Can North Atlantic sea ice anomalies account for Dansgaard-Oeschger climate signals?

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This is publication no. XXXX from the Bjerknes Centre for Climate Research.
ABSTRACT

North Atlantic sea ice anomalies are thought to play an important role in the abrupt Dansgaard-Oeschger (D-O) cycles of the last glacial period. This model study investigates the impacts of changes in North Atlantic sea ice extent in glacial climates to help provide geographical constraints on their involvement in D-O cycles. Based on a coupled climate model simulation of the Last Glacial Maximum (21 ka), the Nordic Seas and western North Atlantic (broadly, south of Greenland) are identified as two plausible regions for large and persistent displacements of the sea ice edge in the glacial North Atlantic. Sea ice retreat scenarios targeting these regions are designed to represent ice cover changes associated with the cold-to-warm (stadial-to-interstadial) transitions of D-O cycles. The atmospheric response to sea ice retreat in the Nordics Seas and in the western North Atlantic are tested individually and together using an atmospheric general circulation model. Nordic Seas ice retreat causes 10° C of winter warming and a 50% increase in snow accumulation at Greenland Summit; concomitant ice retreat in the western North Atlantic has little additional effect. The results suggest that displacements of the winter sea ice edge in the Nordic Seas are important for creating the observed climate signals associated with D-O cycles in the Greenland ice cores.
1. Introduction

Sea ice is an important element in the glacial climate system due to its pivotal role in the surface heat, moisture and momentum budgets of the polar regions. The presence of sea ice lowers surface temperature by insulating the atmosphere from the ocean heat reservoir and by increasing surface albedo. Sea ice also affects the vertical structure of the upper ocean, alters deep water formation sites, and regulates the availability of moisture for building continental ice sheets (Hebbeln et al. 1994; Rind et al. 1995).

Sea ice is thought to be a principal player in Dansgaard-Oeschger (D-O) cycles, the millennial scale climate cycles that are prominent features of North Atlantic proxy records throughout much of the last ice age (Dansgaard et al. 1993; Bond et al. 1993; Taylor et al. 1993; Grootes and Stuiver 1997; Sachs and Lehman 1999; NGRIP Members 2004). The interstadial phase of the D-O cycle is a warm period in the mid- to high-latitude Atlantic region, while the stadial phase is a cold period. A remarkable feature of the D-O cycle is the abrupt transition between stadial (cold) and interstadial (warm) conditions, manifest in the Greenland ice cores by an annual mean warming of up to 16°C in a matter of decades (Lang et al. 1999; Landais et al. 2004a,b; Huber et al. 2006; Sánchez Goñi et al. 2008), and a 50–100% increase in snowfall (Cuffey and Clow 1997; Andersen et al. 2006).

Displacements of the sea ice edge have been proposed as a key ingredient of D-O cycles (Broecker 2000; Gildor and Tziperman 2003) because of their strong influence on regional temperature and their potential to occur rapidly in response to relatively weak forcing via processes such as the ice-albedo feedback (Curry et al. 1995). Reconstructing past sea ice variability is difficult, but some observational evidence exists to support the idea that changing sea ice extent plays an important role in D-O-related abrupt climate changes. For example, deuterium excess measurements in the Greenland ice cores exhibit rapid transitions associated with D-O cycles, and these transitions have been interpreted as north/south shifts of the moisture source for Greenland snow due to the retreat or advance of North Atlantic sea ice (Masson-Delmotte et al. 2005; Jouzel et al. 2007). Furthermore, a benthic δ¹⁸O record from a high resolution marine core on the Norwegian slope
indicates more sea ice formation during cold stadial phases of D-O cycles than during warm interstadials (Dokken et al. 2010).

The exact nature of sea ice’s involvement in D-O cycles is the subject of much current research. One important question is what mechanisms drive these displacements of the sea ice edge. Wind- and ocean-driven Arctic sea ice variability in today’s climate exists on times scales from weeks to decades (e.g., Fang and Wallace 1994; Venegas and Mysak 2000; Deser et al. 2002), but the millennial pacing of D-O cycles requires long-lived changes in sea ice cover. One hypothesis attributes the cycles to an internal oscillation of the ocean thermohaline circulation (OTC) that causes abrupt shifts in ocean heat transport into the North Atlantic, thereby altering sea ice and regional climate (Broecker et al. 1985). A recent multi-proxy analysis of a high-resolution marine core on the Norwegian slope suggests an alternative hypothesis: a series of interactions between the ocean and the marine and terrestrial cryosphere alternately builds and erodes a fresh surface halocline in the Nordic Seas over millennial times scales, promoting the presence or absence of sea ice cover and a toggling of climate between cold stadials and warm interstadials (Dokken et al. 2010). During stadials, the ocean continues to supply heat from more southerly latitudes, but this heat is stored beneath the surface halocline where it remains inaccessible to the atmosphere. The subsurface ocean experiences a warming, also seen in other North Atlantic marine cores and in climate models under certain freshwater hosing scenarios (Knutti et al. 2004; Rasmussen and Thomsen 2004; Ruhlemann et al. 2004; Cheng et al. 2007; Mignot et al. 2007). The halocline is maintained throughout the stadial owing in large part to high rates of sea ice formation, a process that rejects brine in dense, sinking plumes and encourages stratification (Dokken et al. 2010). Eventually, the subsurface becomes warm enough to destabilize the water column. The surface halocline is eroded, and the release of stored heat in combination with other hypothesized changes in heat flux convergence in the high latitudes produce warm, ice-free ocean conditions through the duration of the interstadial.

We aim to assess what type of sea ice anomalies can create the signatures of cold stadials and warm
interstadials recorded in paleoclimate archives. Previous modelling experiments using atmospheric general circulation models (GCMs) suggest that winter ice edge displacements produce temperature and accumulation changes consistent with observations from the Greenland ice cores, while summer ice edge displacements produce too much of an accumulation effect and not enough of a temperature effect (Li et al. 2005). The Li et al. (2005) study supports other observational and modelling studies that identify abrupt North Atlantic climate events as primarily a wintertime phenomenon (Renssen and Isarin 2001; Denton et al. 2005), but several caveats concerning the modelling results must be mentioned. First, the ocean boundary conditions used in the study were based on a glacial sea surface temperature (SST) and sea ice reconstruction (CLIMAP; Crowley 2000) that is now known to contain substantial errors, particularly in regions such as the tropical Pacific and North Atlantic (see synthesis in Li and Battisti 2008). Second, the prescribed sea ice edge displacement patterns were somewhat arbitrary and highly simplified.

