et al., 2017) and conventionally attributed to the balance between the input of CO$_2$ from ocean crust production, volcanism, metamorphism, and organic carbon weathering and the output of CO$_2$ from silicate weathering and organic carbon burial (Foster et al., 2017; Kent & Muttoni, 2013; Walker et al., 1981; Weisheit & Erba, 2004). During the 1990s, however, the view of constant equable warmth and sluggish oceanic circulation during the Cretaceous was questioned, and much more...
variable climatic conditions were suggested for this time period (e.g., Kemper, 1987; Miller et al., 2005; Mutterlose et al., 2003; Price et al., 2000; Weissert & Lini, 1991). Even if warm climates in the high latitudes dominated much of the Cretaceous (to early Eocene) and warmth may have been caused by high CO₂, short intervals of more restricted glacializations probably occurred (at least) on Antarctica (Miller et al., 2005). These authors propose a vision of Earth’s cryospheric evolution that reconciles warm, generally ice-free poles with cold snaps that resulted in glacio-eustatic lowerings of 20–40 m, probably driven by orbital forcing (Matthews & Frohlich, 2002; DeConto & Pollard, 2003; for general review and references, see Haq, 2014).

The availability of nutrients and elevated atmospheric CO₂ may have stimulated primary production, which is thought to be one of the main factors (besides bottom-water stagnation) causing anoxic conditions in large parts of the oceans (e.g., Arthur et al., 1984, 1987; de Graciansky et al., 1984; Demaison & Moore, 1980; Erbacher et al., 2001; Ohkouchi et al., 2015; Schlanger & Jenkyns, 1976). These major episodes are known as “oceanic anoxic events” (Schlanger & Jenkyns, 1976) and represent major perturbations in the global climate and ocean system defined by massive organic carbon (OC) burial and black shale formation in marine environments. Oceanic anoxic events occurred on a regional to worldwide scale with the most prominent and widespread ones in the Toarcian, the Valanginian, the Aptian-Albian, the Cenomanian-Turonian, and the Coniacian-Santonian (Arthur et al., 1984, 1987; Brumsack, 1980; de Graciansky et al., 1984; Erba et al., 2004, 2015; Erba et al., 2019; Gambacorta et al., 2016; Méhay et al., 2009; Ohkouchi et al., 2015; Rais et al., 2007; Schlanger & Jenkyns, 1976; Stein et al., 1986). Upper Jurassic to Cretaceous black shales were also deposited in all the large circum-Arctic sedimentary basins (e.g., North Alaskan Basin, Mackenzie Delta Basin, Sverdrup Basin, Barents Shelf, and western Siberia) that are highly productive in terms of gas and oil (e.g., Bakke et al., 1998; Dixon et al., 1992; Leith et al., 1992; Lenniger et al., 2014; Spencer et al., 2011).

A dominantly warm climate probably existed in the Northern High Latitudes during Middle to Late Cretaceous times as indicated by the presence of deciduous trees and leaves with characteristic morphologies, at 80–85°N (Herman & Spicer, 1996; Parrish & Spicer, 1988; Spicer & Herman, 2010), the presence of crocodiles beyond 60°N (Markwick, 1998), and most specifically, the discovery of champosaurs (cold-blooded reptiles; closest living relatives are crocodiles) in the Turonian of the Sverdrup Basin (Axel Heiberg Island) at 72°N paleolatitude (Tarduno et al., 1998; Huber, 1998; for summary, see O’Regan et al., 2011). Besides indication for a relatively warm (sub-) Arctic climate during Cretaceous times, there are also signals pointing to cold climatic conditions. In particular for parts of the Early Cretaceous, hints of Icehouse conditions or at least cool climates in the higher latitudes were found. Glendonites (pseudomorphs of the low-temperature hydrated form of calcium carbonate, ikaiite) found in lower Valanginian and upper Aptian sediments from the Sverdrup Basin in Arctic Canada (70–80°N paleolatitude) imply that Early Cretaceous seawater temperatures were at times close to freezing (Kemper, 1987). Almost certainly these cooler temperatures record global changes because, at least in the case of the late Aptian, coeval glendonites are also known from the Southern Hemisphere, found in the Eromanga Basin in Australia at a paleolatitude of 65°S (De Lurio & Frakes, 1999; Frakes & Francis, 1988). Furthermore, Frakes and Francis (1988) postulate at least seasonally cold ocean temperatures and limited polar ice caps for the Early Cretaceous from the occurrence of ice-rafted deposits in Siberia, Australia, and Spitsbergen. Nannofossil assemblages from offshore mid-Norway and the Barents Sea also suggest cold conditions (Mutterlose et al., 2003; Mutterlose & Kessels, 2000). Based on oxygen stable isotope paleothermometry of well-preserved Middle Jurassic to Lower Cretaceous belemnites from Kong Karls Land, Svalbard, cool high-latitude marine paleotemperatures of <8 °C were estimated for the early to middle Valanginian, which may be compatible with the formation of high-latitude ice (Ditchfield, 1997).

What about marine sedimentary records from the Arctic Ocean representing information about climatic conditions during (late Jurassic to) Cretaceous times, that is, at times when the Arctic Ocean was probably much more isolated from the world ocean in terms of deep-water connection (Figure 8)? The isolation of the Arctic Ocean at those times was an important prerequisite for the development of widespread anoxic/euxinic conditions during the Late Jurassic/Early Cretaceous (e.g., Spencer et al., 2011). As already mentioned above, most of the sedimentological, micropaleontological, and geochemical data available for paleoenvironmental reconstructions of the Arctic realm during this time interval are from petroleum exploration drill holes from the marginal seas such as the Barents Sea and along the Norwegian continental
In the southern Barents Sea, for example, OC-rich sediments of Late Jurassic/Early Cretaceous age were recovered, characterized by OC contents of up to about 35% (Figure 8; Langrock et al., 2003a). High hydrogen-index values of 400–600 mg HC/g C obtained by Rock-Eval pyrolysis (Langrock et al., 2003a) and maceral compositions of the sediments dominated by lipid-rich organic matter derived from mainly marine but also brackish and freshwater algae ("alginites") as well as fine-grained "liptodetrinites" (Figure 8) support a type II kerogen characterization. The excellent preservation of the labile, easily degradable organic matter particles and the significant amounts of freshwater algae argue for deposition in a restricted shallow basin very close to land rather than an upwelling regime (Arthur et al., 1987; Taylor et al., 1998). The increasing OC content in the late Volgian section was most likely caused by a combination of increasing preservation by bottom-water anoxia, coupled with periods of increasing primary production as well as a low and decreasing dilution of clastic material, related to a marked rise in sea level (Figure 8; Langrock et al., 2003a). During the Late Jurassic and Early Cretaceous, such anoxic conditions proposed for large parts of the proto-Arctic Ocean, probably extended toward the south into the Norwegian-Greenland Sea as shown in geochemistry and microscopy data sets from black shale sequences recovered in drill cores along the Norwegian shelf (Figure 8; Langrock et al., 2003b; Langrock & Stein, 2004). The high abundance of autochthonous small-sized pyrite frambooids indicative for an anoxic water column (i.e., under euxinic conditions), such as typical for the modern Black Sea (cf. Wilkin et al., 1997), the carbon-sulfur-iron margin (Figure 8), whereas data from the central Arctic Ocean are restricted to a few short sediment cores (Figure 9; see below).
relationships (cf. Hofmann et al., 2000; Lückge et al., 1996), the high concentrations of Mo, and the Re/Mo ratios also support widespread euxinic conditions (Langrock et al., 2003a, 2003b; Lipinski et al., 2003).

In addition, patchily distributed, coarse-grained minerals (apparently quartz) are found, which are 60 to 200 μm in diameter (Figure 8). These minerals could not have been transported along with the claystone facies. Here a possible mechanism for the input of these large sand grains was ice rafting. Episodes of a relatively cool climate were already suggested for the earliest Cretaceous Boreal realm as described above. Hence, episodic (or even seasonal) IRD input might have occurred in the late Volgian Barents Sea (Figure 8) at a paleo-latitude of almost 55°N (Langrock et al., 2003a).

In the central Arctic Ocean, Cretaceous sediments were only cored at three locations on the Alpha Ridge: cores FL-437 and FL-533 taken during the drift of the ice-island T-3 from 1963 to 1974 and Core CESAR-6 of the Canadian Expedition to Study the Alpha Ridge (CESAR) in 1983 (Figure 9a; Thiede et al., 1990). In 2004, an about 3-m-thick interval of Campanian very dark gray clayey mud and silty sands with OC values of about 1–2% was recovered in the lowermost part of the ACEX record from Lomonosov Ridge (Backman et al., 2006; Stein, 2007). During the FRAM-2014/2015 Ice Drift Expedition (Kristoffersen et al., 2016; Kristoffersen & Tholfsen, 2016), a 300-m-thick sequence of flat-lying Cenozoic sediments overlying dipping (probably) Mesozoic sediments, separated by an unconformity, were recognized in a seismic reflection profile across Lomonosov Ridge close to the North Pole (the age classification proposed by the authors is based on comparison with the ACEX drilling results). A gravity corer was dropped down where the potential for exposure of sedimentary rocks below the unconformity was high, and, as result, a short core (FRAM14/15-2) containing a “20 cm thick section of very cohesive, light cream colored sediments – something we have not seen so far from the Arctic Ocean” was recovered (Kristoffersen & Tholfsen, 2016; see Figure 9a for core
location). These sediments are almost pure siliceous oozes composed of well-preserved diatoms, radiolarians, silicoflagellates, etc. (first observations by R. Stein, unpubl. data, 2019). More detailed work by micropaleontologists on the species identification and age determination (Campanian/Maastrichtian? or younger?) are in process now; thus, no further discussion is presented at this stage.

At Core CESAR-6, siliceous, well-laminated Campanian-Maastrichtian sediments with excellently preserved microfossils were retrieved (Jackson et al., 1985). Very similar sediments of Campanian age were also recovered at Core Fl-437: a yellowish laminated siliceous rich in diatoms, ebrideans, silicoflagellates, and archaeomonads with sparse fish remains (Clark et al., 1986; Dell’Agnese & Clark, 1994). More recently, Davies et al. (2009), Davies et al., 2011 analyzed in detail the laminated section of Core CESAR-6 using “backscattered electron imagery” and identified a seasonal succession of two distinct lamina types that represent an annual lamina couplet (Figure 9b). Type A of laminae consists of diatom resting spores representing flux from the spring bloom, and type B of laminae consists of diatom vegetative cells representing flux from production during the stratified months of the Arctic summer (Davies et al., 2011). In some of these lamina couplets, coarse-grained terrigenous particles (mainly quartz) indicative of ice-rafting (IRD) were found (Figure 9b). The diatom laminae point to a strong seasonality, that is, to ice-free summers, whereas the presence of IRD is interpreted as a signal for sea ice that occasionally may have occurred during severe winters (Davies et al., 2009; Davies et al., 2011).

In Core Fl-533, the lowermost 67 cm of the 348-cm-long sequence comprise OC-rich black mud (black shales) of early Maastrichtian age (Clark et al., 1986; Firth & Clark, 1998). These black shales are characterized by very high OC contents of up to almost 16% and hydrogen index values of about 250 to 350 mg HC/g OC (Figure 9c), indicating an immature, mixed marine-terrigenous type of organic matter with major contributions from marine/aquatic algae. That means, both OC content and composition were totally different in comparison to those within the Quaternary deposits (Stein et al., 2001; Stein & Macdonald, 2004). Whereas the OC buried and preserved in Quaternary sediments is mainly refractory terrigenous matter, the OC found in the Cretaceous sediments mainly consists of labile algae-type organic matter. These data suggest that the Cretaceous Arctic Ocean basin may have been an important sink for algae-type OC (and CO₂) and, thus, of importance for the global climate system of that time.

In order to get some information about the Cretaceous SST of the central Arctic Ocean, Jenkyns et al. (2004) used the TEX86 approach (Schouten et al., 2002; see section 4) and analyzed black shale samples from Core Fl-533 for membrane lipids of marine Crenarchaeota (Figure 9c). The TEX86 results indicate an average Arctic Ocean SST of 15 ± 1 °C at a paleolatitude of about 80°N for that part of the early Maastrichtian represented by the black muds (Figure 9c). Considering the diatom and IRD data from the CESAR-6 Core (see above), this SST of about 15 °C should certainly represent summer conditions. A summer SST of about 15 °C and the presence of sea ice during winter may represent conditions similar to those described for the Eocene central Arctic Ocean (see section 5.2).

5.2. The Early Cenozoic Transition: Euxinic Conditions and Onset of Arctic Sea Ice

For the Arctic Ocean, information about the early Cenozoic climate history is quite limited (cf. Figure 7). A large amount of information is based on paleobotanic studies of circum-Arctic terrestrial outcrops. As reported from several localities around the Arctic Ocean such as Axel Heiberg Island, Ellesmere Island, Mackenzie Delta, Alaska, and the New Siberian Islands, floral and faunal assemblages indicate exceptionally warm and humid conditions during the late Paleocene-early Eocene (Basinger et al., 1994; Dolezych et al., 2018; Eberle et al., 2010; Eldrett et al., 2009; Greenwood et al., 2010; Salpin et al., 2018; Suan et al., 2017; West et al., 2015). Multiproxy data from early Eocene deltaic plain sediments of the Mackenzie Delta (Arctic Canada near 75°N), for example, indicate annual air temperatures averaging 21–22 and 10–14 °C in winter (Salpin et al., 2018). Late Paleocene mean annual shallow-marine temperatures between 11 and 22 °C were recorded from mollusk stable isotopes determined in sediments from the Prince Creek Formation (Alaska, 80°N; Brice et al., 1996). Direct information about paleoenvironmental conditions in the central Arctic Basin prior to the ACEX drilling in 2004 was coming from a single short sediment core obtained from the drifting ice island T3 in 1969 (Core FL-422; see Figure 9a for location). The sediments of this core representing part of the middle Eocene time interval are laminated, rich in siliceous microfossils, especially diatoms and silicoflagellates, and characterized by alternating layers of vegetative cells and resting spores, interpreted as evidence for strong seasonality, upwelling of nutrient-rich waters, and warm surface-
water temperatures (Bukry, 1984; Clark, 1974; Dell’Agnese & Clark, 1994). As mentioned in the previous section, similar sediments with excellently preserved siliceous microfossils were also found in Core FRAM14/15-2 recovered from Lomonosov Ridge during the FRAM-2014/2015 Ice Drift Expedition (Kristoffersen & Tholfsen, 2016; Kristoffersen et al., 2016; for core location, see Figure 9a). However, the age of this sedimentary section (i.e., Campanian/Maastrichtian vs. Eocene) has not been determined yet.

