

Sea ice variability in the Nordic Seas over Dansgaard–Oeschger climate cycles during the last glacial – A biomarker approach

Henrik Sadatzki

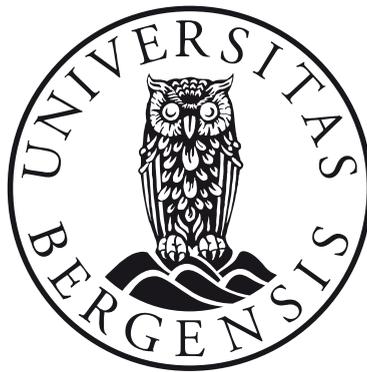
Thesis for the Degree of Philosophiae Doctor (PhD)
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Scientific environment

The research leading to this dissertation was carried out at the Department of Earth Science and the Bjerknes Centre for Climate Research, University of Bergen, Norway. The PhD project benefitted from close collaborations with the NORCE Norwegian Research Centre, Bergen, Norway, and with the Alfred Wegener Institute Helmholtz Centre for Polar and Marine Research, Bremerhaven, Germany, where the bulk of laboratory work was performed. This PhD study was part of the *Arctic Sea Ice and Greenland Ice Sheet Sensitivity* (Ice2Ice) project and the research leading to these results has received funding from the European Research Council under the European Union's Seventh Framework Programme (FP7/2007-2013) / ERC grant agreement n° 610055. Main supervisor of this PhD thesis was Eystein Jansen (University of Bergen) and co-supervisors were Trond M. Dokken (NORCE Norwegian Research Centre) and Rüdiger Stein (Alfred Wegener Institute).



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“The top of the world is turning from white to blue in summer as the ice that has long covered the north polar seas melts away. This monumental change is triggering a cascade of effects that will amplify global warming and could destabilize the global climate system.”

Peter Wadhams, September 26, 2016

Abstract

The Arctic sea ice cover is in fast transition. Resolving past sea ice fluctuations and its link with abrupt climate change might be key for a better understanding of yet unknown climatic consequences of future Arctic sea ice loss. The last glacial period was marked by recurring abrupt climate changes, referred to as Dansgaard–Oeschger (D–O) climate cycles. These D–O climate cycles and in particular the associated abrupt warming transitions by up to 15°C over Greenland happening within years or decades might have been linked to shifts in sea ice cover in the Nordic Seas.

This PhD thesis aims at resolving and constraining the largely unknown millennial-scale sea ice variability in the Nordic Seas and its pivotal role for abrupt climate changes during the D–O cycles based on empirical proxy data evidence. Novel sea ice reconstructions are mainly based on the sedimentary abundances of the sea ice algae biomarker IP₂₅ and open-water phytoplankton biomarkers.

This thesis includes two multi-decadal to centennial-scale biomarker sea ice records from the southern and central Norwegian Sea covering the time period ~30–40 thousand years ago, which reveal unprecedented insights into the nature of glacial sea ice fluctuations during D–O cycles (Papers 1 and 2). A comparison of these biomarker sea ice records with LOVECLIM model output data of sea ice cover (Paper 1) and a new bromine-enrichment sea ice record from the RECAP ice core (East Greenland) (Paper 2), sheds light on the mechanisms and timing of rapid sea ice shifts with respect to abrupt Greenland climate changes. A third biomarker sea ice record from the Eirik Drift south of Greenland elucidates the sea ice cover and export in the northwestern North Atlantic ~30–40 thousand years ago (Paper 3). This thesis also comprises a calibration based on a robust linear correlation between the sea ice index PIP₂₅ in (sub-)Arctic surface sediments and modern spring sea ice concentration, which allows a quantification of past sea ice changes (Paper 2).

The results presented in this thesis provide hitherto unknown details of spatiotemporal changes in glacial sea ice cover and tephra-assisted links to climate recorded in Greenland ice cores. Substantial rapid sea ice reductions and ocean overturning in the Norwegian Sea shaped the abrupt cold-to-warm D–O climate transitions, following a more gradual initial sea ice retreat. This reveals insights into sea ice-related feedbacks for abrupt D–O climate shifts and advances our understanding of abrupt transitions in the coupled ocean-sea ice-climate system during the last glacial.

List of publications

Paper 1

Sadatzki, H., Dokken, T. M., Berben, S. M. P., Muschitiello, F., Stein, R., Fahl, K., Menviel, L., Timmermann, A., Jansen, E. Sea ice variability in the southern Norwegian Sea during glacial Dansgaard–Oeschger climate cycles. *Manuscript in review for Science Advances*

Paper 2

Sadatzki, H., Maffezzoli, N., Dokken, T. M., Simon, M. H., Berben, S. M. P., Fahl, K., Kjær, H. A., Spolaor, A., Stein, R., Vallelonga, P., Vinther, B. M., Jansen, E. Rapid sea ice reduction in the Nordic Seas and abrupt warming over Greenland during the last glacial. *Manuscript in preparation*

Paper 3

Sadatzki, H., Griem, L., Dokken, T. M., Ninnemann, U., Stein, R., Fahl, K., Jansen, E. Evidence of deep *Zoophycos* burrowing and an enhanced glacial sea ice cover from the Eirik Drift south of Greenland. *Manuscript in preparation*

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1 Introduction

1.1 Opening remarks

The current Arctic sea ice loss, as pictured by Peter Wadhams' quote in the beginning of this thesis, has led the scientific community to pay an increased attention to sea ice observation and to strive for a better understanding of the role of sea ice in the Earth's climate system. The Intergovernmental Panel on Climate Change (IPCC) has highlighted in the latest, fifth Assessment Report that the annual average sea ice extent in the Arctic has decreased over the period 1979–2012 at a rate of 3.5–4.1 % per decade, which corresponds to a loss of 0.45–0.51 million square kilometers per decade (IPCC, 2013, Chapter 4 Observations: Cryosphere). It has been proposed that the diminishing sea ice has played a key role in the amplified warming in the Arctic over the last decades, compared to the global average temperature (Screen and Simmonds, 2010). It is believed that positive feedback mechanisms associated with sea ice decline, through reducing the albedo and enhancing ocean–atmosphere heat exchange, will lead to further rapid warming and sea ice loss in the future, potentially invoking severe consequences for polar ecosystems, Greenland ice sheet stability, and the global climate system (Screen and Simmonds, 2010). Although climate model simulations are improving, the majority of model simulations underestimate the Arctic sea ice decline that is observed since 1979 (Stroeve et al., 2007; IPCC, 2013, Chapter 4 Observations: Cryosphere). The reason for this may be an incomplete representation of feedback mechanisms in the coupled ocean-ice-climate system, an underestimation of sea ice thinning, and an underestimation of sea ice drift and export from the Arctic in the climate models (Boé et al., 2009; Rampal et al., 2011). This means that projections of future Arctic sea ice loss might be too conservative, that the Arctic Ocean may become ice-free in summer before the end of the twenty-first century, and that our knowledge of the implications of sea ice loss for future climate change remains incomplete (Boé et al., 2009; Rampal et al., 2011).

For a better and more comprehensive understanding of the role of sea ice in the climate system, the detection of sea ice fluctuations in the past and investigation on how these were related to large-amplitude and abrupt climate changes are crucial. Albeit “...it's really almost miraculous that we know anything about past climate” (Wallace Broecker, in Broecker, 2010, Preface, page ix), the Paleoceanography and Paleoclimatology communities are increasingly capable of resolving the details of

past changes in the coupled ocean-ice-climate system. Empirical evidence from proxy reconstructions and model simulations enable us to unravel abrupt ocean circulation, sea ice and climate changes in the past, which allows constraining the boundary conditions, rates and dimensions of natural climate variability, providing potential analogues for evaluating future climate projections. In the light of the current Arctic sea ice loss, the large ERC Synergy project Ice2Ice, a part of which is the present PhD thesis, puts forth a multi-disciplinary team effort to tackle the pressing question of the cause and future implications of past abrupt climate change recorded in Greenland ice cores and the wider Northern Hemisphere. Ice2Ice project members jointly work towards testing and evaluating the main hypothesis, being that Arctic and sub-Arctic sea ice cover exerts important controls on past and future Greenland temperature and ice sheet variation. A key target of this complex endeavor is a detailed investigation and understanding of hitherto unresolved ocean-sea ice feedbacks and mechanisms involved in the recurring abrupt Greenland warming events during the last glacial period ~10–100 thousand years ago, the so-called Dansgaard–Oeschger (D–O) events. This PhD thesis in particular aims at resolving the nature and timing of glacial changes in sub-Arctic sea ice cover and unraveling the role of sea ice in the abrupt climate transitions of the D–O events.

1.2 Arctic and sub-Arctic sea ice

Sea ice is an important component in the climate system as it affects Earth's radiation budget, ocean–atmosphere heat, moisture and gas exchange, ocean circulation, and marine biological productivity (Dieckmann and Hellmer, 2010). In 2018, the Arctic sea ice cover varied between 12.6 million square kilometers in March and 3.3 million square kilometers in September (Fig. 1) and about 15 % of the world's oceans are at least temporarily covered by sea ice (National Snow and Ice Data Center, <https://nsidc.org/>, accessed October 2018). While the sea ice cover has an enormous spatial extent, it only is a very thin layer of a few meters covering the ocean. This makes sea ice extremely vulnerable to small oceanic or atmospheric perturbations, meaning that a relatively small forcing can lead to dramatic sea ice shifts at a high pace, with substantial implications for the climate system (Dieckmann and Hellmer, 2010). With this and the current Arctic sea ice decline in mind, Arctic summer sea ice is classified as a policy-relevant tipping element in the climate system and considered

as a critical feedback element for future and past abrupt climate change (Lenton et al., 2008).

Sea ice reflects 50–70 % of the incoming radiation, while the ocean surface reflects only 6 % (National Snow and Ice Data Center, <https://nsidc.org/>, accessed October 2018). This ~8–12 times higher ability of sea ice to reflect solar radiation (albedo), compared to ocean water, means that the extensive Arctic sea ice cover controls the reflection and absorption of solar energy and thus has a substantial influence on Arctic temperatures. Moreover, sea ice acts as an efficient thermal insulator, separating the cold atmosphere at high latitudes from warmer ocean waters underneath, which prevents ocean–atmosphere heat and gas exchange, evaporation, and surface-ocean mixing due to reduced wind forcing (Dieckmann and Hellmer, 2010). Diminished surface-ocean mixing and associated lowered nutrient concentrations, together with a limited light availability, cause a substantially reduced phytoplankton production underneath the permanent, partially snow-covered, sea ice cover in the Arctic Ocean. Nevertheless, the sea ice itself forms a habitat for some specialized algae living within or at the bottom of the sea ice, such as sea ice diatoms

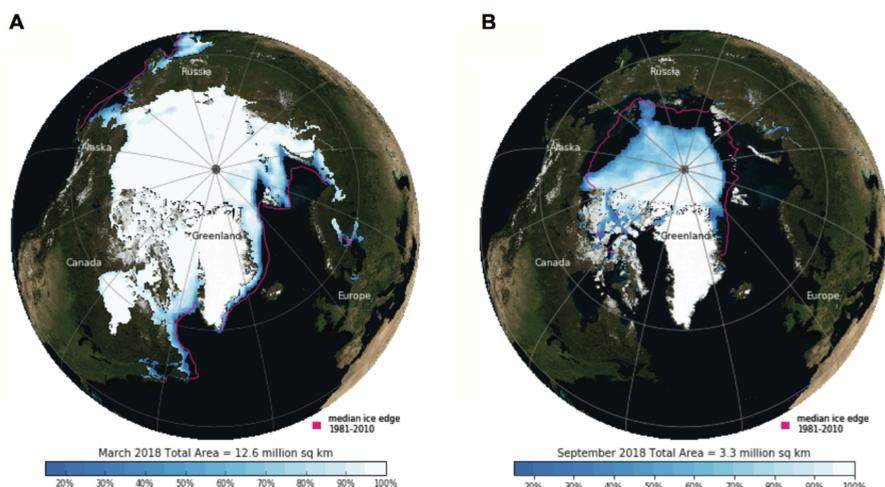


Figure 1 Modern Arctic sea ice concentrations for (A) March 2018 and (B) September 2018. Total sea ice areas are indicated at the bottom of each panel. Pink lines indicate the median ice edge for the period from 1981 to 2010, for March and September in (A) and (B), respectively. Maps are based on monthly average data of Sea Ice Index, Version 3 (Fetterer et al., 2017) from the National Snow and Ice Data Center (accessed October 2018).

belonging to the genera of *Haslea* and *Pleurosigma* (Arrigo, 2010; Brown et al., 2014). Ice algae production causes a yellow-brownish color of the bottom side of sea ice and forms a basic component for the marine ecology and carbon cycling in the Arctic (Horner and Alexander, 1972; Arrigo, 2010). Sea ice algae blooms occur in spring, when the light comes back, and are most pronounced in Arctic and sub-Arctic marginal seas with seasonal sea ice retreat (Arrigo, 2010; Xiao et al., 2015).

The seasonally ice-free, large shelf areas north of Canada, Alaska, Eurasia, and east of Greenland are considered as sea ice factories, thus as areas where sea ice is formed (Dieckmann and Hellmer, 2010). Sea ice forms when seawater is freezing in winter, producing ice crystals and small droplets of accumulated sea salts (so-called brine). As the sea ice ages, the brine drains out to the underlying ocean through channels within the sea ice, which increases the salinity and thus density of the seawater, causing it to sink (Dieckmann and Hellmer, 2010). Brine rejection associated with sea ice formation thus affects local, vertical water masses movement, and through this deep-water formation process also influences the larger-scale ocean circulation (Dieckmann and Hellmer, 2010). Aged sea ice forms a freshwater source that, when it melts, reduces the surface ocean salinity and density, contributing to enhanced surface stratification. The latter process includes seasonal sea ice melting, which fosters surface stratification and recurring winter sea ice formation on site, and melting of sea ice that drifted away from its formation area. With his *Fram* expedition (between 1893 and 1896) Fridtjof Nansen unambiguously showed that sea ice drifts from the shelf areas, through the Arctic Ocean, into the Nordic Seas, driven by winds and surface currents (Nansen, 1902). Intendedly frozen into sea ice near the New Siberian Islands, the *Fram* drifted with the sea ice and arrived northwest of Svalbard three years later. The Fram Strait (between Svalbard and Greenland) is the main gateway through which sea ice is exported from the Arctic Ocean to the Nordic Seas and downstream to the North Atlantic (Smedsrud et al., 2011). Hence, sea ice is transported far away from its formation area before it melts and thereby forms an important freshwater source that can affect surface stratification and ocean circulation on a larger scale.