Li et al. (2005) established seasonality constraints on D-O-related displacements of the sea ice edge, pointing to large winter changes and small summer changes as a necessary condition. The goal of the present work is to develop some intuition for the geographical nature of these sea ice anomalies. Because there are still very few reconstructions of past sea ice conditions, we adopt a modelling approach using a combination of coupled and atmosphere-only simulations to extend the work of Li et al. (2005). Coupled model simulations are examined to establish robust characteristics of the North Atlantic sea ice cover in glacial times. Sea ice retreat scenarios are created based on these glacial sea ice distributions, making the results of this study more applicable to D-O events than those of Li et al. (2005), in which the retreat scenarios were defined in a somewhat ad hoc fashion. While geographically more realistic, the scenarios remain idealized in the sense that where sea ice retreats, the sea surface temperatures are set to the freezing point, and sea surface conditions elsewhere (outside of the North Atlantic) are not altered. We focus on two issues in particular: (1) What are the characteristics of the North Atlantic sea ice cover in glacial climates, and in which regions is it susceptible to exhibiting large anomalies? (2) Is it possible to place geographical
constraints on D-O related displacements of the ice edge in these regions? In answering the second question, we wish to evaluate where sea ice anomalies must occur to have a substantial climate impact consistent with the D-O signals recorded in Greenland rather than to establish a precise D-O retreat scenario that produces an exact match to the ice core records.

Section 2 contains a description of the simulations and model data sets. In section 3, we analyse coupled model simulations of the LGM to identify the Nordic Seas and western North Atlantic as regions of interest for North Atlantic winter sea ice variability. Section 4 examines surface climate and circulation changes associated with sea ice perturbations in these two regions using a series of atmosphere-only experiments. Section 5 presents a discussion of pending issues related to the response of the atmosphere to North Atlantic sea ice perturbations and implications for D-O cycles. Finally, concluding remarks are presented in section 6.

2. Data and methods

We use sea ice data from coupled climate model simulations of Last Glacial Maximum (LGM; 21 ka) climate from the second phase of the Paleoclimate Model Intercomparison Project (PMIP2). The models included in the analysis are listed in Table 1. For the LGM, the models are forced in compliance with PMIP2 protocol: 21 ka insolation; atmospheric greenhouse gas concentrations based on ice core measurements (Flückiger et al. 1999; Dällenbach et al. 2000; Monnin et al. 2001); atmospheric aerosols at preindustrial values; land ice and coastlines from the ICE-5G reconstruction (Peltier 2004) corresponding to 120 m sea level depression at LGM. The PMIP2 control simulations for the pre-industrial (PI) climate are also used for comparison. General features of the simulated climates are presented in Braconnot et al. (2007). The LGM simulations exhibit good agreement with key changes indicated by proxy data such as a weakening of the monsoons, cooling of the tropics, cooling of the North Atlantic region and Eurasian continent, and a southeastward extension of the North Atlantic storm track (Kageyama et al. 2006; Braconnot et al. 2007;
Laîné et al. 2008; Otto-Bliesner et al. 2009), although cooling in polar regions is underestimated (Masson-Delmotte et al. 2006). Complete documentation on the models and experimental set up is available at the PMIP2 website (http://pmip2.lsce.ipsl.fr).

One hundred years of monthly mean sea ice data from one of the models, the Community Climate System Model (CCSM3) developed at the National Center for Atmospheric Research (NCAR), is used to better understand the distribution and natural variability of North Atlantic sea ice at the LGM. The model and its performance in a present day set up is documented in Collins et al. (2006a), and the LGM simulation is described in Otto-Bliesner et al. (2006). Briefly, the relevant configuration of CCSM3 comprises the primitive equation Community Atmosphere Model version 3 (CAM3) at T42 horizontal resolution with 26 hybrid coordinate vertical levels (Collins et al. 2006b); a land model with land cover and plant functional types, prognostic soil and snow temperature and a river routing scheme (Dickinson et al. 2006); the NCAR implementation of the Parallel Ocean Program (POP) on a 320×384 displaced-pole grid (nominal horizontal resolution of 1°) with 40 vertical levels (Smith and Gent 2002); and a dynamic-thermodynamic sea ice model on the same grid as the ocean model (Briegleb et al. 2004). The model has also been used in a range of paleoclimate studies, in particular to examine the climate response to freshwater hosing in glacial and deglacial climates compared to the present climate (Cheng et al. 2007; Bitz et al. 2007; Liu et al. 2009; Otto-Bliesner and Brady 2010).

The climatological sea ice cover in the CCSM3 LGM simulation compares well with reconstructions from this time period (Li and Battisti 2008), and so is used as the starting point in designing sea ice scenarios representing the stadial and interstadial phases of the D-O cycle. CAM3, the atmospheric component of CCSM3, is forced with these scenarios to investigate the climate response to sea ice perturbations in key regions of the North Atlantic. The CAM3 experiments are initialized in January with output from the CCSM3 LGM simulation, and results are based on 30 years of monthly mean output after discarding 2 years of spinup. The National Snow and Ice Data Center’s sea ice index (detrended) generated from passive
microwave satellite data (Fetterer et al. 2009) is used to calculate Northern Hemisphere sea ice extent and variability in the 1978–2008 observing period for comparison with model simulations.

3. Coupled model simulations of Northern Hemisphere LGM sea ice

The objective of this section is to identify the regions where we expect large excursions of the winter sea ice edge in glacial climates. The analysis is based mainly on CCSM3 coupled model output because there are very few long, proxy-based reconstructions of glacial sea ice extent. We first provide an overview of Northern Hemisphere sea ice in all available LGM simulations from PMIP2, then focus on a more detailed analysis of the CCSM3 simulation.

a. PMIP2 model intercomparison

The PMIP2 model participants listed in Table 1 simulate a range of sea ice thickness and concentration distributions in the LGM climate (Figure 1). In the Northern Hemisphere, the higher resolution climate models produce annual mean ice areas from over $6.4 \times 10^{12}$ m$^2$ to $10.4 \times 10^{12}$ m$^2$, with mean Arctic ice thicknesses from 5.2 m to 15.7 m (Table 2). The two models with greatest mean ice thickness (CNRM-CM33 and ECBILTCLIO, the latter of which is the only low resolution intermediate complexity model of the group) produce on the order of double the volume of Northern Hemisphere sea ice compared to the other models, and exhibit relatively weak seasonal cycles in ice area and volume (Figure 1g and Table 2).

Because the reconstruction of past sea ice extent remains a difficult and unresolved challenge (Kucera et al. 2005), it is unclear how best to evaluate the reliability of these simulation results. All the higher resolution models perform quite well in simulating the mean Arctic sea ice cover over the 1981–2000 satellite observing period (Arzel et al. 2006); compared to the National Snow and Ice Data Center data set (Comiso 2002), the errors in March and September sea ice extent are <10% with the exception of certain cases such
as CCSM3 in March (+16%) and HadCM3M2 in September (-18%). Northern Hemisphere sea ice area in the PMIP2 control (PI) simulations ranges from $7.1-14.7 \times 10^{12} \text{m}^2$ (Table 2) and also compares reasonably well with estimates from the 1978–2008 observing period (approximately $12 \times 10^{12} \text{m}^2$, not detrended). However, good validation with observations does not necessarily indicate that these models have a better representation of sea ice processes, or that they will produce a more realistic evolution of sea ice in past or future scenarios.

