



"Impact of ice sheet meltwater fluxes on the climate evolution at the onset of the Last Interglacial"

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Abstract

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Impact of ice sheet meltwater fluxes on the climate evolution at the onset of the Last Interglacial

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Abstract

Large climate perturbations occurred during Termination II when the ice sheets retreated from their glacial configuration. Here we investigate the impact of ice sheet changes and associated freshwater fluxes on the climate evolution at the onset of the Last Interglacial. The period from 135 to 120 kyr BP is simulated with the Earth system model of intermediate complexity LOVECLIM v.1.3 with prescribed evolution of the Antarctic ice sheet, the Greenland ice sheet and the other Northern Hemisphere ice sheets. Variations in meltwater fluxes from the Northern Hemisphere ice sheets lead to North Atlantic temperature changes and modifications of the strength of the Atlantic meridional overturning circulation. By means of the interhemispheric see-saw effect, variations in the Atlantic meridional overturning circulation also give rise to temperature changes in the Southern Hemisphere, which are modulated by the direct impact of Antarctic meltwater fluxes into the Southern Ocean. Freshwater fluxes from the melting Antarctic ice sheet lead to a millennial time scale oceanic cold event in the Southern Ocean with expanded sea ice as evidenced in some ocean sediment cores, which may be used to constrain the timing of ice sheet retreat.

1 Introduction

Understanding the climate and ice sheet evolution during past warm periods in the history of the Earth may provide important insights for projections of future climate and sea-level changes. The growing amount of paleo-reconstructions for the Last Interglacial period (e.g. Govin et al., 2012; Capron et al., 2014) in combination with improved model simulations of this most recent warm period (e.g. Bakker et al., 2013; Lunt et al., 2013; Langebroek and Nisancioglu, 2013; Loutre et al., 2014) make it an interesting target for studying the coupled climate-ice sheet system.

According to reconstructions, the Last Interglacial (LIG, from ~ 130 –115 kyr BP) was characterised by global annual mean surface temperature of up to 2°C above the pre-

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5 industrial (e.g. Turney and Jones, 2010; Capron et al., 2014) and a sea-level high stand of 6–9 m above the present day (Kopp et al., 2009; Dutton and Lambeck, 2012). As the penultimate glacial maximum was at least as severe as the Last Glacial Maximum in both hemispheres (EPICA community members, 2004; Svendsen et al., 2004), this
10 implies a large amplitude glacial-interglacial transition in terms of temperature and ice sheet configuration. At the onset of the LIG, a rapid warming of $\sim 10^\circ\text{C}$ from the preceding cold state is recorded in deep Antarctic ice cores (Masson-Delmotte et al., 2011) to have occurred between ~ 135 and 130 kyr BP. Current ice core records from the Greenland ice sheet (GrIS) do not extend long enough back in time to cover the entire penultimate deglaciation and associated warming (NEEM community members, 2013), but a similar timing and magnitude of warming compared to the Antarctic can be reconstructed for sea surface temperatures off the West European margin (Sánchez Goñi et al., 2012). The warming is closely related with an ice sheet retreat in both hemispheres. Despite large uncertainties in reconstructions, the global sea-level stand at 135 kyr BP
15 of as low as -80 m (Grant et al., 2012) is indicative of the large amount of freshwater that entered the ocean in form of meltwater from the retreating ice sheets over a relatively short period. Aside from determining the amplitude of sea-level changes, which is the focus of many studies (e.g., Robinson et al., 2011; Stone et al., 2013), the associated climate impacts and possible feedbacks on the ice sheet evolution of this
20 freshwater forcing are an important element for a process understanding of the coupled climate-ice sheet changes at that time.

25 A climatic mechanism that is thought to be directly related to changes in the Northern Hemisphere (NH) ice sheet freshwater fluxes (FWF) is the interhemispheric see-saw effect (Stocker, 1998) that links Southern Hemisphere (SH) warming to a weakening of the Atlantic meridional overturning circulation (AMOC). If active during the onset of the LIG, NH ice sheet melting during Termination II would have been the cause for a substantial AMOC weakening and NH cooling, while reduced interhemispheric heat transport caused a gradual SH warming (Stocker and Johnson, 2003). The see-saw mechanism was evoked to explain part of the peak Antarctic warming during the

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LIG (e.g. Holden et al., 2010), even though some Southern Ocean (SO) warming was shown to be possible with orbital forcing alone (without NH freshwater forcing), and has been speculated to have caused increased Antarctic ice shelf melting and West Antarctic ice sheet (WAIS) retreat (Duplessy et al., 2007). The retreat of the WAIS, which is believed to have been grounded at the edge of the continental shelf during the penultimate glaciation, generated a large anomalous flux of freshwater into the SO. Such freshwater forcing could have had a substantial influence on the SO configuration in terms of sea ice extent and ocean circulation as shown in model experiments for the last deglaciation (Menviel et al., 2011), for future global warming scenarios (Swingedouw et al., 2008) and for the present day (Bintanja et al., 2013). The impact of increased Antarctic FWF is thought to consist of a surface ocean freshening, stratification of the surface ocean and cooling, which promotes sea ice growth (e.g. Bintanja et al., 2013) and reduced Antarctic Bottom Water (AABW) formation (Menviel et al., 2011). Recently, Golledge et al. (2014) suggested that such mechanism might also have provided a feedback on Antarctic ice sheet (AIS) retreat for meltwater pulse 1A during the last termination, by promoting warming of the mid-depth ocean waters that provide additional heat for melting ice shelves.

