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Warm Nordic Seas delayed glacial inception in Scandinavia

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Warm Nordic Seas delayed glacial inception in Scandinavia

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Abstract

We simulate the last glacial inception, 115 000 years ago, with a three dimensional thermomechanical ice sheet model of the Northern Hemisphere, forced by a comprehensive coupled climate model. High oceanic heat transport into the Nordic Seas prevents large scale ice growth over Scandinavia. Glacial inception in the region starts on the highest mountains in the south when sea surface temperatures in the Nordic Seas are reduced by at least 3 °C. Ice growth in Northern Scandinavia requires a cooling by at least 4 °C. This is in good agreement with marine proxy data from the Nordic Seas and North Atlantic as well as available terrestrial data. This study thus provides a physical understanding and revised timing of the first glacier advance over Scandinavia.

1 Introduction

During the past 800 000 years, climate has been relatively cold with ice sheets covering large parts of the Northern Hemisphere continents most of the time. At the last glacial maximum, 21 000 years before present, an ice sheet existed over North America with a volume equivalent to the present day Antarctic ice sheet, decreasing global sea level by 74 m (Peltier, 2004). Eurasia was covered with a ice volume equal to 2.5 times the Greenland ice sheet, or 17 m sea level equivalent. Additional 25–30 m of sea level equivalent accumulated on the Antarctic ice sheet (Lambeck and Chappell, 2001) for a total sea level decrease of 120 to 130 m relative to present (Waelbroeck et al., 2002).

While extensive work addresses the extent and volume of ice sheets at their maximum, the erosive effect of the ice masses themselves removed most of the sediments holding information about glacial inception. Thus, little is known about advancing glaciers at the end of the last interglacial, the epoch when climate for the last time was similar to present day. Geological data suggests that ice first grew over the Canadian Arctic islands, Labrador, northern continental Canada, continental western Siberia and Eurasian Arctic islands (Andrews and Barry, 1978; Mangerud and Svendsen, 1992;

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Clark et al., 1993; Svendsen et al., 2004; Lambeck et al., 2006).

The last glacial maximum Fennoscandian ice sheet is thought to be the result of ice masses spreading west and southward from mountain nucleation sites in western Siberia, the Eurasian Arctic islands, and probably on Svalbard. However, an independent glaciation over Scandinavia started soon after the first global sea level decrease (Mangerud et al., 1981). This first ice growth was apparently rapid and large because the glacio-isostatic depression in south-western Norway was as fast as the global sea level decrease due to ice growth elsewhere (Mangerud, 1991). The ice front reached the fjord within a few kilometers from an excavation site at Fjøsanger, close to Bergen, allowing the deposition of ice-rafted detritus in marine sediments off the coast (Baumann et al., 1995; Risebrobakken et al., 2005, 2007).

Despite these reconstructions, the history of early Scandinavian glaciation is inconclusive (Mangerud, 2004). Coastal ice limits are known for only four localities in southern Norway, and the earliest limit is based on a single record. Dating is problematic because the time period of interest is beyond the applicability of radiocarbon dating. Even in younger sediments only intervals between glacier advances contain enough organic material to be used for this method. Instead, the chronology is based on correlation with other regions that can be dated by other means. Evidence for early glacier advances and retreats in Southwestern Scandinavia is contrasted by speleothems from karst caves north of the Arctic circle that suggest ice-free conditions until approximately 95 ka (Lauritzen, 1984, 1986; Mangerud, 1991).

