**Climate
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Discussions**

Heinrich event 1: an example of dynamical ice-sheet reaction to oceanic changes

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Abstract

Heinrich events, identified as enhanced ice-rafted detritus (IRD) in North Atlantic deep sea sediments (Heinrich, 1988; Hemming, 2004) have classically been attributed to Laurentide ice-sheet (LIS) instabilities (MacAyeal, 1993; Calov et al., 2002; Hulbe et al., 2004) and assumed to lead to important disruptions of the Atlantic meridional overturning circulation (AMOC) and North Atlantic deep water (NADW) formation. However, recent paleoclimate data have revealed that most of these events probably occurred after the AMOC had already slowed down or/and NADW largely collapsed, within about a thousand years (Hall et al., 2006; Hemming, 2004; Jonkers et al., 2010; Roche et al., 2004), implying that the initial AMOC reduction could not have been caused by the Heinrich events themselves. Here we propose an alternative driving mechanism, specifically for Heinrich event 1 (H1), by which North Atlantic ocean circulation changes are found to have strong impacts on LIS dynamics. By combining simulations with a coupled climate model and a three-dimensional ice sheet model, our study illustrates how reduced NADW and AMOC weakening lead to a subsurface warming in the Nordic and Labrador Seas resulting in rapid melting of the Hudson Strait and Labrador ice shelves. Lack of buttressing by the ice shelves implies a substantial ice-stream acceleration, enhanced ice-discharge and sea level rise, with peak values 500–1500 yr after the initial AMOC reduction. Our scenario modifies the previous paradigm of H1 by solving the paradox of its occurrence during a cold surface period, and highlights the importance of taking into account the effects of oceanic circulation on ice-sheets dynamics in order to elucidate the triggering mechanism of Heinrich events.

1 Introduction

A major effort has been devoted in the last decade in order to understand rapid glacial climate variability as registered in many climatic archives. Greenland ice core records indicate that the last glacial period was punctuated by more than 20 abrupt warmings

larger than 10 K (Dansgaard-Oeschger events) followed by progressive cooling (Dansgaard et al., 1993; Grootes et al., 1993). As revealed by the study of marine sediment cores in the North Atlantic, six of the temperature minima in Greenland were also coeval with unusual amounts of ice rafted debris (IRD) originating primarily from the areas around Hudson Bay (Bond et al., 1992). Several mechanisms have been proposed to explain these anomalous ice discharge events, known as Heinrich events. The first considers these to be internal oscillations of the Laurentide ice sheet (LIS) associated with alterations of basal conditions (MacAyeal, 1993; Calov et al., 2002). A sudden break-up of ice shelves has also been implicated via atmospheric warming (Hulbe et al., 2004) or tidal effects (Arbic et al., 2004).

Evidence for strongly reduced NADW formation during Heinrich events (Sarnthein et al., 1994) has led to the interpretation that massive iceberg discharge caused important disruptions in the Atlantic Ocean circulation. Yet, recent paleoclimate data have revealed that during H1 (ca. 17.5 kyr BP) peak IRD discharge from the LIS occurred several hundred years after the AMOC had slowed down or largely collapsed (Hall et al., 2006). Furthermore, H1 and earlier H events show the largest IRD peaks occurring several hundred years after the onset of the cold period (Hemming, 2004; Jonkers et al., 2010; Roche et al., 2004), suggesting that the initial AMOC reduction could not have been caused by the Heinrich events themselves.