In the present work, we study the effect of evolving ice sheet boundary conditions on the climate, by simulating the climate evolution at the onset and over the course of the LIG with an Earth system model of intermediate complexity (EMIC). The model is forced with realistic ice sheet boundary conditions from offline simulations of high-resolution models of the Antarctic and Greenland ice sheets and reconstructions of the remaining NH ice sheets. The model and experimental setup are described in Sects. 3 and 4, respectively, followed by results (Sects. 5, 6 and 7), their discussion in Sect. 8 and conclusions (Sect. 9).

2 Model description

We use the EMIC LOVECLIM version 1.3, which includes components representing the atmosphere, the ocean and sea ice, the terrestrial biosphere and the ice sheets, cf. Fig. 1. The model has been utilised in a large number of coupled climate-ice sheet studies (e.g. Driesschaert et al., 2007; Swingedouw et al., 2008; Goelzer et al., 2011, 2012a; Loutre et al., 2014) and is described in detail in Goosse et al. (2010). In this study, the climate components are forced by time-evolving ice sheet boundary conditions that are updated every full year. The ice sheet models provide changes in topography, ice sheet extent (albedo) and spatially and temporally variable FWF.

2.1 Northern Hemisphere ice sheet forcing

We have little geomorphological evidence for NH ice sheet evolution during Termination II since it was mostly destroyed by the re-advance leading to the Last Glacial Maximum (LGM). Therefore, a reconstruction of NH ice sheet evolution (except for the Greenland ice sheet) is made by remapping the post-LGM retreat according to the same benthic $\delta^{18}\text{O}$ value (Lisiecki and Raymo, 2005), assumed as an indicator of the global ice volume. The post-LGM retreat is reconstructed from a large number of geomorphological constraints on ice sheet extent that have been digitized from the literature (Dyke and Prest, 1987; Dyke et al., 2002; Clague and James, 2001; Mayewski et al., 1981; Andersen, 1981; Landvik et al., 1998; Mangerud et al., 2002; Svendsen et al., 1999) and ice sheet surface elevation has been reconstructed by assuming a plastic flow relation for the ice sheets. A LGM sea-level contribution of the NH ice sheets relative to the present day of -110 m is assumed and translates into a similar magnitude for the penultimate glacial maximum (Lisiecki and Raymo, 2005). Due to the assumed analogy, different configurations of the Northern Hemisphere ice sheets between last and penultimate glaciation (e.g. Obrochta et al., 2014) are not represented in these reconstructions. The resulting boundary conditions consist of a chronology of ice mask and surface elevation changes over the entire LIG period (Fig. 2).

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NH freshwater fluxes were estimated from the same method by using derived volume changes as input to a continental runoff routing model (Goelzer et al., 2012b) to identify the magnitude and location of meltwater fluxes to the ocean (Fig. 3a), as already described in some detail in Loutre et al. (2014). Support for the derived chronology of NH ice sheet evolution and their FWF can be found in records of ice-rafted detritus (IRD) from the subpolar North Atlantic that show similar variability during the deglaciation and in particular a last IRD peak ~ 128 kyr BP preceding low IRD levels throughout MIS 5e (Kandiano et al., 2004; Oppo et al., 2006).

2.2 Simulations of the Greenland and Antarctic ice sheets

For the present study, the climate components are partially forced by results from stand-alone simulations of the Greenland and Antarctic ice sheets. The configuration of both ice sheet models and the forcing interface follows the description in Goose et al. (2010) with the following exceptions. Forcing for the ice sheet models is derived from scaling present-day observations of temperature and precipitation with indices based on ice core records, as is standard for long-term paleo ice sheet modelling (e.g. Huybrechts, 1990; Letreguilly et al., 1991; Zweck and Huybrechts, 2005; Greve et al., 2011; Stone et al., 2013). For the GrIS the forcing record was created following Fürst et al. (2015) by combining a synthesised Greenland $\delta^{18}\text{O}$ record derived from Antarctica Dome C using a bipolar seesaw model (Barker et al., 2011) with the NEEM temperature reconstruction (NEEM community members, 2013) between 128.44 and 120 kyr BP. The Barker $\delta^{18}\text{O}$ record is converted to temperature with a constant factor $T = 1.5 \cdot (\delta^{18}\text{O} + 34.83)$. Positive temperature anomalies of the NEEM record are scaled by a factor 0.6 to fulfil constraints on maximal ice sheet retreat from Camp Century and Dye3 ice core locations that are assumed to have been ice covered during the LIG.

The AIS forcing is derived directly from the Antarctica Dome C record (EPICA community members, 2004). Furthermore, both ice sheet models are forced by changes in global sea-level stand based (for the GrIS) on the benthic deep-sea record of Lisiecki and Raymo (2005) and (for the AIS, where the sea-level changes are the dominant

forcing) on a more recent sea-level reconstruction using Red Sea data (Grant et al., 2012). The chronology of the latter may be expected to be more accurate since, unlike for the first, ice volume is independent of deep-sea temperatures in its reconstruction. Finally, the Antarctic ice sheet model is run at a horizontal resolution of 20×20 km instead of $10 \text{ km} \times 10 \text{ km}$ (as in the standard configuration and for the Greenland ice sheet model) due to computational constraints for the relatively long duration of the LIG simulation.