There are several observational constraints that help to discriminate between the LGM simulations of these models. CNRM-CM33 and ECBILTCLIO appear to be the least reasonable models: they have extremely thick ice in the Arctic (15.7 m and 9.5 m annual mean), and perennial sea ice in the Nordic Seas. The latter is not in agreement with marine proxy data, which suggest the presence of at least seasonally open ocean in the Nordic Seas at the LGM (Meland et al. 2005; de Vernal et al. 2005; Kucera et al. 2005). Consistent with the proxy data, the summer Nordic Seas are entirely ice-free in two of the models, CCSM3 and HadCM3M2, while IPSL-CM4 and MIROC3.2 show a strip of open ocean confined to the Norwegian coast. All the models simulate sea ice concentrations close to 100% year round in the Arctic Basin, but show considerable differences in the low-concentration (<50%) regions of the western North Atlantic and western North Pacific. Consistent with proxy reconstructions, CCSM3 and HadCM3M2 exhibit the most winter sea ice in these regions, particularly in the Atlantic off the coast of North America, where one would expect favourable conditions for sea ice formation due to freshwater input and cold air advection off the margin of the Laurentide ice sheet. (Note, however, that a model bias could also be contributing to the western North Atlantic ice cover in CCSM3, which exhibits overly extensive sea ice in the Labrador Sea for the present climate (Jochum et al. 2008).) Hence, we accept the CCSM3 and HadCM3M2 as plausible representations of the sea ice extent at LGM, and further explore the former model to help guide the design of sea ice retreat scenarios that might be expected during glacial times.
b. **Maintenance and variability of North Atlantic sea ice in CCSM3**

We turn our attention to North Atlantic sea ice as simulated by CCSM3 to gain insight into its glacial distribution, and to identify regions where it can be sensitive to forcing, and hence likely to undergo large excursions. The processes controlling the hypothesized D-O sea ice anomalies are obviously not captured in this type of equilibrium coupled simulation, but we use the simulated climatology to guide the design of the sea ice perturbation experiments in the next section. These analyses serve also to illustrate that the North Atlantic sea ice cover can exhibit substantially altered behaviour in a different climate.

Annual mean sea ice thickness and concentration distributions are shown in Figure 1a. Of particular interest is the seasonal sea ice cover in the Nordic Seas and western North Atlantic. In winter, both regions exhibit extensive sea ice; in summer, ice retreats back towards the east coasts of Greenland and North America, leaving the Nordic Seas and western North Atlantic ice-free.

Figure 2 shows the movement, growth and melt of North Atlantic sea ice as JFM seasonal averages. These months were chosen to represent the heart of the Northern Hemisphere winter growth season: sea ice growth peaks in February, while the net gain in sea ice volume peaks in December/January. The growth term includes congelation (basal) and frazil (surface ocean layer) ice growth, and snow-ice formation by flooding of snow-covered ice. The melt term includes basal and lateral melt, determined largely by ocean-sea ice heat flux, and top melt (negligible during winter), determined largely by atmosphere-sea ice heat flux.

As in today’s climate, growth occurs in the Arctic basin and along high latitude coastlines, where persistent offshore winds drive sea ice away from the land to produce polynya-like conditions. High growth rates in the Nordic Seas are consistent with the presence of winter sea ice cover there, and thermodynamic processes account for over 90% of the ice volume tendency over the JFM season. Dynamics plays a more important role for the western North Atlantic ice tongue extending off the coast of North America. Growth rates are relatively small in the interior of the ice tongue (Labrador Sea and central North Atlantic), and melt dominates at the southern and eastern edges. Considerable ice growth is seen in Baffin Bay and Davis Strait,
and ice velocities indicate advection out from Davis Strait and south with the East Greenland Current to feed the ice tongue. In fact, the ice volume tendency due to thermodynamic processes is on average negative over the JFM season, and it is the large positive dynamical tendency that maintains the sea ice cover in the western North Atlantic.

Winter melt takes place along the ice edge, and in several locations in the interior of the Labrador Sea and the Nordic Seas (Figure 2, bottom panel). This melt occurs primarily at the base of the sea ice, although there is a substantial contribution from lateral melt along the southern portion of the East Greenland Current, and at the southeastern edge of the ice tongue (not shown). Basal and lateral melt are controlled by ocean-sea ice heat flux, hence winter melt is attributed mainly to ocean heat transport at the LGM, much as was found for the present climate in an earlier version (CCSM2) of the NCAR coupled model (Bitz et al. 2005). However, as noted by Bitz et al. (2005), large offshore ice growth and ice advection are important influences on the position of the ice edge in certain regions (such as the glacial western North Atlantic).

In addition to the sea ice cover being more extensive, other seasonal aspects are altered in the LGM simulation compared to the present day. For example, interannual sea ice variability in sea ice extent is mostly a wintertime phenomenon in the LGM simulation (Figure 3, bottom), whereas both ice area and sea ice volume exhibit substantial summer variability in observations (Deser et al. 2000) and in the simulations of present day climate (Figure 3, top). The interannual sea ice variability shown here arises from internal atmosphere-ocean processes in the coupled model, and it is unlikely that these are the same processes controlling the long-lived ice edge displacements in the proposed D-O mechanism. It is nevertheless informative that shifts in the seasonality of variability are possible in different climates. And despite the disparate time scales, the dominance of winter variability is consistent with paleoclimate evidence and modelling results (Renssen and Isarin 2001; Denton et al. 2005; Li et al. 2005) indicating that winter season ice anomalies were instrumental in producing abrupt climate change in the North Atlantic region during the last glaciation and the early Holocene.
Geometry is a chief cause for this shift in seasonality. The maps in Figure 3 show that variations in sea ice concentration occur in the marginal ice zones, the part of the ice cover that is affected by ocean processes such as swells and waves. Variability is small close to the pole, where relatively thick ice is more or less constantly present. If the sea ice edge is long (e.g., located far from the coast with excursions into open ocean), the marginal ice zone is extensive, and there is more opportunity for differences in ice extent from one year to the next. The LGM simulation has very little marginal ice zone in summer compared to winter, and consequently very little summer variability in sea ice area. In today’s climate, summer sea ice is restricted mostly to the Arctic Ocean, but there is a considerable amount of marginal ice because much of the ice cover at the periphery of the Arctic basin is relatively thin.

Like seasonality, the spatial signature of sea ice variability can also change in different climates. Today, sea ice variability has a dipole pattern with opposing centres of action in the Davis Strait/Labrador Sea region and the Nordic Seas from weekly (Fang and Wallace 1994) to seasonal (Walsh and Johnson 1979) time scales. The temporal variability of this dipole pattern is coupled to the North Atlantic Oscillation (NAO), the leading mode of Northern Hemisphere sea level pressure variability (Rogers and van Loon 1979; Fang and Wallace 1994). In the LGM simulation, sea ice variability in the Nordic Seas (NS) and western North Atlantic (WAtl) is not strongly related. The simultaneous correlation of winter sea ice area in the NS and WAtl regions is less than 0.3 (<10% shared variance); similar correlation values between the Nordic Seas and western North Atlantic are found with the latter leading by up to 2 years. Figure 4 shows composites of JFM sea ice concentration for the 25 years (i.e., 25% of the total 100 years) with the least (category “−”) and most (category “+”) sea ice cover in the two regions. The bar codes indicate that, one-third of the time, the regions experience anomalously low or high sea ice cover simultaneously; the other two-thirds of the time, the anomalously low or high ice conditions in the regions share no particular relationship. The maps in Figure 4 reinforce the point, with the NS− and NS+ composites showing weak concentration anomalies in the western North Atlantic, and the WAtl− and WAtl+ composites showing weak concentration anomalies.
in the Nordic Seas.