The identification of additional petrological changes in IRD indicates that for some of the Heinrich layers, the initial increase in IRD flux is associated with icebergs of European origin predating the LIS surges (Hemming, 2004). Such precursor events have been suggested to play a mechanistic role in the initiation of the AMOC reduction (Hall et al., 2006) as well as in the LIS collapse (Grousset et al., 2000). Ocean-ice-sheet interactions including sea-level rise (Levermann et al., 2005) and subsurface temperature warming (Mignot et al., 2007) as a result of NADW reduction have been proposed both to amplify the initial AMOC reduction and the breakup of ice shelves. Lack of buttressing by the ice-shelves would result in substantial ice-stream acceleration leading to increased iceberg production and, thus, to the proper Heinrich event (Alvarez-Solas

et al., 2010b; Hulbe, 2010; Flückiger et al., 2006; Shaffer et al., 2004). This hypothesis is supported by observations in Antarctica that illustrate the relevance of ocean-ice-sheet interactions for understanding recent changes in ice stream velocities (Rignot et al., 2004; Scambos et al., 2004). Here these ideas are assessed quantitatively by investigating the potential effects of oceanic circulation changes on LIS dynamics at the time of H1.

2 Model setup and experimental design

We combined results of simulations with the climate model CLIMBER-3α (Montoya et al., 2005; Montoya and Levermann, 2008) and the GRISLI three-dimensional ice-sheet model of the Northern Hemisphere (Ritz et al., 2001; Peyaud et al., 2007).

Concerning CLIMBER-3α, the starting point is a simulation of the Last Glacial Maximum (LGM, ca. 21 kyr before present (BP)). The forcing and boundary conditions follow the specifications of the Paleoclimate Modelling Intercomparison Project Phase II (PMIP2, http://pmip2.lsce.ipsl.fr), namely: changes in incoming solar radiation, reduced greenhouse gas concentration (since our model only takes CO_2 into account, an equivalent atmospheric CO_2 of 167 ppmv concentration was used to account for the lowered CH_4 and N_2O atmospheric CO_2 concentration), the ICE-5G ice-sheet reconstruction (Peltier, 2004) and changes in land-sea mask consistent with the latter, and an increase of 1 psu to account for the ∼120 m sea-level lowering. Vegetation and other land-surface characteristics as well as the river-runoff routing were unchanged with respect to the present-day control run (Montoya et al., 2005). Due to the coarse resolution of its atmospheric component, the surface winds simulated by the model are not adequate to force the ocean. For experiments representing modest deviations with respect to the preindustrial climate, an anomaly model was implemented in which the wind-stress anomalies relative to the control run are computed and added to climatological data (Montoya et al., 2005). This approach, however, is not appropriate for a considerably different climate such as that of the LGM. Recently, the sensitivity of the

glacial AMOC to the wind-stress strength was investigated by integrating the model to equilibrium with the Trenberth et al. (1989) surface wind-stress climatology multiplied globally by varying factors $\alpha \in [0.5, 2]$ (Montoya and Levermann, 2008). The simulated LGM AMOC strength was found to increase continuously with surface wind-stress up to $\alpha_c \equiv 1.7$. In this wind-stress regime, NADW takes place south of the Greenland-Scotland ridge. At $\alpha = \alpha_c \equiv 1.7$ a threshold, associated with a drastic AMOC increase of more than 10 Sv and a northward shift of deep water formation north of the Greenland-Scotland ridge (GSR), was found. Thus, for $\alpha = \alpha_c \equiv 1.7$ the model exhibits two steady states, with weak and strong AMOC as well as GSR overflow, respectively. The strong AMOC state (LGM1.7-strong) is associated with a stronger North Atlantic current and poleward heat transport, reduced sea-ice cover in the North Atlantic and increased surface temperatures relative to LGM1.7-weak (see also Montoya and Levermann, 2008).

Although the CLIMAP (1976) sea-surface temperature reconstruction indicates that the Nordic Seas were perennially covered with sea-ice during the LGM, more recent data suggest instead that this region was seasonally ice-free (Hebbeln et al., 1994; De Vernal et al., 2006). Thus, our LGM1.7-strong climate simulation is in better agreement with these data and provides a better representation of the LGM climate than LGM-1.7weak, and is herein taken as the starting point for all simulations.