To embed the dynamic GrIS simulation in the other NH boundary conditions, the geometric evolution of the Greenland ice sheet masks out prescribed changes where ice is present. In that case, the prescribed ice sheet evolution and associated FWF are not limited by the present-day configuration of the Greenland ice sheet as in Loutre et al. (2014). With activated GrIS and AIS evolution, their dynamically calculated FWF (Fig. 3b) replace the background freshwater flux from runoff over land calculated by the land model. The ice sheet evolution is illustrated in Fig. 2 for the modelled Greenland ice sheet embedded in the NH reconstruction (top) and the Antarctic ice sheet (bottom).

In our setup, the combined NH sea-level contributions (including Greenland) and Antarctic sea-level contribution fall within the 67% confidence interval of probabilistic sea-level reconstructions (Kopp et al., 2009) for the first peak in sea-level contributions and the following period ($\sim 124\text{--}120$ kyr BP). In both cases, the final 20 m rise in hemispheric sea-level contributions before is steeper and occurs $1\text{--}2$ kyr earlier compared to the reconstructions, which is consequently also the case for the rise in global sea-level at the onset of the LIG. When assuming an additional peak contribution of glaciers (0.42 ± 0.11) and thermal expansion of the ocean (0.4 ± 0.3) as given by Masson-Delmotte et al. (2013), the assumed ice sheet evolution in our setup reproduces well the average sea-level contribution between 125 and 120 kyr BP from the best estimate of Kopp et al. (2009), but does not represent the multi-peak structure of global sea-level contribution during the LIG as suggested by Kopp et al. (2009, 2013).

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2.3 Initialisation

The goal of our initialisation technique is to prepare a climate model state for the transient simulations starting at 135 kyr BP exhibiting a minimal coupling drift. Both the Greenland and Antarctic ice sheet models are integrated over the preceding glacial cycles and the entire LIG in stand-alone mode. The climate model is then initialized to a steady state with ice sheet boundary conditions, greenhouse gas forcing and orbital parameters for the time of coupling (135 kyr BP). In this way, when LOVECLIM is integrated forward in time for transient experiments, the climate component is already relaxed to the ice sheet boundary conditions and exhibits a minimal model drift in unforced control experiments (not shown).

3 Experimental setup

All simulations are forced by time-dependent changes in greenhouse gas (GHG) concentrations and insolation running from 135 until 120 kyr BP (Fig. 4). The radiative forcing associated with the reconstructed GHG levels (Petit et al., 1999; Pépin et al., 2001; Raynaud et al., 2005; Loulergue et al., 2008; Spahni et al., 2005) is below preindustrial values for most of this period and barely exceeds it at ~ 128 kyr BP. The changes in the distribution of insolation received by the Earth are dynamically computed from the changes in the orbital configuration (Berger, 1978) and represent the governing NH forcing during peak LIG conditions aside from evolving ice sheet boundary conditions. In the following, we will compare results of the reference experiment with all ice sheet boundary conditions evolving in time (Reference) to experiments in which the ice sheet boundary conditions are partially fixed to the pre-industrial configuration (Table 1). To disentangle the effects of the individual ice sheets, the experiments noGfwf (suppressed GrIS freshwater fluxes) and noAGfwf (suppressed FWF from both AIS and GrIS) are complemented by two predecessor experiments with fixed AIS and GrIS and evolving NH boundary conditions (noAG), as well as a climate experiment forced by

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insolation and GHG changes only with all ice sheet boundary conditions fixed (noIS). The latter two experiments correspond to the allLR and IOnly experiments from Loutre et al. (2014).

4 Temperature evolution at the onset of the LIG

As shown by Loutre et al. (2014), including the forcing from the NH ice sheets in terms of configuration and FWF is crucial to simulate the onset of the LIG temperature increase and amplitude variations with LOVECLIM v.1.3 more in line with proxy records. This helps to partially overcome problems of EMICs (and general circulation models) to simulate the strong temperature contrasts inferred from proxy reconstructions (Bakker et al., 2013; Lunt et al., 2013). The increased amplitude of temperature changes in simulations including NH ice sheet boundary conditions is both due to albedo and elevation changes and to a larger extent due to the implied freshwater forcing from the NH ice sheets (Loutre et al., 2014). Therefore, these experiments are complemented in the present work by model runs that additionally include changes in ice sheet configuration and FWF from the Greenland and Antarctic ice sheets.

The temperature evolution (Fig. 5) before 127 kyr BP is in both hemispheres strongly influenced by the ice sheet boundary conditions and in particular by the freshwater forcing from the ice sheets. The experiments including FWF from the NH ice sheets (noAG and Reference) clearly show temperature variations on the multi-millennial time scale in both hemispheres following variations in ice sheet freshwater input (cf. Fig. 3). Differences in the temperature evolution between noAG and Reference are small in the NH, where the additional freshwater flux from Greenland is small compared to the other sources. In the SH, by contrast, a large perturbation arises around 130 kyr BP, when FWF from the AIS peak. Global mean and hemispheric mean temperatures are similar in all runs after ~127 kyr BP, when the ice sheets have largely reached their interglacial configuration and their FWF remain similar, except for the GrIS, which is retreating until ~120 kyr BP but accounts for only a small contribution. The location of

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largest freshwater induced temperature variations in the NH is the North Atlantic between 40 and 80° N, where changes in the AMOC are the cause for a perturbation of the northward oceanic heat transport and temperature changes which are further amplified by sea ice-albedo and insulation feedbacks. Greenland experiences maximum summer warming in the Reference experiment around 125 kyr BP of less than 3°C over remaining ice covered central Greenland, but marginal warming in the north is up to 10 and up to 14°C over southern margins over a then ice free tundra. These strong warming trends in the ice sheet periphery are due to a combination of elevation changes and local albedo changes, confined to the immediate region of ice sheet lowering and retreat. In the SH, the largest temperature perturbations linked to both NH and SH freshwater fluxes occur in the SO. The largest warming over the ice sheet itself is simulated over the WAIS and is mainly a consequence of the local elevation changes as the ice sheet retreats. However, mainly due to the marine based character of the WAIS, albedo changes are much more limited compared to Greenland as the retreating ice sheet surface is mostly replaced by sea ice.