In summary, there are two main masses of sea ice in the glacial North Atlantic. The Nordic Seas ice distribution is governed primarily by in situ growth and melt in the CCSM3 simulation of LGM climate, whereas the western North Atlantic ice tongue owes its existence in large part to the import of ice formed in Baffin Bay and Davis Strait and advected southeastward into open ocean. Variability in the LGM sea ice extent is mainly a wintertime phenomenon and it is concentrated in these two regions, albeit uncorrelated in time.

4. Climate impacts of North Atlantic sea ice perturbations

In this section, we construct scenarios for hypothetical sea ice anomalies associated with D-O cycles and examine their climate impacts in CAM3, the atmospheric component of CCSM3. The scenarios are based on CCSM3’s simulated LGM sea ice cover, with alterations to two regions that are most likely to exhibit large displacements of the ice edge – the Nordic Seas and western North Atlantic – during winter months to represent possible stadial and interstadial conditions. The use of prescribed sea surface temperature (SST) and sea ice conditions is a common modelling strategy for testing the response of the atmosphere to changes in the surface ocean. In this case, the changes (sea ice only) are attributed to a mechanism for D-O cycles (Dokken et al. 2010) involving ocean processes that are not represented by the model (i.e., halocline formation, brine rejection, freshwater input from land ice). Thus, our experiments provide insight into the the climate impacts of sea ice anomalies in two regions of the North Atlantic, but not into what creates and maintains the sea ice anomalies themselves.

We first construct a stadial (cold) glacial state with extensive sea ice in the North Atlantic, and all other boundary conditions (insolation, greenhouse gases, ice sheets, land/ocean geometry) corresponding to 21 ka, as specified in PMIP2. The stadial simulation (Figure 5a) has identical sea surface conditions to the CCSM3 simulation in the western North Atlantic ice tongue region year-round, but higher winter sea
ice concentrations in and south of the Nordic Seas. We then perturb this state by removing sea ice in the Nordic Seas and the western North Atlantic during the winter (December to May) months such that there is a smooth transition from the November ice cover to the April ice cover (Figure 5e). Sea surface conditions for the scenarios are shown in Figure 5; scenario lowiceNS represents a retreat of sea ice in the Nordic Seas, scenario lowiceW Atl represents a retreat of sea ice in the western North Atlantic, and scenario lowice represents a retreat of sea ice in both regions.

Before examining the temperature response to the sea ice scenarios, we note that the CCSM3’s standard LGM simulation exhibits approximately half the observed annual mean LGM-to-present day temperature difference at Greenland Summit. In fact, this weak LGM Summit cooling is a general feature of the PMIP2 models. In a five-model intercomparison study, the model-to-data discrepancy in Greenland cooling was found to range from 6–16 °C, with the CCSM3 performing better than all but HadCM3M2 (Masson-Delmotte et al. 2006). Masson-Delmotte et al. (2006) examine some possible explanations for the model-to-data discrepancy and find that the lowering of the summit due to topographic smoothing (for example, in CCSM3, the model Summit at LGM is approximately 250 m lower than today’s Summit elevation of 3200 m) and shifts in the seasonality of precipitation produce small – and, in some cases, unhelpful – adjustments to the simulated Summit temperatures. They speculate that the discrepancy could be linked to missing factors such as the lack of dust changes, vegetation feedbacks, soil feedbacks, and the associated impacts on atmospheric heat transport to Greenland.

The model’s response to sea ice retreat will be compared to the Greenland ice core records of D-O warmings. A comprehensive review of ice core parameters that vary in concert with D-O cycles can be found in Wolff et al. (2009). The three proxy records of particular interest here are: 1) abrupt annual mean temperature changes derived from fractionated gases trapped in the ice (Severinghaus et al. 1998); 2) annual accumulation changes derived from ice layer thickness corrected by ice flow models (Alley et al. 1993); and 3) an accumulation-weighted temperature signal derived from the isotopic composition of the ice itself.
(Jouzel et al. 1997). This last data set is one of the best known ice core proxies, but also perhaps one of the most difficult to interpret as it is affected by a host of other factors such as changes in moisture source, moisture transport path, seasonality of snowfall and the strength of the atmospheric inversion (Cuffey et al. 1994; Johnsen et al. 1995; Masson-Delmotte et al. 2005). As noted in the introduction, the goal of these scenario experiments is to determine where sea ice anomalies must occur in order to have a substantial climate impact that is consistent with these ice core records; it is not to establish one definitive retreat scenario that produces an exact match to the ice core records.

We focus on the large scale features of the atmospheric response to sea ice retreat with some attention given to the response at Greenland Summit, defined as the two highest model grid boxes in Greenland (with topographic smoothing, the mean elevation is 2957 m above sea level compared to today’s elevation of 3200 m). Comparisons are also made at several other locations on Greenland (see details in Table 3 caption) where there is a clear signature of D-O cycles in existing ice cores. While the D-O-related isotope signal at these sites amplifies from south to north, differences in moisture source are thought to be partly responsible, and thus the spatial pattern of the associated temperature signal remains unclear (Johnsen et al. 2001; Wolff et al. 2009).

In the sea ice retreat scenarios, there is in general an increase in mean surface temperature and a reduction in its interannual variability over the North Atlantic (Figure 6b–d). The warming is quite localized to the region where sea ice is removed, is relatively shallow (up to about 700 hPa), and decays from south to north over Greenland (Table 3). Temperature is most sensitive to sea ice changes in the Nordic Seas. Removing sea ice in this region alone (scenario lowiceNS) causes over 10 °C of wintertime warming and 5 °C of annual mean warming in the Summit region (Table 3); sea ice retreat in the western North Atlantic alone (scenario lowiceWAtl) has less than half the effect on Summit temperature and accumulation as sea ice retreat in the Nordic Seas alone (scenario lowiceNS). If the Nordic Seas are already ice-free, then the additional retreat of western Atlantic sea ice has little effect, producing a statistically significant but small warming, amounting
to less than half a degree Celsius even in the winter season (compare scenarios lowiceNS and lowice).