Today, large-scale seasonal sea ice retreat occurs in the Barents Sea and Labrador Sea (Fig. 1), where ocean currents provide enhanced oceanic heat flux that can erode the

sea ice cover. These areas are also among the “hot spots” where the overall sea ice decline occurred over the last decades (Lind et al., 2018). The Arctic sea ice loss is clearly visible and most pronounced in September sea ice extent, as recorded by satellites between 1979 and 2018, with an average rate of about 12.8 % per decade (Fig. 2). However, sea ice reduction is also affecting the winter sea ice extent, sea ice thickness, and the area of perennial sea ice cover in the Arctic (IPCC, 2013, Chapter 4 Observations: Cryosphere). Based on a linear relationship observed between decreasing sea ice and increasing carbon dioxide concentration in the atmosphere, it has been proposed that radiative forcing associated with rising greenhouse gas concentrations and consequently rising temperatures may be a trigger of the sea ice decline (Notz and Stroeve, 2016). Nonetheless, the recent sea ice decline may also largely be driven by an enhanced oceanic northward heat advection observed in the Nordic Seas (Serreze et al., 2007; Spielhagen et al., 2011; Årthun et al., 2012; Zhang, 2015; Årthun et al., 2017). Sea ice loss leads to both increased heat absorption due to a weakened albedo effect and to increased exposure of warmer ocean waters to the high latitude atmosphere in the Arctic. Both processes result in an amplified and larger warming in the Arctic compared to the global average (Screen and Simmonds, 2010), a phenomenon that is referred to as Polar amplification and illustrates the **climatic relevance of sea ice as a positive feedback mechanism**.

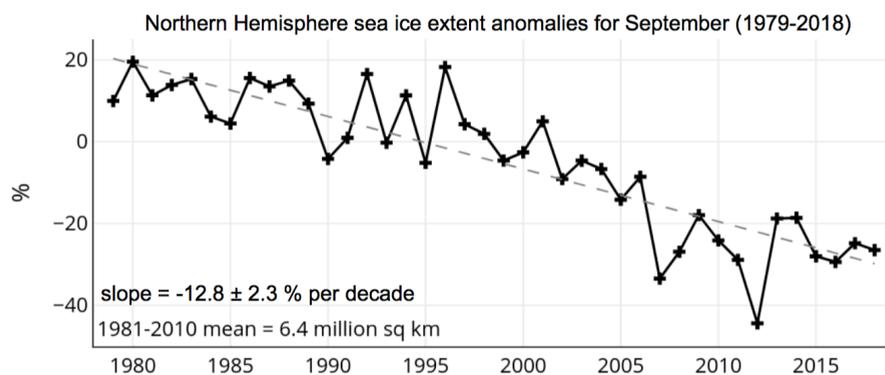


Figure 2 Northern Hemisphere sea ice extent anomalies during September for the period from 1979 to 2018. Monthly sea ice extent anomalies are plotted as percent difference between the extent for September of each year and the mean extent for September based on data from January 1981 to December 2010. The dashed grey line indicates the trend line based on a simple linear regression (or the slope). Figure is from the National Snow and Ice Data Center, https://nsidc.org/data/seaice_index/ (accessed October 2018).

1.3 Oceanography of the Nordic Seas

The Nordic Seas, including the Norwegian Sea, Greenland Sea and Iceland Sea, play an important role for the North Atlantic and global ocean circulation and climate. It is the region where warm surface waters originating from the tropical Gulf Stream deliver heat to northern high latitudes up to the Arctic sea ice edge, and is also a key causal element for the mild climate conditions in Northern Europe and Scandinavia (Hansen and Østerhus, 2000). The ocean overturning in the Nordic Seas contributes about 6 Sv ($10^6 \text{ m}^3/\text{s}$) of deep overflow waters to the North Atlantic Deep Water (Hansen and Østerhus, 2000). The Nordic Seas are thus a key region for oceanographic processes controlling the Atlantic Meridional Overturning Circulation (AMOC), or the global thermohaline circulation, or Wallace Broecker's Great Ocean Conveyor (Broecker, 1987; Broecker, 1991; Broecker, 2010).

Today, strong gradients in sea surface temperature and salinity characterize the Nordic Seas, separating the warmer and more saline Atlantic domain in the eastern Nordic Seas from the cooler and fresher Polar domain in the western Nordic Seas (Fig. 3A and B). The Polar Front and Arctic Front mark the transition between the perennially or seasonally sea ice-covered Polar domain and the annually ice-free Atlantic domain in the central and eastern Nordic Seas (Fig. 3) (Hopkins, 1991). The contrasting surface ocean conditions are largely controlled by the two major surface ocean currents in the region, one being the North Atlantic Current transporting warm and saline Atlantic waters into the eastern Nordic Seas, and the other one being the East Greenland Current transporting cold, fresher and ice-covered waters from the Arctic Ocean to the northwestern North Atlantic (Fig. 3C) (Hopkins, 1991; Hansen and Østerhus, 2000).

The North Atlantic Current can be considered as extension of the Gulf Stream and transports warm and saline waters from the subtropical North Atlantic to the Norwegian Sea, which in part continues to flow as Norwegian Atlantic Current to the Fram Strait and Barents Sea (Fig. 3) (Mauritzen, 1996). As it flows northward into the Norwegian Sea, these warm and saline surface waters gradually cool and thereby lose buoyancy. This process of surface water densification due to heat loss to the atmosphere causes part of the inflowing waters to sink and recirculate as intermediate or deep waters in the Nordic Seas and Arctic Ocean (Mauritzen, 1996). In addition,

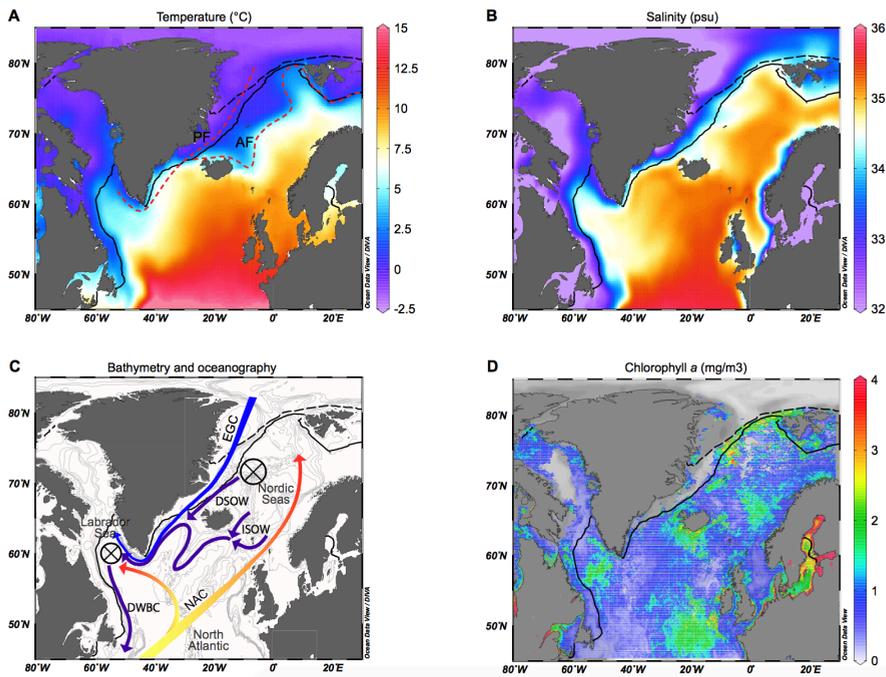


Figure 3 Modern oceanographic and environmental conditions in the Nordic Seas and northern North Atlantic. (A) Annual average sea surface temperature and (B) annual average sea surface salinity, both based on the World Ocean Atlas (WOA) 13 dataset and averaged between 1955 and 2012. (C) Bathymetry and oceanography. NAC = North Atlantic Current, EGC = East Greenland Current, ISOW = Iceland Scotland Overflow Water, DSOW = Denmark Strait Overflow Water, DWBC = Deep Western Boundary Current. Circled crosses indicate deep convection cells in the Nordic Seas and Labrador Sea. (D) Chlorophyll *a* concentration during June 2012 based on long-term multi-sensor time-series of satellite ocean-colour data using the version 3-1 dataset produced by the Ocean Colour project of the ESA Climate Change Initiative (CCI) (<http://catalogue.ceda.ac.uk/uuid/9c334fbe6d424a708cf3c4cf0c6a53f5>) (Sathyendranath et al., 2018). Red dashed lines in (A) indicate the Polar Front (PF) and Arctic Front (AF). Black lines in (A)–(D) mark the modern sea ice extent during September (dashed) and March (solid), averaged between 1981 and 2010 (<http://nsidc.org>; Fetterer et al., 2017). Maps were produced with the Ocean Data View software (Schlitzer, 2016).

open ocean deep convection occurs near the sea ice edge in the Greenland Sea (and Labrador Sea), where new deep water is formed (Fig. 3C and Fig. 4) (Hansen and Østerhus, 2000). The newly formed and re-circulated deep waters cross the Greenland-Scotland Ridge as Denmark Strait Overflow Water and Iceland-Scotland Overflow Water, and then flow along sea floor morphology into the North Atlantic (Fig. 3C). As Deep Western Boundary Current, the deep waters flow along the eastern

and southern Greenland margins into the Labrador Sea and, together with deep water formed in the Labrador Sea, form the North Atlantic Deep Water (Dickson and Brown, 1994). The northward flow at the surface and the reverse flow at depth in the northern North Atlantic essentially constitute the upper and lower limb of the AMOC, with deep-water formation in the Nordic Seas and the Labrador Sea being main drivers of the Great Ocean Conveyor (Broecker, 1987; Broecker, 1991).

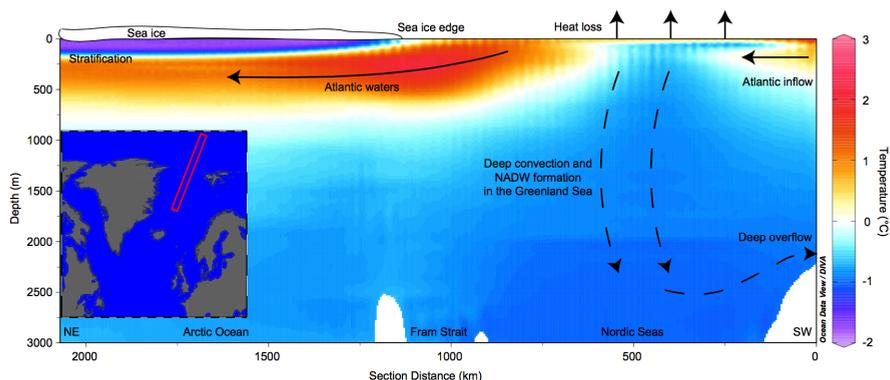


Figure 4 Cross-section of modern oceanographic conditions in the Arctic Ocean and Nordic Seas and across the sea ice edge near the Fram Strait. The cross-section shows annual average water temperature (color-coded) based on the World Ocean Atlas (WOA) 13 dataset and averaged between 1955 and 2012. Major oceanographic features discussed in the text are indicated. Figure was produced with the Ocean Data View software (Schlitzer, 2016).

The modern sea ice edge in the western and northern Nordic Seas essentially marks the boundary between contrasting environmental and oceanographic conditions of the Atlantic and Polar domains. While the phytoplankton production is generally reduced under the Arctic sea ice cover, it is stimulated in open waters of the Norwegian Sea but also at the sea ice edge where seasonal sea ice retreat occurs during summer, as reflected by the chlorophyll *a* concentration in surface waters (Fig. 3D). Furthermore, north of the sea ice edge in the Fram Strait the water column structure is substantially different from that in the Nordic Seas. The warm Atlantic inflow into the Nordic Seas continues into the Arctic Ocean at water depths of ~200–800 m beneath the sea ice cover (Fig. 4) (Mauritzen, 1996). This warm subsurface layer of Atlantic waters in the Arctic Ocean is sandwiched between cooler waters filling the deep basin and very cold and fresh waters at the surface, reflecting the strong surface stratification. A strong salinity gradient (halocline) causing this surface stratification compensates for

the vertical temperature inversion that would act against stratification, and prevents melting of sea ice from the oceanic heat reservoir below (Aagaard et al., 1981). This clearly illustrates the role of sea ice and its underlying surface stratification in effectively insulating the cold high-latitude atmosphere from relatively warmer ocean waters in the Arctic today, compared to oceanic heat being released to the atmosphere in the Nordic Seas (Fig. 4).

The relevance of (sub-)Arctic sea ice for climate and the importance of deep-water formation in the Nordic Seas for global ocean circulation suggest that **the Achilles heel of past abrupt climate change may be located in the Nordic Seas**, following a phrasing by Stephan Rahmstorf (Rahmstorf, 2004). Only minor (freshwater) perturbations of the system can cause non-linear, abrupt changes between different ocean circulation and climate modes in the North Atlantic, as proposed for the glacial D–O cycles (Ganopolski and Rahmstorf, 2001).

1.4 Dansgaard–Oeschger climate cycles

The abrupt climate fluctuations of the D–O cycles during the last glacial comprise some of the most pronounced and puzzling examples of climate change. The millennial-scale D–O climate cycles were first documented in ice core records from Greenland and are named after the ice core pioneers Willi Dansgaard and Hans Oeschger. Thereafter, they were found in numerous ocean and climate records from around the world (e.g., Voelker, 2002). Their importance for a better understanding of the climate system has been widely recognized, for example by the IPCC (IPCC, 2013, Chapter 5 Information from Paleoclimate Archives). Abrupt climate change occurs when the climate system is forced to cross a threshold, at which the system transitions into a new state, be it forced naturally or anthropogenically (Alley et al., 2003). The abrupt climate change of the D–O cycles in the past and the one possibly expected for the future (IPCC, 2013, Chapter 12 Long-term Climate Change: Projections, Commitments and Irreversibility) might share similarities in terms of abruptness as well as the underlying dynamics in the coupled ocean-sea ice-climate system leading to the occurrence of abrupt climate change. Notably, despite wide-ranging and long-term efforts in studying the D–O cycles, the mechanisms and

feedbacks involved in the abrupt climate transitions of the D–O cycles remain a matter of debate.