With the ACEX drilling in 2004, a thick sequence of upper Cretaceous to Quaternary sediments became available for paleoclimatic reconstructions (Figure 10; Backman et al., 2006; Moran et al., 2006). These sediments are characterized by prominent changes in sediment composition and texture, suggesting drastic paleoenvironmental changes through time (Backman et al., 2006; Stein, 2007). Whereas the lower half of the ACEX sequence (Units 2 to 4) mainly consists of dark gray silty clay and biosiliceous ooze characterized by high TOC values of 1% to >5%, the upper half of the ACEX sequence (subunits 1/1 to 1/4) is composed of silty clay with very low TOC contents of <0.5% (Figure 11a), that is, values very similar to those known from upper Quaternary records determined in gravity cores from the Lomonosov Ridge and representing modern-type oxic deep-water conditions (Stein et al., 2001). In subunit 1/5 (about 193 to 199 m of composite depth (mcd)) characterized by distinct gray/black color bandings (“Zebra Unit”), TOC maxima of 7% to 14.5%

Figure 10. Age-depth diagram and main lithological units of the ACEX section (O’Regan, 2011, supplemented), based on the biostratigraphically derived age model by Backman et al. (2008) and the alternate chronology based on osmium isotopes (Poirier & Hillaire-Marcel, 2011). Depth of first occurrence of IRD between 240 and 260 mcd are marked as light blue horizontal bar. On top (a) International Bathymetric Chart of the Arctic Ocean (Jakobsson et al., 2012) representing modern boundary conditions with connections to the world ocean; (b) paleogeographic/paleobathymetric reconstruction for the late early Miocene. BSG = Barents Sea Gateway; JM = Jan Mayen Microcontinent; KR = Knipovich Ridge; MJ = Morris Jessup Rise; NS = Nares Strait; YP = Yermak Plateau (Jakobsson et al., 2007), and (c) Paleogeography of the Arctic region for the early middle Eocene during the phase of high biosilica production and preservation and euxinic conditions (50–45 Ma; Stickley et al., 2009). The location of the ACEX site is marked as a red circle.
were measured (Stein, 2007). The Eocene section (about 200 to 385 mcd) characterized by high (algae-type) organic carbon content is mainly related to euxinic conditions predominant at that time when the Arctic Ocean was quite isolated from the world ocean (Figures 10 and 11; Backman et al., 2006; Stein et al., 2006, Stein, Weller, et al., 2014). The transition from euxinic conditions to the modern-oxic conditions occurred during Oligocene-Miocene times and is related to the opening of Fram Strait, allowing the inflow of oxygen-rich deep water from the Atlantic into the Arctic Ocean (Jakobsson et al., 2007).

The exact timing of this prominent environmental change, that is, 17.5 versus 32 Ma, however, is still under debate and depends on the age model that is used, that is, Backman et al. (2008) versus Poirier and Hillaire-Marcel (2011); Figure 10). Backman et al. (2008) stratigraphic framework of the Cenozoic ACEX sequence is based on biostratigraphic, cosmogenic isotope, magneto- and cyclostratigraphic data, and they proposed a major hiatus spanning the time interval from 44.4 to 18.2 Ma. When following this age model, a large part of the climate history covering the transition from Greenhouse to Icehouse is missing in the ACEX

Figure 11. (a) Record of total organic carbon (TOC) contents as determined in the composite ACEX sedimentary sequence (Stein, 2007). Data on lithological units (LU) and core recovery (Rec.) from Backman et al. (2006). Age models AM-1 (Backman et al., 2008) and AM-2 (Poirier & Hillaire-Marcel, 2011) are shown. The “Zebra” Unit (Unit 1/5) marks the transition from an euxinic ocean to an oxygenated ocean that occurred when the formerly isolated Arctic Ocean became connected to the world ocean via Fram Strait (cf. Figure 10). Depending on the age model, this transition is in the middle Miocene (17.5 Ma) or Oligocene (32 Ma). Cam = Campanian; LP = late Paleocene; Mid Mioc. = middle Miocene; L.Pl. = late Pleistocene. (b) Record of terrigenous sand (150–250 μm) interpreted as ice-rafted debris (IRD); blue circles mark occurrence of pebble-sized IRD (based on St. John, 2008). The onset of glaciation is marked as light blue bar. Photographs show coarse fraction >250 μm (IRD) in two samples from 240.39 and 21.4 m composite depth (mcd), respectively.
sequence (Figure 7b). That means, the late Eocene-Oligocene time interval characterized by major ice growth on Antarctica, global sea level drop, and possible growth of Northern Hemisphere ice sheets (Lear et al., 2000; Miller et al., 1987, 2005; Zachos et al., 2008) is not represented in the ACEX records. Poirier and Hillaire-Marcel (2011) report Rhenium-Osmium (Re-Os) isochron ages and complementary Os-isotope measurements, together with carbon and nitrogen data, that give information on the redox state of the sediment during deposition. Based on their new data, these authors challenged the existence of a major hiatus between subunits 1/6 and 1/5 (Figure 10). That means, they proposed an alternate age model sequence that closes the gap in the ACEX record, resulting in a more condensed sedimentary section with three to five times lower mean sedimentation rates (0.2–0.8 cm/kyr) between about 49 and 12 Ma (Figure 10). The alternate age model, of course, would significantly modify the reconstructions of the tectonic evolution of Lomonosov Ridge and the paleoecological Arctic Ocean history including the reconstruction of the early onset of sea ice in the Arctic Ocean during the Eocene. It questions the timing (i.e., early Miocene), duration, and mechanism (i.e., the opening of Fram Strait) for the sedimentological and geochemical transition recorded in the ACEX sequence (cf. Jakobsson et al., 2007).

Whereas it is widely accepted that near the Eocene/Oligocene boundary, large ice sheets first appeared on Antarctica, coincident with decreasing atmospheric carbon dioxide concentrations (Figure 7, “Oi-1 glaciation”; e.g., Shackleton & Kennett, 1975; Kennett & Shackleton, 1976; Miller et al., 1987; Lear et al., 2000; Zachos et al., 2001, Zachos et al., 2008; Escutia et al., 2011, 2014), the database for the onset of glacial conditions in the Northern High Latitudes is much more limited. Prior to the ACEX record, it was indirectly inferred from sub-Arctic IRD records in the Norwegian-Greenland Sea, Iceland Sea, and Irminger Sea and Fram Strait area that Northern Hemisphere Glaciation began at about 14 Ma (e.g., Thiede et al., 1998), that is, about 20 Ma later than the first appearance of large ice sheets on Antarctica. With the ACEX drilling results, however, the date of Northern Hemisphere cooling and the early onset of sea ice was pushed back into the Middle Eocene when for the first time significant amounts of IRD occur in ACEX sedimentary sequence at about 250-m core depth (Figure 11b; Backman et al., 2006, Backman et al., 2008; Moran et al., 2006; St. John, 2008; Stickley et al., 2009; Darby, 2014; Tripati & Darby, 2019). This early onset of Arctic sea-ice formation and the contemporaneous decrease in SST are clearly reflected in the ACEX micropaleontological (Stickley et al., 2009) and biomarker records (Stein et al., 2014).

The first occurrence of sea-ice related diatoms, contemporaneously with IRD, was at about 47–46 (when using the ACEX age model of Backman et al., 2008) or 44–41 Ma (when using the alternate chronology of Poirier & Hillaire-Marcel, 2011; “Age Model 2”; Figures 10 and 12b). Iceberg transport was probably also present in the middle Eocene, as indicated by mechanical surface-texture features on quartz grains from this interval (St. John, 2008; Stickley et al., 2009). An early onset/intensification of NHG during Eocene times is also supported by IRD records from the Greenland Basin ODP Site 913, indicating high input of ice-rafted grains from Greenland sources related to an extended Greenland Ice Sheet (Eldrett et al., 2007; Tripati et al., 2008; Tripati & Darby, 2019). These findings suggest that Earth's transition from the Greenhouse to the Icehouse world was probably bipolar, which points to greater control of global cooling linked to changes in greenhouse gases in contrast to tectonic forcing (Backman et al., 2006; Backman et al., 2008; DeConto et al., 2008; Moran et al., 2006; Stickley et al., 2009; Tripati & Darby, 2019).

The cooling trend and onset of sea-ice formation recorded in the ACEX IRD and diatom records, coinciding with the global post-EECO cooling trend (Zachos et al., 2008), are also reflected in the ACEX SST record (Figure 12a; Stein, Weller, et al., 2014). Prior to about 47.1 Ma (44.2 Ma according to Age Model 2), warm SSTs between 18 and 26 °C prevailed, interrupted by a prominent, short-lived cooling to 5–10 °C near 47.5 Ma (45.6 Ma according to Age Model 2). At about 46.3 Ma (41.7 Ma according to Age Model 2), summer SST dropped to <17 °C (range 8–17 °C), more or less coinciding with a significant increase in IRD (St. John, 2008; Stickley et al., 2009; Figure 12). An absolute SST minimum of 6–8 °C was reached at about 44.8 Ma (37.5 Ma according to Age Model 2), followed by a short but prominent warm phase with a SST around 17 °C near 44.5 Ma (36.8 Ma according to Age Model 2). This warming event is characterized by the absence of IRD (Figure 12), interpreted to reflect lack of sea ice. Apart from this warm event, however, widespread sea ice seems to be the more typical phenomenon of the Arctic Ocean after about 45.0 Ma (38.1 Ma according to Age Model 2; Figures 11 and 12).
Overall, the reconstructed central Arctic Ocean SST values remain surprisingly high, even in the upper part of the ACEX record. The apparent paradox of having such a warm SST and sea ice, however, can be explained. As the alkenone-producing coccolithophorids need sunlight for photosynthesis, the alkenone SST represents the summer SST. Then, considering the strong seasonal temperature variability of >10 °C during the early-middle Eocene (Basinger et al., 1994; Wolfe, 1994; Greenwood & Wing, 1995; Suan et al., 2017; see Weller & Stein, 2008, for more detailed discussion of ACEX alkenone data and references), favorable conditions for sea-ice formation may have occurred during winter time. That means, after about 46.3 Ma (or 41.7 Ma according to Age Model 2), the environmental conditions in part of the Arctic Ocean might have been similar to that observed in the modern Baltic Sea where summer SSTs >15 °C and winter SSTs <1 °C with sea-ice formation are typical (Krause, 1969; Schmelzer & Holfort, 2015; Wüst & Brögmus, 1955).

In contrast to the biomarker-based reconstructions clearly indicating a middle Eocene seasonal Arctic sea-ice cover (see above), Darby (2014) postulated ephemeral formation of even perennial sea ice in the Arctic Ocean during the middle Eocene. Darby’s argument is based on the occurrence of specific coarse Fe grains (IRD) indicative of a distal source area (such as North America or Siberia). Taking the IRD with a North American or East Siberian origin found in the ACEX sediments and using modern sea-ice drift trajectories and velocities, Darby concluded that more than 1 year was needed to transport the sediments entrained in sea ice to the ACEX location. Hence, sea ice must have survived the summer melting season to reach the ACEX Site, indicating a perennial sea-ice cover. The onset of deposition of such Fe grains at the ACEX site (Figure 12c) is almost contemporaneous with the drop in SST (Figure 12a) as well as the appearance of IRD and sea-ice diatoms (Figure 12b). Thus, if present, phases of perennial sea ice should have been more the exception in the middle Eocene, whereas seasonal sea ice should have been the rule.
Based on a simulation of ice-drift pattern and velocities under permanent and seasonal ice conditions, Tremblay et al. (2015) have demonstrated that sea-ice drift was probably significantly faster under warmer climatic conditions with less or much thinner sea ice than today and that IRD from the North American or East Siberian shelves may reach the North Pole and even beyond within first-year ice. Thus, they conclude that the presence of ice-rafted sediment of North American and East Siberian origin at the North Pole (and at the ACEX site) is not a definitive indication of a perennial sea-ice cover in the Arctic Ocean. This may imply that one main assumption of Darby (2014) should be regarded critically.

### 5.3. Ice-Free Summers in the Central Arctic Ocean During the Late Miocene: An Analogue for the Near Future?

Late Miocene climatic conditions significantly warmer than today have been reconstructed from marine and terrestrial proxy records from around the globe (Pound et al., 2011; Zachos et al., 2008; Zhang et al., 2014). Quantitative SST proxy data from the High Arctic were restricted to a few terrestrial records (Utescher et al., 2015; Wolfe, 1994). Stein et al. (2016) complemented these studies with the first late Miocene SST and sea-ice records for the central Arctic Ocean that were obtained from a sediment core recovered at the steep lower slope of Lomonosov Ridge close to the Siberian continental margin where older sediments are

---

Figure 13. (a) Proxy evidence from Core PS87/106 (for location, see Figure 14) for late Miocene Arctic Ocean climate conditions (from Stein et al., 2016): alkenone-based summer sea-surface temperature (SST), concentrations of alkenones, dinosterol and brassicasterol as proxy for primary productivity, concentrations of terrigenous sterols (sum of campesterol and β-sitosterol), concentrations of sea-ice proxy IP25, and sea-ice index PdinoIP25 (cf. section 4; for further background and calculation, see Stein et al., 2016). (b) The seasonal sea-ice cycle and related principal processes controlling productivity and carbon flux in the central Arctic Ocean during the late Miocene (Stein et al., 2016, after Wassmann, 2011, supplemented). MIZ = Marginal Ice Zone. The dark period, height of the Sun, and changing thickness of snow and ice over the year as well as phytoplankton, zooplankton, and sea-ice productivity are shown. IP25 values for the different seasons are indicated.
outcropping near the seafloor (Stein, 2015; Core PS87/106: 81°12.76′N, 141°10.47′E, water depth 1,472 m; see Figure 14a for location). In this core, a hiatus was identified at 3.7-m core depth that separates overconsolidated upper Miocene sediments from the overlying upper Quaternary sediments (Stein et al., 2016). The late Miocene age for the lower part of the sedimentary sequence was identified by the common occurrence of the acritarch Decahedrella martinheadii (cf. Matthiessen, Brinkhuis, et al., 2009; Schreck et al., 2012). Reworking of these late Miocene palynomorphs is excluded based on the excellent preservation of the delicate palynomorph specimens.