The GRISLI ice-sheet model is nowadays the only one able to properly deal with both grounded and floating ice on the paleo-hemispheric-scale, since it explicitly calculates the Laurentide grounding line migration, ice-streams velocities and ice-shelves behaviour. Inland ice deforms according to the stress balance using the shallow ice approximation (Morland, 1984; Hutter, 1983). Ice shelves and dragging ice shelves (ice streams) are described following MacAyeal (1989). This 3-D ice-sheet–ice-shelf model has been developed by Ritz et al. (2001) and validated over Antarctica (Ritz et al., 2001; Philippon et al., 2006; Alvarez-Solas et al., 2010a) and over Fennoscandia (Peyaud et al., 2007). A comprehensive description of the model is given by these authors. In order to obtain realistic Northern Hemisphere ice sheets at the time of H1, herein GRISLI was forced throughout the last glacial cycle by the climatic fields

resulting from scaling the climate anomalies simulated by the CLIMBER-3α model for LGM and present conditions by an index derived from the Greenland GRIP δ^{18}O ice core record (Dansgaard et al., 1993) (Supplement). This method has been used in many studies to simulate the evolution of the cryosphere during the last glacial cycle (Charbit et al., 2007). The simulated Northern Hemisphere ice-sheet characteristics for 18 kyr BP (Fig. 1) show good agreement with reconstructions in terms of volume and geographical distribution, and it agrees remarkably well with these in terms of ice-stream locations (Winsborrow et al., 2004).

3 Oceanic subsurface warming, ice-shelf collapse and ice-streams acceleration

Our results show that the ice retreat first started over Fennoscandia between 20 and 18 kyr BP. Melting of the Fennoscandian ice sheet resulted in enhanced freshwater flux (sea level rise equivalent of around 2 m) into the Nordic Seas. To assess the impact of the latter on the North Atlantic ocean circulation, several experiments were carried out by imposing comparable freshwater fluxes on the glacial simulation with the climate model. Freshwater fluxes with a fixed amplitude of 0.2 Sv with varying duration (Δt) between 10 and 100 yr were added between 61° N–63° N and 6° W–5° E, representing a sea-level rise between ca. 0.2 and 2 m. In the glacial simulation, NADW takes place in the Nordic and Labrador Seas (not shown). For the weakest freshwater flux perturbations ($\Delta t \leq 20$ yr), NADW was reduced everywhere, but for $\Delta t > 20$ yr, it was inhibited everywhere north of 50° N, thereby increasing sea ice extent and leading to the formation of a strong halocline with presence of warmer subsurface waters, especially in the Nordic Seas (Fig. 1). This simulated pattern fully agrees with marine proxies in Nordic Seas (Clark et al., 2007; Dokken and Jansen, 1999).

This subsurface temperature anomaly (Fig. 1c) propagates on advective timescales (within a few decades; see Supplement animations) toward the Labrador Sea. To investigate its potential effects on the LIS, we carried out two main sets of cryospheric experiments in which the climate fields (surface air temperature and precipitation) of

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the state with weakened NADW were used to force the GRISLI ice sheet/ice shelf model. In the first case, subsurface temperature changes associated with changes in the ocean circulation were taken into account, while in the second case, these were neglected. The comparison between both simulations allows us to isolate and quantify the effects of the oceanic forcing on the LIS dynamics.

In the first case, the enhanced heat flux from the ocean to the ice due to subsurface warming, induces an increase of the basal melt below the Labrador ice shelf (Fig. 1). The reduced shelf thickness increases the calving rate substantially. The breakup of the large ice shelf is very fast (within decades), resulting in a first pronounced peak of ice discharge (from iceberg calving) and freshwater flux into the ocean (from basal melting) (Fig. 4, blue). The ice shelf disintegration has dynamical implications far inland. Ice streams located at the mouth of Hudson Strait and south of Greenland were buttressed by the Labrador ice shelf embayment. Removing this buttressing effect by the ice shelf disintegration results in a sudden acceleration of flow in these ice streams. Comparable to recent observations on the Antarctic Peninsula after the breakup of the Larsen B ice shelf, ice velocities in the coastal LIS increase by a factor 4, shifting from ca. $1000\,\mathrm{m\,yr^{-1}}$ to $4000\,\mathrm{m\,yr^{-1}}$ (Fig. 2).