5 Influence of ice sheet meltwater fluxes

To study the role of the different freshwater contributions from the ice sheets in more detail and evaluate their importance for the climate evolution, we compare additional simulations where FWF from the Greenland and Antarctic ice sheets are partially suppressed relative to the Reference experiment (Fig. 6). The Antarctic ice sheet forcing leads to considerable changes in the Southern Hemisphere, but has very little impact on the NH temperature evolution. Conversely, variations in the NH and GrIS freshwater forcing on millennial time-scales imply temperature changes in the SH on the background of general LIG warming. Differences between the experiments in the AMOC evolution (Fig. 6b) are largely explained by whether FWF from the NH ice sheets and the Greenland ice sheet are included or not. The additional effect of the FWF from the GrIS is limited compared to the large impact of the general NH ice sheet forcing and

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consists of an additional weakening of the AMOC. It is most pronounced during periods of AMOC recovery and after 130 kyr BP, when melting of the GrIS beyond its present day configuration sets in. The simulated evolution of AMOC strength is in good agreement with paleo evidence based on $\delta^{13}\text{C}$ data (Bauch et al., 2012) and in particular with a recent reconstruction based on chemical water tracers (Böhm et al., 2015). The timing of Heinrich Stadial 11 and the variations in AMOC strength thereafter are well captured by our reference simulation, which gives independent credibility to our NH ice sheet reconstructions. The evolution of NH sea ice area (Fig. 6c) shows maxima at times of AMOC minima and vice versa and is closely linked to NH surface temperature variations (cf. Fig. 5b) by modifying the heat exchange between ocean and atmosphere. The largest sea ice area between 135 and 130 kyr BP is simulated in the Reference experiment, which also exhibits the lowest AMOC strength of all experiments.

The situation in the SH is more complex as surface temperature and sea ice evolution are influenced both by freshwater forcing from the Antarctic ice sheet and also by the FWF in the NH. The AMOC variability gives rise to changes in the SH through the so-called interhemispheric see-saw effect (Stocker, 1998). The SH begins to warm as the NH cools due to modified oceanic heat transport across the equator. Minima in SH temperature (cf. Fig. 5c) and maxima in SH sea ice area (Fig. 6d) are therefore associated with maxima in AMOC strength. The additional effect of including GrIS freshwater forcing is consequently also felt in a warmer SH with less sea ice formation. However, the influence of GrIS freshwater fluxes and consequential AMOC variations on the SH temperature appears to be mostly limited to the beginning of the experiment between ~ 135 and 131 kyr BP. It could be speculated that this is related to the larger extent of the SH sea ice in a colder climate, making the system more sensitive due to an increased potential for sea ice-albedo and insulation feedbacks. We also note that modelled periods of increased NH freshwater fluxes, reduced AMOC strength and higher SH temperatures are roughly in phase with periods of steeper increase in GHG concentrations (cf. Fig. 4b), in line with evidence from marine sediment proxies that indicate that CO_2 concentration rose most rapidly when North Atlantic Deep Water

shoaled (Ahn and Brook, 2008). Since GHGs and NH freshwater fluxes are (independently) prescribed in our experiments, the described in-phase relationship lends further credibility to our NH ice sheet reconstruction.

The FWF from Antarctic ice sheet melting (Fig. 6f) increases the SO sea ice area by freshening and stratifying the upper ocean waters, which in turn leads to lower surface temperatures. In our experiments, the increased freshwater flux from the retreating AIS between 131 and 129 kyr BP is in phase with a period of transient AMOC strengthening, which leads to a combined effect of surface cooling and sea ice expansion in the SO.

The formation of Antarctic Bottom Water (AABW) is strongly controlled by salinity and sea ice area (and therefore temperature) of the polar surface waters and hence by both Antarctic and indirectly by NH freshwater fluxes. The AABW formation (Fig. 6e) is stronger for saltier and colder surface conditions and therefore strongest in case noAGfwf, where FWF are suppressed from the AIS (saltier) and the GrIS (colder). For a similar Antarctic freshwater forcing, the AABW formation is stronger for a larger SH sea ice area. Including Antarctic FWF (in noAGfwf) leads to a generally weaker AABW formation (compared to noGfwf) as surface waters become fresher. These relationships imply also that a stronger decrease in AABW formation, associated with decreased CO₂ uptake by the ocean can be found for periods of steeper increase in prescribed radiative forcing. Again, this appears to support consistency in timing between prescribed radiative and NH ice sheet forcing in our modelling.

6 Temperature evolution in the Southern Ocean

The amplitude of variations in SH surface temperature at the millennial time scale induced by NH freshwater fluxes is the strongest in the SO, where anomalies can be amplified by sea ice-albedo and insulation feedbacks. This is also the region that experiences the largest temperature change due to FWF from the Antarctic ice sheet itself.