For the reconstruction of the late Miocene Arctic Ocean sea-ice and SST conditions, Stein et al. (2016) carried out biomarker analyses of the lower 1.3-m-thick section of Core PS87/106 using the IP_{25} sea-ice biomarker approach together with alkenone-based U^K_{37} index determinations (cf. Figure 6c). The results demonstrated for the first time that the late Miocene central Arctic Ocean was relatively warm with SSTs of about 5 °C and ice-free conditions during summer, whereas sea ice occurred during spring and autumn/winter (Figure 13). That means, in contrast to the modern situation characterized by a perennial sea-ice cover (cf. Figure 3) and SST values <0 °C, a seasonal sea-ice coverage was predominant in the late Miocene.

Although at a first sight the sedimentary section of Core PS87/106 only represents a short snapshot of late Miocene Arctic climate, more detailed information about the late Miocene climate on a regional to even global scale can be obtained from this record (Stein et al., 2016). Based on the biomarker data, this late Miocene section of Core PS87/106 probably represents one almost complete (glacial/interglacial) cycle with extended and reduced spring sea-ice conditions (Figure 13a). Using mean sedimentation rates of about 3.2 cm/kyr as calculated independently from close-by gravity cores (Stein et al., 2001), the duration of this cycle is about 40 kyr, that is, very similar to the 41-kyr obliquity cycle. Hence, the record may represent just one obliquity cycle with ice-free conditions during summers in both the cold ("glacial") as well as the warm ("interglacial") phase (Figure 13). As Core PS87/106 is of late Miocene age, ice-free summer conditions should have occurred in the central Arctic Ocean during the warmer middle Miocene to early late Miocene time interval a fortiori (Stein et al., 2016).

A seasonal sea-ice cover typical for this time interval is also supported by the occurrence of dinoflagellates and agglutinated benthic foraminifers (Cronin et al., 2008; Matthiessen, Brinkhuis, et al., 2009). In contrast to the biomarker-based reconstructions as well as the micropaleontological data, an Arctic Ocean perennial sea-ice cover from middle Miocene onward has been proposed by Darby (2008) and Krylov et al. (2008), based on provenance studies of IRD in ACEx sediments described for the middle Eocene situation (see above) and the use of modern sea-ice drift trajectories and velocities. Under warmer climatic conditions with less sea ice, however, this should be interpreted cautiously.

Further support for late/middle Miocene ice-free conditions during the summer months comes from climate simulations with CO₂ concentrations of 450 ppm, using a coupled atmosphere-ocean general circulation model (Stein et al., 2016; for details and background of the atmosphere-ocean general circulation model, see Knorr et al., 2011; Knorr & Lohmann, 2014). The simulated mean August SST as well as mean sea-ice concentrations for March, June, and September/August for a high (450 ppm) CO₂ level indicate (1) a winter season with a closed sea-ice cover in the central Arctic Ocean decreasing toward the marginal seas, (2) spring season with sea-ice concentrations of 20–60%, and (3) a summer season with ice-free conditions and SSTs >0 °C (Figure 14a), an almost perfect match with the proxy records (Figure 13a; Stein et al., 2016). When doing the model simulation with a relatively low CO₂ level of 278 ppm, however, summer SSTs are below 0 °C, and sea ice still occurs in the central Arctic Ocean during the summer season; that is, there is a mismatch between proxy and model results (Stein et al., 2016). Such lower CO₂ concentrations of 200–300 ppm have been reconstructed in earlier studies (e.g., Pagani et al., 2005), whereas more recent studies have resulted in higher CO₂ proxy estimates of 300–450 ppm for the late-middle Miocene (Tripati et al., 2009; Zhang et al., 2013; cf. also Ruddiman, 2010).

Extrapolating the Mauna Loa atmospheric CO₂ concentration record (https://www.co2.earth/) and the related temperature increase, CO₂ concentrations of 450 ppm and global warming to about 2 °C above the preindustrial level, respectively, will be reached in about 25 years, and the central Arctic Ocean will become ice-free during summer (Figure 14b; Niederdrenk & Notz, 2018; Notz & Stroeve, 2018; Stroeve & Notz, 2018). That means, sea-ice conditions similar to those reconstructed for the late Miocene may occur in about 2–3
decades. Although the sea‐ice conditions might be similar, however, the rate of change was quite different between both situations. Whereas the recent change from a permanent to a seasonal central Arctic Ocean sea‐ice cover (strongly driven by anthropogenic forcing; cf. Notz & Stroeve, 2016) proceeds over a few decades, the corresponding past (natural) change probably occurred over thousands of years.

5.4. A 100‐Ma Summary Record of Arctic Ocean SST and Sea‐Ice Evolution

In the following, the existing central Arctic Ocean SST data are summarized and related to multiproxy records of the global and middle to low latitude climate evolution throughout the last 100 Ma, including the global benthic isotope stack, CO₂ concentrations and SST records, and the Arctic sea ice history (Figure 15; Table 1). Most of the middle to low latitude records used in this study are from ODP sites. The locations of these sites are shown in a distribution map of modern annual SST (Figure 16). A similar
Paleoclimate data from the Arctic that in addition also considers terrestrial climate information was published by O’Regan et al. (2011). The summary presented in Figure 15 highlights where data are available, where the gaps are, and—based on this—where the perspectives and needs for future activities are.

Late Cretaceous SST records based on $\delta^{18}$O and TEX$_{86}$ proxy data from several localities show, independently of their environmental and diagenetic setting, a remarkable similarity, demonstrating that the climatic trends are probably global in nature (e.g., Huber et al., 1995; Jenkyns et al., 1994; O'Brien et al., 2017). There is a general warming trend during the Cenomanian reaching peak values at the Cenomanian/Turonian boundary, followed by a decline through the Coniacian to
| Location/area | Expedition, core | Epoch | Age (Ma) | lat. | SST (°C) | Proxy | Reference |
|---------------|-----------------|-------|----------|------|---------|--------|-----------|
| Arctic Ocean  | FL533           | Cenomanian/Turonian | 93 | 80°N | 24 | Extrapolated | Jenkyns et al. (2004) |
|               | Alpha Ridge FL533| Santonian | 85 | 80°N | 19 | Extrapolated | Jenkyns et al. (2004) |
|               | Alpha Ridge FL533| Campanian | 82 | 80°N | 16 | Extrapolated | Jenkyns et al. (2004) |
|               | Alpha Ridge FL533| Maastrichtian | 70 | 80°N | 15 | Extrapolated | Jenkyns et al. (2004) |
|               | Alpha Ridge FL533| Campanian | 73 | 80°N | 14 | Extrapolated | Jenkyns et al. (2004) |
|               | Alpha Ridge FL533| Maastrichtian | 56 | 80°N | 13 | Extrapolated | Jenkyns et al. (2004) |
| Lomonosov Ridge | IODP Exp 302, ACEX | PETM | 58 | 87.9°N | 23 | TEX86 | Sluijs et al. (2006), Sluijs et al., 2008 |
|               | Lomonosov Ridge | Lowermost Eocene | 55 | 87.9°N | 17 | TEX86 | Sluijs et al. (2006), Sluijs et al., 2008 |
|               | Lomonosov Ridge | Early Eocene (ETM2) | 53.5 | 87.9°N | 27 | TEX86 | Sluijs et al. (2009) |
|               | Lomonosov Ridge | Early Eocene | 50 | 87.9°N | 13 | TEX86 | Sluijs et al. (2008) |
|               | Lomonosov Ridge | Early Eocene (Azolla) | 48.6 | 87.9°N | 9 | TEX86 | Brinkhuis et al. (2006) |
|               | Lomonosov Ridge | Early Eocene | 48.5 (48.4) | 87.9°N | 13 | TEX86 | Brinkhuis et al. (2006) |
|               | Lomonosov Ridge | Middle Eocene | 47.1 (44.2) | 87.9°N | 23 | TEX86 | Stein, Weller, et al. (2014) |
|               | Lomonosov Ridge | Middle Eocene | 46.3 (41.7) | 87.9°N | 18 | Uk37 | Stein, Weller, et al. (2014) |
|               | Lomonosov Ridge | Middle Eocene | 45.0 (41.1) | 87.9°N | 10 | Uk37 | Stein, Weller, et al. (2014) |
|               | Lomonosov Ridge | Middle Eocene | 44.1 (40.6) | 87.9°N | 8 | Uk37 | Sangiorgi et al. (2008) |
|               | Lomonosov Ridge | Middle Eocene | 43.0 (39.1) | 87.9°N | 16 | Uk37 | Stein, Weller, et al. (2014) |
|               | Lomonosov Ridge | Middle Miocene (Eoc.) | 18.2 (16.1) | 87.9°N | 10 | Uk37 | Weller and Stein (2008) |
|               | Lomonosov Ridge | Late Miocene | 11.5 | 87.9°N | 10.5 | Uk37 | Stein et al. (2016) |
|               | Lomonosov Ridge | Late Miocene | 9.1 | 87.9°N | 15.5 | Uk37 | Stein et al. (2016) |
|               | Lomonosov Ridge | Late Miocene | 7.5 | 87.9°N | 8.9 | Uk37 | Stein et al. (2016) |
|               | Lomonosov Ridge | Late Miocene | 6 | 87.9°N | 5.4 | Uk37 | Stein et al. (2016) |
|               | Lomonosov Ridge | Late Miocene | 6 – 9 | 81.2°N | 4 – 7 | TEX86 | Sangiorgi, Brumsack, et al. (2008) |
|               | Yermak Plateau ODP Leg 151, Site 910 | Early Pliocene | 5 | 80°N | 9 | TEX86 | Knies et al. (2014) |
|               | Yermak Plateau ODP Leg 151, Site 910 | Late Pliocene | 2.5 | 80°N | 14 | TEX86 | Knies et al. (2014) |

**Table 1**: Compilation of sea surface temperatures of Late Cretaceous and Cenozoic time intervals from high (Arctic Ocean), middle, and low latitudes, shown in Figure 15.
Table 1
(continued)

| Location/area          | Expedition, core         | Epoch            | Age (Ma) | (Paeo-) lat. | SST (°C) | Proxy | Reference                |
|------------------------|--------------------------|------------------|----------|--------------|---------|-------|--------------------------|
| Rockall Plateau/North Atlantic | ODP Leg 162, Site 982 | Middle Miocene   | 16       | 55°N         | 28      | Uk37  | Herbert et al. (2016)    |
| Rockall Plateau/North Atlantic | ODP Leg 162, Site 982 | Late Miocene     | 12       | 55°N         | 26      | Uk37  | Herbert et al. (2016)    |
| Rockall Plateau/North Atlantic | ODP Leg 162, Site 982 | Late Miocene     | 6        | 55°N         | 18      | Uk37  | Herbert et al. (2016)    |
| Rockall Plateau/North Atlantic | ODP Leg 162, Site 982 | Early Pliocene   | 4.5      | 55°N         | 20      | Uk37  | Herbert et al. (2016)    |
| Rockall Plateau/North Atlantic | ODP Leg 162, Site 982 | Late Pliocene    | 2.5      | 55°N         | 14      | Uk37  | Herbert et al. (2016)    |
| Iceland Plateau/North Atlantic | ODP Leg 151, Site 907   | Late Miocene     | 12       | 70°N         | 20      | Uk37  | Stein et al. (2016)      |
| Iceland Plateau/North Atlantic | ODP Leg 151, Site 907   | Late Miocene     | 6        | 70°N         | 9       | Uk37  | Stein et al. (2016)      |
| Iceland Plateau/North Atlantic | ODP Leg 151, Site 907   | Early Pliocene   | 4.5      | 70°N         | 11      | Uk37  | Herbert et al. (2016)    |
| Low latitudes          |                          |                  |          |              |         |       |                          |
| Pacific and Atlantic oceans | Composite record        |                  | 93       | 0–30°N       | 36      | $\delta^{18}$O, TEX86  | O’Brien et al. (2017) |
| Pacific and Atlantic oceans | Composite record        |                  | 82       | 0–30°N       | 32      | $\delta^{18}$O, TEX86  | O’Brien et al. (2017) |
| Tanzania/East African margin | Composite record        |                  | 70       | 0–30°N       | 28      | $\delta^{18}$O, TEX86  | O’Brien et al. (2017) |
| Tanzania/East African margin | Composite record        |                  | 55       | 20°S         | 31      | TEX86 | Pearson et al. (2007)   |
| Tanzania/East African margin | Composite record        |                  | 50       | 20°S         | 34      | TEX86 | Pearson et al. (2007)   |
| Equatorial Western Atlantic | ODP Leg 154, Site 925   | Late Eocene      | 40       | 5°N          | 28      | Uk37, TEX86 | Liu et al. (2009) |
| Equatorial Western Atlantic | ODP Leg 154, Site 925   | Oligocene        | 33       | 5°N          | 28      | Uk37, TEX86 | Liu et al. (2009) |
| Equatorial East Pacific | ODP Leg 138, Site 850   | Late Miocene     | 12       | 1°N          | 28      | Uk37  | Herbert et al. (2016)   |
| Equatorial East Pacific | ODP Leg 138, Site 850   | Late Miocene     | 6        | 1°N          | 27      | Uk37  | Herbert et al. (2016)   |
| Equatorial East Pacific | ODP Leg 138, Site 850   | Early Pliocene   | 4.5      | 1°N          | 26      | Uk37  | Herbert et al. (2016)   |
| Equatorial East Pacific | ODP Leg 138, Site 850   | Late Pliocene    | 2.5      | 1°N          | 24      | Uk37  | Herbert et al. (2016)   |