The duration of this process is considerably longer than for the ice shelf disintegration which caused it. The force balance change, associated with the absence of longitudinal stresses previously exerted by the ice shelf against the continental edges, propagates inland along the ice streams up to their source (located at Hudson Bay in the case of the Hudson Strait ice stream). The ice discharge reaches a maximum at the mouth of the Hudson Strait ice stream around $700\,\mathrm{yr}$ after the beginning of the subsurface warming in the Labrador Sea (Fig. 3), corresponding to the second peak in iceberg discharge into the Atlantic Ocean (Fig. 4).

Large portions of the eastern LIS, where ice dynamics are mainly controlled by the above mentioned ice streams, suffer an important reduction in their thickness (more than $500\,\mathrm{m}$ in the Hudson Bay/Strait area), illustrating the relevance of considering the dynamic coupling between ice streams and ice shelves. Note that when neglecting

oceanic temperatures changes (Fig. 3, black) or when a constant basal melting rate is applied (Supplement) the ice sheet model does not generate any self-sustained ice discharge. As noted above, this is a critical point for the triggering mechanism of Heinrich Events.

4   Conclusions

To summarize, we propose that H1 was triggered by warm North Atlantic subsurface waters resulting from reduced NADW formation. Note that under this mechanism the iceberg discharge configuring H1 is not responsible for the initial NADW reduction. However, the associated freshwater discharge from the H1 event could further impact deep water formation, eventually leading to its shutdown. Here we have shown that a previously weakened meridional oceanic circulation is needed to create the subsurface water anomalies that will perturb ice shelves and therefore trigger the required ice surges.

Although the focus here is on H1, the initial requirement is potentially valid for all six Heinrich Events, given the fact that they all occur during a cold stadial period. The mechanisms that led the ocean into a stadial condition during the other Heinrich events are not discussed here. For H1 we assume, as suggested by proxies (Hall et al., 2006), that the early deglaciation of the Fennoscandian ice sheet resulted in enhanced freshwater fluxes to the North Atlantic, forcing the ocean into a state with weak Atlantic overturning and NADW south of Iceland, similar to a stadial period. The assumption under which the ocean needs to shift into a stadial condition as a precursor for triggering Heinrich Events solves the paradox raised by previous studies (Bond and Lotti, 1995; Shaffer et al., 2004; Clark et al., 2007; Alvarez-Solas et al., 2010b). Under this new mechanism, the dynamic ocean-ice-sheet interaction leads to both cold surface conditions and warm subsurface waters, which is crucial for ice shelves breakup. Reducing their buttressing effect induces a large iceberg discharge and an ice-stream acceleration that tranlates into up to 2 m of sea level rise, with a maximum rate of

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4 mm yr^{-1} (the same order of magnitude as the present-day anthropically-induced rise, with all effects included) only by dynamical reaction of the Laurentide ice sheet.

Our results provide a new consistent mechanism to trigger H1 composed of a sequence of events from initial subsurface warming of the ocean to the final massive ice purge well after the initial NADW reduction, in agreement with data.

Supplementary material related to this article is available online at: http://www.clim-past-discuss.net/7/1567/2011/cpd-7-1567-2011-supplement.zip.

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Ice Thickness (m)   t=18 Ky BP   Ice Velocities (m/yr)

1 5 10 100 250 500 1000 2000 3000 4000 5000   0 2 5 10 20 50 100 200 500 1000 2000

