Oceanic processes as potential trigger and amplifying mechanisms for Heinrich events

Jacqueline Flückiger,[1] Reto Knutti,[2] and James W. C. White[1]

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[1] Marine sediments recorded a series of Heinrich events during the last glacial period, massive ice surges that deposited prominent layers of ice-rafted debris in the North Atlantic. Here we explore oceanic mechanisms that can potentially trigger and amplify the observed ice calving events. Simulations of abrupt glacial climate change with a coupled ocean-atmosphere-sea ice model show a substantial regional sea level rise in the North Atlantic in response to a collapse of the Atlantic meridional overturning circulation (MOC). The increased heat uptake of the global ocean after the MOC collapse leads to an additional rise in global sea level. We hypothesize that these sea level changes have the potential to destabilize Northern Hemisphere ice shelves and ice sheets and to trigger ice surges. Sea level rise due to ice calving and subsurface ocean warming provides two positive feedback mechanisms contributing to further destabilization of ice shelves and ice sheets.

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1. Introduction

[2] The last glacial period was punctuated by millennial-scale climatic variations, documented most prominently in Greenland ice cores [*Dansgaard et al.*, 1993]. Abrupt warming took place over Greenland and the North Atlantic region every 1000 to 3000 years, followed by a gradual cooling over centuries to millennia before the warm interstadial periods, the so-called Dansgaard-Oeschger (D-O) events, abruptly ceased. In between the warm phases, during the stadial periods, the North Atlantic climate and most parts of the Northern Hemisphere experienced cold climate conditions. Characteristics for several of these stadials are massive ocean sediment deposits, layers of ice rafted debris (IRD) in the North Atlantic, indicating large input of icebergs from the Laurentide and from ice sheets around the Nordic Seas [*Heinrich*, 1988; *Bond et al.*, 1993; *Bond and Lotti*, 1995; *Hemming*, 2004]. The time periods of ice calving from the Laurentide are referred to as Heinrich (H) events [*Bond et al.*, 1993]. H event stadials are accompanied by a significant warming in the south as documented in ice cores from Antarctica [*Johnsen et al.*, 1972; *Blunier and Brook*, 2001] and by a sea level rise in the order of 10 to 30 m [*Shackleton et al.*, 2000; *Yokoyama et al.*, 2001; *Chappell*, 2002; *Siddall et al.*, 2003]. Further, δ^{13}C and ^{231}Pa/^{230}Th measurements along marine cores from the North Atlantic [*Shackleton et al.*, 2000; *Elliot et al.*, 2002; *McManus et al.*, 2004] indicate a strongly reduced ventilation because of a reduced or probably even halted North Atlantic Deep Water (NADW) formation [*Marchal et al.*, 1998]. H events occur at the end of Bond cycles, which are long-term cooling cycles that consist of several progressively colder D-O events [*Bond et al.*, 1993]. While H events occurred about every 7 kyr, stadial periods were more frequent. Non-Heinrich stadials lack the large input of IRD from the Laurentide ice sheet but still show IRD input from the ice sheets adjacent to the Nordic Seas [*Bond and Lotti*, 1995; *Elliot et al.*, 1998; *Dokken and Jansen*, 1999; *van Kreveld et al.*, 2000], indicating more frequent but much less intense ice calving from the east Greenland, Icelandic and Fennoscandinavian ice sheets than from the Laurentide. This is supported by the sea level record, which shows no or only small sea level changes during non-Heinrich stadials [*Shackleton et al.*, 2000; *Siddall et al.*, 2003]. During non-Heinrich stadials the NADW formation was probably reduced and shifted southward, the meridional overturning was weaker and shallower but did not collapse [*Ganopolski and Rahmstorf*, 2001; *Elliot et al.*, 2002]. An overview of paleorecords from ice cores and marine sediments is given in Figure 1.