Note. Location/areas, expeditions/cores, epochs and age, paleolatitudes, mean SST values, used proxies, and references are listed. For the ACEX data, ages listed in brackets are based on Age Model 2 (Poirier & Hillaire-Marcel, 2011); the other ages are based on Backman et al. (2008). Sites included in the Late Cretaceous composite record from low latitudes are shown in Figure 16.
Campanian/Maastrichtian, also reflected in the SST records from the boreal English Chalk (Jenkyns et al., 1994; Jenkyns et al., 2004) and the Boreal Chalk Sea (Stevns-1/Denmark; Thibault et al., 2016; Figure 15; for location, see Figure 9a). Using the English Chalk SST record that extends into the lowest Maastrichtian, Jenkyns et al. (2004) generated a tentative middle to Late Cretaceous paleotemperature curve for the Arctic Ocean, taking the Maastrichtian TEX$_{86}$-SST value from Core FL-533 as a calibration point (15 °C; Figure 9c) and assuming that the latitudinal gradient did not greatly change over the time period between about 70 and 100 Ma. Based on this (still tentative) approach, average Arctic Ocean SST probably reached values of 20 to 25 °C during much of the Cenomanian-Turonian time interval, decreasing to about 13 °C in the uppermost Campanian (Figure 15; Jenkyns et al., 2004). Under these colder (summer) temperature conditions, winter SST values were probably below 0 °C, allowing some sea-ice formation as suggested from the occasional occurrence of IRD (Figure 9b). During the peak warm Cenomanian-Turonian, conversely, the Arctic Ocean was ice-free year-round as also supported by terrestrial records (O’Regan et al., 2011; see section 5.1). Comparing this high-latitude record with Late Cretaceous SST values from the low latitudes <30°, this suggests a Northern Hemisphere pole-equator temperature gradient of about 1–15 °C (Figure 15; Wilson & Opdyke, 1996; Jenkyns et al., 2004; O’Brien et al., 2017).

Unfortunately, for the upper Maastrichtian/Paleocene interval coinciding with reduced atmospheric CO$_2$ concentrations, no SST and sea-ice proxy data are available from the central Arctic Ocean. The occurrence of glendonites and erratics in otherwise fine-grained sediments of Paleocene/early Eocene age found on Svalbard requires near-freezing temperatures and the presence of at least seasonal sea ice (Figure 15; Spielhagen & Tripati, 2009). If seasonal sea ice occurred around Svalbard, then it should have occurred in the central Arctic Ocean during that time interval a fortiori.

Peak SST values of about 23–27 °C were reached during the PETM, ETM2, and EECO intervals, more or less contemporaneously with maximum atmospheric CO$_2$ concentrations (Figure 15; Sluijs et al., 2006, Sluijs et al., 2009; Weller & Stein, 2008; Brassell, 2014; Stein, Weller, et al., 2014). Under these extreme warm conditions, the Arctic Ocean was ice-free throughout the year. The pole-equator temperature gradient was very similar to that estimated for the Cenomanian/Turonian Thermal Maximum. For a long time, modelling studies failed to reproduce such extreme conditions, that is, the very high Arctic Ocean SST values and the more equable global climate with a reduced pole-equator temperature gradient (e.g., Shellito et al., 2003; Tindall et al., 2010). Huber and Caballero (2011) have demonstrated, however, that with a suitably large radiative forcing, the pattern of high-latitude amplification reconstructed by proxies can be largely, but not perfectly, reproduced.

When interpreting these SST data, one should have in mind that the early Eocene data including the PETM, ETM2, and Azolla events is based on TEX$_{86}$ values (Brinkhuis et al., 2006; Sluijs et al., 2006; Sluijs et al., 2008; Sluijs et al., 2009), whereas the EECO values are based on U$_{37}^{K}$ (Stein, Weller, et al., 2014). Unfortunately no U$_{37}^{K}$-SST values are available for most of the EECO due to the absence of alkenones in this interval (cf.
There are, however, some TEX\textsubscript{86} based SST values (Brinkhuis et al., 2006; Sluijs et al., 2008) that show surprisingly low values of 8–13 °C for the EECO; that is, they are about 10 °C lower than the \( U^K_{17} \)-SST values from close-by depth intervals (Figure 15; Table 1), a difference that is very similar to the Arctic seasonal temperature variability (Basinger et al., 1994; Greenwood & Wing, 1995; Suan et al., 2017; Wolfe, 1994). Thus, one possible explanation for this discrepancy could be that these TEX\textsubscript{86} SST values represent more the winter SST (for further details and discussion, see Weller & Stein, 2008), further supporting year-round ice-free conditions in the central Arctic Ocean under this extreme warmth.

With the post-EECO decrease in CO\textsubscript{2}, the Arctic Ocean SST values decreased, following the global trend (Figure 15; Zachos et al., 2008). The decrease in Arctic Ocean SST values generally follows that found in subpolar records from the Northern and Southern hemispheres (for location of sites, see Figure 16); that is, Site 913 (Liu et al., 2009) and Site 1172 (Bijl et al., 2009), however, is much larger than that of the low latitudes (Site 925; Liu et al., 2009; Bijl et al., 2009). This results in a significantly increased pole-equator temperature gradient to >20 °C, which is, a value similar to that of today. If the Eocene cooling was predominantly controlled by a decrease in atmospheric greenhouse gas concentration that cooled both the high and the low latitudes (Huber & Sloan, 2001; Zachos et al., 2008), additional processes are required to explain the relative stability of tropical SSTs and the more significant cooling at higher latitudes (Bijl et al., 2009). That means, high-latitude climate feedbacks must have had a more important influence on climate change than previously thought. Differences in cloud/water vapor as well as ice-albedo effects have been mentioned as potential positive-feedback mechanism for middle Eocene cooling (Bijl et al., 2009).

The general coincidence of onset and/or extension of sea ice with major SST cooling steps may support the importance of sea ice formation and related ice-albedo effects for the exceptional drop in Arctic Ocean SST. Based on the current (still incomplete) data set, sea ice started to occur when the (summer) SST values decreased below a threshold of about 17 °C, as indicated by the onset of IRD and sea-ice diatoms in the ACEX record (Figures 12 and 15). The datum of this prominent climatic change is at about 46.3 or 42 Ma, depending on the age model that is used.

Despite the ongoing debate about the exact age of the onset of Arctic sea ice, the new semi-quantitative sea-ice and SST records are an important contribution to the ongoing and controversial debate about the reconstruction of the early (pre-Quaternary) Arctic Ocean sea-ice cover. Within this debate, the distinction between seasonal and perennial sea ice is critical, because year-round sea ice in the central Arctic implies very different climate feedback mechanisms (i.e., Earth’s albedo and heat exchange conditions) than an environment with ice-free conditions during summer.

Between about 40 and 10 Ma, the most prominent steps of global climate change on the pathway from Greenhouse into Icehouse conditions are unfortunately not documented in the Arctic SST record so far, that is, the Eocene/Oligocene cooling event near 34 Ma characterized by a drastic drop in atmospheric CO\textsubscript{2} and the onset of massive glaciations on Antarctica (Escutia et al., 2011, 2014; Kennett & Shackleton, 1976) and the mid-Miocene cooling near 14 Ma with a further drop in atmospheric CO\textsubscript{2} and expansion of ice sheets on Antarctica (Shevenell et al., 2004; Tripati et al., 2009; Figure 15). A drastic cooling at the Eocene/Oligocene boundary in the high northern and southern latitudes is, however, supported from the SST records from Site 913 (75°N) and Site 511 (55°S), both showing a sharp contemporaneous drop in SST near 34 Ma (Figure 15; Liu et al., 2009). The mid-Miocene cooling event is reflected in the Northern High Latitudes by a significant increase in IRD flux in the Nordic seas, related to extended ice sheets on Svalbard and Greenland (Knies & Gaina, 2008; Tripati & Darby, 2019).

For the middle-late Miocene, quantitative information about the central Arctic Ocean is very limited as outlined in the previous section. Warm climatic conditions occurred in the Middle Miocene as reflected in the alkenone-based summer SSTs of 10–13 °C determined in a small number of ACEX sediment samples (Figure 15). For the late Miocene, proxy data as well as modeling results indicate ice-free conditions during the summer season (Stein et al., 2016). The latitudinal SST gradient seems to be about 15 °C when comparing the ACEX data with SST data from sites 907 (70°N), 982 (55°N), and 850 (1°N; Herbert et al., 2016; Stein et al., 2016), that is, lower than the modern one (Figure 15). With the middle-late Pliocene cooling, coinciding with a drop in atmospheric CO\textsubscript{2} (Pagani et al., 2010), SST values at the Arctic, subarctic, middle and low-latitude sites contemporaneously decreased. As the drop in temperatures is much more drastic in the higher
latitudes, the pole-equator temperature gradient increased to >20 °C (Figure 15). With the cooling trend, Arctic sea-ice cover became more severe.

For the Pliocene central Arctic Ocean, the available data point to strong temporal and spatial variabilities of the sea-ice cover rather than a stable perennial ice cover (Matthiessen, Knies, et al., 2009; Polyak et al., 2010). Contemporaneously with the late Pliocene global cooling (Figure 15), Northern Hemisphere ice sheets extended onto the shelves (Duk-Rodkin et al., 2004; Knies et al., 2009). Since that time probably also an increased input of meltwaters from ice sheets and low-salinity waters from the Pacific Ocean into the Arctic Ocean via Bering Strait may have caused a freshening of Arctic surface waters facilitating sea-ice formation (Haug et al., 2001; Matthiessen, Knies, et al., 2009; Polyak et al., 2010). Based on sedimentological and micropaleontological data, sea-ice conditions similar to those of today with a perennial sea-ice cover in the central Arctic Ocean probably became more dominant during the last about 1 Ma (Dipre et al., 2018). So far, however, additional multiproxy records distributed throughout the Arctic Ocean are needed.
for a more detailed reconstruction of the history of perennial versus seasonal sea ice and ice-free intervals during the past several million years (Matthiessen, Knies, et al., 2009; Polyak et al., 2010).

6. Outlook: The Challenge for Future Arctic Ocean IODP Drilling

The paleoceanographic and paleoclimate results obtained from ACEX studies were unprecedented and—in combination with additional punctual information from a few short sediment cores—give important insight into the long-term Arctic climate and sea-ice history on its pathway from Greenhouse to Icehouse conditions. Here especially the early onset of seasonal sea ice in the central Arctic Ocean during the middle Eocene and the presence of ice-free conditions during summers throughout the middle-late Miocene have to be highlighted. However, major questions related to the climate history of the Arctic Ocean during early Cenozoic times remain unanswered, partly due to the major mid-Cenozoic hiatus (or strongly reduced sedimentation rates) and partly to the poor recovery of the ACEX record (Backman et al., 2006; Backman et al., 2008; Poirier & Hillaire-Marcel, 2011). In order to decipher the pre-Quaternary climate (and sea-ice) history of this unique and sensitive, but still not well-known region on Earth, long continuous sedimentary records obtained by scientific drilling are needed. Such records are planned to be recovered within a new IODP drilling campaign on the southern Lomonosov Ridge, actually scheduled for late summer/early autumn 2021 (IODP Expedition 377 “Arctic Ocean Paleooceanography (ArcOP); http://www.ecord.org/expedition377/). Based on new seismic and coring data obtained during Polarstern Expedition PS87 in 2014 (Stein, 2015) and Polarstern Expedition PS115/2 in 2018 (Stein, 2019), several locations for potential drill sites have been proposed and further optimized (Figure 17; Stein et al., 2015; Stein, 2019). As primary drill site, Site LR-11A

Figure 18. Schematic illustration of Arctic sea ice cover and circum-Arctic ice sheets during MIS6. (a) Tentative reconstruction of MIS 6 ice shelves and ice sheets. Orange arrows indicate ice flow inferred from geophysical mapping and white arrows hypothesized ice flow. The Eurasian Late Saalian Ice Sheet (MIS 6) is based on Svendsen et al. (2004); the North American Ice Sheet shown is the Late Wisconsinan based on Dyke et al. (2002) and England et al. (2009), assuming that the late Illinios Ice Sheet (MIS 6) covered at least this area in the Arctic region (figure from Jakobsson et al., 2010, supplemented). (b) Cartoon of an ice shelf covering the entire central Arctic Ocean with flow lines generalized from mapped glacial landforms (Jakobsson et al., 2016). AP = Arlis Plateau; CB = Chukchi Borderland; LR = Lomonosov Ridge; MJ = Morris Jesup Rise; YP = Yermak Plateau; x–x′ highlights the crest of the Lomonosov Ridge. (c) Ice sheet configuration (Ehlers & Gibbard, 2007; Jakobsson et al., 2014), including the extended ice sheet on the East Siberian continental margin (Niessen et al., 2013), and sea-ice cover proposed by Stein et al. (2017). Strong katabatic winds related to the ice sheets (shown tentatively as stippled black arrows) were probably responsible for ice-free polynya-type conditions off the major ice sheets, causing phytoplankton and sea-ice algae productivity. Such conditions were recorded in biomarker records from cores PS2138 and PS2757 (see Stein et al., 2017 for further details). Large light blue arrow indicates the Transpolar Drift (TPD; figure from Stein et al., 2017, supplemented). In (a) to (c), the area of the proposed ArcOP Expedition is shown as a red rectangle.
has been selected where about 230 m of Plio-Pleistocene, 460 m of Miocene, and >200 m of Oligocene-Eocene may be recovered.