The last two decades brought only minor advances from new field observations. Numerical ice sheet modeling could improve the understanding of this period but most studies focus on the primary inception regions over North America and western Siberia (Marshall and Clarke, 1999; Zweck and Huybrechts, 2005; Calov et al., 2005a; Kubatzki et al., 2006; Charbit et al., 2007; Peyaud et al., 2007). This is probably due to the lack of constraints on timing and extent of ice from data, but also because the steep Scandinavian topography and variable climate are a challenge for the realistic representation of glaciers. The interaction of land based ice with floating ice shelves like in the Barents

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Sea remains an issue (Peyaud et al., 2007). Moreover, recent studies reveal a major reorganization of ocean currents in the region during glacial inception (Risebrobakken et al., 2005, 2007; Born et al., 2010a,b), and ocean circulation is thought to have a crucial impact on glaciers in the maritime climate of Scandinavia. The importance of ocean circulation changes has been shown in the study by Calov et al. (2005b), where an artificially weakened Atlantic meridional overturning circulation facilitates ice sheet growth in southern Scandinavia due to lower temperatures, while less ice grows further north as a consequence of reduced precipitation.

Prolonged interglacial warmth in the North Atlantic, despite decreasing high latitude insolation and decreasing sea levels indicating progressing ice growth on land, is reported from a number of marine sediment cores (Ruddiman and McIntyre, 1975; McManus et al., 2002). Further north in the Nordic Seas, recent studies show that the gradual climate cooling due to decreasing insolation after the last interglacial was interrupted by a period of transient warmth (Risebrobakken et al., 2007; Born et al., 2010a,b). As a direct result of weaker summer insolation, Arctic sea ice grows thicker and ice export in the East Greenland Current intensifies. The freshwater transport by sea ice into the subpolar North Atlantic reduces deep convection in the northwest Atlantic and thereby effectively changes the density structure of the region. This causes a nonlinear response of the subpolar gyre that amplifies freshening by sea ice. The weakening of the subpolar gyre results in a higher fraction of subtropical waters in the Atlantic inflow into the Nordic Seas and thus warming along the path of the Norwegian Atlantic Current.

As a result, ocean heat transport into the Nordic Seas increased between 119 ka and 115 ka. The anomalous warmth ends with rising insolation and an estimated temperature decrease of about 3 °C as inferred from planktic foraminiferal data. Ice-rafted detritus indicates that ice growth over Scandinavia started at about the same time. However, dating of terrestrial sediments involves uncertainties of several millennia and the better-dated marine sediment records are ambiguous about the location of glacier advances. Furthermore, neither of these records provide information on whether the

presence of a warm ocean favored glaciation due to enhanced moisture supply or if the relatively warm climate caused a delay in glacial inception.

In this study we suggest that a well documented transient strengthening of the Norwegian Atlantic Current delayed large scale glacier advance over Scandinavia until after 115 000 years ago (115 ka), followed by a rapid accumulation of ice. The climate and ice sheet models used are described in Sect. 2. Section 3 presents a simulation of land ice at 115 ka and investigates the impact of ocean heat transport variability on ice growth. Section 4 quantifies the reduction of sea surface temperature necessary to accumulate ice over Scandinavia. We discuss and conclude in Sect. 5 and 6.

2 Model description and experiments

The numerical ice sheet model used in this study is SICOPOLIS (Simulation COde for POLythermal Ice Sheets), a three dimensional thermomechanical ice sheet model (Greve, 1997). It simulates ice rheology and mass balance on a regular grid, using a stereographic projection centered on the North Pole (Fig. 1). The horizontal resolution is set to 40 km and there are 90 vertical layers: 80 equidistant layers in cold ice and 10 representing a layer of ice at the pressure melting point. The model employs the shallow ice approximation, neglecting longitudinal stress gradients. As a consequence of this simplification, floating ice shelves and ice streams cannot be accurately simulated. This introduces only a negligible error for the simulation of the Northern Hemisphere at a time of relatively low ice volume.

Mass balance at the upper boundary of the model is computed on the basis of monthly surface temperature and total precipitation. The snow fraction of monthly precipitation is estimated as a linear function between -10 and 7°C , with all precipitation falling as snow below that range and all as rain above. Only solid precipitation contributes to accumulation. Rain runs off immediately. Calculation of melt is based on a positive degree day model (Reeh, 1991) with parameters $\beta_s = 3 \text{ mm K}^{-1}$ for snow and $\beta_s = 12 \text{ mm K}^{-1}$ for ice. A constant geothermal heat flux of 55 mW/m^2 is assumed at

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the lower boundary.