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In order to study the effect of Antarctic FWF in more detail, we also analysed the oceanic temperature evolution at different levels south of 63° S (Fig. 7). The effect of the AIS freshwater flux in the reference experiment becomes visible in the sea surface temperature after 132 kyr BP (Fig. 7a) as a cooling due to stratification and sea ice expansion (Fig. 7c). At the same time, the subsurface ocean warms (Fig. 7b) as heat is trapped under the stratified surface waters and expanding sea ice area. When the FWF decline towards the end of the AIS retreat around 128 kyr BP, sea ice retreats again and the heat is released to the atmosphere, where it generates an overshoot in SST compared to the experiment with constant Antarctic freshwater fluxes (noAGfwf). The largest effect of this heat buffering is found in winter in regions of strongest warming in the Bellingshausen Sea and off the Gunnerus ridge adjacent to Dronning Maud Land. The maximum sea-ice extent in the SH (Fig. 7c) occurs at the time of largest surface cooling at 129.5 kyr BP.

As a further consequence, the time of maximum annual mean surface air temperature (defined as MWT for Maximum Warmth Timing; Bakker et al., 2013) in the SO differs by several thousand years between experiments. Including Antarctic FWF leads to an earlier MWT (by up to 2 kyr) in large parts of the SO south of 60° S and in the central and eastern parts of the Atlantic sector of the SO up to 40° S. Conversely, a later MWT (by up to 3 kyr) is found in the Indian and Pacific sectors of the SO north of 60° S when Antarctic FWF is accounted for. In the Reference experiment and noGfwf, the MWT lies relatively homogeneously between 129 and 128 kyr BP for the entire SO south of 45° S and falls together with the overshoot in SST after the peak input of Antarctic FWF (Fig. 8). The observed changes of the MWT in the SO due to the additional Antarctic freshwater input can therefore in either way be understood as a shift towards the time when heat from the mid-depth ocean buffer is released to the surface.

The freshwater induced surface cooling at the onset of the LIG appears to be superficial and relatively short lived and of clearly different signature compared e.g. to the Antarctic cold reversal during the last deglaciation. The cooling event is indeed not recorded in our modelled temperature evolution over central East Antarctica, in line

with a lack of its signature in Antarctic ice core records for that time period (Petit et al., 1999; EPICA community members, 2004). A sea ice expansion during Termination II together with an *oceanic* cold reversal around 129.5 kyr BP is however recorded in some deep-sea sediment cores, where the composition of planktonic diatoms suggests meltwater as the primary cause (Bianchi and Gersonde, 2002; Cortese and Abelmann, 2002).

7 Discussion

Despite remaining uncertainties in the timing of ice sheet retreat during Termination II, we find several lines of evidence in support of our ice sheet reconstructions and the associated climatic signatures. The NH ice sheet reconstruction shows some similarity with the IRD signal recorded in North Atlantic sediment cores (Kandiano et al., 2004; Oppo et al., 2006), while the simulated evolution of the AMOC strength (Fig. 6a) is in good agreement with a recent reconstruction based on chemical water tracers (Böhm et al., 2015). The combination of NH and SH sourced freshwater forcing variations produces a stronger decrease in AABW formation, associated with decreased CO₂ uptake by the ocean for periods of steeper increase in prescribed radiative forcing, in line with evidence from marine sediment proxies that indicate that CO₂ concentration rose most rapidly when North Atlantic Deep Water shoaled (Ahn and Brook, 2008). Our modelling results furthermore suggest that the major Antarctic ice sheet retreat from its glacial configuration could be constrained by an oceanic cold event recorded in several SO sediment cores around Antarctica (Bianchi and Gersonde, 2002; Cortese and Abelmann, 2002). As a schematic sensitivity test to uncertainties in the overall glacial Antarctic ice sheet volume, we have performed one more experiment identical to Reference except for Antarctic FWF scaled to 50% of their reference value. The resulting magnitude of the SO cold event and overshoot is lower but exhibits the same timing and spatial expression as in the reference case. The described mechanisms

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limited feedback on ice sheet retreat, similar to what has been suggested by Golledge et al. (2014) for meltwater pulse 1A during the last termination. However, the oceanic warming recorded in our model is not strong and the duration of the perturbation does not appear to be long enough for a sustained impact on the retreat of the ice sheet. Furthermore, the peak in freshwater flux appears when the ice sheet has already retreated considerably and WAIS grounding lines are located mostly on the continental shelves, more protected from the warm water build-up in the mid-depth ocean. A large-scale marine ice sheet retreat of the likely less vulnerable East Antarctic ice sheet sectors (Mengel and Levermann, 2014) appears particularly unlikely, given the environmental forcing at the time apparent in our modelling results. However, in depth studies of these interactions require detailed coupled simulations of the entire ocean-ice sheet system.

8 Conclusions

We have presented a transient simulation of Termination II and the Last Interglacial period with realistic ice sheet boundary conditions from reconstructed Northern Hemisphere ice sheets and detailed stand-alone simulations of the Greenland and Antarctic ice sheets. Our results show that the temperature evolution at the onset of the Last Interglacial was in both hemispheres considerably influenced by meltwater fluxes from the decaying ice sheets. While Antarctic freshwater fluxes lead to strong perturbations of the Southern Ocean, Northern Hemisphere freshwater fluxes have an influence both on the Northern and Southern Hemisphere temperature evolution through the oceanic see-saw effect. The importance of additional freshwater input from the Greenland ice sheet during Termination II is small compared to the much larger fluxes from the other NH ice sheets and becomes more important only later during the Interglacial when it is the only remaining ice sheet contributing freshwater fluxes to the North Atlantic. In the Southern Hemisphere, anomalous freshwater input from the Antarctic ice sheet leads to an episode of surface freshening, increased stratification and sea ice cover and consequently reduced ocean heat loss to the atmosphere, with temporary heat build-up

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in the mid-depth ocean. We argue that the surface ocean cooling associated with this event may be used to constrain an early Antarctic retreat when matched with similar signatures evident in some deep-sea sediment cores from the Southern Ocean.