Sea ice retreat also produces an increase in snow accumulation in Greenland (Table 3). Net accumulation (accumulation minus ablation) is calculated from the snow hydrology component of the land model (Figure 6e–h). A proper assessment of the ice sheet mass balance requires that the atmospheric fields be downscaled to adequately high spatial resolutions, then used to force an ice sheet model that includes a good representation of ice dynamics. However, the accumulation field can give a sense of the expected mass balance changes when sea ice retreats. As with temperature, accumulation is more affected by retreat of sea ice in the Nordic Seas alone (50% increase in the annual mean in the Summit region) than in the western North Atlantic alone (11% increase in the annual mean in the Summit region); and once the Nordic Seas ice cover has been removed (scenario lowiceNS), the further removal of western North Atlantic sea ice (scenario lowice) produces no additional accumulation increase (Table 3). The accumulation changes associated with Nordic Seas retreat are on the low end of observational estimates, which show a 50–100% increase at Summit and NorthGRIP during D-O warmings (Cuffey and Clow 1997; Andersen et al. 2006).

The Greenland warming in these sea ice retreat scenarios is somewhat less than estimates of up to 16°C from proxy data (Lang et al. 1999; Landais et al. 2004a,b; Huber et al. 2006; Sánchez Goñi et al. 2008), and up to 7°C in the previous model study of Li et al. (2005). We note, however, that the 16°C warming pertains to one of the largest warmings recorded in a D-O cycle (event #19), and that the range of abrupt warmings spans from 7–16°C (see Wolff et al. 2009, and references therein). As for the comparison to Li et al. (2005), there are several key differences in experimental setup that contribute to the larger warming response in that study. First, Li et al. (2005) used an old version of the model (CCM3) that allows for only 100% ice concentration or 0% ice concentration in each grid box, whereas the new version (CAM3) allows for partial sea ice cover over a grid box such that sea ice retreats from non-100% cover in most places. Second, the old experiments included increases in SSTs at locations where sea ice is removed, whereas the current experiments keep SSTs at the freezing point. If we increase SSTs above the freezing point where
sea ice is removed, more atmospheric warming is achieved (not shown). Practical limits on how much SSTs may rise are ultimately set by the ocean’s heat supply to the region. In the proposed D-O mechanism, the stadial–to-interstadial change in sea ice is part of a coupled, transient response of the climate system associated with the disappearance of the Nordic Seas halocline and the opening of the subsurface ocean heat reservoir to the atmosphere. Benthic foraminifera data show evidence of subsurface heat storage during stadials (possible warming of up to 5–6 °C in several hundred metres of the upper ocean), although there remains some debate over the exact interpretation of the benthics (Dokken et al. 2010). The existence of subsurface warming when there is extensive sea ice in the North Atlantic is also demonstrated in coupled climate model simulations, which produce a maximum warming of several degrees to 8 °C at depths of up to 1500 m in the Nordic Seas, with additional weaker warming extending below the surface cooling into the tropical and south tropical Atlantic (Knutti et al. 2004; Cheng et al. 2007; Mignot et al. 2007).

The weaker response upon removal of sea ice in the western North Atlantic compared to the Nordic Seas can be viewed in terms of the circulation changes and local diabatic changes produced in each case. Sea ice retreat in the Nordic Seas (scenario lowiceNS) causes a decrease in sea level pressure along the northern flank of the Icelandic low, implying a strengthening of the easterly flow over the Nordic Seas (Figure 7). This is consistent with Byrkjedal et al. (2006), who tested slightly different glacial ocean boundary conditions, and find an increase in warm easterly flows at Summit when the Nordic Seas are ice-free. These sea level pressure changes are accompanied by a local increase in tropospheric eddy kinetic energy consistent with more synoptic activity (not shown). When sea ice retreats in the western North Atlantic (scenario lowiceWS), only the western flank of the Icelandic low deepens, producing low level circulation changes that do not greatly impact Summit.

Modelling studies investigating the atmospheric response to sea ice variability in today’s climate and to future Arctic sea ice decline have found a large-scale atmospheric response that projects strongly onto the NAO, the leading pattern of Northern Hemisphere atmospheric variability (Alexander et al. 2004; Mag-
nusdottir et al. 2004; Seierstad and Bader 2009; Honda et al. 2009). At the LGM, the leading pattern of atmospheric variability exhibits an NAO-like dipole structure and appears to be governed by the same physics as its modern counterpart, but its spatial signature is distorted and shifted (Pausata et al. 2009). The sea level pressure response to sea ice retreat in the Nordic Seas (scenario lowiceNS) projects onto this altered leading pattern of variability (compare Figure 2 of Pausata et al. (2009) with Figure 7b). Sea ice retreat in the western North Atlantic (scenario lowiceWAtl) has a weaker sea level pressure response signature (Figure 7a) that bears no similarity to the leading pattern of variability. Despite differences in the mean climate states, these results are consistent with the study of Magnusdottir et al. (2004), who find that atmospheric circulation is more sensitive to sea ice anomalies in the Barents Sea than those in the Labrador Sea in today’s climate. Aloft, both glacial retreat scenarios produce wave-like patterns in geopotential height, reminiscent of Figure 11 in Byrkjedal et al. (2006), suggesting a linear response to midlatitude thermal anomalies on a sphere (Hoskins and Karoly 1981).

In terms of local diabatic changes, the effect of sea ice retreat at Greenland Summit is related to (a) whether the atmosphere feels the influence of a cold ice-covered surface, warm ice-free ocean, or something in between (partially ice-free) at its bottom boundary, and (b) how the atmosphere distributes the (changed) heat flux when the bottom boundary changes. First, much of the western North Atlantic ice cover exists in low concentrations (indicated by the area between the 15% and 50% ice concentration contours in Figure 8), allowing some heat exchange between the ocean and atmosphere even when sea ice is present. Hence, the change in ocean-atmosphere heat flux from the stadial scenario to the retreat scenarios is moderate in the case of western North Atlantic retreat compared to Nordic Seas retreat. Second, temperature advection is largest where the mean near-surface winds have a large component perpendicular to the sea ice edge (i.e., perpendicular to the strongest temperature gradients). These locations are primarily associated with the Nordic Seas rather than the western North Atlantic (Figure 9, top). When sea ice retreats, the convergence of heat due to changes in temperature advection increases over southeastern Greenland (Figure 9, bottom),
and this occurs whether the western North Atlantic ice tongue remains or retreats (Figure 9d, f).

To summarize, the scenario experiments presented here indicate that temperature in the vicinity of the Greenland Summit region is more sensitive to sea ice retreat in the Nordic Seas than in the western North Atlantic.

5. Discussion

It is well known that variability in the large scale atmospheric circulation causes sea ice variability over a wide range of time scales in observational studies (Fang and Wallace 1994). For example, a weak negative feedback of winter sea ice on the NAO has been detected in modelling experiments (Magnusdottir et al. 2004) and observations (Strong et al. 2009), operating on lags of up to several weeks. The question of whether sea ice has a similar restoring negative feedback or rather an amplifying positive feedback on the atmosphere in glacial climates could be crucial to evaluating the role of sea ice in D-O cycles.