Climate data is provided by the Institut Pierre Simon Laplace coupled model version 4 (IPSL CM4). This comprehensive climate model comprises ocean, sea ice, atmosphere and land surface components (Marti et al., 2010). It has previously been used to simulate the last glacial inception and was validated against the available proxy data to simulate the transient warmth in the Nordic Seas at 115 ka (Braconnot et al., 2008; Born et al., 2010a,b). In addition to the coupled model experiments, we carried out a series of sensitivity experiments using only the atmosphere component of the model (LMDZ 3.3) with modified sea surface conditions taken from the coupled model equilibrium. In order to investigate the effect of weaker oceanic heat transport, sea surface temperatures in the Arctic Ocean, Nordic Seas and North Atlantic were decreased by fixed values of 1, 2, 3 and 4 °C. Sea ice area was adjusted accordingly, increasing coverage linearly from 0% at 0 °C water temperature to 100% at -1.8 °C. The simulations were ran into quasi-equilibrium for 20 years.

3 Glacial inception over Scandinavia and ocean heat transport

Forcing the ice sheet model with fixed 115 ka boundary conditions from the coupled climate model yields additional ice growth compared to a preindustrial control experiment (Fig. 1). Ice grows on northern continental Canada, Labrador, the Canadian Archipelago and western Siberia. Ice caps on Iceland and Baffin Island are larger than in the preindustrial control. However, the Scandinavian peninsula remains ice-free. Present-day ice caps in Scandinavia are below model resolution and thus not simulated.

Mass balance variability of extra-tropical glaciers is usually controlled by changes in ablation due to anomalous summer temperatures. Ice melting is a highly non linear function of temperature and easily exceeds accumulation once a threshold in temperature is reached. Typical variability in accumulation is smaller in all but the most marine climates, as has been shown for glaciers in southern Norway (Nesje and Dahl, 2000).

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In order to investigate further the robustness of an ice-free Scandinavia in a relatively cold 115 ka climate, the ice sheet model is forced by a composite climate based on the 40 years (10%) with coldest summers (April to September) over Scandinavia of the last 400 years of the coupled model experiment. Mass balance in the mountains of Scandinavia is close to positive under these conditions (Fig. 2). However, this is not sufficient to start ice accumulation.

Although the temperature criterion was defined over land (see red box in Fig. 3), the temperature difference between the 40 years of warmest and coldest summers extends over the ocean tracing the sea ice edge. This suggests that sea ice and ocean heat transport into the Nordic Seas plays an important role for decadal variability over Scandinavia at 115 ka. Indeed, the correlation of air temperature over Scandinavia and ocean heat transport across the Greenland Scotland ridge is high with a coefficient $r = 0.81$ for ocean heat transport leading by 1 year (Fig. 4). The amplitude of surface air temperature variability over Scandinavia is about 1 K.

4 How much sea surface cooling is needed to start Scandinavian glaciation?

Despite low insolation levels at 115 ka and ice growth at similar latitudes over North America and Siberia, climate is too warm over Scandinavia to support ice growth (Fig. 1). This result is robust and no ice grows even in years with cold summers. Ocean heat transport, probably through its impact on the sea ice edge, plays an important role for surface air temperature and land ice growth over Scandinavia (Fig. 4).

A series of experiments with artificially lowered sea surface temperatures in the North Atlantic, Nordic Seas and Arctic Ocean was carried out. Sea ice area was extended according to the modified temperatures. At a reduction of 3 °C relative to the 115 ka coupled model experiment, a permanent ice cap develops over southern Norway, approximately at the location of present-day Jostedalsglacier (Fig. 5). For sea surface temperatures 4 °C below the coupled model average for 115 ka, ice also grows over northern Norway. Ice caps grow also outside Scandinavia, most notably on

the islands in the Barents and Kara Seas and in western and far-eastern continental Siberia.