The D–O cycles are reflected by numerous high-frequency and large-amplitude, millennial-scale fluctuations in the $\delta^{18}\text{O}$ measured in ice cores from Greenland, which punctuated the last glacial period ~10–110 thousand years ago (Fig. 5A) (e.g., Dansgaard et al., 1993; NGRIP members, 2004). With the discovery of the D–O cycles in the mid 1980s, paleoceanographers and paleoclimatologists directed their focus towards millennial-scale ocean and climate variability during the last glacial period (Broecker, 2010). Before, the main target was to identify and understand the climatic imprint of cyclic changes in the Earth’s orbital configuration with respect to the sun, which have been described by and named after Milutin Milankovitch (Milankovitch, 1930). Northern Hemisphere summer insolation varies at periodicities of ~100,000 years (related to eccentricity), ~41,000 years (related to obliquity) and ~20,000 years (related to precession), orbital insolation cycles that are referred to as Milankovitch cycles. The combined effect of these insolation changes on the climate system is reflected, for example, in the global benthic foraminiferal $\delta^{18}\text{O}$ curve indicating the waxing and waning of Northern Hemisphere continental ice sheets during the last 150 thousand years (Figs. 5B and 5C) (Lisiecki and Raymo, 2005). Benthic $\delta^{18}\text{O}$ records serve as stratigraphic reference records of orbital-scale climate variations to define the Marine Isotope Stages (MIS), where cold glacials (large ice volume – increased benthic $\delta^{18}\text{O}$) have an even number and warm interglacials (smaller ice volume – decreased benthic $\delta^{18}\text{O}$) have an odd number (Fig. 5B) (e.g., Prell et al., 1986). The abrupt D–O climate changes were particularly frequent in times of an intermediate insolation and an intermediate continental ice volume such as during MIS 3 (Fig. 5) (despite its odd number, MIS 3 is considered as part of the last glacial, but it is distinct from the full glacial MIS 2 and MIS 4 in terms of continental ice volume/sea level). The intermediate continental ice volume of MIS 3 corresponds to an average sea level that was about 70–90 m below the modern level, while sea level during the Last Glacial Maximum (~20 thousand years ago) was about ~120 m below the modern level (Waelbroeck et al., 2002; Siddall et al., 2003; Siddall et al., 2008).

The D–O cycles comprise recurring climate variations between cold Greenland

stadials (GS), intervals with decreased ice core $\delta^{18}\text{O}$, and warmer Greenland interstadials (GI), intervals with increased ice core $\delta^{18}\text{O}$ (Fig. 5A). The abrupt jumps in the ice core $\delta^{18}\text{O}$ marking the GS–GI transitions occur within decades and largely reflect atmospheric warming over the Greenland ice sheet (e.g., Dansgaard et al., 1993; NGRIP members, 2004), which are referred to as D–O events. Independent air temperature estimates based on nitrogen isotope ratios of the air in gas bubbles included in the ice core, suggest that the abrupt D–O warming amounts to 5–16.5°C (Kindler et al., 2014). After the abrupt warming, Greenland temperatures gradually decrease throughout GI, and drop abruptly back to the stadial level at GI–GS transitions, leading to a characteristic saw-tooth shape of the glacial ice core $\delta^{18}\text{O}$ record (Fig. 6A). Other parameters measured in the ice cores also reveal the millennial-scale D–O variability, providing insights into other climate factors apart from air temperature (e.g., Seierstad et al., 2014). For example, deuterium excess and

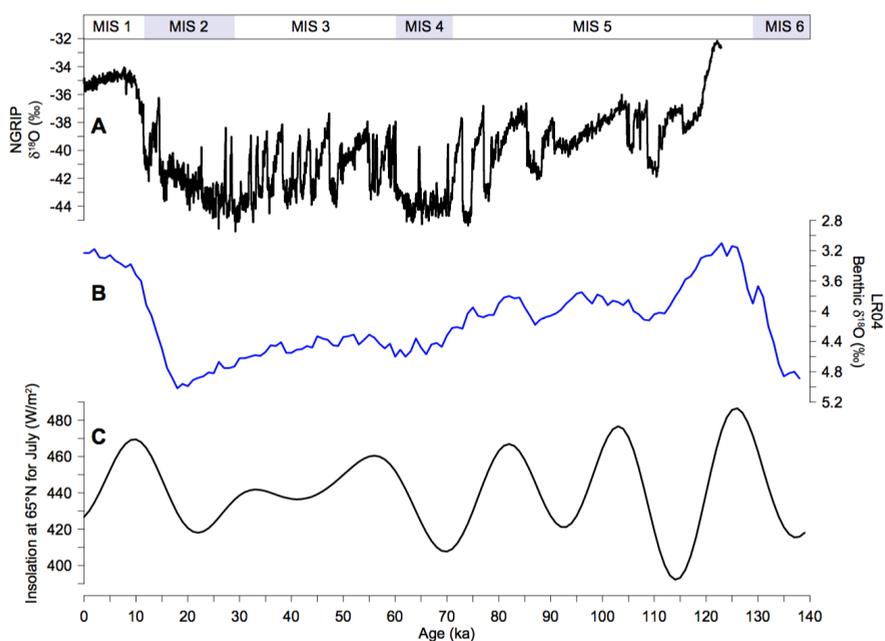


Figure 5 Late Pleistocene records of (A) $\delta^{18}\text{O}$ of the NGRIP ice core from Greenland (NGRIP members, 2004), (B) global stack of benthic foraminiferal $\delta^{18}\text{O}$ (Lisiecki and Raymo, 2005), and insolation at 65°N for July (Berger and Loutre, 1991) for the last 140 thousand years. Marine Isotope Stages (MIS) are numbered at the top and peak glacial stages are shaded. All records are plotted on their own age scale from the original publications.

dust concentration in Greenland ice cores suggest that the moisture source area for precipitation in Greenland and the large-scale atmospheric circulation changed dramatically over the course of the D–O cycles (Masson-Delmotte et al., 2005; Ruth et al., 2007). Based on sub-annually resolved records and precise temporal constraints using annual layer counting in the ice core records, it has been proposed that a rapid northward shift of the moisture source and an increase in snow accumulation rate happened within decades or only a few years at the onset of a D–O event (Steffensen et al., 2008; Thomas et al., 2009).

High-resolution paleoceanographic records from North Atlantic sediment cores also document the millennial-scale variability during the last glacial, akin to Greenland's D–O cycles (Voelker, 2002). Based on evidence of planktic foraminifer assemblages and planktic foraminiferal $\delta^{18}\text{O}$, it has been suggested that surface ocean conditions in the North Atlantic were cooler and fresher during GS, and warmer and more saline during GI (e.g., Bond et al. 1993; McManus et al., 1999; van Kreveld et al., 2000; Sarnthein et al., 2001; Elliot et al., 2002; Jensen et al., 2018a). This is, for example, reflected by the relative abundance of the polar planktic foraminifer *Neogloboquadrina pachyderma* (sinistral) (Fig. 6B) (van Kreveld et al., 2000). Records of ice rafted debris (IRD – large clastic grains that are too large to have been transported by winds or currents) have revealed that the North Atlantic was characterized by an enhanced presence of icebergs during GS compared to GI, indicating a periodic iceberg (and thus freshwater) discharge from surrounding glacial ice sheets (e.g., Bond et al. 1993; van Kreveld et al., 2000; Sarnthein et al., 2001). Furthermore, glacial North Atlantic sediments revealed very pronounced and distinct IRD layers, characterized by high detrital carbonate content and extremely low or absent foraminifer abundance, first identified by Hartmut Heinrich (Heinrich, 1988). These layers are interpreted as reflecting massive discharges of iceberg armadas (and freshwater) from the Laurantide Ice Sheet in North America to the North Atlantic, associated with particularly cold sea surface temperatures (Heinrich, 1988; Andrews and Tedesco, 1992; Bond and Lotti, 1995; Hemming, 2004). These events are named Heinrich events and have been found in some but not all of the GS, in those GS which have the longest durations and precede the most pronounced GI; notably, the ice core $\delta^{18}\text{O}$ record does not reflect a deterioration of 'normal' stadial Greenland temperature during Heinrich events (Fig. 6A).

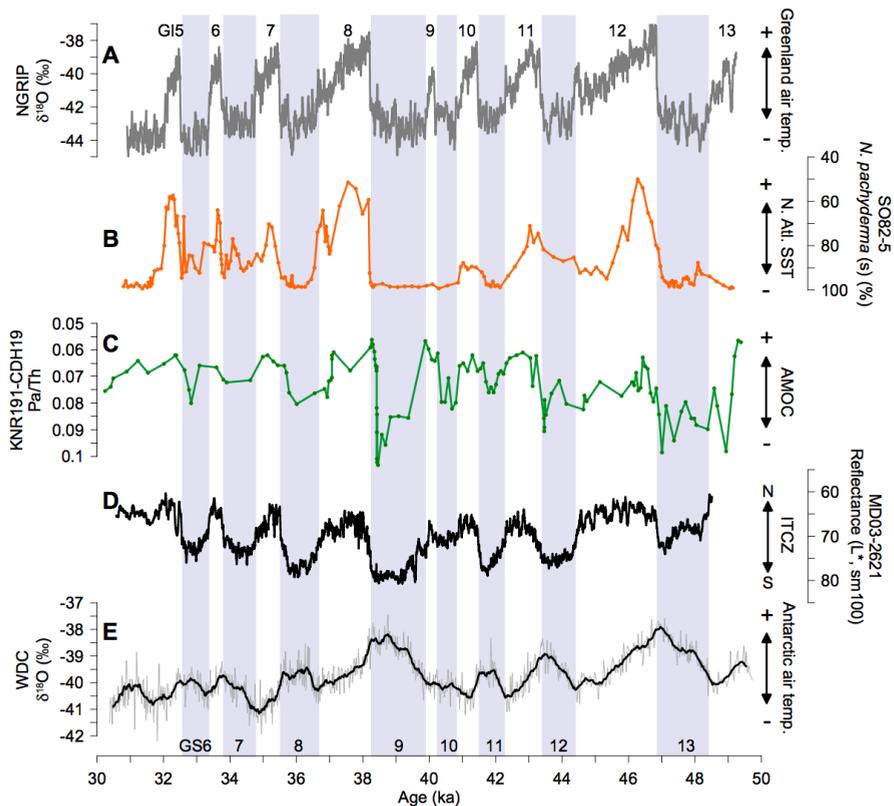


Figure 6 Proxy records of glacial millennial-scale variability related to Dansgaard–Oeschger climate cycles between 30 and 50 thousand years ago. (A) $\delta^{18}\text{O}$ of the NGRIP ice core from Greenland (NGRIP members, 2004). (B) Relative abundance of the polar planktic foraminifera *Neogloboquadrina pachyderma* (sinistral) in core SO82-5 from the northern North Atlantic (van Kreveld et al., 2000). (C) Pa/Th of bulk sediment from core KNR191-CDH19 from the Bermuda Rise (Henry et al., 2016). (D) Sediment total reflectance (L^*) from core MD03-2621 from the Cariaco Basin (Deplazes et al., 2013). (E) $\delta^{18}\text{O}$ of the WAIS Divide ice core (WDC) from the West Antarctic Ice Sheet (WAIS) (WAIS Divide Project Members, 2015). All records are plotted on their own age scale from the original publications. Greenland interstadials (GI) are numbered at top, Greenland stadials (GS; shaded bars) at bottom.

In addition to the climate proxy records from the northern North Atlantic, a record of alkenone biomarkers showed that sea surface temperatures in the subtropical North Atlantic also varied in unison with Greenland’s climate variability during the last glacial (Sachs and Lehman, 1999). This illustrates a coupling between temperatures over Greenland with surface ocean temperatures in the entire North Atlantic as an element of the D–O climate variability. Proxy records such as benthic $\delta^{13}\text{C}$ and

sedimentary Pa/Th from the deep subtropical Atlantic revealed that the vertical water mass structure and the lateral water mass transport varied on millennial timescales during the last glacial and deglacial (Fig. 6C) (McManus et al., 2004; Henry et al., 2016; Waelbroeck et al., 2018). These proxy records suggest that the AMOC was enhanced during GI, reduced during all GS, and weakest during the Heinrich events (Henry et al., 2016). Changes in northward transport of warm and saline surface waters to the northern North Atlantic, associated with AMOC variations, is consistent with the proxy records of glacial surface ocean conditions in the North Atlantic (van Kreveld et al., 2000; Sarnthein et al., 2001). Hence, surface and deep ocean proxy records support the hypothesis that variations in the Great Ocean Conveyor, or AMOC, were the pacemaker of the glacial millennial-scale climate variability recorded in Greenland ice cores (Fig. 7) (Broecker et al., 1985; Broecker, 1991).

Numerous marine and terrestrial climate records from around the globe have also revealed the glacial millennial-scale variability (Voelker, 2002). For example, the light reflectance record of a sediment cores from the Cariaco Basin north of Venezuela indicates changes in the relative contributions of terrigenous and biogenic components in the sediment (Fig. 6D) (Peterson et al., 2000; Deplazes et al., 2013). These reflectance changes might thus indicate lateral movements of the tropical rainfall belt and thus changes in the position of the Intertropical Convergence Zone (ITCZ) (Deplazes et al., 2013). The ITCZ is considered Earth's thermal equator, where northeasterly and southeasterly trade winds converge and intense insolation causes moist air to rise, resulting in heavy rainfall in the tropics. Accordingly, as a response to the differences in heat distribution across the hemisphere during D–O cycles the average position of the northern extent of the ITCZ and associated rainfall (including seasonal shifts) were centered over the Cariaco Basin and its catchment area during GI and shifted southward during GS (Peterson et al., 2000; Deplazes et al., 2013). Other important climate archives documenting the D–O variability in phase with Greenland climate changes are speleothems (carbonate cave deposits). The speleothem $\delta^{18}\text{O}$ records from China and South America suggest millennial-scale changes in strength and sources of monsoon rainfall, supporting that the ITCZ shifted between a more northerly position during warm GI and a more southerly position during cold GS on a global scale (Wang et al., 2001; Kanner et al., 2012).