As we have already stated, sea ice anomalies in the D-O mechanism proposed by Dokken et al. (2010) are maintained by interactions between different components of the glacial climate system, most importantly the ocean and the marine and terrestrial cryosphere. There are a number of ways in which the atmosphere might reinforce the North Atlantic climate response to stadial or interstadial sea ice conditions. Modelling studies suggest that extratropical thermal forcing produces atmosphere-ocean changes that extend to the tropics. Specifically, high latitude cooling – either due to AMOC shutdown (Vellinga and Wood 2002; Zhang and Delworth 2005; Broccoli et al. 2006), or forced exclusively in the high latitudes by the presence of sea ice or land ice sheets (Chiang et al. 2003; Chiang and Bitz 2005) – has been shown to induce a southward shift of the Atlantic intertropical convergence zone (ITCZ), which in turn can strengthen the winter Hadley cell as well as the (thermally-driven) subtropical component of the jet (Lee and Kim 2003). The result would be a strong, steady and relatively southward-lying jet that transports less heat into the North Atlantic region. Conversely, upon retreat of sea ice, there would be a northward migration of the ITCZ, a weakening of the
Hadley cell, and a shift towards a more poleward-lying, eddy-driven jet. The boost in atmospheric heat transport and mechanical stirring would further break up the North Atlantic ice cover, and thus provide a positive feedback to the ice retreat in the Nordic Seas.

Such processes are beyond the scope of the modelling strategy adopted here, in which atmosphere-ocean feedbacks are not represented. However, the uncoupled atmospheric GCM experiments we have performed indicate that, in glacial climates, sea ice retreat produces a more variable jet while sea ice advance produces a steadier jet. For example, in retreat scenario lowice, there is an increase in the interannual variability of 300-hPa winds of about +10% in DJF despite the fact that the climatological mean wind field is not much affected (Figure 10, top panels). The pattern of jet variability, with maxima straddling the jet axis, indicates more frequent north-south excursions of the jet over the Atlantic sector when it is ice-free. In addition, the leading mode of interannual jet variability from empirical orthogonal function (EOF) analysis of 300-hPa wind explains slightly more of the total variance and has larger amplitude (Figure 10, bottom panels). This result is consistent with the modelling study of Byrkjedal et al. (2006), who find that reducing sea ice in the glacial Nordic Seas leads to increased sea level pressure variability.

6. Concluding remarks

The climate impacts of sea ice anomalies in the glacial North Atlantic have been investigated using coupled climate model simulations of Last Glacial Maximum (LGM) climate as well as a series of sea ice perturbation experiments performed with an atmospheric general circulation model. The main findings of this work are summarized below.

- Coupled climate models simulate a range of sea ice distributions for the LGM climate (Figure 1 and Table 2). Of the six PMIP2 models examined here, all exhibit a perennially ice-covered Arctic. Only CCSM3 and HadCM3M2 simulate the seasonally open Nordic Seas inferred from paleoclimate proxy

- In the CCSM3 simulation of LGM climate, two main masses of sea ice exist: one in the Nordic Seas, where the ice distribution is governed primarily by in situ growth and melt, and one in the western North Atlantic, where the ice distribution owes its existence in large part to the import of ice formed in Baffin Bay and Davis Strait (Figure 2). Other seasonal and spatial aspects of the sea ice cover are altered relative to the present day CCSM3 simulation, such as the appearance of a pronounced wintertime maximum in interannual variability (Figure 3).

- Surface climate signals recorded at Greenland Summit are influenced more strongly by Nordic Seas ice anomalies than by western North Atlantic ice anomalies (Figure 6 and Table 3).

Collectively, our experiments suggest that the abrupt retreat of Nordic Seas ice is an effective way to produce large, abrupt warmings and accumulation increases on Greenland, consistent with those observed during transitions from the stadial to intersadial phase of D-O cycles (although the simulated response is not as strong as the largest transitions in the proxy records). Our results are also likely to be relevant for understanding a different class of near-global abrupt climate changes known as Heinrich events. Heinrich events involve massive iceberg discharges from the North American ice sheet, and produce prominent and systematic climate changes including a cooling of the Northern Hemisphere Atlantic region, a strengthening of winter monsoon winds, and shifts in ocean circulation (see Hemming 2004, and references therein). Heinrich events occur during the stadial phases of certain selected D-O cycles, but these stadials are not distinguishable as more extreme than others in the Greenland ice cores. An explanation for this discrepancy could be that the icebergs associated with Heinrich events primarily affect the North Atlantic ice-rafting zone, a belt of latitudes (40°–55°N) situated well south of the Nordic Seas. If Greenland is indeed most sensitive to the Nordic Seas, it is possible that surface ocean variability in other regions of the North Atlantic could impact global climate despite remaining largely undetected in Greenland ice cores.
Acknowledgments.

We wish to thank T. Dokken, S. Hemming and K. Nisancioglu for insightful discussions; M. Michelsen and J.-Y. Peterschmitt for assistance with model simulations and data; N.G. Kvamstø, B. Tremblay and three anonymous reviewers for helpful comments on the manuscript. We acknowledge the international modelling groups and the Laboratoire des Sciences du Climat et de l’Environnement (LSCE) for providing and archiving the PMIP2 model data. The PMIP2/MOTIF Data Archive is supported by CEA, CNRS, the EU MOTIF project (EVK2-CT-2002-00153) and the Programme National d’Etude de la Dynamique du Climat (PNEDC). This research is part of the Norwegian Research Council ARCTREC project, and was funded by the Comer Fellowship Program (CL, DSB) and National Science Foundation grant ATM-0502204 (DSB, CMB).
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PMIP2 coupled models. The sponsoring institutions of the models are: CCSM3, National Center for Atmospheric Research, USA; CNRM-CM33, Méteo-France/Centre National de Recherches Météorologiques, France; ECBILTCLIO, Royal Netherlands Meteorological Institute (KNMI), Netherlands; HadCM3M2, UK Met Office, UK; IPSL-CM4, Institut Pierre Simon Laplace, France; MIROC3.2, Center for Climate System Research (University of Tokyo), National Institute for Environmental Studies, and Frontier Research Center for Global Change (JAMSTEC), Japan.

PMIP2 coupled model control (pre-industrial) and Last Glacial Maximum (LGM) sea ice conditions for the Northern Hemisphere (NH), North Atlantic (NAtl) and Arctic (defined as poleward of 80°N). The annual mean and seasonal range are shown for each variable. Model descriptions appear in Table 1.

Temperature and net accumulation on Greenland in CAM3 sea ice experiments. Winter season temperature $T_{DJF}$ (°C), annual mean temperature $T_{ANN}$ (°C), and net annual accumulation $acc_{ANN}$ (mm y$^{-1}$) averaged over two model grid points near the core sites of Greenland Summit (mean coordinates 71.2°N, 38.5°W, 2957 m), DYE-3 (mean coordinates 65.6°N, 43.6°W, 2333 m), NorthGRIP (mean coordinates 75.3°N, 42.2°W, 2612 m) and Camp Century (mean coordinates 76.7°N, 61.9°W, 815 m). The columns marked $\Delta$ show the difference between each scenario and the stadial case. The 95% confidence interval is $\leq 0.02$ °C for temperature and $\leq 0.8$ mm y$^{-1}$ for accumulation, making all differences significant at the 0.05 level.
<table>
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<th>Ocean</th>
<th>Sea ice†</th>
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† EVP = elastic-viscous plastic dynamics, FD = free drift dynamics (neglects internal ice force), ITD = ice-thickness distribution, VP = viscous plastic dynamics; number of layers refers to temperature layers in ice/snow; heat reservoir for solar radiation (Semtner 1976) or brine pockets (Bitz and Lipscomb 1999)
Table 2. PMIP2 coupled model control (pre-industrial) and Last Glacial Maximum (LGM) sea ice conditions for the Northern Hemisphere (NH), North Atlantic (NAtl) and Arctic (defined as poleward of 80°N). The annual mean and seasonal range are shown for each variable. Model descriptions appear in Table 1.