These results show that sea surface temperatures have strong impact on the mass balance in Scandinavia. However, the fact that the mass balance is positive when sea surface temperatures are low is not trivial. In addition to more favorable conditions due to generally colder climate and weaker ablation, a cold sea surface also reduces moisture supply and accumulation. In the experiments presented here, the ablation effect dominates the mass balance change in most regions, especially over Southern Scandinavia and the Northern British Isles, where synoptic weather systems carry heat taken up from the Atlantic Ocean (Fig. 6). However, despite large changes, the mass balance remains negative in most regions.

The effect of the colder sea surface and larger sea ice area on accumulation is limited to coastal regions, most notably southeast Greenland and to some extent western Scandinavia (Fig. 6, lower). Changes in accumulation are about an order of magnitude weaker than changes in ablation and dominate exclusively in already glaciated areas of the unperturbed experiment (see Fig. 1). In these regions, summer temperatures are cold enough such that less snow melts in summer than accumulates in winter. Consequently, lower accumulation impacts the annual mass balance. In regions where summer melting exceeds accumulation, a small change in accumulation is unlikely to result in a positive mass balance because melting rates are usually much larger than accumulation.

5 Discussion

Combining results from a comprehensive coupled climate model and a three-dimensional land ice model, ice growth at the end of the last interglacial was investigated. Simulated ice nucleation regions at 115 ka include the Canadian Arctic, Labrador and Siberian islands, in good agreement with reconstructions (Clark et al., 1993; Svendsen et al., 2004). In particular, glacier advances from Siberian islands

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were found to be important for the build-up of the Kara Sea ice sheet (Möller et al., 2007).

No ice is found over Scandinavia with 115 ka climate forcing. Even when forcing the ice model only with years of particularly cold summers, ice did not accumulate. However, ice accumulation was found sensitive to ocean heat transport into the Nordic Seas (Fig. 4). This implies a relation between sea surface temperature and Scandinavian mass balance that can be used to help constrain the timing of the last glacial inception with high-resolution marine proxy data. Available terrestrial data for Scandinavia has much lower temporal resolution. Sea surface temperatures of the Nordic Seas need to cool by at least 3 °C from the 115 ka average for inception over southern Scandinavia. A 4 °C cooling induces ice growth over northern Scandinavia. Cooling also has a positive effect on other known nucleation sites like western Siberia, the Siberian islands and Beringia (Brigham-Grette, 2001; Brigham-Grette et al., 2001; Svendsen et al., 2004).

As shown by proxy data and models, strong ocean heat transport into the Nordic Seas between 119 ka and 115 ka was important for climate evolution of the last glacial inception (Risebrobakken et al., 2007; Born et al., 2010b). The climate model used in this study was validated against the available proxy data to simulate this feature (Born et al., 2010a). Layers of ice-rafted detritus, particles too big to be transported other than by icebergs detached from ice sheets, are found in marine sediments of the Nordic Seas following a period of relatively warm conditions and immediately after a 3 °C drop in sea surface temperatures off the central Norwegian coast at 115 ka (Baumann et al., 1995; Risebrobakken et al., 2005, 2007), in good agreement with simulations presented here. This is also supported by terrestrial data from southwestern Scandinavia (Mangerud et al., 1981; Mangerud, 1991, 2004). However, the ice sheet model suggests that a larger temperature decrease is necessary to induce ice growth in northern Scandinavia. This is consistent with reconstructions from speleothems in this region that suggest significant ice growth only after 95 ka (Lauritzen, 1984, 1986).

Numerous studies of the last glacial cycle show that the large ice sheets of the glacial maximum nucleated over North America and western Siberia (Marshall and

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Clarke, 1999; Siegert et al., 2001; Zweck and Huybrechts, 2005; Calov et al., 2005a; Kubatzki et al., 2006; Charbit et al., 2007; Peyaud et al., 2007). Glacier advance over Scandinavia early during the last glacial cycle is regarded as of secondary importance to the continental-scale glaciation. However, the mechanism presented here, a rapid glacier advance over Scandinavia at 115 ka, is consistent with this previous work. Decreasing heat transport into the Nordic Seas at 115 ka might also influence the primary nucleation sites around the Barents and Kara Seas and thereby the evolution of the Eurasian ice sheet later during the glacial.