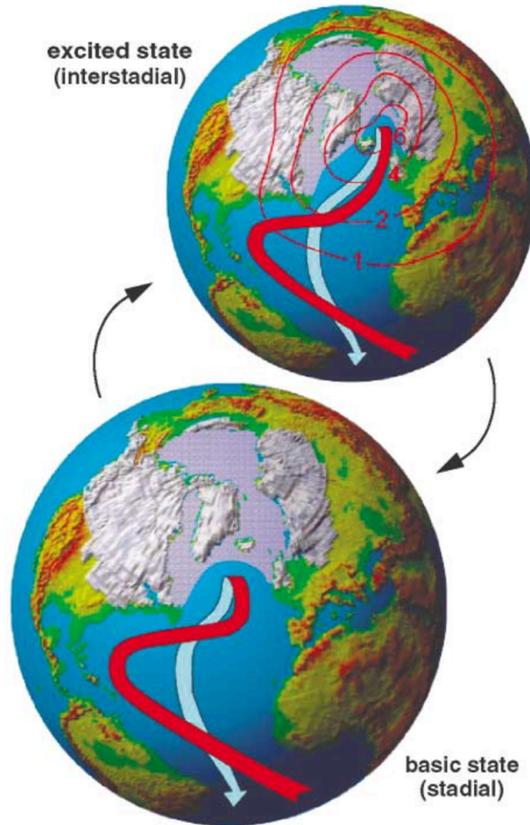


Figure 7 Schematic of the two glacial circulation modes and climate states after Ganopolski and Rahmstorf (2001) (from Ganopolski and Rahmstorf, 2002). Bottom: the stable ‘cold’ or ‘stadial’ mode. Top: the unstable ‘warm’ or ‘interstadial’ mode. The contours illustrate surface air temperature anomaly compared to the ‘stadial’ state. Surface current flowing north (red) and deep current flowing south (light blue) indicates the AMOC or Atlantic part of the Great Ocean Conveyor. Continental ice sheets are based on the reconstruction of Peltier (1994).

Furthermore, $\delta^{18}\text{O}$ and other records from Antarctic ice cores also reveal the glacial millennial-scale climate variability (WAIS Divide Project Members, 2015; Markle et al., 2017), illustrating its global significance. The high-resolution $\delta^{18}\text{O}$ record of the West Antarctic Ice Sheet (WAIS) Divide ice core (WDC), however, illustrates that the temperature variability in Antarctica was asynchronous to that in Greenland (Fig. 6E) (WAIS Divide Project Members, 2015). It has been estimated that on average abrupt Greenland warming led the onset of Antarctic cooling by 218 ± 92 years, while

Greenland cooling led the onset of Antarctic warming by 208 ± 96 years (WAIS Divide Project Members, 2015). This offset temperature evolution between the Northern and Southern Hemispheres has led to the notion of the thermal bipolar seesaw (Stocker and Johnsen, 2003; Stenni et al., 2011). There is a consensus that both latitudinal shifts of the ITCZ as well as the offset temperature variations in Greenland and Antarctica, are linked to millennial-scale changes in AMOC and associated meridional heat transport (Wang et al., 2001; Barker et al., 2009; Deplazes et al., 2013; WAIS Divide Project Members, 2015; Waelbroeck et al., 2018).

It has been found that the recurring D–O signals in some intervals, but far from all, can be aligned by a pacing period of $\sim 1,470$ years in $\delta^{18}\text{O}$ records from Greenland ice cores (Schulz, 2002; Rahmstorf, 2003) and in marine proxy records from the North Atlantic and Nordic Seas (Bond and Lotti, 1995; Dokken and Jansen, 1999; van Kreveld et al., 2000). It has been argued that this $\sim 1,470$ year periodicity in the climate cycles might be related to the sun's activity (Braun et al., 2005; Braun and Kurths, 2010). Braun et al. (2005) demonstrated that a superposition of two cycles of freshwater input the North Atlantic with periods near 87 and 210 years can result in $\sim 1,470$ year climate cycles. The periods near 87 and 210 years resemble those of the Gleissberg and DeVries solar cycles, respectively, mimicked to force the freshwater flux in the simulation (Braun et al., 2005). This frequency conversion between forcing and response might be linked to the nonlinear behavior of the thermohaline ocean circulation system, illustrating that the $\sim 1,470$ year climate cycles might have originated from solar variability, even though solar variability itself does not reveal a $\sim 1,470$ year cycle (Braun et al., 2005). The fact that the $\sim 1,470$ year cycle is absent in many time periods casts doubt on this narrative. A global climate model simulation with a low-resolution, simplified model by Ganopolski and Rahmstorf (2001) illustrated that a low-amplitude cyclic freshwater forcing could cause a rapid shift from a 'cold' circulation mode with deep-water formation south of Iceland to a 'warmer' circulation mode with deep-water formation in the Nordic Seas, where the simulated Greenland temperature rise resembled that observed in ice core records (Fig. 7). The authors argued that the glacial ocean circulation was an excitable system and abrupt changes resulted from weak periodic forcing and stochastic fluctuations of the freshwater flux to the North Atlantic (Ganopolski and Rahmstorf, 2001; Ganopolski and Rahmstorf, 2002). The rapid shift between the two different

circulation modes in response to a more gradual freshwater forcing, but also the abrupt Greenland D–O climate transition itself, illustrate a strongly non-linear behavior of the glacial ocean–climate system and calls for a threshold at D–O climate transitions (Broecker, 2000; Ganopolski and Rahmstorf, 2001). The existence of such a “flipping” non-linear behavior is not robustly found in more complex models, except under large freshwater forcing (Stouffer et al., 2006).

It has been argued that the weaker stadial circulation mode and the stronger interstadial circulation mode coexisted under intermediate-sized Northern Hemisphere ice sheets as during MIS 3, and that shifts between them could not only be triggered by variable freshwater flux to the North Atlantic, but also by small changes in ice sheet height and/or atmospheric carbon dioxide (CO₂) (Zhang et al., 2014; Zhang et al., 2017). The transition from a weak to strong AMOC mode could have resulted from a slight change of the intermediate ice sheets (equivalent to a sea level drop by less than 2 m), which would have shifted the northern westerly winds northward; this would have affected the subpolar and subtropical gyre circulation such that less sea ice was exported to the northeastern North Atlantic, permitting strong convective deep-water formation and a strong AMOC (Zhang et al., 2014). The required millennial-scale changes in ice sheet height and thus sea level appear realistic for MIS 3, but are poorly constrained by empirical evidence (Siddall et al., 2003; Siddall et al., 2008). Atmospheric CO₂, on the other hand, is well known from bubbles of ancient air included in Antarctic ice cores and varied between about 195 and 225 parts per million (ppm) on millennial timescales during MIS 3 (Ahn and Brook, 2008; Bauska et al., 2018). It has been suggested that an atmospheric CO₂ rise, as observed during Heinrich events, might have affected the atmospheric moisture transport across Central America, modulating the freshwater budget of the North Atlantic and deep-water formation (Zhang et al., 2017). Hence, this mechanism provides an alternative explanation of transitions between the weaker stadial and stronger interstadial AMOC mode, at least when the CO₂ rise was large enough (Zhang et al., 2017).

With the aim to elucidate the underlying dynamics of the globally distributed D–O climate variability, Menviel et al. (2014) presented a transient hindcast model simulation of multiple D–O cycles with MIS 3 boundary conditions. Forced by variable freshwater fluxes to the North Atlantic, this model simulation reproduced

numerous climate signals of both D–O cycles and Heinrich events, as seen in proxy records from around the globe (Menviel et al., 2014). They concluded that ice sheet calving and subsequent changes in AMOC were the main drivers of the millennial-scale glacial climate variability (Menviel et al., 2014). Hence, model simulations, together with proxy evidence of IRD in the North Atlantic during GS, suggest that ice sheet dynamics and related periodic freshwater perturbations in the North Atlantic might have played an important role in abrupt D–O climate transitions, even though icebergs were not the trigger of cold GS (Ganopolski and Rahmstorf, 2001; Menviel et al., 2014; Zhang et al., 2014; Bond et al., 1993; van Kreveld et al., 2000; Sarin et al., 2001; Barker et al., 2015). Notably, all model-based studies mentioned above also referred to shifts in the North Atlantic sea ice cover and ocean–sea ice–atmosphere feedbacks that amplify the abrupt climate transition in Greenland (see 1.5) (Ganopolski and Rahmstorf, 2001; Menviel et al., 2014; Zhang et al., 2014; Zhang et al., 2017).

Unforced model simulations also produce abrupt Greenland warming and cooling events which spontaneously arise in these runs and resemble the D–O cycles (Kleppin et al., 2015; Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018). Based on a preindustrial climate simulation, Kleppin et al. (2015) argued that stochastic atmospheric forcing, affecting the subpolar gyre circulation, sea ice cover, and meridional heat transport in the northwestern North Atlantic, triggers a climate warming transition. On the other hand, an unforced model simulation under Last Glacial Maximum boundary conditions has shown that abrupt Greenland warming can result from subsurface thermohaline instability underneath the sea ice cover in the northwestern North Atlantic (Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018). In this model simulation, the subsurface thermohaline instability causes a rapid opening of a super-polynya, reinvigoration of ocean convection and heat release to the atmosphere (see 1.5) (Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018). Besides these internal atmospheric or oceanic forcings, or coupled ice sheet–ocean interactions, volcanic eruptions have also been discussed as potential trigger mechanism of the D–O climate cycles (Baldini et al., 2015).

In summary, the D–O variability during the last glacial is well known from numerous climate and ocean records from around the world, and climate models can reproduce

the main features observed in the proxy records. However, the exact forcing mechanism of the D–O cycles and the origin of the ~1,470 year pacing of some of the D–O events are still not fully understood. Nevertheless, there appears to be a consensus that the **strongly non-linear, abrupt climate transitions of the D–O cycles involved ocean circulation changes and sea ice-related feedbacks** in the northern North Atlantic, which will be further outlined in the next chapter.

1.5 The importance of sub-Arctic sea ice for Dansgaard–Oeschger cycles

The importance of sea ice variability in past abrupt climate change has been acknowledged since the early days of the field of Paleoclimatology, for example, by Ruddiman and McIntyre (1981) and Broecker et al. (1985). Moreover, its role as feedback mechanism in the climate system has been intensively studied for decades (e.g., Manabe and Stouffer, 1988; Schiller et al., 1997). Several studies have proposed that changes in the sea ice cover in the northern North Atlantic and Nordic Seas may form a key element explaining the non-linear behavior, striking abruptness and large amplitude of the D–O climate transitions recorded in Greenland ice cores (see 1.4) (Dansgaard et al., 1993; Alley et al., 1993; Broecker, 2000; Timmermann et al., 2003; Gildor and Tziperman, 2003; Denton et al., 2005; Li et al., 2005; Li et al., 2010; Petersen et al., 2013; Dokken et al., 2013). Novel sea ice proxy methods have been proposed and developed during the last years, which allow for direct and semi-quantitative reconstructions of sea ice fluctuations in the past (Belt et al., 2007; Belt and Müller, 2013). Accordingly, investigation of linkages between ocean circulation, sea ice and abrupt climate change in the past – for example during the D–O cycles – has increasingly become a “hot topic” in Paleoclimatology Research (e.g., Hoff et al., 2016).

Multiple proxy records suggest that during cold GS the surface ocean in the Nordic Seas was highly stratified by a surface freshwater lid, similar to the conditions observed in the Arctic Ocean today (e.g., Rasmussen et al., 1996; Rasmussen and Thomsen, 2004; Dokken et al., 2013). Both proxy and model data suggest that subsurface waters at intermediate depths were warmer during GS compared to GI and today, insulated from the atmosphere by a sea ice cover that is sustained by the freshwater layer (Fig. 8) (Rasmussen and Thomsen, 2004; Friedrich et al., 2010;

Dokken et al., 2013; Ezat et al., 2014). Hence, the relatively fresh polar surface waters, the Polar and Arctic Fronts and probably also the sub-Arctic sea ice cover may have extended to the south of the Nordic Seas or even to the North Atlantic during GS (Fig. 8) (Rasmussen and Thomsen, 2004; Dokken et al., 2013; Hoff et al., 2016). This also suggests that there might not have been active open-ocean deep convection in the Nordic Seas during GS, but probably in the North Atlantic (Rasmussen et al., 1996; Dokken and Jansen, 1999; Rasmussen and Thomsen, 2004; Dokken et al., 2013; Ezat et al., 2014). On the other hand, proxy reconstructions indicate that there was active surface and deep-ocean mixing in the Nordic Seas during GI, implying a northward shifted position of the Polar and Arctic fronts and a reduced sea ice cover, roughly similar to today (Fig. 8) (Rasmussen and Thomsen, 2004; Dokken et al., 2013). Deep ocean temperatures at >1,200 m water depth in the Nordic Seas were about 2–4°C lower during GI than during GS, suggesting a removal of the stadial subsurface oceanic heat reservoir linked to deep ocean convection (Ezat et al., 2014). Hence, the two different scenarios of oceanographic and sea ice conditions in the Nordic Seas during GS and GI, as illustrated in Figure 8, are consistent with reconstructed variations in AMOC and associated northward heat transport to the northern North Atlantic, and with the idea of two glacial circulation modes of the Great Ocean Conveyor during D–O cycles (neglecting the third weakest mode proposed for Heinrich events) (Fig. 7) (Broecker et al., 1985; Broecker, 1991; van Kreveld et al., 2000; Ganopolski and Rahmstorf, 2001; Henry et al., 2016).