The primary objectives of ArcOP Expedition 377 (http://www.ecord.org/expedition377/) are mainly based on what we learned from the results of the ACEX Expedition (Backman et al., 2006; Backman & Moran, 2008, 2009; Stein, Weller, et al., 2014, Stein et al., 2015). Some of the key scientific themes and hypotheses to be tested using material and data of the ArcOP Expedition are described shortly in the following:

• History of Arctic Ice Sheets, Sea Ice, and Global Climate

An overall key theme is the detailed reconstruction of the history of the circum-Arctic ice sheets and the sea ice in relationship to the global climate record. Key questions still to be answered include the following: Did the Arctic Ocean climate follow the global trend shown in Figure 15? Are the Early Eocene Climate Optimum (poorly recovered in the ACEX record) and the middle Miocene warmings also reflected in Arctic Ocean records? Did extensive glaciations (such as the Oi-1 Glaciation) develop synchronously in both the Northern and Southern Hemispheres (Zachos et al., 2001; Zachos et al., 2008)? Did major East Siberian ice sheets occur during late Pliocene-Pleistocene as recently proposed by Niessen et al. (2013)? What are the related scale and timing of short- and long-term sea-level changes? What is the past variability of Cenozoic sea ice in terms of frequency, extent, and magnitude—a pressing scientific question—even after the accomplishment of the first ACEX campaign? Within this debate, the distinction between seasonal and perennial sea ice is critical, because year-round sea ice in the central Arctic implies very different climate feedback mechanisms, that is, Earth’s albedo and heat exchange conditions, than an environment with ice-free conditions during summer.

As reconstructed from bathymetric, seismic/acoustic, and sediment core data as well as satellite and aerial photographs and geological field work on land, the Quaternary glacial history of the Arctic Ocean is characterized by the repeated build-up and decay of circum-Arctic ice sheets on the continental shelves, the development and disintegration of ice shelves, and related changes in ocean-circulation patterns and sea ice cover (Ehlers & Gibbard, 2007; Hughes et al., 1977; Jakobsson et al., 2014; Niessen et al., 2013; Svendsen et al., 2004). There is, however, still an ongoing and partly controversial debate about the timing
and extent of maximum glaciations (Figure 18). A comprehensive circumpolar overview of glacial landforms, stratigraphies, and chronologies and their interpretation in terms of glacial history is given by Jakobsson et al. (2014), summarizing the current state of knowledge and identifying key questions arising from this synthesis. Based on new evidence of ice‐shelf groundings on bathymetric highs in the central Arctic Ocean, Jakobsson et al. (2016) recently proposed an extended thick ice shelf covering the entire central Arctic Ocean (Figure 18b) and dated it to MIS 6 (~140 ka). Numerical simulations of a kilometer‐thick Arctic ice shelf seem to be consistent with ice grounding observations (Gasson et al., 2018). Biomarker proxy records from sediment cores recovered at the Siberian and Barents Sea continental margin, however, indicate at least occasionally open‐water conditions during MIS 6, that is, an ice edge situation, that allowed phytoplankton and ice algae production as well as increased flux of terrigenous matter (Figure 18c; Stein et al., 2017). Such polynya‐like open‐water conditions probably resulted from strong katabatic winds in front of the ice sheet (Figure 18c). One possible but still somewhat speculative explanation for these discrepancies could be a temporal succession of different scenarios, indicating a very dynamic system. During maximum glaciation, an extended thick ice shelf causing ice‐shelf grounding (ice rise) with no ice algae production may have occurred, followed by an ice shelf decay and retreat causing polynya‐like conditions with sea ice algae and phytoplankton production triggered by strong katabatic winds during the latest MIS 6 (Stein et al., 2017). A continuous Cenozoic Arctic Ocean sedimentary record representing Quaternary and Neogene/Paleogene cold and warm periods will allow us to test the different hypotheses related to changes in Arctic ice sheets and sea ice through time.

- History of Arctic River Discharge

The proximal location of the proposed ArcOP sites relative to the Siberian margin allows a detailed study of the history of Arctic river discharge and its paleoenvironmental significance. In this context, the Pliocene...
closure of the Isthmus of Panama and/or the Neogene uplift of the Tibetan Plateau and the Mongolian Plateau might have been of particular interest. These plate tectonic processes might have triggered enhanced discharge rates of Siberian rivers and changed the freshwater balance of the Arctic's surface waters, considered to be an important factor for the formation of Arctic sea ice and onset of major glaciations (Figure 19; Driscoll & Haug, 1998; Haug et al., 2001; Wang, 2004). Furthermore, a record of the Siberian river discharge per se might give important insight into the evolution of continental climate in the hinterland through Miocene/Pliocene times.

- High-Resolution Characterization of the Pliocene Warm Period in the Arctic

During the Pliocene warm period, SST in several ocean basins were substantially warmer (Haywood et al., 2005; Lawrence et al., 2006; Marlow et al., 2000), and global mean surface temperature was estimated to be at least 3 °C higher than today (Haywood & Valdes, 2004). How did the Arctic Ocean evolve during the Pliocene warm period and subsequent cooling? How do the marine climate records correlate with terrestrial records obtained from the Siberian Lake El'gygytgyn (Figure 20; Melles et al., 2012; Brigham-Grette et al., 2013)?

- The "Hiatus Problem"

What is the cause of the major hiatus spanning the late Eocene to early Miocene time interval in the ACEX record (based on the original age model of Backman et al., 2008)? Does this hiatus in fact exist, or is it rather an interval of extremely reduced sedimentation rate as proposed by Poirier and Hillaire-Marcel (2011; Figure 10)? If there is a major hiatus, is it related to the subsidence history of Lomonosov Ridge (cf. O’Regan et al., 2008)? Was the hiatus a response to increased bottom water currents during the opening of surface and deep water connections via the Fram Strait (cf. Moore & the Expedition 302 Scientists, 2006)? Additional long sedimentary records from the Arctic Ocean with good microfossil recovery are needed to solve these obvious problems in chronology and related paleoenvironmental and tectonic reconstructions. An improvement of the chronology is also an important prerequisite for calculating rates of climate change, a fundamental aspect to better understand the processes controlling past and present climate change and to distinguish between natural and anthropogenic forcings.

The outcome of the second Arctic Ocean IODP drilling campaign will certainly help to improve our understanding of the complex ocean-atmosphere-ice system in the polar high northern latitudes and its role in the past, modern, and future global climate. In addition, this expedition will provide us the second long continuous sedimentary section representing the Cenozoic time interval to be used for detailed reconstructions of the long-term evolution of the Arctic paleoclimate from the early Cenozoic Greenhouse to the late Cenozoic Icehouse conditions. That means, this will certainly be another major step forward in Arctic Ocean paleoclimate research. Nevertheless, further scientific Arctic drilling expeditions have to follow during the coming years in order to fully understand the role of the Arctic Ocean in the Earth’s system and its change through time (Coakley & Stein, 2010; Stein, 2011). Some active proposals are already in the IODP system, but they are not ready to be scheduled yet (Figure 20). More site survey data and review actions are still needed. Furthermore, some key areas for reconstruction of the early Arctic Ocean climate and also tectonic history are still missing, for example, the Alpha Ridge where Cretaceous sedimentary sections are almost outcropping at the seafloor (cf. section 5.1, Figure 9).

Acknowledgments
This publication is a contribution to the Research Programme FACES II, Topic 3 (The Earth system from a polar perspective: Data, modeling and synthesis) of the Alfred Wegener Institute Helmholtz Centre for Polar und Marine Research (AWI). The study used samples and data provided by AWI (grants AWI-PS17_01 and AWI-PSI15/2_1) and the IODP Program. Many thanks to the reviewers, Helmut Weissert (ETH Zurich, Switzerland), and Matt O’Regan (Stockholm University, Sweden), for numerous constructive suggestions for improving the manuscript.

References
Aagaard, K., & Carmack, E. C. (1985). Thermohaline circulation in the Arctic Mediterranean seas. Journal of Geophysical Research, 90(C3), 4833–4846. https://doi.org/10.1029/JC090iC03p04833
Aagaard, K., Swift, J. H., & Carmack, E. C. (1989). The role of sea ice and other fresh water in the Arctic circulation. Journal of Geophysical Research, 94(C10), 14,485–14,498. https://doi.org/10.1029/JC094iC10p14485
ACIA (2004). Impacts of a warming Arctic: Arctic climate impact assessment (139 pp.). Cambridge University Press, Cambridge. (http://www.acia.uaf.edu)
ACIA (2005). Arctic climate impact assessment (1042 p.). Cambridge: Cambridge University Press.
Alsén, H., Regnery, J., Ashchenazi-Polivoda, S., Mellijsen, A., Ron-Yankovich, L., Abramovich, S., et al. (2013). Sea surface temperature record of a Late Cretaceous tropical Southern Tethys upwelling system. Palaeogeography, Palaeoclimatology, Palaeoecology, 392, 350–358. https://doi.org/10.1016/j.palaeo.2013.09.013
AMAP (2017). Snow, water, ice and permafrost in the Arctic (SWIPA) 2017. Oslo, Norway: Arctic Monitoring and Assessment Programme (AMAP). xiv + 269 pp.
Paleoceanography and Paleoclimatology

Anisimov, O. A., Shiklomanov, N. I., & Nelson, P. E. (1997). Effects of global warming on permafrost and active-layer thickness: Results from transient general circulation models. Global and Planetary Change, 15(3-4), 61–77. https://doi.org/10.1016/S0921-8181(97)00009-X

Arthur, M. A., Dean, W. E., Stow, D. A. V. (1984). Models for the deposition of Mesozoic-Cenozoic fine-grained organic-carbon-rich sediments in the deep sea. In D. A. V. Stow & D. J. W. Piper (Eds.), Fine-grained sediments: Deep-water processes and facies. Geol. Soc. London, Spec. Publ. (15, pp. 527–560). London.

Arthur, M. A., Schlanger, S. O., & Jenkyns, H. C. (1987). The Cenomanian-Turonian Oceanic anoxic event, ii. Paleoceanographic controls on organic-matter production and preservation. In J. Brooks & A. Fleet (Eds.), Marine Petroleum Source Rocks, Geol. Soc. London, Spec. Publ. (Vol. 26, pp. 401–420). London.

Backman, J., Jakobsson, M., Frank, M., Sangiorgi, F., Brinkhuis, H., Stickley, C., et al. (2008). Age model and core-seismic integration for the Cenozoic Arctic coring expedition sediments from the Lomonosov Ridge. Paleoceanography, 23, PAIS03. https://doi.org/10.1029/2007PA001476

Backman, J., & Moran, K. (2008). Introduction to special section on Cenozoic Paleoceanography of the Central Arctic Ocean. Paleoceanography, 23, PAIS01. https://doi.org/10.1029/2007PA001516

Backman, J., & Moran, K. (2009). Expanding the Cenozoic paleoceanographic record in the central Arctic Ocean: IODP Expedition 302 synthesis. Central European Journal of Geoscience, 1, 157–175.

Bakke, T., Hameedi, J., Kimstach, V., Macdonald, R., Melnikov, S., Robertson, A., et al. (1998). Chapter 10, Petroleum hydrocarbons. In A. Backman, J., & Moran, K. (Eds.), Expanding the Cenozoic paleoceanographic record in the central Arctic Ocean: IODP Expedition 302 synthesis. Central European Journal of Geoscience, 1, 157–175.

Belt, S. T., Smik, L., Köseoğlu, D., Knies, J., & Hüs, K. (2019). Evidence from silicoflagellate microfossils for the influence of Eocene fresh surface waters in the Eocene Arctic Ocean. Nature Geoscience, 12(9), 689–693. https://doi.org/10.1038/s41561-019-0360-4

Belt, S. T., Massé, G., Rowland, S. J., Poulin, M., Michel, C., & LeBlanc, B. (2007). A novel chemical fossil of palaeo sea ice: IP25. Nature Geoscience, 1(6), 341–344. https://doi.org/10.1038/nge0876

Belt, S. T., Massé, G., Rowland, S. J., Poulin, M., Michel, C., & LeBlanc, B. (2007). A novel chemical fossil of palaeo sea ice: IP25. Nature Geoscience, 1(6), 341–344. https://doi.org/10.1038/nge0876

Belt, S. T., Smik, L., Köseoğlu, D., Knies, J., & Hüs, K. (2019). A novel biomarker-based proxy for the spring phytoplankton bloom in Arctic and sub-arctic settings—HBI IP25. Earth and Planetary Science Letters, 523, 115703. https://doi.org/10.1016/j.epsl.2019.06.038

Bjil, P. K., Schouten, S., Slijes, A., Reichert, G.-J., Zachos, J. C., & Brinkhuis, H. (2009). Early Palaeogene temperature evolution of the southwest Pacific ocean. Nature, 409, 776–779.

Brassell, S. C. (2014). Climatic influences on the Paleogene evolution of alkenones. Paleoceanography, 29, 255–272. https://doi.org/10.1002/2013PA002576

Brassell, S. C., Eglinton, G., Marlowe, I. T., Pflaumau, U., & Sarntche, M. (1986). Molecular stratigraphy: A new tool for climate assessment. Nature, 320(6058), 129–133. https://doi.org/10.1038/320129a0

Brice, K. L., Arthur, M. A., & Marinovich, L. J. (1996). Late Palaeocene Arctic Ocean shallow-marine temperatures from mollusk stable isotopes. Paleoceanography, 11, 241–249.

Brigham-Grette, J., Melleis, M., Minsky, P., Andreev, A., Tarasov, P., DeConto, R., et al. (2013). Pliocene warmth, extreme polar amplification, and stepped Pleistocene cooling recorded in NE Russia. Science, 340(6139), 1421–1427. https://doi.org/10.1126/science.1233137

Brown, J., Ferrians, O. J. J., Heginbottom, J. A., & Melnikov, E. S. (1997). International Permafrost Association circum-Arctic map of permafrost and ground ice conditions, Scale 1:10,000,000. U.S. Geol. Surv.