While the climate model simulations have been discussed extensively in earlier publications (Braconnot et al., 2008; Born et al., 2010a,b), several aspects of the ice sheet model need clarification here. Relatively large glaciers are present in Scandinavia today, but are not shown in the control experiment (Fig. 1). This is because the model resolution of 40 km is too coarse to resolve these glaciers. Resolution can become critical for the representation of ice masses in areas of steep topography, because glacier nucleation often occurs on the highest peaks and propagates to lower elevations. It has been proposed to parameterize this effect by calculating the mass balance at the highest present day elevation within each grid cell if it is higher than the simulated ice sheet (Zweck and Huybrechts, 2005). While this approach preserves information that would otherwise be lost due to averaging the topography over the grid size, it neglects the effect of steep topographic gradients of sub-grid scale. Steep slopes typically accompany the occurrence of high peaks and limit the spatial extent of glaciers originating from high elevations. Ice is quickly transported to lower elevations and thus effectively removed. Accounting for the full sub-grid height distribution, including the presence of deep valleys and fjords, requires a more sophisticated approach and a parametrization of alpine glaciers (Marshall and Clarke, 1999). This, however, is beyond the scope of the present study. We acknowledge that the finite model resolution might impact the findings presented here quantitatively but not qualitatively.

The shallow ice approximation employed in our model inhibits the simulation of floating ice shelves that are believed to have played a significant role for the evolution of

the Eurasian ice sheet as ice grew across the Barents and Kara Seas. However, the first glaciation of Scandinavia was the result of local accumulation of ice and the results presented here are thus not affected by the limitation to grounded ice. The offline coupling scheme used does not communicate elevation and albedo changes back to the climate model and thus neglects potentially important feedback mechanisms. However, simulated elevation changes are small due to limited ice accumulation. Albedo changes are minor because accumulation regions of the ice sheet model are already covered with snow in the climate model.

6 Conclusions

Our modeling results are in good agreement with proxy data and the main findings are:

- (a) Strong ocean heat transport into the Nordic Seas between 119 ka and 115 ka delayed large scale ice growth over Scandinavia.
- (b) Scandinavian glacial inception requires a 3°C drop in sea surface temperatures, in good agreement with marine and terrestrial proxy data.
- (c) The positive effect of strong oceanic heat transport on ablation dominates the positive effect of enhanced evaporation from the warm ocean surface.
- (d) At inception, ice first started growing in the high mountains of Southern Scandinavia.

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Warm Nordic Seas delayed glacial inception in Scandinavia

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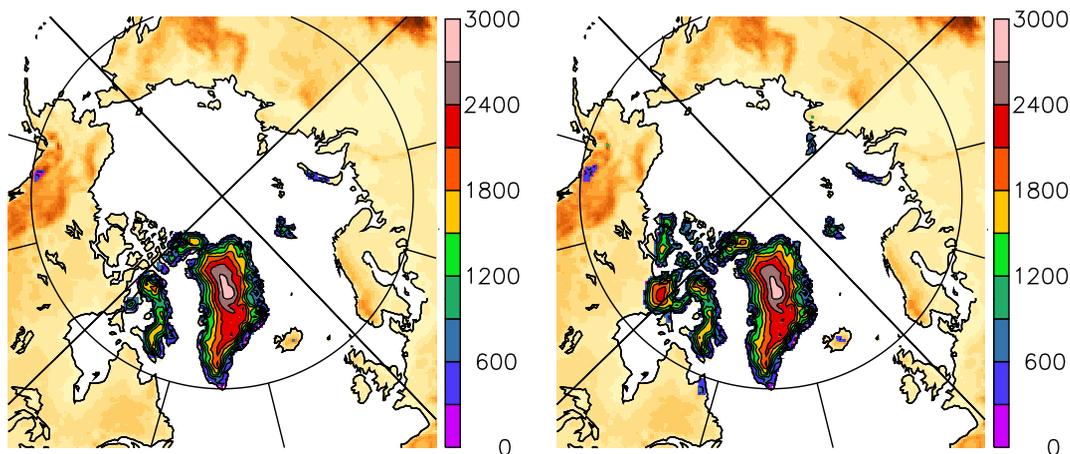