−2.0 −1.0 −0.5 −0.0 1.0 2.0 3.0 4.0 5.0 10.0

Subsurface temperature anomaly (K)
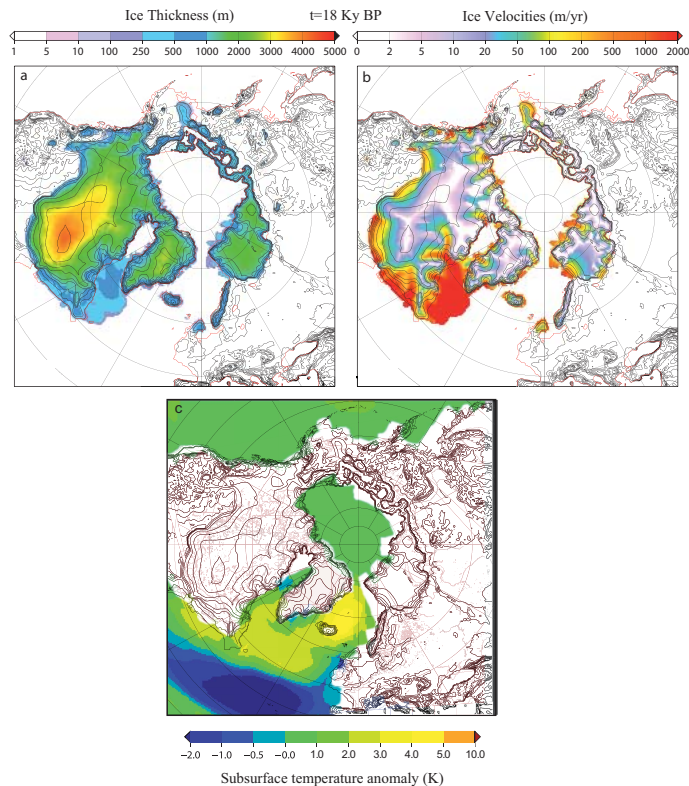
Fig. 1. Northern Hemisphere ice sheets simulated by the GRISLI model at 18 Kyr BP, prior to Heinrich event 1, in terms of ice thickness **(a)** and ice velocities **(b)**, and subsurface (550–1050 m) mean annual temperature anomaly (in K) in response to the shutdown of Nordic Seas deep water formation **(c)**. This temperature anomaly and the corresponding ice-shelf basal melting has been considered during the period 18–17 Kyr BP.
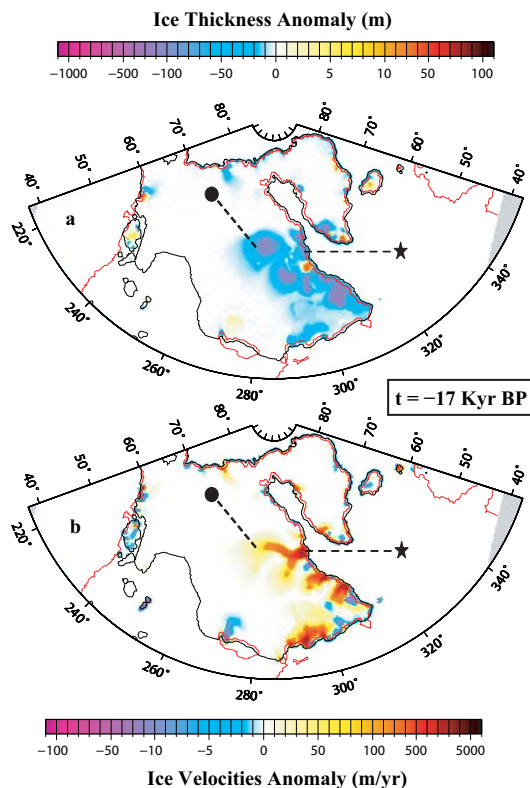
Fig. 2. Ice thickness (in m) and velocity (in m yr^{-1}) anomalies of the Greenland and Laurentide ice sheets when accounting for the effects of the oceanic circulation changes (implying an