Brown, T. A., Belt, S. T., Tatarak, A., & Mundy, C. J. (2014). Source identification of the Arctic sea ice proxy IP25. Nature Communications, 5(1), 14197. https://doi.org/10.1038/ncomms5197

Bruckquardt, H. J. (1988). Geochemistry of Cretaceous black shales from the Atlantic Ocean (DSDP Legs 11, 14, 36 and 41). Chem. Geol., 31, 1–25. https://doi.org/10.1016/0009-2541(80)90064-9

Bukry, D. (1984). Paleoceanic paleoceanography of the Arctic Ocean is constrained by the middle or late Eocene age of USGS Core FR-422: Evidence from silicoflagellates. Geology, 12(4), 199–201. https://doi.org/10.1130/0091-7613(1984)12<199:PPATAO>2.0.CO;2

Cavaliere, D., Gleiser, P., Parkinson, C. L., Comiso, J. C., & Zwally, H. J. (1997). Observed hemispheric asymmetry in global sea ice changes. Science, 278(5340), 1104–1106. https://doi.org/10.1126/science.278.5340.1104

Cavaliere, D., & Parkinson, C. L. (2012). Arctic sea ice variability and trends, 1979–2010. Cryosphere, 6(4), 881–889. https://doi.org/10.5194/ce-6-881-2012
Johannessen, O. M., Bengtsson, L., Miles, M. W., Kuzmina, S. I., Semenov, V. A., Alekseev, G. V., et al. (2004). Arctic climate change, observed and modelled temperature and sea-ice variability. Tellus, 56(4), 328–341.

Johannessen, O. M., Shalina, E. V., & Miles, M. W. (1999). Satellite evidence for an Arctic Sea ice cover in transformation. Science, 286(5446), 1937–1939. https://doi.org/10.1126/science.286.5446.1937

Jokat, W., Geisler, W., Coakley, B., Eagles, G., Eismann, H., Gebhardt, C., et al. (2015). Seismic investigations across the Lomonosov Ridge. In R. Stein (Ed.), The Expedition PS87 of the research vessel Polarstern to the Arctic Ocean in 2014, Reports on Polar and Marine Research (Vol. 688, pp. 49–96). Bremerhaven: Alfred Wegener Institute.

Jørgensen, M. T., Racine, C. H., Walters, J. C., & Osterkamp, T. E. (2001). Permafrost degradation and ecological changes associated with a warming climate in central Alaska. Climatic Change, 48(4), 551–579. https://doi.org/10.1023/A:1005667424292

Kanao, M., Suvorov, V. D., Toda, S., & Tsuboi, S. (2015). Seismicity, structure and tectonics in the Arctic region. Geoscience Frontier, 6(5), 665–677. https://doi.org/10.1016/j.gsf.2014.11.002

Kemper, E. (1987). Das Klima der Kreidezeit. Geologisches Jahrbuch A, 96, 5–185.

Kemett, J. P., & Shackleton, N. J. (1976). Oxygen isotopic evidence for the development of the psychrosphere 38 Myr ago. Nature, 260(5531), 513–515. https://doi.org/10.1038/260513a0

Kent, D. V., & Muttoni, G. (2013). Modulation of Late Cretaceous and Cenozoic climate by variable drawdown of atmospheric CO2 from weathering of basaltic provinces on continents draining through the equatorial humit belt. Climate of the Past, 9(2), 525–546. https://doi.org/10.5194/cp-9-525-2013

Kim, J. H., Schouten, S., Hopmans, E. C., Donner, B., & Sinninghe Damsté, J. S. (2008). Global sediment core-top calibration of the TEX86 paleothermometer in the ocean. Geochimica et Cosmochimica Acta, 72(4), 1154–1173. https://doi.org/10.1016/j.gca.2007.12.010

Knies, J., Cabello-Sanz, P., Belt, S. T., Baranow, S., Fietz, S., & Rossel-Mélo, A. (2014). The emergence of modern sea ice cover in the Arctic Ocean. Nature Communications, 5(1), 5608. https://doi.org/10.1038/ncomms6668

Knies, J., & Gaina, C. (2008). Middle Miocene ice sheet expansion in the Arctic—Views from the Barents Sea. Geochemistry, Geophysics, Geosystems, 9, Q0015. https://doi.org/10.1029/2007GC001824

Knies, J., Matthiessen, J., Vogt, C., Laberg, J. S., Igelsten, B. O., Smelror, M., et al. (2009). The Plio–Pleistocene glaciation of the Barents Sea–Svalbard region: A new model based on revised chronostratigraphy. Quaternary Science Reviews, 28(9–10), 812–829. https://doi.org/10.1016/j.quascirev.2008.12.002

Knies, J., Müller, C., Nowaczyk, N., Vogt, C., & Stein, R. (2000). A multiproxy approach to reconstruct the environmental changes along the Eurasian continental margin over the last 150 kyr. Marine Geology, 163(1–4), 317–344. https://doi.org/10.1016/S0025‐3208(99)00106‐1

Knoor, G., Butzin, M., Micheels, A., & Lohmann, G. (2011). A warm Miocene climate at low atmospheric CO2 levels. Geophysical Research Letters, 38, L20701. https://doi.org/10.1029/2011GL048873

Knoor, G., & Lohmann, G. (2014). Climate warming during Antarctic ice sheet expansion at the Middle Miocene transition. Nature Geoscience, 7(5), 376–381. https://doi.org/10.1038/NGEO2119

Koč, N., Janssen, E., & Halldíason, H. (1993). Paleoceanographic reconstructions of surface ocean conditions in the Greenland, Iceland and Norwegian seas through the last 14 ka based on diatoms. Quaternary Science Reviews, 12(2), 115–140. https://doi.org/10.1016/0277‐3791(93)90012‐B

Koppers, A. A. P., Escutia, C., Inagaki, F., Pälike, H., Saffer, D. M., & Thomas, D. (2019). TEX86 paleothermometer in the ocean. Nature Communications, 10, 5608. https://doi.org/10.1038/s41467-019-10630-7

Krause, G. (1969). Ein Beitrag zum Problem der Erneuerung des Tiefenwassers im Arkona-Bekken. Kieler Meeresforschung, 25, 268–271.

Kristoffersen, Y., 1990. Eurasian basin, in: Grantz, A., Johnson, L., Sweeny, J.F. (Eds.), Petroleum Society. 42. https://doi.org/10.1029/2007PA001497

Kristoffersen, Y. & Tholfsen, A., 2016. Ice drift station FRAM-2014/2015 weekly reports. NERC Technical Reports No. 365, Nansen Environmental and Remote Sensing Center, Bergen, Norway, 333 pp.

Kristoffersen, Y., & Tholfsen, A., 2016. Ice drift station FRAM-2014/2015 weekly reports. NERC Technical Reports No. 365, Nansen Environmental and Remote Sensing Center, Bergen, Norway, 333 pp.

Krylov, A. A., Andreeva, I. A., Vogt, C., Backman, J., Krupskaya, V. V., Grikurov, G. E., et al. (2008). A shift in heavy and clay mineral provenance indicates a middle Miocene onset of a perennial sea-ice cover in the Arctic Ocean. Paleoceanography, 23, PAIS06. https://doi.org/10.1029/2007PA001497

Kwok, R., & Cunningham, G. F. (2015). Variability of Arctic sea ice thickness and volume from CryoSat-2. Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences, 373(2045), 20140517. https://doi.org/10.1098/rsta.2014.0157

Langrock, U., & Stein, R. (2004). Origin of marine petroleum source rocks from the Late Jurassic/Early Cretaceous Norwegian Greenland Sea—Evidence for stagnation and upwelling. Marine and Petroleum Geology, 21(2), 157–176. https://doi.org/10.1016/j.marpetgeo.2003.11.011

Langrock, U., Stein, R., Lipinski, M., & Brumsack, H. J. (2003a). Late Jurassic to Early Cretaceous black shale formation and paleoenvironment in high northern latitudes—Examples from the Norwegian-Greenland-Sea. Paleoceanography, 18(3), n/a. https://doi.org/10.1029/2002PA000867

Langrock, U., Stein, R., Lipinski, M., & Brumsack, H. J. (2003b). Paleoenvironment and sea-level change in the Early Cretaceous Barents Sea—Implications from near-shore marine sапропелes. Geo‐Marine Letters, 23(1), 34–42. https://doi.org/10.1007/s00367-003-0122-5

Larsen, H. C., Saunders, A. D., Cliff, P. D., Beget, J., Wei, W., Spezzaferri, S., & ODP Leg 152 Scientific Party (1994). Seven million years of glaciation in Greenland. Science, 264(5161), 952–955. https://doi.org/10.1126/science.264.5161.952

Lawrence, K. T., Liu, Z., & Herbert, T. D. (2006). Evolution of the eastern tropical Pacific through Plio-Pleistocene glaciation. Science, 312(5770), 79–83. https://doi.org/10.1126/science.1120395

Lea, C. H., Elderfield, H., & Wilson, P. A. (2000). Cenozoic deep-sea temperatures and global ice volumes from Mg/Ca in benthic foraminiferal calcite. Science, 287(5451), 269–272. https://doi.org/10.1126/science.287.5451.269

Leith, T. L., Weiss, H. M., Merk, A., Arhus, N., Elvehakk, G., Embry, A. F., et al. (1992). Mesozoic hydrocarbon source-rocks of the Arctic region. In T. O. Vorren, et al. (Eds.), Arctic geology and petroleum potential, NPF Spec. Publ. (Vol. 2, pp. 1–25). Tromsø: Norwegian Petroleum Society.

Lenniger, M., Norh-Hansen, H., Hills, L. V., & Bjerrum, C. J. (2014). Arctic black shale formation during Cretaceous Oceanic Anoxic Event 2. Geology, 42(9), 799–802. https://doi.org/10.1130/G35732.1
Limoges, A., Ribeiro, S., Weckström, K., Heikilli, M., Zamelczyk, K., Andersen, T. J., et al. (2018). Linking the modern distribution of biogenic proxies in High Arctic Greenland shelf sediments to sea ice, primary production, and Arctic-Atlantic inflow. *Journal of Geophysical Research: Biogeosciences*, 123(3), 760–786. https://doi.org/10.1002/2017JG003840

Lipinski, M., Warning, B., & Brumsack, H.-J. (2003). Trace metal signatures of Jurassic/Cretaceous black shales from the Norwegian Shelf and the Barents Sea. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 190, 459–475. https://doi.org/10.1016/S0031-0182(02)00619-3

Littler, K., Roïsson, S. A., Bown, P. R., Nederbragt, A. J., & Pancost, R. D. (2011). High sea-surface temperatures during the Early Cretaceous Epoch. *Nature Geoscience*, 4(3), 169–172. https://doi.org/10.1038/ngeo1081

Liu, Z., Pagani, M., Zinniker, D., DeConto, R., Huber, M., Brinkhuis, H., et al. (2009). Global cooling during the Eocene-Oligocene climate transition. *Science*, 322(5910), 1187–1190. https://doi.org/10.1126/science.1161088

Lückge, A., Bousafir, M., Lalier-Verges, E., & Littke, R. (1996). Comparative study of organic matter preservation in immature sediments along the continental margins of Peru and Oman, part I: Results of petrographical and bulk geochemical data. *Organic Geochemistry*, 24(4), 437–451. https://doi.org/10.1016/0146-6380(96)00045-9

Macdonald, R. (1996). Awakenings in the Arctic. *Nature*, 380(6572), 286–287. https://doi.org/10.1038/380266a0

Markwick, P. J. (1998). Fossil crocodilians as indicators of Late Cretaceous and Cenozoic climates: Implications for using palaeontological data in reconstructing palaeoclimate. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 137(3–4), 205–271. https://doi.org/10.1016/S0031-0182(98)00108-9

Marlow, J. R., Lange, C. B., Wefer, G., & Rosell-Melé, A. (2000). Upwelling intensification as part of the Pliocene-Pleistocene climate transition. *Science*, 290(5500), 2288–2291. https://doi.org/10.1126/science.290.5500.2288

Marlowe, I. T., Brassell, S. C., Eglinton, G., & Green, J. C. (1984). Long chain unsaturated ketones and esters in living algae and marine sediments. *Organic Geochemistry*, 6, 135–141. https://doi.org/10.1016/0146-6380(84)90034-2

Marlowe, I. T., Brassell, S. C., Eglinton, G., & Green, J. C. (1989). Long-chain alkenones and alkyl alkenoates and the fossil fornic record of marine sediments. *Chemical Geology*, 88(3–4), 349–375. https://doi.org/10.1016/0009-2541(90)90098-R

Martin, J. M., Guan, D. M., Elbaz-Poulat, F., Thomas, A. J., & Gordeev, V. V. (1993). Preliminary assessment of the distributions of some trace elements (As, Cd, Cu, Fe, Ni, Pb and Zn) in a pristine aquatic environment: The Lena River estuary. (Russia). *Marine Chemistry*, 43(1–4), 185–199. https://doi.org/10.1016/0304-4203(93)90024-C

Matthews, R. K., & Frolich, C. (2002). Maximum flooding surfaces and sequence boundaries: Comparisons between observations and orbital forcing in the Cretaceous and Jurassic (65-190 Ma). *Arctic Alpine Research*, 34(4), 353–368. https://doi.org/10.1657/0004-086X-34.4.353

Mathiessen, J., Brinkhuis, H., Poulsen, N., & Smelror, M. (2009). Decadredritta martinheadii Manum 1997—A stratigraphically and palaeoenvironmentally useful Miocene acritarch of the high northern latitudes. *Micropaleontology*, 55, 171–186.