Indeed, climate model simulations have shown that a shift from an extensive sea ice cover to a reduced sea ice cover in the Nordic Seas leads to a ~10°C atmospheric winter warming over Greenland, consistent with results from ice core records (Li et al., 2005; Li et al., 2010). The model simulation also revealed a 50 % snow accumulation increase in Greenland associated with the sea ice reduction, which likewise agrees well with observations from ice core records (Li et al., 2005; Li et al., 2010). Furthermore, sea ice retreat in the Nordic Seas could explain a northward shift in moisture source area for Greenland precipitation, as recorded by the rapid transition in deuterium excess in Greenland ice cores (Masson-Delmotte et al., 2005). This suggests that rapid switches in sea ice cover likely triggered, or at least contributed to the abrupt climate transitions during the D–O cycles (Timmermann et al., 2003; Gildor and Tziperman, 2003). Sea ice reduction may have amplified the D–O

warming at GS–GI transitions by reducing the cooling albedo effect and exposing relatively warmer subsurface water in the Nordic Seas to the atmosphere (see 1.2) (Gildor and Tziperman, 2003). In turn, sea ice expansion may have amplified the rapid cooling recorded in Greenland ice cores at GI–GS transitions through the same feedback mechanisms as described above.

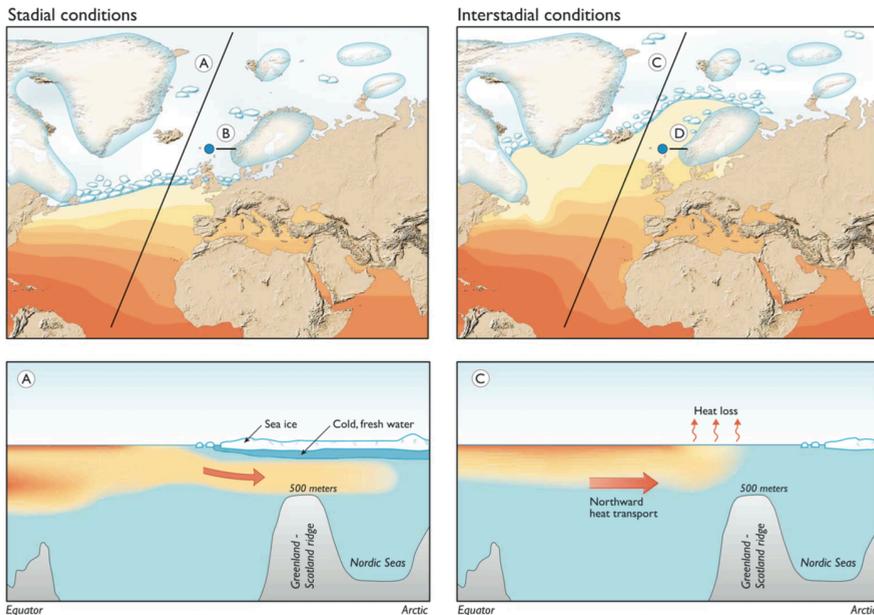


Figure 8 Scheme of hypothesized stadial and interstadial oceanographic and sea ice conditions in the Nordic Seas during glacial Dansgaard–Oeschger climate cycles (from Dokken et al., 2013).

It has been suggested that variations in thermohaline circulation in the North Atlantic, presumably in combination with stochastic freshwater forcing, may have triggered rapid switches in sea ice cover, which in turn amplified the climate signal (Timmermann et al., 2003; Gildor and Tziperman, 2003; Menviel et al., 2014). The meridional heat flux controlling the sea ice cover in the North Atlantic and Nordic Seas may have been controlled, for example, by freshwater flux from continental ice sheets to the North Atlantic (Menviel et al., 2014), a strong salinity difference between the more saline North Atlantic and the fresher Nordic Seas without external freshwater forcing (Peltier and Vettoretti, 2016), and/or the subpolar gyre circulation

in the northwestern North Atlantic and stochastic atmospheric wind forcing thereof (Kleppin et al., 2015). Furthermore, the subsurface oceanic heat reservoir beneath the sea ice, as observed during GS in the Nordic Seas, might have triggered sea ice retreat in the Nordic Seas (Timmermann et al., 2003; Dokken et al., 2013; Jensen et al., 2016; Jensen et al., 2018b). It has been shown that a relatively small increase in subsurface temperature of Atlantic waters flowing into the Nordic Seas can result in a rapid, non-linear destabilization and melting of sea ice, under a relatively small freshwater input (Jensen et al., 2016; Jensen et al., 2018b).

There are a few sea ice proxy reconstructions from sediment cores from the southern Norwegian Sea covering the glacial D–O cycles, based on molecular biomarkers or dinoflagellate cysts (Hoff et al., 2016; Wary et al., 2016). Based on a 80,000 year-long biomarker record with centennial to millennial-scale resolution, Hoff et al. (2016) argued that sea ice disappeared rapidly at the onset of the D–O events (GI), expanded later and peaked in near-perennial occurrence at the onset of the subsequent GS. This reconstruction thus seems to support the stadial and interstadial sea ice scenarios that have been proposed, for example, by Dokken et al. (2013) (Fig. 8). A sea ice record based on dinoflagellate cysts from the southern Norwegian Sea has revealed different trends in sea ice cover, compared to the biomarker record (Wary et al., 2016). The dinoflagellate cyst assemblages indicate that there was intensive winter sea ice formation during GI and reduced sea ice during cold GS (Wary et al., 2016). The available sea ice records are both restricted to the southern periphery of the Nordic Seas, have a limited (centennial-scale) resolution, and seem to reveal opposing trends (Hoff et al., 2016; Wary et al., 2016). Hence, it remains uncertain how the Nordic Seas ice cover actually varied during the glacial D–O cycles. This uncertainty is a motivating factor behind the work performed in this PhD thesis.

Some evidence of larger-scale sea ice fluctuations in the northern North Atlantic comes from halogen and sea salt records of Greenland ice cores (Maffezzoli et al., 2018, in rev.). In particular, a new bromine enrichment record from the RECAP ice core from coastal Eastern Greenland indicates orbital and millennial-scale changes in seasonal sea ice cover in the northern North Atlantic over the last 120 thousand years (Maffezzoli et al., in rev.). A sea ice record from the NEEM ice core in northwestern Central Greenland has similarly revealed millennial-scale sea ice changes, but with

significantly higher temporal resolution than the RECAP ice core record (Spolaor et al., 2016). The NEEM record has been interpreted as reflecting an enhanced perennial sea ice cover in the Canadian Arctic during cold GS and less perennial but more seasonal Arctic sea ice during GI (Spolaor et al., 2016). However, the interpretation of ice core-based sea ice records can be hampered by shifts in the marine source area for moisture and aerosols as well as by changes in aerosol transport distance (Spolaor et al., 2016). Moreover, the temporal resolution of $\sim 1,500$ years for the glacial period in the RECAP ice core, which reflects North Atlantic conditions, is insufficient to properly resolve the D–O cycles (Maffezzoli et al., in rev.). Similar to proxy evidence from sediment cores, ice core-based evidence of sea ice changes in the Nordic Seas and North Atlantic region remains limited and does not unambiguously support the hypothesized stadial and interstadial sea ice scenarios (Fig. 8).

Besides the hypothesized importance of sea ice shifts in the Nordic Seas for D–O climate cycles, model studies have suggested that changes in sea ice cover in the northwestern North Atlantic might have been crucial (Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018). In an unforced model simulation with spontaneously occurring D–O-like climate variations, abrupt Greenland warming resulted from rapid sea ice disappearance in the form of a sudden opening of a super-polynya in the ice-covered northwestern North Atlantic (Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018). It has been argued that the rapid opening of the super polynya results from a subsurface thermohaline convective instability beneath the extended sea ice cover, under a diminishing surface salinity stratification (Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018). The subsurface thermohaline convective instability would cause a mixing up of relatively warm subsurface waters, which would melt the sea ice lid, and subsequently release heat to the atmosphere (Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018), a mechanism similar to that proposed for the Nordic Seas (Dokken et al., 2013). Moreover, the authors suggested that sea ice export from the Arctic Ocean and Nordic Seas to the North Atlantic significantly contributed to gradual freshening in the North Atlantic sub-polar gyre region during GI, eventually leading to a reduction of North Atlantic Deep Water formation at GS–GI transitions (Vettoretti and Peltier, 2018). However, there is no robust sea ice proxy evidence yet that could support or reject the simulated sea ice changes in the northwestern North Atlantic.

Altogether, proxy reconstructions and model simulation seem to support that sea ice fluctuations in the Nordic Seas and North Atlantic and sea ice-related feedback mechanisms amplified some kind of initial oceanographic and/or atmospheric forcing during the glacial D–O cycles, causing – or at least contributing to – the abrupt climate transitions recorded in Greenland ice cores. A recent study also suggested that “early-warning signals” for the abrupt climate transition of the D–O events, identified in the $\delta^{18}\text{O}$ record of the NGRIP ice core, could be physically explained by sea ice fluctuations before the actual onset of a D–O event (Boers, 2018). However, **the nature, timing and exact role of sea ice fluctuations in past abrupt D–O climate changes and underlying mechanisms remain speculative and uncertain**, due to the lack of robust, high-resolution empirical proxy data evidence of the millennial-scale sea ice variability in the Nordic Seas and North Atlantic.

2 Objectives

The primary objective of this PhD project is to reconstruct the glacial millennial-scale sea ice variability in the Nordic Seas by generating biomarker sea ice proxy records that cover several abrupt D–O climate changes between ~30 and ~40 thousand years ago with unprecedented temporal resolution. The goal is to provide robust empirical proxy evidence of glacial sea ice variability in order to evaluate the relevance of past sea ice dynamics and ocean-sea ice-climate feedbacks for abrupt climate change over Greenland and the wider North Atlantic region. The overall **hypothesis** addressed is that sea ice retreat acted as a critical feedback mechanism that shaped the extremely abrupt and large-amplitude atmospheric warming of the D–O events.

To reach these goals and address the hypothesis, the specific objectives are:

- 1) To evaluate and advance the biomarker approach for qualitative and quantitative sea ice reconstructions. This will be based on reconciling new and previously published sea ice biomarker proxy data from modern surface sediments from Arctic and sub-Arctic core-top samples, in comparison with satellite-derived observations of modern sea ice concentration.
- 2) To develop robust chronologies for sediment cores from regions with high sediment accumulation, providing a stratigraphic framework for high-resolution biomarker sea ice records. This will be based on stratigraphic alignment of pertinent marine and ice core proxy records, and in part assisted by tephrochronological constraints and radiocarbon (^{14}C) dating of planktic foraminifera.
- 3) To reconstruct and characterize glacial changes in sea ice cover and phytoplankton productivity in the Nordic Seas using novel molecular biomarker proxy records with multi-decadal to centennial-scale resolution from two key sites in the Norwegian Sea.
- 4) To document the mechanisms and consequences of past sea ice retreat and expansion in the Nordic Seas. This will be achieved by comparing the new paleo-sea ice observations with 1) existing sediment proxy records of surface and deep-ocean temperature and convection in the Nordic Seas, 2) model simulated changes in Nordic Seas ice cover, and 3) ice core proxy records of sea ice in the Nordic Seas/North Atlantic and Greenland climate.

5) To reconstruct and provide insights into the nature and variability of glacial sea ice cover in the northwestern North Atlantic, using biomarker sea ice records from a high-sediment accumulation core site south of Greenland.

3 Approach, material and methods

The methodological approach of this PhD project mainly relies on the identification and quantification of molecular biomarkers in marine sediments, used for reconstructing sea ice conditions and phytoplankton productivity in the Nordic Seas and North Atlantic during the last glacial. The new organic geochemical proxy datasets are supplemented by other sedimentological, micropaleontological and inorganic geochemical data of surface and deep ocean conditions, which were either previously published or newly acquired in the framework of this project.

3.1 Sea ice biomarker approach

Since Simon Belt first proposed the biomarker IP₂₅ as an Ice Proxy (with 25 carbon atoms) (Fig. 9) (Belt et al., 2007), the development and utilization of biomarkers for advanced sea ice reconstruction in Arctic and sub-Arctic regions has increasingly been put forward. Previously, proxy methods used for sea ice reconstructions were based, for example, on the sedimentary abundance of IRD, assemblages of dinoflagellate cysts, diatoms and foraminifera, and the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ of these microfossil shells (Weinelt et al., 2001; Sarnthein et al., 2003; Armand and Leventer, 2010; Polyak et al., 2010; Stein et al., 2012). These proxies allow reconstruction of temperature and other water mass characteristics, but do not directly reflect sea ice conditions (Weinelt et al., 2001). The novel sea ice proxy IP₂₅ is a specific compound that is produced during spring by sea ice diatoms which live in brine channels within the sea ice or underneath but attached to sea ice (see 1.2) (Belt et al., 2007; Thomas and Dieckmann, 2008; Brown et al., 2011). IP₂₅ has been identified in marine sediments from Arctic and sub-Arctic regions covering the last glacial cycle (last 130 thousand years or so), the Pliocene (~4 million years), and the Miocene (>5 million years) and thus appears to be a stable molecule even in deep and old sediments (Stein et al., 2012; Belt and Müller, 2013; Knies et al. 2014; Stein et al., 2016; Hoff et al., 2016). Therefore, IP₂₅ forms a key proxy that can provide unprecedented direct and robust evidence of sea ice conditions in the Nordic Seas during D–O cycles.

IP₂₅ is a highly branched isoprenoid (HBI) monoene (Fig. 9) (Belt et al., 2007). Its production was found to be restricted to the minority of sea ice diatom taxa and specific to few but widespread taxa of the groups of *Haslea* and *Pleurosigma*, that is

Haslea crucigeroides (and/or *Haslea spicula*), *Haslea kjellmanii*, and *Pleurosigma stuxbergii* var. *rhomboides* (Brown et al., 2014). Several studies have mapped the distribution of IP₂₅ in core-top surface sediments in Arctic and sub-Arctic ocean regions, which clearly revealed that an increased sedimentary abundance of IP₂₅ is associated with overlying seasonal sea ice conditions (Müller et al., 2011; Navarro-Rodriguez et al., 2013; Stoyanova et al., 2013; Xiao et al., 2015). In turn, IP₂₅ is lowered and largely absent in surface sediments from regions that experience both open ocean conditions (such as the central Nordic Seas) and a perennial sea ice cover (such as the central Arctic Ocean) (Xiao et al., 2015). This illustrates the ambiguity of IP₂₅ absence in marine sediments, which can suggest both ice-free and perennial sea ice conditions.