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TABLE 3. Temperature and net accumulation on Greenland in CAM3 sea ice experiments. Winter season temperature $T_{\text{DJF}}$ (°C), annual mean temperature $T_{\text{ANN}}$ (°C), and net annual accumulation $\text{acc}_{\text{ANN}}$ (mm y$^{-1}$) averaged over two model grid points near the core sites of Greenland Summit (mean coordinates 71.2°N, 38.5°W, 2957 m), DYE-3 (mean coordinates 65.6°N, 43.6°W, 2333 m), NorthGRIP (mean coordinates 75.3°N, 42.2°W, 2612 m) and Camp Century (mean coordinates 76.7°N, 61.9°W, 815 m). The columns marked $\Delta$ show the difference between each scenario and the stadial case. The 95% confidence interval is $\leq 0.02$ °C for temperature and $\leq 0.8$ mm y$^{-1}$ for accumulation, making all differences significant at the 0.05 level.

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List of Figures

1 Sea ice thickness and extent at LGM simulated by PMIP2 models. (a)–(f) Colour shading shows annual mean sea ice thickness (m). The thick black line shows March ice extent and the thin black line shows September ice extent (both 50% ice concentration). (e) Seasonal cycle of North Atlantic sea. Model descriptions appear in Table 1. (g) Seasonal cycle of Northern Hemisphere sea ice area ($10^{12}$ m$^2$) for each model. For each month, the bars represent models (a)–(f) from left to right, and are colour coded to match the model names in the map panels.

2 Sea ice velocity, growth rate and melt rate in CCSM3 LGM simulation. Top: Speed (cm s$^{-1}$) in colour shading, with ice flow ($\geq$ 2 cm s$^{-1}$) by arrows. Middle: Growth rate (cm d$^{-1}$) including congelation and frazil growth, and snow-ice formation. Bottom: Melt rate (cm d$^{-1}$) including basal, lateral and top melt. In all panels, the black contour shows JFM sea ice extent (15% ice concentration).
3 Seasonal cycle and variability of Northern Hemisphere sea ice area in CCSM3 PD (top) and LGM (bottom) simulations. Plots show the seasonal cycle of simulated Northern Hemisphere ice area, normalized by and expressed as a fraction the annual mean (purple), and of the marginal ice zone (15–50% ice concentration) area \(10^{13} \text{m}^2\); dashed blue). Shading indicates the interdecile range (10th to 90th percentiles; light green) and interquartile range (25th to 75th percentiles; dark green) of ice area, normalized by and expressed as a fraction of the monthly mean. The PD panel also shows observational estimates of Northern Hemisphere ice area (thick grey) and its interdecile range (thin grey) from the National Snow and Ice Data Center’s sea ice index 1978–2008 (detrended), both standardized as described above. Maps show the variance of seasonally averaged sea ice concentration in JFM and JAS (units of sea ice fraction squared). Black contours mark the 15% (thin) and 50% (thick) ice concentrations that define the marginal ice zone.

4 Composites of winters (JFM average) with restricted and extensive sea ice cover in the western North Atlantic (WAtl) and Nordic Seas (NS) in CCSM3 LGM simulation. Maps show averages of the 25 years (i.e., 25% of the total 100 years) with the least (WAtl−, NS−) and most (WAtl+, NS+) JFM sea ice area in each region, expressed as concentration anomalies (%) from the seasonal average. Black contours show the climatological JFM sea ice extent (15% ice concentration). The bar codes provide a schematic indication of the specific years that are included in the 25th (top) and 75th (bottom) percentiles. The vertical axis represents time through the 100-year simulation, with time running from bottom to top. Grey bars on the left side mark years used in the WAtl composites, and grey bars on the right side mark years used in the NS composites; common years appear as long black bars.
Sea ice and sea surface temperature (SST) boundary conditions for CAM3 sea ice experiments. These scenarios are based on the SST and sea ice from the coupled CCSM3 LGM run, but with some additional fabricated alterations to North Atlantic sea ice extent in the winter months as described in the text. (a)–(d) Colours show annual mean SSTs, the thick black line shows March ice extent and the thin black line shows September ice extent (both 50% ice concentration). (e) Seasonal cycle of North Atlantic sea ice area ($10^{12}$ m$^2$) for each scenario. For each month, the bars represent scenarios (a)–(d) from left to right, and are colour coded to match the scenario names in the map panels.

Surface temperature and net accumulation in CAM3 sea ice experiments. Top: 2 m temperature for the stadial (cold) simulation (left) and temperature change in the sea ice retreat scenarios (right panels). Annual mean temperature at Greenland Summit is indicated at the bottom left of each panel. Bottom: Net accumulation over the ice sheets for the stadial simulation (left) and the net accumulation change with sea ice retreat (right panels). In all panels, the thick black contour shows March sea ice extent, the thin black contour shows September sea ice extent (50% ice concentration), and the location of Greenland Summit ($71.16^\circ$N, $38.45^\circ$W, 2957 m) is marked by a dot.

Sea level pressure in CAM3 sea ice experiments. Winter (DJF) sea level pressure (5 hPa contours, black $\leq$ 1030 hPa, grey $\geq$ 1035 hPa) for sea ice retreat scenarios (a) lowiceWAtl and (b) lowiceNS. Colour shading shows the difference compared to sea level pressure in for stadial sea ice conditions.

Surface heat flux in CAM3 sea ice experiments. Annual mean ocean to atmosphere heat flux (W m$^{-2}$) for (a) stadial sea ice conditions and (b) sea ice retreat scenario lowice; (c) stadial – lowice difference. Black contours show the 15% (thin) and 50% (thick) annual mean sea ice concentration.
Temperature advection by surface winds for (left) stadial sea ice conditions, (centre) sea ice retreat scenario lowice and (right) sea ice retreat scenario lowiceNS. (a, c, e) Winter (DJF) advection of 2 m temperature ($T_s$) by surface winds ($\vec{u}_s$) with 30 C day$^{-1}$ reference length shown by the bold horizontal line at the top right of panels. (b, d, e) Convergence of winter (DJF) surface temperature advection field. Contours are drawn at 0.5 C day$^{-1}$ m$^{-1}$ intervals with black denoting convergence and blue denoting divergence; the zero contour has been omitted. In all panels, the thick purple contour shows March sea ice extent and the thin purple contour shows September sea ice extent (50% ice concentration) for the scenario.