Our transient simulations confirm results from earlier studies that stress the importance of ice sheet boundary for the climate evolution at the onset of the LIG. However, most of the freshwater induced changes remain visible for at most 1–2 kyr after cessation of the perturbations, indicative of a relative short memory of the (surface) climate system. Additional effects may arise from climate-ice sheet feedbacks not considered in the present model configuration, which should be investigated in fully-coupled experiments.

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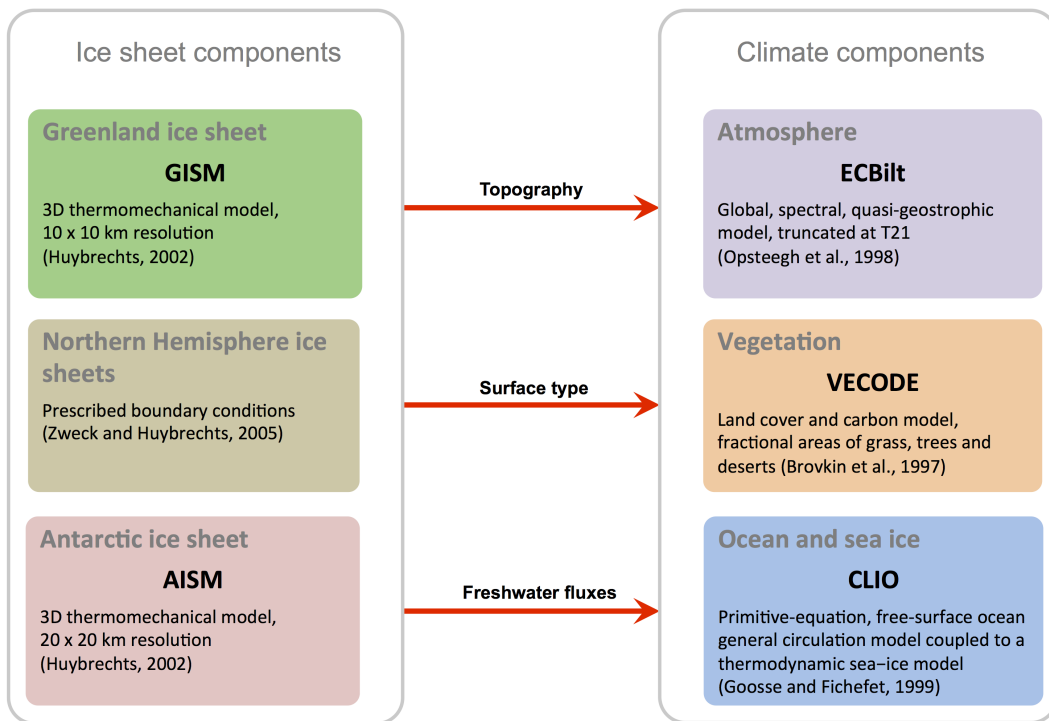


Figure 1. LOVECLIM model setup for the present study including prescribed ice sheet boundary conditions from the Northern Hemisphere, Greenland and Antarctic ice sheets.

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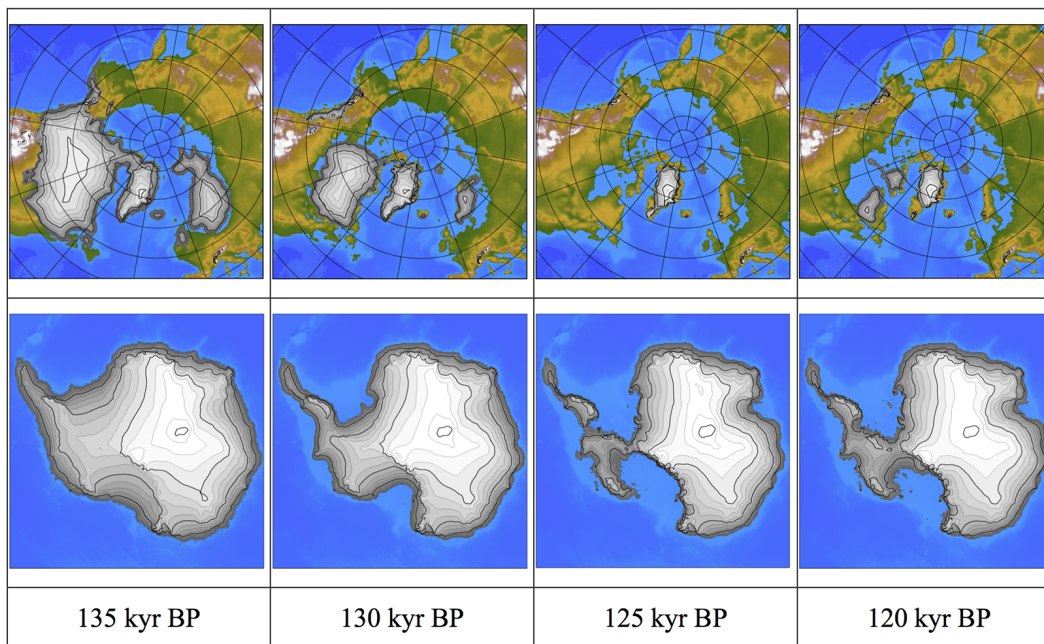


Figure 2. Evolution of reconstructed Northern Hemisphere ice sheets and embedded modelled Greenland ice sheet (top) and modelled Antarctic ice sheet (bottom) used as boundary conditions for the climate model.