oceanic subsurface warming) after one thousand years at 17 Kyr BP. The star and circle indicate the location of the Hudson Strait ice stream mouth and source, respectively.
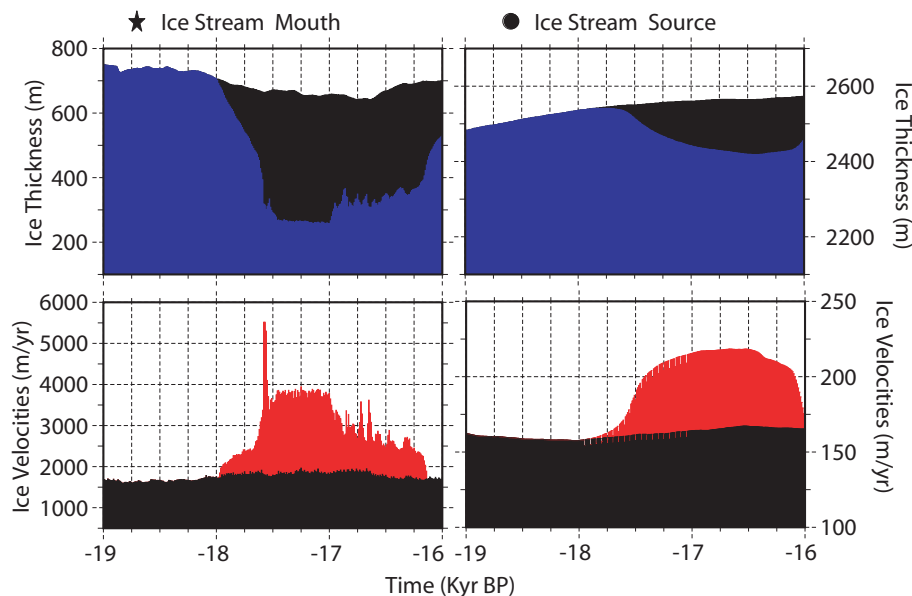
Fig. 3. Evolution of the ice thickness (in m) and velocities (in $m\,yr^{-1}$) for the perturbed simulation (blue and red, respectively) and for the standard simulation without accounting for oceanic circulation changes (black).
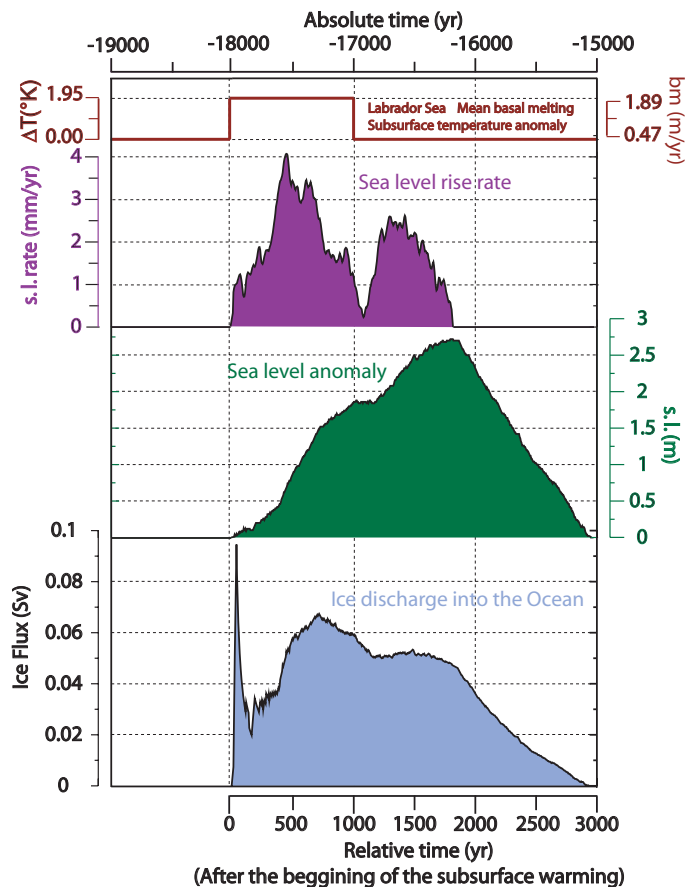
Fig. 4. Labrador Sea subsurface temperature anomaly (in K) and basal melting (in $m\,yr^{-1}$; red curve), sea level rise rate (in $m\,yr^{-1}$), sea level rise (m) and iceberg calving (in Sv), derived from the effects of the oceanic subsurface warming on the dynamic behavior of the Laurentide ice sheet.