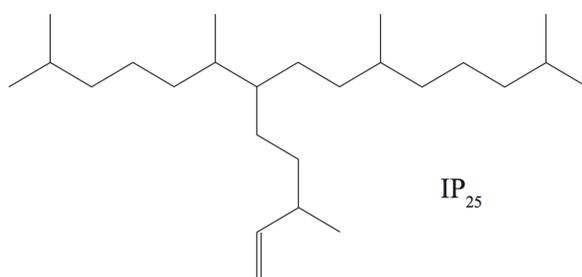


Figure 9 Chemical structure of IP₂₅ (after Belt et al., 2007). The double bond is located at C₂₃-C₂₄.

It has been suggested that combined analyses of IP₂₅ and open-water phytoplankton biomarkers allow for improved sea ice reconstructions (Müller et al., 2009; Müller et al., 2011). Brassicasterol and dinosterol are biomarkers reliably reflecting phytoplankton production and open-water conditions (Boon et al., 1979; Volkman, 1986; Volkman, 2003; Xiao et al., 2015). While dinosterol is produced by marine dinoflagellates (Volkman et al., 1993), brassicasterol is largely synthesized by marine diatoms, but also by marine haptophyte algae (coccolithophores), freshwater diatoms, and perhaps sea ice diatoms (Volkman, 1986; Fahl and Stein, 2012; Belt et al., 2013). Open-water haptophyte algae produce long-chain C₃₇ ketones (alkenones), which also serve as indicator of marine phytoplankton productivity (Volkman et al., 1995). Furthermore, terrigenous biomarkers such as campesterol and β -sitosterol produced

by higher plants on land (Volkman, 1986), can reflect input of terrigenous material to the ocean (similar to IRD), supplementing biomarker records of marine ice algae and phytoplankton production.

Moreover, there are several isomers of HBIs, which have different degrees of unsaturation (different numbers of double bonds) and are thus different from the HBI monene IP₂₅ (which has one double bond) (Belt et al., 2000). While IP₂₅ and the HBI dienes (HBI-II) appear to co-occur associated with seasonal sea ice in the Arctic, IP₂₅ has not (yet) been identified in the Southern Ocean but the HBI-II has been proposed as promising Ice Proxy for the Southern Ocean with 25 carbon atoms (IPSO₂₅) (Belt et al., 2007; Belt and Müller, 2013; Belt et al., 2016). It is believed that HBI trienes (HBI-III) are synthesized by a small number of marine diatom taxa (Belt et al., 2015). Investigation of surface sediment samples from the Barents Sea has revealed that HBI-III is present under open waters, extremely reduced or absent under seasonal and perennial sea ice, and occurs in maximum abundance underneath the marginal ice zone (Belt et al., 2015). Accordingly, HBI-III is hypothesized to represent a potential proxy for retreating sea ice or the marginal ice zone in both the Northern Hemisphere (Belt et al., 2015) and Southern Hemisphere (Collins et al., 2013).

Figure 10 illustrates how combined analyses of sea ice algae and phytoplankton biomarkers in Arctic surface sediments reveal insights into different sea ice conditions. An increased phytoplankton biomarker abundance and contemporaneously lowered/absent IP₂₅ reflect open-ocean conditions; lowered or absent phytoplankton biomarker and contemporaneously lowered or absent IP₂₅ indicate perennial sea ice; and an increased abundance in both types of biomarkers mark seasonal sea ice conditions (Fig. 10) (Müller et al., 2011; Xiao et al., 2015). This combined biomarker approach thus overcomes the ambiguity of the absence of IP₂₅ and enables distinction between open-water and perennial sea ice conditions. Following this, it has been suggested that combining IP₂₅ with an open-water phytoplankton biomarker in the so-called phytoplankton-IP₂₅ index (PIP₂₅), allows for improved, semi-quantitative sea ice reconstruction (Fig. 10) (Müller et al., 2011):

$$\text{PIP}_{25} = [\text{IP}_{25}] / ([\text{IP}_{25}] + [\text{phytoplankton marker}] \times c) \quad (1)$$

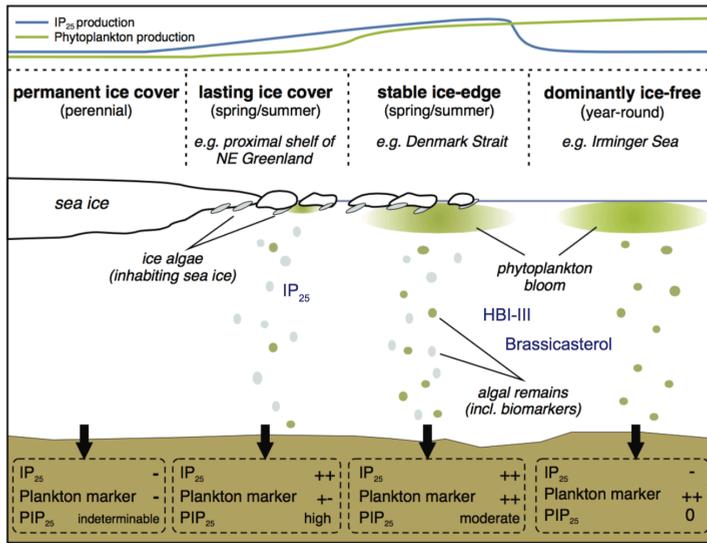


Figure 10 Scheme illustrating ice algae and phytoplankton production under different sea ice conditions during spring/summer (slightly modified, but originally from Müller et al., 2011). The sedimentary abundances of the specific biomarkers (IP₂₅ and phytoplankton marker) and PIP₂₅ values expected for the different sea ice conditions are also indicated.

The balance factor c corresponds to the ratio of mean IP₂₅ and mean phytoplankton biomarker concentration for a given set of samples from a certain area or sediment core. It accounts for often very different concentrations of IP₂₅ and phytoplankton biomarker (Müller et al., 2011). The PIP₂₅ can be estimated using brassicasterol, dinosterol or HBI-III as phytoplankton biomarker. PIP₂₅ values vary between 0 and 1, indicating open-ocean conditions and perennial sea ice cover, respectively (Fig. 10).

Calibration studies using Arctic and sub-Arctic core-top samples have shown a robust positive linear correlation between PIP₂₅ and modern spring sea ice concentration in various Northern Hemisphere ocean regions (Müller et al., 2011; Navarro-Rodriguez et al., 2013; Stoyanova et al., 2013; Xiao et al., 2015; Kolling, 2017), illustrating the great potential of the sea ice biomarker approach. Furthermore, based on results from Barents Sea core-top samples Smik et al. (2016) argued that the dependence of the PIP₂₅ on the balance factor c is smallest, and the positive linear correlation with spring sea ice concentration is strongest, for P_{III}IP₂₅ (using IP₂₅ and HBI-III), as compared to P_BIP₂₅ (using IP₂₅ and brassicasterol). Latest results from widespread Arctic and sub-Arctic core-top samples reveal a similar trend (Kolling, 2017).

3.2 Sediment core material

New biomarker proxy data was measured on a number of marine sediment samples from both core-tops (0–1 cm) of multicores and glacial sections of piston cores from the sub-Arctic and Arctic (Fig. 11; Tab. 1). The main target was to obtain biomarker sea ice records covering several D–O cycles during the last glacial with decadal to centennial-scale resolution. Three high-resolution sea ice records were generated for three key core sites in the southern and central Norwegian Sea (presented in Paper 1 and Paper 2) and in the northwestern North Atlantic (presented in Paper 3) (sites 1–3; Fig. 11; Tab. 1). A variety of glacial sediment samples from core sites at the eastern and southern Greenland margins were additionally tested for their potential of applying the sea ice biomarker approach and providing high-resolution glacial sea ice records (sites 4–7; Fig. 11; Tab. 1). In most cases, however, biomarker extractions and analyses of these samples yielded IP₂₅ (and other biomarker) concentrations below or close to the instrumental detection limit. In addition, multiple core-top samples were analyzed to supplement and evaluate the sea ice biomarker calibration database and allow for improved quantitative sea ice reconstructions during glacial D–O cycles (sites 8–48; Fig. 11; Tab. 1) (presented in Paper 2 and Paper 3).

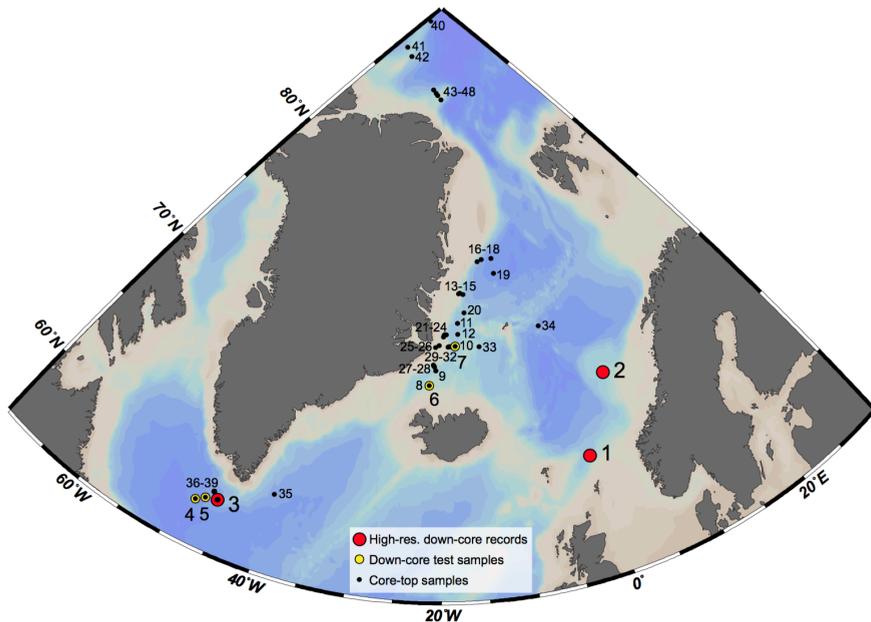


Figure 11 Core sites investigated in this PhD project. The core sites and proxy data produced are detailed in Table 1, according to the numbering shown here. Bathymetric map was produced with the Ocean Data View software (Schlitzer, 2016).

Table 1 Sediment cores and proxy data investigated in this PhD project. Numbers correspond to those shown in Figure 11. 1–3: High-resolution down-core records. 4–7: Down-core test samples. 8–48: Core-top samples.

No.	Core	Latitude (°N)	Longitude (°E)	Water depth (m)	Proxy data
1	MD99-2284	62.375	-0.980	1,500	HBIs (no HBI-III), sterols, TOC
2	MD95-2010	66.684	4.566	1,226	HBIs, sterols, alkenones, TOC
3	GS16-204-23CC	58.229	-45.698	2,270	HBIs, sterols, TOC, pl. ^{14}C , pl. $\delta^{18}\text{O}$
4	GS16-204-21CC	57.712	-48.012	3,335	HBIs, sterols, TOC
5	GS16-204-22CC-A	58.047	-47.039	3,160	HBIs, sterols, TOC
6	GS15-198-36CC	67.861	-21.882	770	HBIs, sterols, TOC
7	GS15-198-38CC	70.127	-17.663	1,610	HBIs, sterols, TOC
8	GS15-198-36MC	67.861	-21.882	770	HBIs, sterols, TOC
9	GS15-198-37MC	68.736	-20.873	1,284	HBIs, sterols, TOC
10	GS15-198-38MC	70.127	-17.663	1,610	HBIs, sterols, TOC
11	GS15-198-39MC	71.471	-17.083	1,699	HBIs, sterols, TOC
12	GS15-198-40MC	70.823	-17.080	1,566	HBIs, sterols, TOC
13	GS15-198-41MC	73.138	-15.679	1,992	HBIs, sterols, TOC
14	GS15-198-42MC	73.188	-16.521	826	HBIs, sterols, TOC
15	GS15-198-43MC	73.242	-16.296	1,251	HBIs, sterols, TOC
16	GS15-198-44MC	74.948	-11.922	1,950	HBIs, sterols, TOC
17	GS15-198-45MC	75.031	-10.947	2,628	HBIs, sterols, TOC
18	GS15-198-46MC	75.003	-8.821	3,266	HBIs, sterols, TOC
19	GS15-198-47MC	74.140	-8.821	3,046	HBIs, sterols, TOC
20	GS15-198-48MC	72.072	-15.751	1,402	HBIs, sterols, TOC
21	GS15-198-49MC	70.828	-19.113	1,292	HBIs, sterols, TOC
22	GS15-198-50MC	70.839	-19.432	942	HBIs, sterols, TOC
23	GS15-198-51MC	70.796	-19.509	712	HBIs, sterols, TOC
24	GS15-198-52MC	70.711	-19.655	401	HBIs, sterols, TOC
25	GS15-198-53MC	70.195	-20.379	311	HBIs, sterols, TOC
26	GS15-198-54MC	69.064	-21.294	660	HBIs, sterols, TOC
27	GS15-198-55MC	69.012	-21.274	942	HBIs, sterols, TOC
28	GS15-198-56MC	68.962	-21.140	1,193	HBIs, sterols, TOC
29	GS15-198-58MC	70.104	-20.987	602	HBIs, sterols, TOC
30	GS15-198-59MC	70.115	-18.832	903	HBIs, sterols, TOC
31	GS15-198-60MC	70.129	-18.644	1,167	HBIs, sterols, TOC
32	GS15-198-61MC	70.123	-18.413	1,423	HBIs, sterols, TOC
33	GS15-198-62MC	70.020	-13.563	1,423	HBIs, sterols, TOC
34	GS15-198-63MC	70.528	-2.756	2,995	HBIs, sterols, TOC
35	GS16-204-19MC	59.814	-39.799	2,675	HBIs, sterols, TOC
36	GS16-204-21MC	57.712	-48.012	3,335	HBIs, sterols, TOC
37	GS16-204-22MC	58.047	-47.039	3,160	HBIs, sterols, TOC
38	GS16-204-23MC	58.229	-45.698	2,270	HBIs, sterols, TOC
39	GS16-204-24MC	58.604	-46.380	2,540	HBIs, sterols, TOC
40	FRAM2014/15-08	89.147	-96.200	2,006	HBIs, sterols, TOC
41	FRAM2014/15-11	87.038	-61.357	1,260	HBIs, sterols, TOC
42	FRAM2014/15-13	86.722	-52.088	960	HBIs, sterols, TOC
43	FRAM2014/15-15-1	85.180	-25.387	1,340	HBIs, sterols, TOC
44	FRAM2014/15-15-2	85.008	-23.540	1,160	HBIs, sterols, TOC
45	FRAM2014/15-15-3	84.872	-22.442	1,750	HBIs, sterols, TOC
46	FRAM2014/15-15-4	84.800	-21.547	1,630	HBIs, sterols, TOC
47	FRAM2014/15-15-5	84.612	-20.238	810	HBIs, sterols, TOC
48	FRAM2014/15-15-6	84.503	-19.880	820	HBIs, sterols, TOC

3.3 Sampling, sample preparation and analysis

Two of the sediment cores (sites 1 and 2) investigated were previously retrieved during *Marion Dufresne* (IMAGES I) cruise MD101 in 1995 and *Marion Dufresne* (IMAGES V) cruise MD114 in 1999. The other piston cores and multicores were retrieved during *G.O. Sars* cruises GS15-198 in 2015 (https://www.bcdc.no/files/bcdc-theme/documents/GS15-198_cruise%20report.pdf) and GS16-204 in 2016 (https://www.bcdc.no/files/bcdc-theme/documents/GS16-204_cruise%20report.pdf) within the framework of the Ice2Ice project. After collection, the sediment cores were stored at 4°C in the facilities of the Department of Earth Science, University of Bergen, and Uni Research Climate (now NORCE Norwegian Research Centre). Sediment cores were sampled with metal spatulas. Sediment samples of ~4–30 g wet weight were transferred into glass vials and immediately frozen at –20°C for several days. Frozen sediment samples were freeze-dried, subsequently weighed to obtain the water content and dry weights of each sample, and then homogenized using an agate mortar.