Mathiessen, J., Knies, J., Nowaczyk, N. R., & Stein, R. (2001). Late Quaternary diatomfossilgate cyst stratigraphy at the Eurasian continental margin, Arctic Ocean: Indications for Atlantic water inflow in the past 150,000 years. *Global and Planetary Change*, 31(1–4), 65–86. https://doi.org/10.1016/S0921-8181(01)00113-8

Mathiessen, J., Knies, J., Vogt, C., & Stein, R. (2009). Pliocene palaeoceanography of the Arctic Ocean and subarctic seas. *Philosophical Transactions of the Royal Society*, 367(1861), 21–48. https://doi.org/10.1098/rsta.2008.0203

McClelland, J. W., Holmes, R. M., Dunton, K. H., & Macdonald, R. W. (2012). The Arctic Ocean estuary. *Estuaries and Coasts*, 35(2), 353–368. https://doi.org/10.1007/s12237-010-9357-3

Méhay, S., Keller, C. E., Bernasconi, S. M., Weissert, H., Erba, E., Bottini, C., & Hochuli, P. A. (2009). A volcanic CO2 pulse triggered the Cretaceous Oceanic Anoxic Event 1a and a biocalcification crisis. *Geology*, 37(9), 819–822. https://doi.org/10.1130/G30100A.1

Melles, M., Brigham-Grette, J., Minyuk, P. S., Nowaczyk, N. R., Wennrich, V., DeConto, R. M., et al. (2012). 2.8 million years of Arctic ice reconstructions in the northern Norwegian Seaway: A key for understanding Late Jurassic to Early Cretaceous paleoenvironments. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 371(3–4), 315–320. https://doi.org/10.1016/j.palaeo.2012.12.007

McConnaughey, T. A., & Stein, A. B. (2001). Age as part of the Pliocene/Pleistocene transition. *Paleoceanography*, 16(1), 1–19. https://doi.org/10.1029/PA020001p00001

Miller, K. G., Wright, J. D., & Browning, J. V. (2005). Visions of ice sheets in a greenhouse world. *Marine Geology*, 217(3–4), 215–231. https://doi.org/10.1016/j.margeo.2005.02.007

Milliman, J. D., & Meade, R. H. (1983). World-wide delivery of river sediment to the oceans. *Journal of Geology*, 91(1), 1–21. https://doi.org/10.1086/628741

Moore, T. C., & the Expedition 302 Scientists (2001). Sedimentation and subsidence history of the Lomonosov Ridge, in Arctic Coring Expedition (ACEX), Proc. Integ. Ocean Drill. Program. 302. 2002. https://doi.org/10.2204/iodp.proc.302.105.2006

Moriarty, P., Meade, R. H., van Storch, J.-L., & Rosell-Melé, A. (1998). Estimation of the ocean surface temperature dataset from the eastern South Atlantic and the global ocean (60°N–60°S). *Geochimica et Cosmochimica Acta*, 62(10), 1757–1772. https://doi.org/10.1016/S0016-7037(98)00097-0

Mueller, R. D., Sirockias, M., Gaia, C., & Roest, W. R. (2008). Age, spreading rates and spreading symmetry of the world’s ocean crust. *Geochemistry, Geophysics, Geosystems*, 9, Q04006. https://doi.org/10.1029/2007GC001743

Mutterlose, J., Brumsack, H., Fölsch, S., Hay, W., Klein, C., Langrock, U., et al. (2003). The Greenland-Norwegian Seaway: A key for understanding Late Jurassic to Early Cretaceous palaeoenvironments. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 181(1), 1010. https://doi.org/10.1016/S0031-0182(01)00625

Mutterlose, J., & Kessels, K. (2000). Early Cretaceous calcareous nanofossils from high latitudes: Implications for palaeobiogeography and palaeoclimate. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 160(3–4), 347–372. https://doi.org/10.1016/S0031-0182(00)00082-1
Mutterlose, J., Malkoe, M., Schouten, S., Sinninghe Damsté, J. S., & Forster, A. (2010). TEX̝{\textsubscript{13}C} and stable δ¹³C paleothermometry of Early Cretaceous sediments: Implications for Paleocene ecology and paleotemperature proxy application. *Earth and Planetary Science Letters*, 298(3–4), 286–298. https://doi.org/10.1016/j.epsl.2010.07.043

Nelson, F. E., Shiklomanov, N. I., & Mueller, G. R. (1999). Variability of active-layer thickness at multiple spatial scales, northcentral Alaska, U.S.A. In *Arctic, Antarctic, and Alpine Research*, 31, 158–165.

Nghiem, S. V., Rigor, I. G., Perovich, D. K., Cline-Clolon, P., Weatherly, J. W., & Neumann, G. (2007). Rapid reduction of Arctic perennial ice. *Geophysical Research Letters*, 34, L19504. https://doi.org/10.1029/2007GL031138

Niederdrnk, A. L., & Noz, D. (2018). Arctic ice sea in a 1.5°C warmer world. *Geophysical Research Letters*, 45(4), 1963–1971. https://doi.org/10.1002/2017GL076159

Nies, H., Harms, I. H., Karcher, M. J., Dethleff, D., Bahe, C., Kuhlmann, G., et al. (1998). Anthropogenic radioactivity in the Nordic Seas—Results of a joint project. *Deutsche Hydrographische Zeitschrift (Geom. J. Hydrogr.)*, 50(4), 313–343. https://doi.org/10.1007/BF02964228

Niessen, F., Hong, J. K., Hegewald, A., Matthiessen, J., Stein, R., Kim, H., et al. (2013). Repeated Pleistocene glaciation of the East Siberian continental margin. *Nature Geoscience*, 6(10), 842–846. https://doi.org/10.1038/NGEO1904

Nørgaard, N., Nørgaard, N., Nies, H., Harms, I. H., Karcher, M. J., Dethleff, D., Bahe, C., Kuhlmann, G., et al. (1998). Anthropogenic radioactivity in the Nordic Seas—Results of a joint project. *Deutsche Hydrographische Zeitschrift (Geom. J. Hydrogr.)*, 50(4), 313–343. https://doi.org/10.1007/BF02964228

O'Regan, M., Williams, C. J., Frey, K. E., & Jakobsson, M. (2011). A synthesis of the long-term trends of Arctic seafloor sedimentation and climate change. *Paleoceanography*, 18(3), 1963. https://doi.org/10.1029/2009PA001781

O'Regan, M., King, J., Backman, J., Jakobsson, M., Pälike, H., Moran, K., et al. (2008). Constraints on the Pleistocene chronology of sediments in the Nordic Seas. *Paleoceanography*, 23, PA1S19. https://doi.org/10.1029/2007PA001551

O'Regan, M., Williams, C. J., Frey, K. E., & Jakobsson, M. (2011). A synthesis of the long-term paleoclimatic evolution of the Arctic. *Oceanography*, 24(3), 66–80. https://doi.org/10.5670/oceanog.2011.57

Pagani, M., Liu, Z., LaRiviere, J., & Ravelo, A. C. (2010). High Earth system climate sensitivity determined from Pliocene carbon dioxide concentrations. *Nature Geoscience*, 3, 27–30. https://doi.org/10.1038/ngeo724

Pagani, M., Zachos, J. C., Freeman, K. H., Tipple, B., & Bohaty, S. (2005). Marked decline in atmospheric carbon dioxide concentrations during the Paleocene. *Science*, 309(5734), 590–603. https://doi.org/10.1126/science.1110063

Park, K., Kang, S. M., & Kim, D. (2018). Contrasting local and remote impacts of surface heating on polar warming and amplification. *American Meteorological Society*, 31, 3155–3166.

Parish, J. T., & Spicer, R. A. (1988). Late Cretaceous terrestrial vegetation: A near-polar temperature curve. *Geology*, 16(1), 22–25. https://doi.org/10.1130/0091-7613(1988)016<0022:LMVATP>2.0.CO;2

Pearson, P. N., & Palmer, M. R. (2000). Atmospheric carbon dioxide concentrations over the past 60 million years. *Nature*, 406(6879), 695–699. https://doi.org/10.1038/35021000

Pearson, P. N., van Dongen, B. E., Nicholas, C. J., Pancost, R. D., Schouten, S., Singano, J. M., & Wade, B. S. (2007). Stable warm tropical climate through the Eocene Epoch. *Geology*, 35(3), 211–214. https://doi.org/10.1130/G23175A.1

Peterson, B. J., Holmes, R. M., McClelland, J. W., Vorosmarty, C. J., Lammers, R. B., Shiklomanov, A. I., et al. (2002). Increasing river discharge to the Arctic Ocean. *Science*, 298(5591), 2171–2173. https://doi.org/10.1126/science.1077445

Petrov, O., Morozov, A., Shokalsky, S., Kashubin, S., Artemieva, I. M., Sobolev, N., et al. (2016). Crustal structure and tectonic model of the Arctic region. *Earth Science Reviews*, 154, 29–71. https://doi.org/10.1016/j.earscirev.2015.11.013

Pflüger, S., Gascard, J.-C., Wolfgeng, I., Mudie, P., & Abelmann, A. (1989). Particle-laden Eurasian Arctic sea ice: Observations from July and August 1987. *Polar Research*, 7(1), 59–66. https://doi.org/10.1111/j.1755-8369.1989.tb00604.x

Pflüger, S. L., Colon, R., Nürnberg, D., Eicken, H., & Rigor, I. (1997). Reconstructing the origin and trajectory of drifting Arctic sea ice. *Journal of Geophysical Research*, 102(C6), 12575–12586. https://doi.org/10.1029/96JC03980

Poirier, A., & Hillaire-Marcel, C. (2011). Improved Os-isotope stratigraphy of the Arctic Ocean. *Geophysical Research Letters*, 38, n/a. https://doi.org/10.1029/2011GL047953

Polnyak, L., Alley, R. B., Andrews, J. T., Brigham-Grette, J., Cronin, T. M., Darby, D. A., et al. (2010). History of sea ice in the Arctic. *Quaternary Science Reviews*, 29(15–16), 1757–1778. https://doi.org/10.1016/j.quascirev.2010.02.010

Pound, M. J., Haywood, A. M., Salzmann, U., Riding, J. B., Hunt, D. J., & Hunter, S. J. (2011). A TORTONIAN (Late Miocene, 11.61–7.25 Ma) global vegetation reconstruction. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 308(1–4), 29–45. https://doi.org/10.1016/j.palaeo.2010.11.029

Prühl, F. G., & Wakeham, S. G. (1987). Calibration of unsaturation pattern in long-chain ketone compositions for palaeotemperature assessment. *Nature*, 330(6146), 367–369. https://doi.org/10.1038/330367a0

Quaternary Science Reviews
Price, G. D., Ruffell, A. H., Jones, C. E., Kalin, R. M., & Mutterlose, J. (2009). Isotopic evidence for temperature variation during the Early Cretaceous (late Ryazanian-mid Hauterivian). *Journal of the Geological Society of London*, 155, 335–343.

Rachold, V., Bobishyanov, D. Y., Grigoriev, M. N., Hubberten, H. W., Junker, R., Kunitsky, V. V., et al. (2007). Nearshore Arctic subsea permafrost in transition. *Eos, 88*(13), 149–150. https://doi.org/10.1029/2007EO130001

Rachold, V., Eicken, H., Gordeev, V. V., Grigoriev, M. N., Hubberten, H.-W., Lisitzin, A. P., et al. (2004). Modern terrigenous organic carbon input to the Arctic Ocean. In R. Stein, & R. W. Macdonald (Eds.), *The organic carbon cycle in the Arctic Ocean*, (pp. 33–56). Heidelberg: Springer Verlag. https://doi.org/10.1007/978-3-642-18912-9_8

Rais, F., Louis-Schmid, B., Bernasoni, S. M., & Weissert, H. (2007). Paleoceanographic and palaeoclimatic reorganization around the Middle–Late Jurassic transition. *Palurogeography, Palaeoclimatology, Palaeoecology*, 251(3-4), 527–546. https://doi.org/10.1016/j.palaeo.2007.05.008

Raymo, M. E., Jansen, E., Blum, P., & Herbert, T. D. (1999). *Proc. ODP*. *Sci. Results*, (Vol. 162). College Station, TX: Ocean Drilling Program.

Raymo, M. E., Rind, D., & Ruddiman, W. F. (1990). Climatic effects of reduced Arctic sea ice limits in the GISS II general circulation model. *Paleoceanography*, 5(3), 367–382. https://doi.org/10.1029/PA005i003p00367

Rekant, P., Cherkashov, G., Vanzei, B., & Kruitsky, P. (2005). Submarine permafrost in the nearshore zone of the southwestern Kara Sea. *Geo-Marine Letters*, 25, 167–182.

Ribeiro, S., Seijr, M. K., Limoges, A., Heikkilä, M., Andersen, T. J., Tallberg, P., et al. (2017). Sea ice and primary production proxies in surface sediments from a High Arctic Greenland fjord: Spatial distribution and implications for palaeoenvironmental studies. *Ambio*, 46, S106–S118.

Robinson, S. A., Dickson, A. J., Pain, A., Jenkyns, H. C., O’Brien, C. L., Farnsworth, A., & Lunt, D. J. (2019). Southern Hemisphere sea surface temperatures during the Cenomanian–Turonian: Implications for the termination of Anoxic Event 2. *Geology*, 47(2), 131–134. https://doi.org/10.1130/1049-2525.1

Romanovski, N. N., Hubberten, H. W., Gavrilov, A. V., Eliseeva, A. A., & Tiperko, G. S. (2005). Offshore permafrost and gas hydrate stability zone on the shelf of the East Siberian Seas. *Geo-Marine Letters*, 25(2-3), 167–182. https://doi.org/10.1007/s00367-004-0198-6

Romanovski, N. N., Hubberten, H.-W., Gavrilov, A. V., Tumskoy, V. E., & Kholodov, A. L. (2004). Permafrost of the east Siberian Arctic shelf and coastal lowlands. *Quaternary Science Reviews*, 23(11-13), 1359–1369. https://doi.org/10.1016/j.quascirev.2003.12.014

Romanovsky, V. E., Smith, S. L., & Christiansen, H. H. (2010). Permafrost thermal state in the polar Northern Hemisphere during the International Polar Year 2007–2009: A synthesis. *Permafrost and Periglacial Processes*, 21(2), 106–116. https://doi.org/10.1002/pp.689

Royer, D. L. (2010). Fossil soils constrain ancient climate sensitivity. *Proceedings of the National Academy of Sciences*, 107(2), 517–518. https://doi.org/10.1073/pnas.0913188107

Ruddiman, W. F. (2010). A paleoclimatic enigma? *Science*, 328(5980), 838–839. https://doi.org/10.1126/science.1188292

Rudels, B., Jones, E. P., Anderson, L. G., & Kattner, G. (1994). On the intermediate depth waters of the Arctic Ocean. *Geophysical Research Letters*, 21(9), 129–132. https://doi.org/10.1029/94GL00482

Sakshaug, E. (2004). Primary and secondary production in the Arctic Seas. In R. Stein, & R. W. Macdonald (Eds.), *The organic carbon cycle in the Arctic Ocean*, (pp. 33–56). Heidelberg: Springer Verlag. https://doi.org/10.1007/978-3-642-18912-9_3

Salpin, M., Schnyder, J., Suan, G., & Suc, J. (2005). Evolution of the Arctic marine biota: A catastrophic trend or a slow change of flow through the Barents and Kara Seas. *Ambio*, 34(2), 130–137. https://doi.org/10.1579/0044-7447-34.2.130

Scourfield, E., Bockheim, J., Canadell, J., Euskirchen, E., Field, C., Goryachkin, S., et al. (2008). Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle. *Bioscience*, 58(8), 701–714. https://doi.org/10.1641/0006-3568(2008)088[0701:VOPCET]2.0.CO;2

Scotese, C. R. (2001). *Atlas of Earth history. PALEOMAP Project*. Arlington: Department of Geology, University of Texas at Arlington. ISBN: 0-9700290-9-9.