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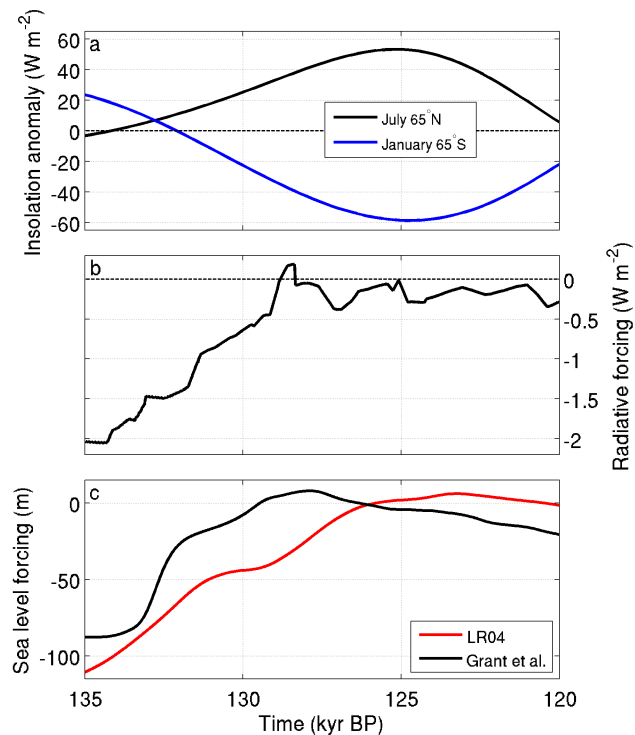


Figure 4. Prescribed model forcing. **(a)** Average monthly insolation anomaly relative to the pre-industrial at 65° N in July (black) and 65° S in January (blue). **(b)** Combined radiative forcing anomaly of prescribed greenhouse gas concentrations (CO_2 , CH_4 , N_2O) relative to the pre-industrial. **(c)** Sea-level forcing for the ice sheet components derived from either oceanic $\delta^{18}\text{O}$ data (Lisiecki and Raymo, 2005) scaled to a global sea-level contrast between LGM and present day of 130 m (red) or derived from a Red Sea relative sea-level record (Grant et al., 2012, black).

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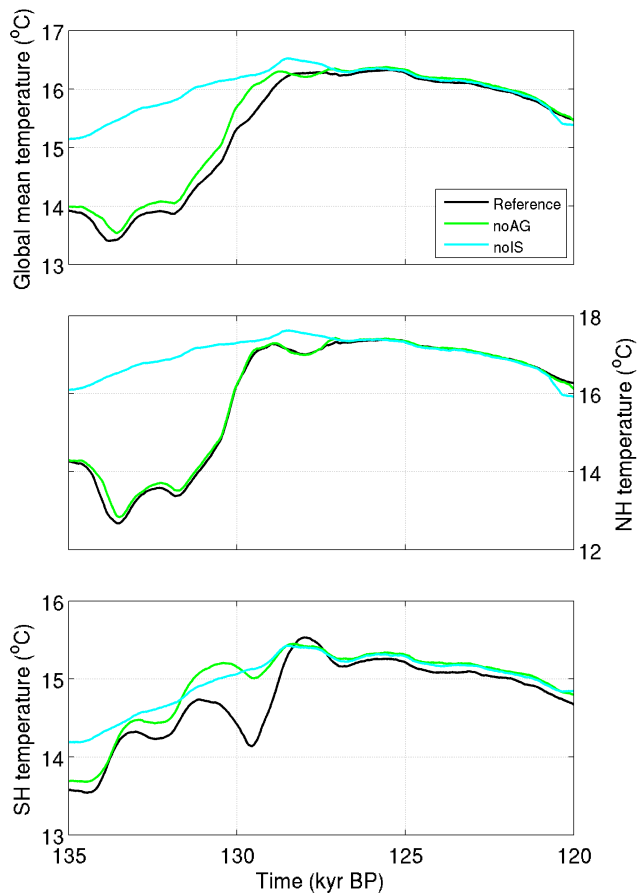


Figure 5. Evolution of global mean and hemispheric mean surface temperature for experiments with different ice sheet forcing included. Curves are smoothed with a running mean of 200 years for better comparison.

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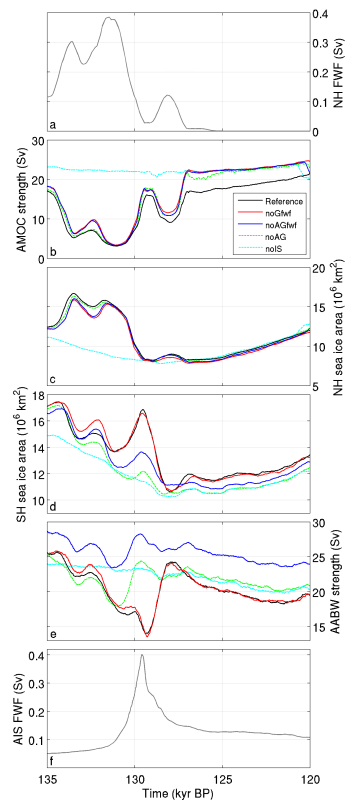


Figure 6. Freshwater forcing and oceanic response characteristics. NH **(a)** and Antarctic ice sheet freshwater fluxes **(f)**, strength of the Atlantic meridional overturning circulation **(b)**, NH sea ice area **(c)**, SH sea ice area **(d)** and strength of Antarctic Bottom Water formation **(e)** for the different experiments with and without freshwater forcing from Greenland, Antarctic and NH ice sheet melting. Curves are smoothed with a running mean of 200 years for better comparison.