Firstly, the total organic carbon (TOC) content was measured on ~100 mg of freeze-dried and homogenized sediment powder using a carbon-sulphur determinator (CS-125, Leco) at the Alfred Wegener Institute, Bremerhaven, after removal of carbonate by adding 500 ml hydrochloric acid to each sample. Secondly, biomarkers were extracted from ~5 g of freeze-dried and homogenized sediment powder either using an accelerated solvent extractor (ASE) or ultrasonication, and dichloromethane:methanol (2:1, v/v) as solvent. Before extraction, various internal standards were added to each sample for quantification purposes. The total extracts were then separated into a hydrocarbon fraction and a sterol fraction using open-column chromatography, and analyzed by gas chromatography/mass spectrometry. All chemical sample preparations and biomarker analyses were performed in the organic geochemical laboratory facilities of the Alfred Wegener Institute. Due to slightly different methodological approaches applied on samples from the different cores, not all of the various biomarkers outlined in section 3.1 were equally obtained for all cores (Tab. 1). Further details on the biomarker extraction and analyses are given in the three manuscripts of this thesis (Papers 1–3).

For supplementary analyses of planktic $\delta^{18}\text{O}$, $\delta^{13}\text{C}$ and ^{14}C on core GS16-204-23CC (Tab. 1), wet sediment samples were washed over a 63 μm sieve, oven-dried at 50°C, and subsequently dry-sieved into different size fractions. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ analyses were performed on 15–20 specimens of the polar planktic foraminifer *Neogloboquadrina pachyderma* (sinistral) using mass spectrometry at the FARLAB, Department of Earth Science, University of Bergen. ^{14}C ages were measured on ~4–14 mg CaCO_3 of *N. pachyderma* (s) using accelerator mass spectrometry at the W. M. Keck Laboratory, University of California, Irvine and at the Beta Analytic radiocarbon laboratory.

More methodological details on biomarker analyses, previously published supplementary proxy records, construction of sediment core chronologies, and statistical analyses are given in the manuscripts (Papers 1–3).

4 Summary of papers

4.1 Paper 1: *Sea ice variability in the southern Norwegian Sea during glacial Dansgaard–Oeschger climate cycles*

In Paper 1 we investigate the nature, timing, and role of sea ice fluctuations in the southern Norwegian Sea for abrupt ocean and climate change during the last glacial period. For this purpose, we present new multi-decadal to centennial-scale resolution sea ice biomarker records from core MD99-2284 covering four Dansgaard–Oeschger climate cycles at 32–40 thousand years ago. The sea ice proxy reconstruction is based on the sea ice algae biomarker IP₂₅, open-water phytoplankton sterols (brassicasterol and dinosterol), and the phytoplankton-IP₂₅ index (PIP₂₅) to derive semi-quantitative sea ice estimates. Our biomarker records indicate a strong seasonal sea ice cover in the southern Norwegian Sea during early cold Greenland stadials and largely open-ocean conditions during peak warm Greenland interstadials. These reconstructed sea ice fluctuations are largely consistent with model output data of sea ice cover in the Norwegian Sea, based on a transient simulation with LOVECLIM (Menviel et al., 2014). A statistical evaluation of the new sea ice records with published proxy records of near-surface temperature and planktic and benthic foraminiferal stable oxygen and carbon isotopes from the same sediment core (Dokken et al., 2013), reveals insights into the timing of sea ice changes with respect to deep ocean convection. We find that initial sea ice reductions at the core site preceded the major reinvigoration of convective deep-water formation in the Nordic Seas and abrupt Greenland warming. In turn, sea ice expansions led the buildup of a deep oceanic heat reservoir. Our findings suggest that the sea ice variability shaped regime shifts between surface stratification and deep convection in the Nordic Seas during abrupt glacial climate changes.

4.2 Paper 2: *Rapid sea ice reduction in the Nordic Seas and abrupt warming over Greenland during the last glacial*

In Paper 2 we evaluate the biomarker approach for sea ice reconstruction and investigate spatio-temporal changes in the glacial sea ice cover in the Nordic Seas, with a focus on the link between large-scale sea ice reductions and abrupt warming in Greenland. Firstly, we present new biomarker sea ice data from several core-tops from the northern, eastern and southern Greenland margins, supplementing the database of published biomarker sea ice data from Arctic and sub-Arctic regions. Compared with satellite-based sea ice observations, our core-top biomarker data support an increased sedimentary IP₂₅ abundance under seasonal sea ice. Furthermore, we establish a calibration curve for down-core biomarker records, based on a robust linear correlation between core-top P_{III}IP₂₅ (combining IP₂₅ and the highly branched isoprenoid triene, HBI-III) and spring sea ice concentration ($R^2=0.87$), following Smik et al. (2016) and Kolling (2017). Secondly, we present three independent sea ice records for the glacial period 32–41 thousand years ago, based on new biomarker sea ice data from core MD95-2010 (central Norwegian Sea), biomarker sea ice data from core MD99-2284 (southern Norwegian Sea), and bromine-enrichment sea ice data from the RECAP ice core (East Greenland). Our sea ice records consistently reveal millennial-scale sea ice variability in the Nordic Seas during several Dansgaard–Oeschger climate cycles. Our independent sea ice records suggest substantial and large-scale rapid sea ice reductions within ~250 years or less at abrupt cold-to-warm Greenland climate transitions, following a more gradual initial disappearance of seasonal sea ice in the southern Norwegian Sea. This indicates that rapid sea ice retreat shaped the threshold-like transition from surface stratification with an extensive sea ice cover to major deep ocean convection in the Norwegian Sea with extended ice-free conditions. Our empirical evidence supports the critical role of sea ice decline as a positive feedback to unleash abrupt and large-amplitude atmospheric warming in Greenland during the last glacial.

4.3 Paper 3: *Evidence of deep *Zoophycos* burrowing and an enhanced glacial sea ice cover from the Eirik Drift south of Greenland*

In Paper 3 we report on new and published magnetic susceptibility, planktic foraminiferal $\delta^{18}\text{O}$ and ^{14}C data of four Eirik Drift sediment cores from 2,220–2,270 m water depth and use these proxy records to constrain the glacial stratigraphy in this highly dynamic sedimentation regime. New lipid biomarker data is presented for one of the cores (GS16-204-23CC) with the aim to investigate the sea ice cover south of Greenland during the last glacial period. The highly consistent magnetic susceptibility records of the cores were aligned in order to place all multi-proxy records on a common depth scale. A chronology is developed for the last 65 thousand years using a ^{14}C -based Bayesian chronology combined with a tuning of glacial meltwater signals in planktic $\delta^{18}\text{O}$ records from the Eirik Drift to pertinent $\delta^{18}\text{O}$ signals in a well-dated reference record from the northern Denmark Strait (Voelker et al., 1998). In two cores we find distinct intervals of several aberrant planktic $\delta^{18}\text{O}$ values and aberrant ^{14}C ages that are up to $\sim 19,000$ ^{14}C years younger than underlying and overlying ^{14}C ages. Following Leuschner et al. (2002) and Küssner et al. (2018), we relate this stratigraphic distortion to deep *Zoophycos* burrowing, which suggests a displacement of foraminifera with ages of $<10,000$ – $12,500$ ^{14}C years by more than 400 cm down into glacial sediments, particularly identified within Greenland Interstadial 8 (~ 38 thousand years ago). Deep burrowing was probably related to declining food supply driven by a major shift towards enhanced sediment winnowing under strengthened bottom currents during the Holocene. Despite the stratigraphic distortion, an increased abundance of the sea ice algae biomarker IP₂₅ in surrounding glacial sediments suggests an enhanced seasonal sea ice cover and/or sea ice export south of Greenland 32–41 thousand years ago, forming an important freshwater source for the glacial North Atlantic. Some millennial-scale sea ice variations may have occurred, but we find no evidence of major open-ocean conditions at the core site during Dansgaard–Oeschger warming events.

5 Synthesis and outlook

5.1 Synthesis

The research presented in this PhD thesis significantly advances the knowledge of past sea ice dynamics in the Nordic Seas during glacial millennial-scale abrupt climate changes. I generated multi-decadal to centennial-scale biomarker sea ice records from the Norwegian Sea and northwestern North Atlantic covering four D–O cycles ~30–40 thousand years ago. The new sea ice proxy results provide detailed and solid empirical evidence that 1) unprecedentedly resolves the nature, timing and role of sea ice variability during the abrupt D–O climate changes, and 2) highlights new pressing issues to be addressed by future research.

The high-resolution sea ice records presented in Paper 1 suggest that glacial sea ice fluctuations did not simply follow air temperature changes recorded in Greenland ice cores. Sea ice variations in the southern Norwegian Sea during D–O cycles were rather asynchronous with Greenland temperature changes. The detailed sea ice reconstruction presented in this thesis allows reconciling lower-resolution sea ice records from the southern Norwegian Sea, which revealed partly contrasting trends in glacial sea ice variability (Hoff et al., 2016; Wary et al., 2016). The previously published sea ice records by and large showed either enhanced sea ice cover during GI and reduced seasonal sea ice cover during GS (Wary et al., 2016) or a near-perennial sea ice cover during GS and reduced sea ice conditions during GI (Hoff et al., 2016). Our high-resolution sea ice records suggest that in the southern Norwegian Sea the seasonal sea ice cover was maximal at GI–GS transitions and during early parts of GS; initial sea ice retreat occurred during late GS; minimum sea ice or largely open-ocean conditions were reached at GS–GI transitions and persisted during peak GI, but only for ~200 years on average; and seasonal sea ice occurrence increased throughout later GI (Paper 1). Another new sea ice record from a core site further north (Paper 2) supports these trends in glacial sea ice variability, but also shows that in the central Norwegian Sea shifts from an extended sea ice cover to a reduced seasonal sea ice cover were more rapid and in phase with the Greenland D–O climate transitions (Paper 2). Hence, the new findings presented in this thesis are in agreement with model simulations (Menviel et al., 2014) and generally support the stadial and interstadial sea ice scenarios proposed by Dokken et al. (2013) (Fig. 8). However, they also indicate that the previously unresolved sea ice fluctuations in the glacial

Nordic Seas were an integral part of a cascade of processes leading to the abrupt climate changes in Greenland.

The results presented in Paper 1 illustrate that variations in seasonal sea ice (IP₂₅) and near-surface temperature in the southern Norwegian Sea were closely linked to each other, but asynchronous with variations in deep-water temperature at ~1,500 m in the Nordic Seas. This indicates that seasonal sea ice retreat in the southern Norwegian Sea was probably coupled with an enhanced advection of warm and saline surface waters from the North Atlantic during late GS (Paper 1). Once the winter sea ice season shortened in the southern Norwegian Sea, convective (winter) mixing of the surface ocean probably strengthened during late GS. However, the intermediate-deep heat reservoir of the Nordic Seas was still unaffected and did not yet release heat to the atmosphere. This points at near-surface thermohaline circulation changes that were accompanied by the rather gradual retreat in seasonal sea ice cover in the southern Norwegian Sea, which preceded the abrupt GS–GI climate transition in Greenland. The ultimate trigger of these changes remains unresolved, but the new sea ice records do not support a mechanism that would lead to a sudden sea ice demise (within years or decades) in the larger Nordic Seas/North Atlantic area, as observed in some model simulations (Vettoretti and Peltier, 2018). Nevertheless, the biomarker proxy evidence of initial seasonal sea ice retreat during later GS supports that “early-warning signals” for abrupt D–O events, identified in the $\delta^{18}\text{O}$ of the Greenland ice core, were caused by a destabilization of the sea ice cover on its way to a bifurcation (Boers, 2018).