Atlantic jet position and variance in CAM3 sea ice experiments. Black contours show winter (DJF) 300-hPa zonal wind (10 ms$^{-1}$ contours starting at 30 ms$^{-1}$). Top: Interannual variance of the wind field, with area-averaged variance over the North Atlantic sector indicated at top left (m$^2$s$^{-2}$). Bottom: Leading pattern of 300-hPa wind variability over the North Atlantic sector (ms$^{-1}$ per std deviation of PC1), with percent of total variance and corresponding area-averaged variance explained by EOF1 indicated at top left. In all panels, the thick purple contour shows March sea ice extent and the thin purple contour shows September sea ice extent (50% ice concentration) for the scenario.
Fig. 1. Sea ice thickness and extent at LGM simulated by PMIP2 models. (a)–(f) Colour shading shows annual mean sea ice thickness (m). The thick black line shows March ice extent and the thin black line shows September ice extent (both 50% ice concentration). (e) Seasonal cycle of North Atlantic sea. Model descriptions appear in Table 1. (g) Seasonal cycle of Northern Hemisphere sea ice area (10^{12} m^2) for each model. For each month, the bars represent models (a)–(f) from left to right, and are colour coded to match the model names in the map panels.
Fig. 2. Sea ice velocity, growth rate and melt rate in CCSM3 LGM simulation. Top: Speed (cm s\(^{-1}\)) in colour shading, with ice flow (≥ 2 cm s\(^{-1}\)) by arrows. Middle: Growth rate (cm d\(^{-1}\)) including congelation and frazil growth, and snow-ice formation. Bottom: Melt rate (cm d\(^{-1}\)) including basal, lateral and top melt. In all panels, the black contour shows JFM sea ice extent (15% ice concentration).
**Fig. 3.** Seasonal cycle and variability of Northern Hemisphere sea ice area in CCSM3 PD (top) and LGM (bottom) simulations. Plots show the seasonal cycle of simulated Northern Hemisphere ice area, normalized by and expressed as a fraction the annual mean (purple), and of the marginal ice zone (15–50% ice concentration) area ($10^{13}$ m$^2$; dashed blue). Shading indicates the interdecile range (10th to 90th percentiles; light green) and interquartile range (25th to 75th percentiles; dark green) of ice area, normalized by and expressed as a fraction of the monthly mean. The PD panel also shows observational estimates of Northern Hemisphere ice area (thick grey) and its interdecile range (thin grey) from the National Snow and Ice Data Center’s sea ice index 1978–2008 (detrended), both standardized as described above. Maps show the variance of seasonally averaged sea ice concentration in JFM and JAS (units of sea ice fraction squared). Black contours mark the 15% (thin) and 50% (thick) ice concentrations that define the marginal ice zone.
Fig. 4. Composites of winters (JFM average) with restricted and extensive sea ice cover in the western North Atlantic (WAtl) and Nordic Seas (NS) in CCSM3 LGM simulation. Maps show averages of the 25 years (i.e., 25% of the total 100 years) with the least (WAtl−, NS−) and most (WAtl+, NS+) JFM sea ice area in each region, expressed as concentration anomalies (%) from the seasonal average. Black contours show the climatological JFM sea ice extent (15% ice concentration). The bar codes provide a schematic indication of the specific years that are included in the 25th (top) and 75th (bottom) percentiles. The vertical axis represents time through the 100-year simulation, with time running from bottom to top. Grey bars on the left side mark years used in the WAtl composites, and grey bars on the right side mark years used in the NS composites; common years appear as long black bars.
Fig. 5. Sea ice and sea surface temperature (SST) boundary conditions for CAM3 sea ice experiments. These scenarios are based on the SST and sea ice from the coupled CCSM3 LGM run, but with some additional fabricated alterations to North Atlantic sea ice extent in the winter months as described in the text. (a)–(d) Colours show annual mean SSTs, the thick black line shows March ice extent and the thin black line shows September ice extent (both 50% ice concentration). (e) Seasonal cycle of North Atlantic sea ice area ($10^{12} \text{ m}^2$) for each scenario. For each month, the bars represent scenarios (a)–(d) from left to right, and are colour coded to match the scenario names in the map panels.
Fig. 6. Surface temperature and net accumulation in CAM3 sea ice experiments. Top: 2 m temperature for the stadial (cold) simulation (left) and temperature change in the sea ice retreat scenarios (right panels). Annual mean temperature at Greenland Summit is indicated at the bottom left of each panel. Bottom: Net accumulation over the ice sheets for the stadial simulation (left) and the net accumulation change with sea ice retreat (right panels). In all panels, the thick black contour shows March sea ice extent, the thin black contour shows September sea ice extent (50% ice concentration), and the location of Greenland Summit (71.16° N, 38.45° W, 2957 m) is marked by a dot.
Fig. 7. Sea level pressure in CAM3 sea ice experiments. Winter (DJF) sea level pressure (5 hPa contours, black $\leq 1030$ hPa, grey $\geq 1035$ hPa) for sea ice retreat scenarios (a) lowiceWAtl and (b) lowiceNS. Colour shading shows the difference compared to sea level pressure in for stadial sea ice conditions.
Fig. 8. Surface heat flux in CAM3 sea ice experiments. Annual mean ocean to atmosphere heat flux (W m$^{-2}$) for (a) stadial sea ice conditions and (b) sea ice retreat scenario lowice; (c) stadial − lowice difference. Black contours show the 15% (thin) and 50% (thick) annual mean sea ice concentration.
Fig. 9. Temperature advection by surface winds for (left) stadial sea ice conditions, (centre) sea ice retreat scenario lowice and (right) sea ice retreat scenario lowiceNS. (a, c, e) Winter (DJF) advection of 2 m temperature ($T_s$) by surface winds ($\vec{u}_s$) with 30 C day$^{-1}$ reference length shown by the bold horizontal line at the top right of panels. (b, d, e) Convergence of winter (DJF) surface temperature advection field. Contours are drawn at 0.5 C day$^{-1}$ m$^{-1}$ intervals with black denoting convergence and blue denoting divergence; the zero contour has been omitted. In all panels, the thick purple contour shows March sea ice extent and the thin purple contour shows September sea ice extent (50% ice concentration) for the scenario.
Fig. 10. Atlantic jet position and variance in CAM3 sea ice experiments. Black contours show winter (DJF) 300-hPa zonal wind (10 ms$^{-1}$ contours starting at 30 ms$^{-1}$). Top: Interannual variance of the wind field, with area-averaged variance over the North Atlantic sector indicated at top left (m$^2$s$^{-2}$). Bottom: Leading pattern of 300-hPa wind variability over the North Atlantic sector (ms$^{-1}$ per std deviation of PC1), with percent of total variance and corresponding area-averaged variance explained by EOF1 indicated at top left. In all panels, the thick purple contour shows March sea ice extent and the thin purple contour shows September sea ice extent (50% ice concentration) for the scenario.