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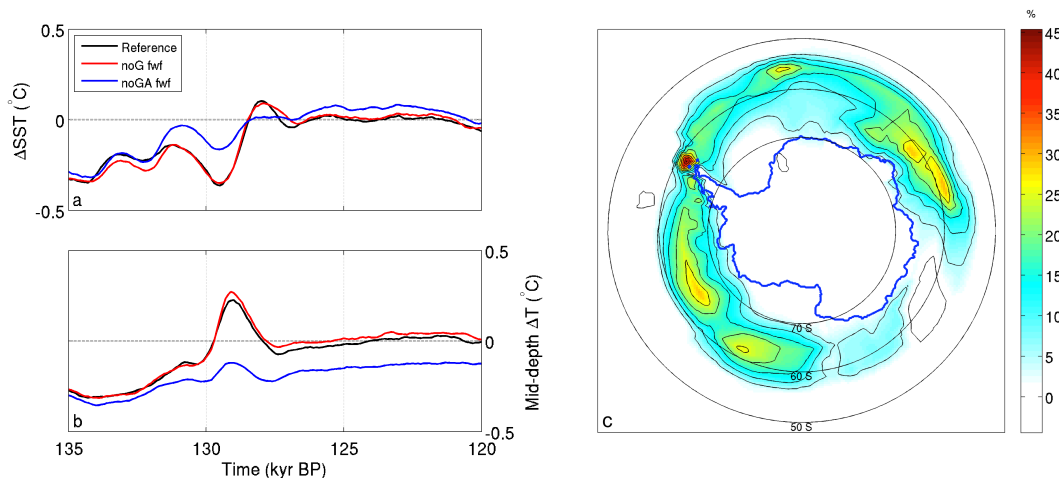


Figure 7. Evolution of annual mean sea surface temperature **(a)** and mid-depth (485–700 m) ocean temperature **(b)** anomalies relative to the pre-industrial in close proximity to the Antarctic ice sheet (south of 63° S). **(c)** Meltwater related changes in annual mean sea ice area at 129 kyr BP from differences between experiments Reference and noAGfwf in per cent. The blue contour outlines the observed ice-shelf edge and grounded ice margin of the present-day Antarctic ice sheet for illustration. All curves **(a, b)** are smoothed with a running mean of 200 years for better comparison.

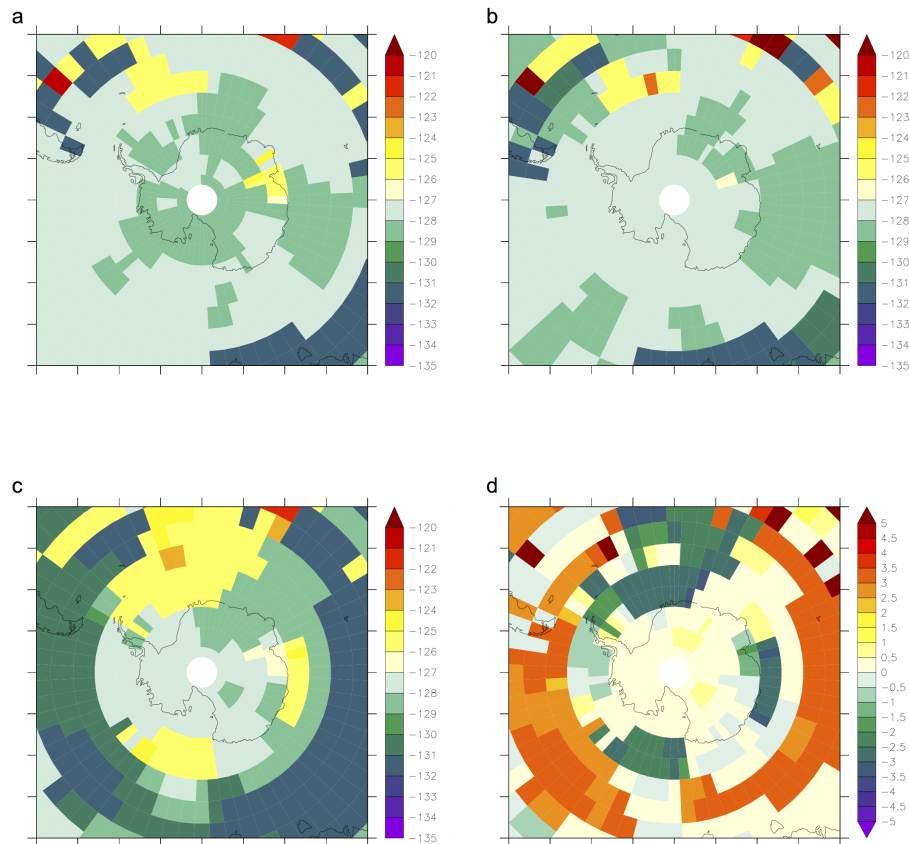


Figure 8. Time of maximum surface air temperature (MWT) in kyrBP for experiments Reference (a), noGfwf (b) and noAGfwf (c) and difference in MWT between experiments noGfwf and noAGfwf (d) in kyr, showing the shift of the MWT when Antarctic freshwater fluxes are included.

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