Fig. 1. Modelled ice thickness (in m) for present day (left) and 115 ka (right). Preindustrial ice area and thickness are well reproduced. Besides the Greenland ice sheet, smaller ice caps are visible in the Canadian Arctic, Spitsbergen, Iceland, Novaya Zemlya, and south eastern Alaska. Most important differences with 115 ka forcing is additional ice growth in northern continental Canada, Labrador, western Siberia and the Eurasian Arctic islands.

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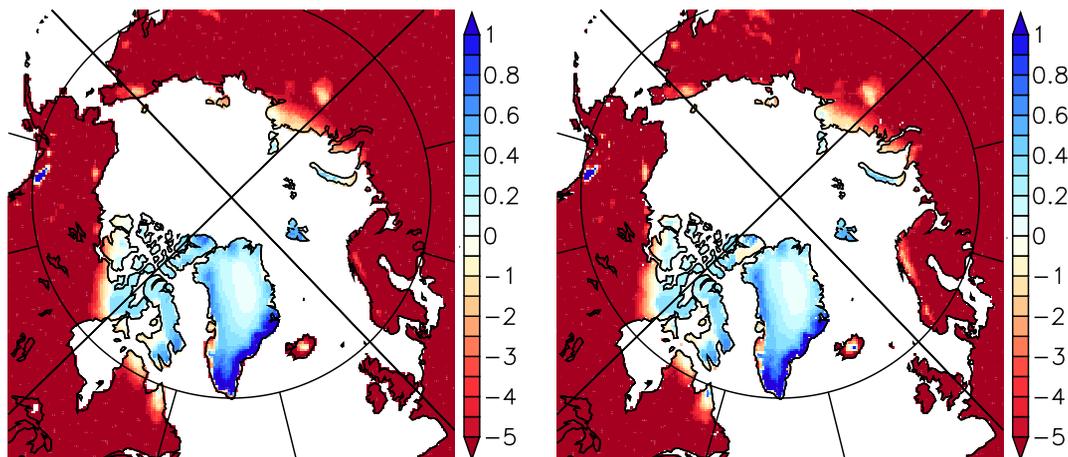


Fig. 2. Left: Mass balance of the 115 ka control experiment, forced by a 100 year average of the climate model. Right: As above, but forced by an average of the 40 years of coldest summers over Scandinavia in the same model experiment (see Fig. 3).

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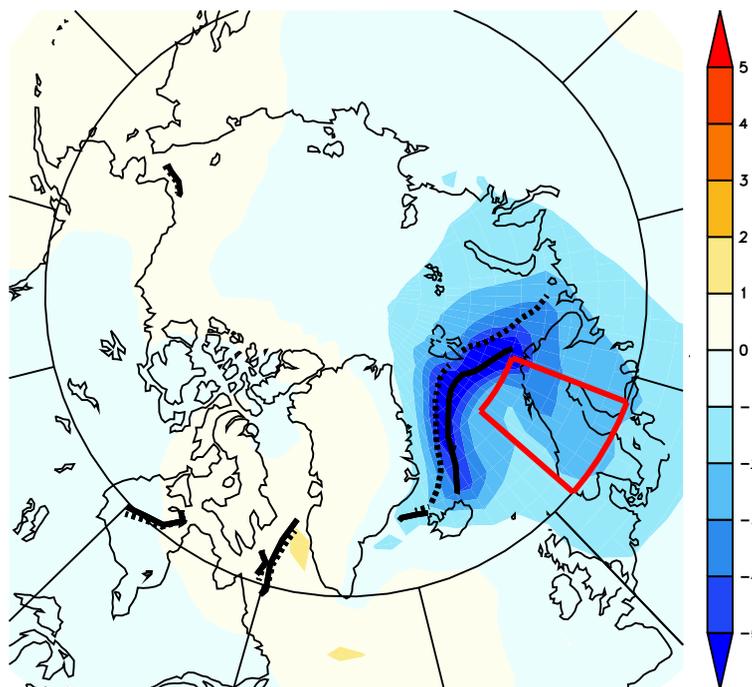