In fact, the sea ice records of core MD95-2010 from the central Norwegian Sea and that of the RECAP ice core presented in Paper 2 unambiguously resolve the rapid large-scale sea ice decline reflecting the bifurcation point at GS–GI transitions. This rapid large-scale sea ice decline, which occurred within ~250 years or less in the proxy records, provides unprecedented and robust evidence of a rapid sea ice switch mechanism that was proposed to shape the abrupt atmospheric warming of a D–O event (Timmermann et al., 2003; Gildor and Tziperman, 2003). The GS–GI bifurcation point was not only marked by the rapid large-scale sea ice decline, but also by coeval major reinvigoration of deep ocean convection in the Nordic Seas and an abrupt shift from Polar to warmer Atlantic surface conditions in the southern

Norwegian Sea (Paper 2). This, happening after the initial sea ice retreat and associated enhanced Atlantic inflow in the southern Norwegian Sea, reinforces that thermohaline circulation changes or other changes in the northward extent of the upper limb of the AMOC may have triggered the rapid switches in sea ice cover (Timmermann et al., 2003; Gildor and Tziperman, 2003; Menviel et al., 2014). The rapid sea ice decline in the central Norwegian Sea, documented and constrained in this PhD thesis, implies an amplification of atmospheric warming through a decreased albedo effect and increased exposure of relatively warm subsurface and surface-ocean waters to the atmosphere (Gildor and Tziperman, 2003). Our new sea ice proxy evidence thus ultimately underpins the hypothesis that sea ice retreat acted as a critical feedback mechanism that shaped the extremely abrupt and large-amplitude atmospheric warming of Greenland's D–O events.

The results of this thesis also indicate that production and export of sea ice may have acted as important freshwater source for the glacial Nordic Seas and northern North Atlantic. The sea ice record from the southern Norwegian Sea, presented in Paper 1, illustrates that seasonal sea ice was present and increased throughout GI, in agreement with previous findings (Hoff et al., 2016; Wary et al., 2016). This trend of increasing seasonal sea ice appears to have initiated with a previously unresolved intra-interstadial sea ice expansion event, occurring after the peak warmth in Greenland (Paper 1). The increasing seasonal sea ice cover presumably resulted from freshwater input from the nearby Scandinavian and British ice sheets (Dokken et al., 2013; Alvarez-Solas et al., in rev.) and may have fostered surface stratification, which in turn reduced deep convection in the Nordic Seas. This was most likely one important factor causing the gradual atmospheric cooling over Greenland during GI. On the other hand, the rapid shift from a reduced to an extended sea ice cover in the central Norwegian Sea illustrates the presence of a threshold response or bifurcation point marking the GI–GS cooling transitions (Paper 2). Hence, our new data also support that rapid large-scale sea ice expansion rapidly increased the albedo, reduced ocean–atmosphere heat exchange in the Nordic Seas, and thus amplified the atmospheric cooling over Greenland. This testifies not only the role of sea ice-related feedbacks in amplifying the abrupt cooling recorded at GI–GS transitions in Greenland ice cores, but also the importance of sea ice for the buildup of an intermediate-deep heat reservoir in the Nordic Seas.

The high-resolution sea ice record from the Eirik Drift, presented in Paper 3, reveals an enhanced seasonal sea ice presence south of Greenland ~31–41 thousand years ago. This may reflect a strengthened EGC and/or greater southward extension of the EGC, associated with increased sea ice export to the Labrador Sea during glacial D–O climate cycles, compared to today (Paper 3). The results of this thesis thus support previous suggestions that increased sea ice export from the Arctic Ocean and Nordic Seas to the northwestern North Atlantic might have contributed to a surface freshening and reduced deep convection in the North Atlantic during the last glacial (Vettoretti and Peltier, 2018). Although the glacial sea ice export to the Labrador Sea might have varied on millennial timescales, the exact nature and timing of these variations remain poorly constrained. The new sea ice record from a core site ~200 km south of Greenland’s southern tip does not reveal indications of Atlantic-like open-ocean conditions, as found for the Norwegian Sea during D–O events (Paper 3). However, our proxy evidence does neither support nor reject the suggestion of a super polynya in the northwestern North Atlantic, which was found to lead to abrupt Greenland warming in a model simulation (Vettoretti and Peltier, 2016).

Furthermore, the results presented in this thesis illuminate both processes that hamper robust age models of sediment cores and ways to improve those age models. In paper 3 we argue that deep burrowing, probably by the *Zoophycos* producer, leads to displacement of younger foraminifer shells by >400 cm into older sediments. This results in a negative radiocarbon age offset of up to >19,000 years and distorted geochemical proxy signals, a stratigraphic distortion with previously unreported dimensions (Küssner et al., 2018). The severe stratigraphic distortion caused by deep burrowing, now documented for glacial sediment sections from the Eirik Drift crest, adds to the processes preventing reliable ¹⁴C-based age models. In general, ¹⁴C-based age models appear insufficient to precisely constrain the D–O variability ~30–40 thousand years ago in marine proxy records, as both the analytical uncertainty and the uncertainty in the ¹⁴C calibration curves for this time period are in the order of several hundreds of years (Reimer et al., 2013). Moreover, calibration of glacial ¹⁴C ages is biased by variable and mostly unknown reservoir ages of seawater, which must be expected especially in northern high-latitude ocean regions characterized by varying sea ice and ocean convection conditions. Nevertheless, results of this thesis illustrate that robust age models for sediment cores from the Nordic Seas and northwestern

North Atlantic can be generated by stratigraphic alignment or tuning of sediment core and ice core proxy signals (Papers 1–3). We could show that identification and geochemical characterization of cryptotephra layers can independently validate and improve tuning-based age models (Paper 1).

Based on such thorough tuning-based chronologies, we demonstrate in Paper 2 that reconstructed trends in sea ice cover in the Norwegian Sea are consistent with those found in the RECAP ice core. With the well constrained ice core chronology and local sea ice signals from the sediment core records, this thesis presents the most comprehensive and detailed sea ice reconstruction available for glacial abrupt D–O climate transitions (Paper 2). Moreover, core-top biomarker data included in Paper 2 support and extend the previously observed strong linear relationship between the $P_{III}IP_{25}$ and spring sea ice concentration (Smik et al., 2016; Kolling, 2017). This enabled an advanced, quantitative reconstruction of changes in spring sea ice concentration in the central Norwegian Sea over the course of the D–O cycles. Accordingly, findings presented in this thesis suggest that the rapid sea ice decline that amplified the abrupt Greenland warming of a D–O event may correspond to a ~50 % reduction of spring sea ice concentration at this particular core site in the central Norwegian Sea (Paper 2).

5.2 Outlook

I demonstrated that present and past sea ice conditions in sub-Arctic regions can be thoroughly reconstructed using sea ice algae and phytoplankton biomarkers from core-top and down-core samples, respectively. In particular, the $P_{III}IP_{25}$ index (IP_{25} combined with HBI-III) reveals promising results with respect to quantitative sea ice reconstructions, that is spring sea ice concentration (Paper 2; Smik et al., 2016; Kolling, 2017). The new biomarker data presented in this thesis supplement the collection of Arctic and sub-Arctic core-top biomarker data. This extended database can be used to reassess the full potential of the different biomarkers for sea ice reconstruction and in particular advance the applicability of HBI-III and $P_{III}IP_{25}$ for sea ice margin and *quantitative* sea ice reconstructions (following Belt et al., 2015; Smik et al., 2016; and Kolling, 2017). Furthermore, the sea ice biomarker data presented in this thesis will serve as invaluable reference proxy database for the

development of new sea ice proxies such as from ancient DNA. The aDNAPROX project (led by Stijn De Schepper) is currently pursuing this attempt at the NORCE Norwegian Research Centre, which aims at calibrating ancient DNA signals in core-top samples against sea ice data from biomarkers, dinocysts and satellite observations. This might allow distinguishing between proxy signals from organisms reflecting locally formed sea ice and drifted sea ice, which is crucial in EGC regions and currently difficult with existing sea ice proxy methods. In general, parallel analyses of different sea ice proxies in core-tops but also down-core records are required to test the robustness of each of the different sea ice proxies for sea ice reconstructions.

The results presented in this thesis reveal the glacial millennial-scale sea ice variability in the Norwegian Sea in great detail and provide preliminary insights into the sea ice variability in the northwestern North Atlantic. Biomarker analyses of a few test samples from glacial section in sediment cores from the eastern and southern Greenland margins indicate that sea ice reconstruction in regions influenced by the EGC are difficult, probably due to a generally lowered phytoplankton production in polar waters. Nevertheless, robust high-resolution sea ice records from the northeastern, northern and northwestern North Atlantic are desirable in order to investigate the larger-scale spatiotemporal sea ice variations during the glacial D–O cycles. This is particularly important to constrain the maximum sea ice extent in the glacial North Atlantic and evaluate the significance of the largely unknown state of sea ice cover and deep convection in the Labrador Sea, and testing the idea of a super polynya in the Irminger Sea, which leads to abrupt Greenland warming in a model simulation (Vettoretti and Peltier, 2016; Vettoretti and Peltier, 2018). Sea ice records from the northern North Atlantic might also help constraining the glacial source area for aerosols and moisture precipitated in Greenland, which would improve the understanding of signals in sea ice records and other climate records from Greenland ice cores.

In addition to further sea ice proxy records, future proxy records reflecting specific ocean conditions in the Nordic Seas and North Atlantic are needed to better constrain the dynamics involved in ocean circulation and sea ice changes during D–O climate cycles. This is crucial to better resolve and constrain the physical mechanisms that are found to potentially have caused the D–O ocean and climate variability in climate

model simulations (see 1.4 and 1.5). For example, compound-specific stable hydrogen isotopes (δD) on biomarkers, such as alkenones or sterols, might reflect changes in sea surface salinity (Schouten et al., 2006; Sachs et al., 2018), which could help better assessing the timing of freshwater input from the Scandinavian and British ice sheets and its role for sea ice expansion during GI. Furthermore, decadal to centennial-scale records of planktic and benthic ^{14}C from the Nordic Seas could reveal further insights into surface and deep ocean mixing (Grootes, 2014). Such high-resolution ^{14}C records covering the D–O cycles ~30–40 thousand years ago might document a strengthening in convective (winter) mixing in the southern Norwegian Sea, suggested to be associated with the initial sea ice retreat during late GS (Paper 2). This would be one important step in better resolving the mechanism causing the initial sea ice retreat during late GS, which intriguingly also occurred during Heinrich Event 4 (late GS9), when AMOC is believed to have been significantly reduced or even halted (Henry et al., 2016).

To better resolve and understand the link between changes in AMOC and sea ice variability in the Nordic Seas, it is crucial to establish extremely robust and high-precision chronologies for sediment core records from the Nordic Seas and for existing and upcoming Pa/Th records from the North Atlantic. In this regard, the chronology of core MD99-2284 – where the high-resolution sea ice record from the southern Norwegian Sea is from – will be independently supported and slightly refined by further crypto-tephra layers that can be geochemically and stratigraphically tied to tephra layers in the Greenland ice cores (Paper 2). This will enable us to constrain with even higher precision the timing and duration of shifts in sea ice cover (and other surface and deep ocean parameters) with respect to abrupt climate changes in Greenland ice core records. A robust chronology and the new high-resolution sea ice records from the Norwegian Sea may also be used to unravel how sea ice shifts in the Nordic Seas during D–O cycles were related to changes in the stability of the Greenland Ice Sheet. For this purpose, it would be necessary to link the sea ice records from the eastern Nordic Seas with IRD records from the western Nordic Seas, for example on the basis of consistent benthic $\delta^{18}O$ variations reflecting intermediate and deep water temperature changes. In addition, high-resolution sea level reconstructions with independent robust chronologies may allow for better

constraining the evolution of ice sheet instability and freshwater input during D–O cycles.

On the one hand, the present PhD thesis illustrates the great potential of biomarkers from Norwegian Sea sediment cores for detailed investigation of sea ice fluctuations during the last glacial. Also, the results of this thesis demonstrate that reinvigoration of deep convection in the Nordic Seas was accompanied by a rapid large-scale sea ice decline and abrupt atmospheric warming over Greenland during the D–O events ~30–40 thousand years ago. On the other hand, previous studies showed that paleoceanographic proxy records from core sites MD95-2010 and MD99-2284 reveal major oceanographic changes during deglacial climate transitions (Dokken and Jansen, 1999; Bakke et al., 2009). However, the deglacial sea ice conditions in the climatically important central Norwegian Sea are largely unexplored. Therefore, investigation of the evolution of the Norwegian Sea ice cover from the Last Glacial Maximum over the deglacial period into the early Holocene forms a key target that future research should address. Biomarker sea ice records might reveal unprecedented insights into the role of sea ice for an extremely poorly ventilated deep Arctic Mediterranean during the last glacial (Thornalley et al., 2015). Moreover, the role of sea ice retreat as a positive feedback mechanism for abrupt Greenland climate change, as resolved in this thesis for the glacial D–O events, might be investigated with respect to well-constrained AMOC changes, abrupt Greenland warmings and northward moisture source shifts ~10–18 thousand years ago (McManus et al., 2004; Steffensen et al., 2008). The question arises whether reinvigoration of deep convection and concomitant rapid sea ice retreat in the Nordic Seas might have contributed to deglacial abrupt atmospheric CO₂ rises by 10–15 ppm during times of abrupt Greenland warming, the source of which has yet to be constrained (Marcott et al., 2014).

Finally, the results of this PhD thesis might be relevant with respect to the current and future Arctic sea ice decline and climate change. It can be noted that the sea ice decline and the transition from Arctic-like surface stratification to Atlantic-like open-ocean convection, which our proxy records suggest to have happened in the Norwegian Sea at the onset of a D–O event, closely resemble observations from the modern Barents Sea (Lind et al., 2018). The reduction in (winter) sea ice occurrence

observed in the modern Barents Sea has been linked to an increased oceanic heat transport, that is, a strengthening and warming of the inflow of Atlantic waters into the Barents Sea (Årthun et al., 2012). This ‘Atlantification’ of the Barents Sea has also been linked with enhanced vertical mixing, increased upward heat fluxes, and the Arctic warming hotspot (Lind et al., 2018). It remains open, however, to what extent the sea ice decline and ‘Atlantification’ of the Arctic will expand and accelerate in the future, and how the rates of these changes and its consequences for the Greenland ice sheet, sea level rise and global climate change will eventually compete with those of the dramatic D–O climate events in the past.

6 References

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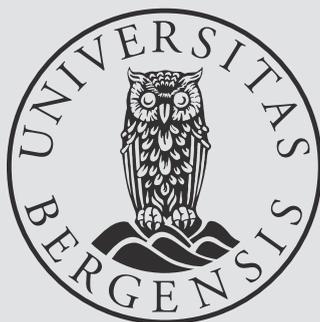
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