Screen, J. A., & Deser, C. (2019). Pacific Ocean variability influences the time of emergence of a seasonally ice-free Arctic Ocean. *Geophysical Research Letters*, 46(4), 2222–2231. https://doi.org/10.1029/2018GL081393
Seidenkrantz, M. S. (2013). Benthic foraminifera as palaeo sea-ice indicators in the subarctic realm—Examples from the Labrador Sea-Baffin Bay region. Quaternary Science Reviews, 79, 135–144. https://doi.org/10.1016/j.quascirev.2013.03.014

Semiletov, I. P., Savelieva, N. L., Weller, G. E., Pipko, I. L., Fugach, S. P., Guvik, A. Y., & Vasilyevkaya, L. N. (2000). The dispersion of Siberian river flows into coastal waters: Meteorological, hydrological and hydrochemical aspects. In E. L. Lewis (Ed.), The Freshwater Budget of the Arctic Ocean, (pp. 323–366). Dordrecht: Kluwer Academic Publishers. https://doi.org/10.1007/978-94-011-4132-1_15

Serreze, M. C., & Barry, R. G. (2011). Processes and impacts of Arctic amplification: A research synthesis. Global and Planetary Change, 77(1-2), 85–96. https://doi.org/10.1016/j.gloplacha.2011.03.004

Serreze, M. C., Holland, M. M., & Stroeve, J. (2007). Perspectives on the Arctic’s shrinking sea-ice cover. Science, 315(5818), 1533–1536. https://doi.org/10.1126/science.1139426

Serreze, M. C., Walsh, J. E., Chapin, F. S., Osterkamp, T., Dyurgerov, M., Romanovsky, V., et al. (2000). Observational evidence of recent change in the northern high-latitude environment. Climatic Change, 46(1-2), 159–207. https://doi.org/10.1023/A:1005504031923

Shackleton, N. J., & Kennett, J. P. (1975). Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon isotope analysis in DSDP Sites 277, 279, and 281. Initial Reports - Deep Sea Drilling Project, 29, 743–755.

Shellito, C. J., Sloan, L. C., & Huber, M. (2003). Climate model sensitivity to atmospheric CO2 levels in the early-middle Paleogene. Palaeogeography, Palaeoclimatology, Palaeoecology, 183(1), 113–123. https://doi.org/10.1016/S0031-0182(02)00718-6

Shepherd, G. E., Müller, R. D., & Seton, M. (2013). The tectonic evolution of the Arctic since Pangea breakup: Integrating constraints from surface geology and geophysics with mantle structure. Earth Sciences Reviews, 124, 148–183. https://doi.org/10.1016/j.earscirev.2013.05.012

Shevenell, A. E., Kennett, J. P., & Lea, D. W. (2004). Middle Miocene Southern Ocean cooling and Antarctic cryosphere expansion. Science, 305(5691), 1766–1770. https://doi.org/10.1126/science.1100061

Sigmond, M., Pyne, J. C., & Swart, N. C. (2018). Ice-free Arctic projections under the Paris Agreement. Nature Climate Change, 8(5), 404–408. https://doi.org/10.1038/s41558-018-0124-y

Singh, R. K., Maheshwari, M., Oza, S. R., & Kumar, R. (2013). Long-term variability in Arctic sea surface temperatures. Polar Science, 7(3-4), 233–240. https://doi.org/10.1016/j.polar.2013.10.003

Sinninghe Damsté, J. S., van Bentum, E. C., Reichart, G.-J., Pross, J., & Schouten, S. (2010). A CO2 decrease-driven cooling and increased latitudinal temperature gradient during the mid-Cretaceous Arctic Anoxic Event 2. Earth and Planetary Science Letters, 293(1–2), 97–103. https://doi.org/10.1016/j.epsl.2010.02.027

Sluijs, A., Röhl, U., Schouten, S., Brumstak, H.-J., Sangiorgi, F., Sinninghe Damsté, J.-S., & Brinkhuis, H. (2008). Arctic late Palaeocene–early Eocene palaeoenvironments with special emphasis on the Palaeocene-Eocene Thermal Maximum (Lomonosov Ridge, Integrated Ocean Drilling Program Expedition 302). Palaeoceanography, 23, PA1S11. https://doi.org/10.1029/2007PA001495

Sluijs, A., Schouten, S., Donders, T. H., Schoon, P. L., Röhl, U., Reichart, G.-J., et al. (2009). Warm and wet conditions in the Arctic region during Eocene Thermal Maximum 2. Nature Geoscience, 2(11), 777–780. https://doi.org/10.1038/ngeo668

Sluijs, A., Schouten, S., Pagani, M., Woltering, M., Brinkhuis, H., Damsté, J. S. S., et al., & the Expedition 302 Scientists (2006). Subtropical Arctic Ocean temperatures during the Palaeocene/Eocene thermal maximum. Nature, 441(7093), 610–613. https://doi.org/10.1038/nature04668

Smik, L., Cabeo-Sanz, P., & Belt, S. T. (2016). Semi-quantitative estimates of paleo Arctic sea ice concentration based on source-specific highly branched isoprenoid alkenes: A further development of the PIP25 index. Organic Geochemistry, 92, 63–69. https://doi.org/10.1016/j.orggeochem.2015.12.007

Spencer, A. M., Embry, A. F., Gastil, D. L., Stoupakova, A. V., & Sorensen, K. (Eds) (2011). Arctic petroleum geology. Geological Society, London, Memoirs, 35, 818 pp(1), 1–15. https://doi.org/10.1144/M35.1

Spicer, R. A., & Herman, A. B. (2010). The Late Cretaceous environment of the Arctic: A quantitative reassessment based on plant fossils. Palaeogeography, Palaeoclimatology, Palaeoecology, 295(3-4), 423–442. https://doi.org/10.1016/j.palaeo.2010.02.025

Spinlacher, R., Baumann, K.-H., Erlenkeuser, H., Nowaczyk, N. R., Noergaard-Pedersen, N., Vogt, C., & Weisell, D. (2004). Arctic Ocean deep-sea record of Northern Eurasian ice sheet history. Quaternary Science Reviews, 23(11-13), 1455–1483. https://doi.org/10.1016/j.quascirev.2003.12.015

Spinhagen, R. F., & Bauch, H. A. (2015). The role of Arctic Ocean freshwater during the past 200 ky. Arktos, 1(1), 18. https://doi.org/10.1007/s41063-015-0013-9

Spinhagen, R. F., & Tripati, A. (2009). Evidence from Svalbard for near-freezing temperatures and climate oscillations in the Arctic during the Palaeocene and Eocene. Palaeoceanography, Palaeoclimatology, Palaeoecology, 278(1-4), 48–56. https://doi.org/10.1016/j.palaeo.2009.04.012

Sreen, G., Kaleschke, L., & Heygster, G. (2008). Sea ice remote sensing using AMSR-E 89 GHz channels. Journal of Geophysical Research, 113, C02S03.

Srivastava, S. P., Arthur, M. A., Clement, B., et al. (1989). Proc. ODP, Sci. Results, (Vol. 105). College Station, TX: Ocean Drilling Program.

St. John, K. (2008). Cenozoic ice-raising history of the Central Arctic Ocean: Terrigenous sands on the Lomonosov Ridge. Palaeogeography, Palaeoclimatology, Palaeoecology, 23, PA1S05. https://doi.org/10.1016/2007PA001483

Stein, R. (2007). Upper Cretaceous/lower Tertiary black shales near the North Pole: Organic-carbon origin and source-rock potential. Marine and Petroleum Geology, 24(2), 67–73. https://doi.org/10.1016/j.marpetgeo.2006.10.002

Stein, R. (2008). Arctic Ocean sediments: Processes, proxies, and palaeoenvironment, Developments in Marine Geology, (Vol. 2). Amsterdam: Elsevier. 587 pp

Stein, R. (2011). The great challenges in Arctic Ocean palaeoceanography. IOP Conference Series: Earth and Environmental Science (Vol. 14). IOP Publishing Ltd. https://doi.org/10.1088/1755-1315/14/1/012001

Stein, R. (Ed) (2015). The Expedition PS107 of the research vessel Polarstern to the Arctic Ocean in 2014. Reports on Polar and Marine Research 688. Bremerhaven: Alfred Wegener Institute for Polar and Marine Research 688, 273 pp. http://epic.awi.de/37728/1/PS107_0688_2015.pdf

Stein, R. (2017). From Greenhouse to Icehouse: The late Mesozoic-Cenozoic Arctic Ocean sea-ice and climate history. Polarforschung, 87, 61–78.

Stein, R. (2019). The Expedition PS115/2 of the research vessel POlarStERN to the Arctic Ocean in 2018. Reports on Polar and Marine Research. Bremerhaven: Alfred Wegener Institute for Polar and Marine Research 728, 249 p. https://doi.org/10.2312/BzPM_0728_2019, https://epic.awi.de/id/eprint/49226/
Stein, R., Blackman, D., Inagaki, F., & Larsen, H.-C. (Eds.) (2014). Earth and life processes discovered from subseafloor environment—A decade of science achieved by the Integrated Ocean Drilling Program (IODP). In Series Developments in Marine Geology, (Vol. 7). Amsterdam/New York: Elsevier. 800 pp

Stein, R., Bouscsein, B., Fahl, K., García de Oteyza, T., Knies, J., & Niessen, F. (2001). Accumulation of particulate organic carbon at the time of the last glacial maximum: Evidence from the IODP Arctic Coring Expedition (ACEX). In R. Stein, D. Blackman, F. Inagaki, & H. Larsen (Eds.), Earth and life processes discovered from subseafloor environment—a decade of science achieved by the Integrated Ocean Drilling Program (IODP), Series Developments in Marine Geology, (Vol. 7, pp. 259–293). Amsterdam/New York: Elsevier.

Stein, R., Fahl, K., Backman, I., Brinkhuis, H., Morán, K., & Pilík, E. (2014). Cenozoic Arctic Ocean climate history: Some highlights from the IODP Arctic Coring Expedition (ACEX). In R. Stein, D. Blackman, F. Inagaki, & H.-C. Larsen (Eds.), Earth and life processes discovered from subseafloor environment—a decade of science achieved by the Integrated Ocean Drilling Program (IODP), Series Developments in Marine Geology, (Vol. 7, pp. 303–316). https://doi.org/10.1016/978-0-08-098822-1

Stein, R., Fahl, K., Gierz, F., Niessen, F., & Lohmann, G. (2017). Arctic Ocean sea ice cover during the penultimate glacial and the last interglacial. Nature Communications, 8, 373. https://doi.org/10.1038/s41467-017-00552-1

Stein, R., Fahl, K., & Müller, J. (2012). Proxy reconstruction of Arctic Ocean sea ice history: “From IRD to IP25”. Polarforschung, 82, 37–71. doi:10.1007/s40432-012-0001

Stein, R., Fahl, K., Schreck, M., Knorr, G., Niessen, F., Forwick, M., et al. (2016). Evidence for ice-free summers in the late Miocene central Arctic Ocean. Nature Communications, 7, 11148. https://doi.org/10.1038/ncomms11148

Thiede, J., Clark, D. L., & Hermann, S. (1990). Late Mesozoic and Cenozoic paleoceanography of the northern polar oceans, in: The geology of North America, (pp. 427–458). The Arctic Ocean Region: Vol. L

Thiede, J., Myhre, A. M., Firth, J. V., Johnson, G. L., & Ruddiman, W. F. (Eds.) (1996). A synthesis. Research synthesis. Proc. ODP, Sci. Results, (Vol. 151). Texas (Ocean Drilling Program): College Station.

Thiede, J., Myhre, A. M., Firth, J. V., Johnson, G. L., & Ruddiman, W. F. (Eds.) (1996). Climate of the Past, (Vol. 1), 103180. https://doi.org/10.1016/j.epsl.2009.12.049

Thiede, J., Clark, D. L., & Hermann, S. (1990). Late Mesozoic and Cenozoic paleoceanography of the northern polar oceans, in: The geology of North America, (pp. 427–458). The Arctic Ocean Region: Vol. L

Tregubov, P. V., Chugunov, V. N., & Gorbounov, V. A. (2013). Arctic permafrost: A challenge within the global permafrost region. Global Biogeochemical Cycles, 27, GB2032. https://doi.org/10.1002/gbc.2003327

Tremblay, L.-B., Smith, R. R. W., & Tung, C. (2017). The role of oceanic CO2 in the regulation of glacial-interglacial climate cycles. Frontiers in Marine Science, 4, 75. https://doi.org/10.3389/fmars.2017.00075

Tremblay, E. J., Albarède, F., & Zeebe, R. E. (2016). Carbonate precipitation as a biogeochemical sink during subaerial exposure. Nature, 539, 374–379. https://doi.org/10.1038/nature19730

Tremblay, E. J., Albarède, F., & Zeebe, R. E. (2016). Carbonate precipitation as a biogeochemical sink during subaerial exposure. Nature, 539, 374–379. https://doi.org/10.1038/nature19730