Fig. 3. Difference in annual average surface air temperature, 40 years of coldest minus 40 years of warmest summers, in °C (colors). Contours show the annual average 50% sea ice edge for the years of coldest (solid) and warmest (dotted) summers. Red box shows region where summer surface air temperature over land is taken as selection criterion.

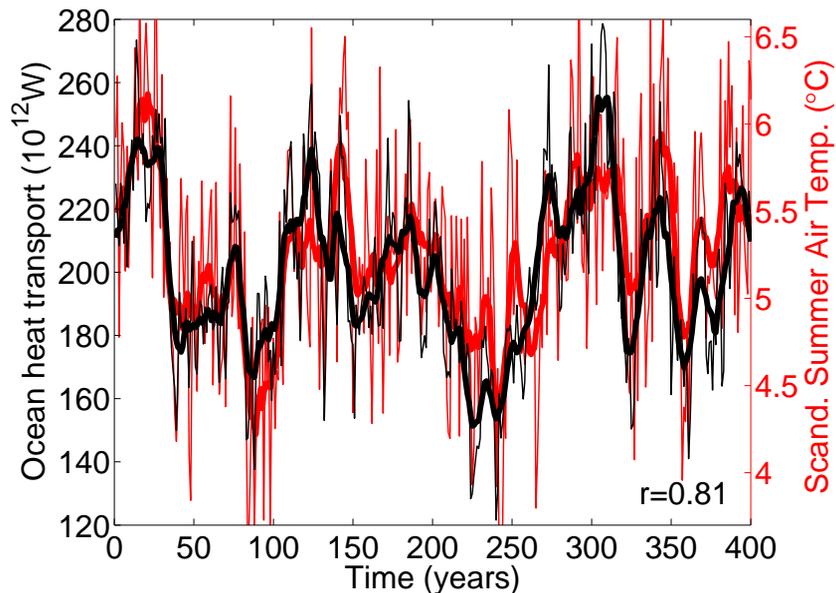


Fig. 4. Time series of ocean heat transport into the Nordic Seas (black) and Scandinavian summer surface air temperature (red, see red box in Fig. 3), both filtered with a 11 years running average. Thin lines show unfiltered time series. Curves correlate with $r = 0.81$ and ocean heat transport leads by one year.

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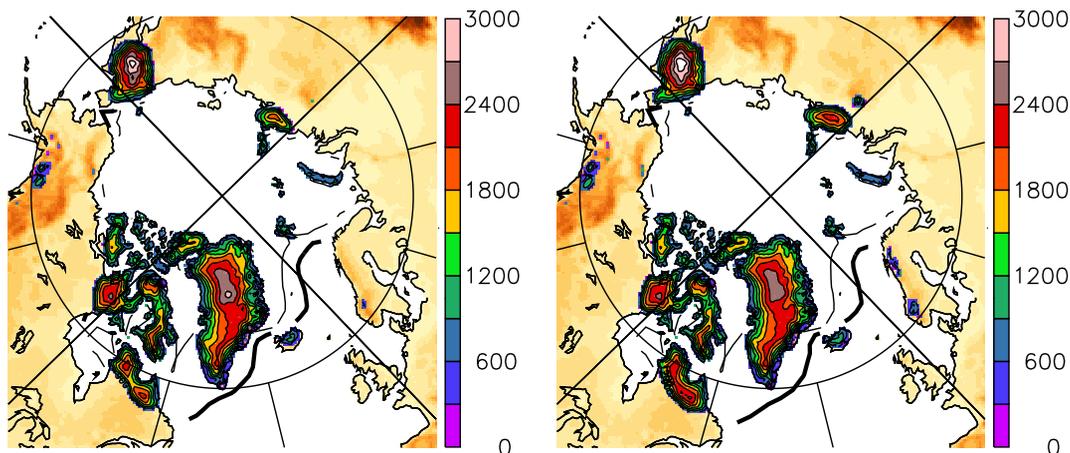


Fig. 5. Ice thickness for experiments with sea surface temperatures reduced by 3°C (left) and 4°C (right) from the coupled model 115 ka average, in m. The 50% annual average sea ice extent is shown for the respective experiment (thick contour) and the unperturbed 115 ka experiment (thin contour). With 3°C colder sea surface, ice starts to grow in the region of present-day Jostedalbreen in southern Norway. Cooling the sea surface by 4 K allows additional ice growth over northern Scandinavia.

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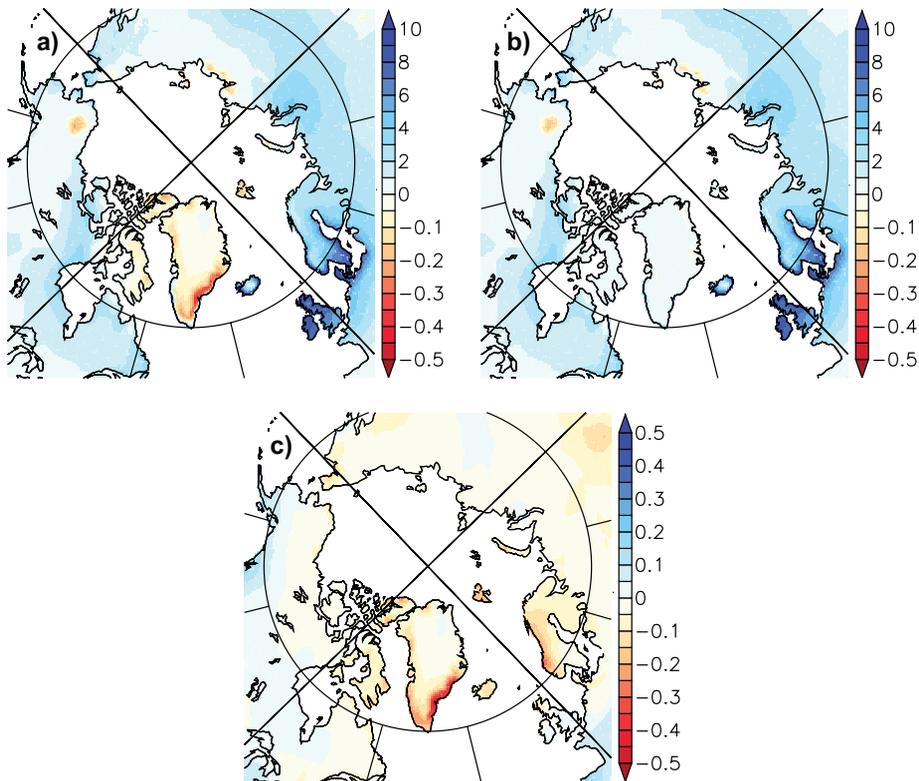


Fig. 6. Mass balance difference (m/yr) between atmosphere only experiment with 3°C lower sea surface temperatures and the control run **(a)**, decomposed into melt **(b)** and precipitation differences **(c)**. Lower sea surface temperatures strongly facilitate ice growth in most regions by decreasing melting. The negative effect of weaker moisture supply is limited and dominates over melt only in areas that are already glaciated in the unperturbed 115 ka experiment. Note the different scale in the lower panel.

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