[3] Conceptual, box models and models of intermediate complexity are capable of simulating many aspects of the abrupt climate events during glacial periods, including the timing and amplitude of the temperature and precipitation patterns, dynamical aspects, and changes in the ocean circulation and tracer distributions [e.g., *Marchal et al.*, 1998; *Ganopolski and Rahmstorf*, 2001; *Schmittner et al.*, 2002; *Schulz et al.*, 2002; *Schmittner et al.*, 2003; *Stocker and Johnsen*, 2003; *Weaver et al.*, 2003; *Knutti et al.*, 2004; *Shaffer et al.*, 2004].

[4] Of particular interest for this paper are the onset and the propagation of H events. Several mechanisms were proposed to explain the onset of H events (see *Hemming* [2004] for an overview), such as for example ice sheet internal processes [*MacAyeal*, 1993; *Papa et al.*, 2006], ice shelf build up and collapse [*Hulbe*, 1997; *Hulbe et al.*, 2004; *Alley et al.*, 2006] and internal oscillations in the ice

[1]Institute of Arctic and Alpine Research, University of Colorado, Boulder, Colorado, USA.
[2]Climate and Global Dynamics, National Center for Atmospheric Research, Boulder, Colorado, USA.

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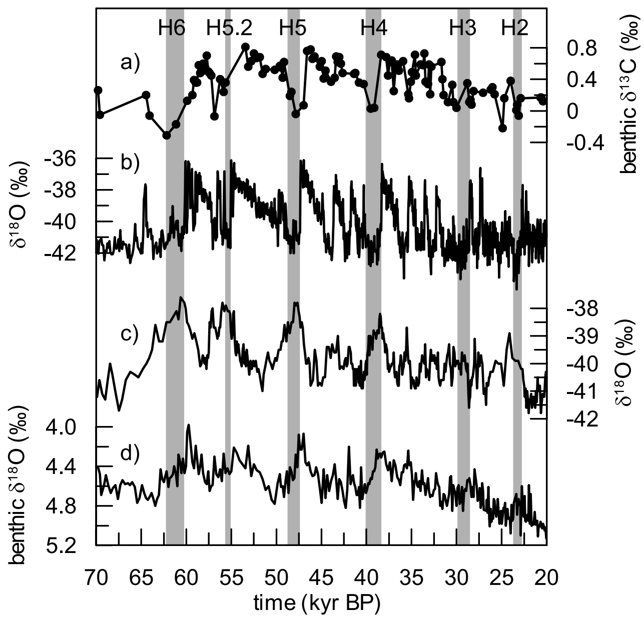

Figure 1. Overview of paleodata over part of the last glacial period: (a) Benthic $\delta^{13}C$ from core MD95-2042 from the Iberian margin off southern Portugal [*Shackleton et al.*, 2000]. Low values indicate reduced ventilation of the North Atlantic Deep Water. Water isotope reconstruction from (b) the Greenland Ice Core Project (GRIP) ice core, Greenland [*Dansgaard et al.*, 1993], and (c) the Byrd ice core, Antarctica [*Johnsen et al.*, 1972], proxies for Greenland and Antarctic temperatures, respectively. (d) Benthic $\delta^{18}O$ from core MD95-2042, which is assumed to be partially a proxy for sea level variations [*Shackleton et al.*, 2000]. Age scales are synchronized to GRIP [*Shackleton et al.*, 2000; *Blunier and Brook*, 2001], and all data are plotted on the GRIP SS09sea timescale [*Johnsen et al.*, 2001]. Shaded bands indicate Heinrich event stadials. The corresponding Heinrich events are labeled at the top of Figure 1 [*Sarnthein et al.*, 2001].

sheet–climate system [*Calov et al.*, 2002]. The causal relation of D-O and H events, however, remains less clear. A possible phase lock mechanism between H events and D-O cycles through the dependence of accumulation rate on temperature was proposed by *Schulz et al.* [2002] in a conceptual model. All those hypotheses are able to explain many aspects of H events, but they all have uncertainties associated with them. They all require to some extent assumptions that cannot be tested at this stage with the available data, and thus leave room for alternative hypotheses and mechanisms.

[5] An oceanic mechanism to explain the onset of H events was discussed by *Shaffer et al.* [2004] based on results from a zonally averaged climate model with idealized Atlantic and Southern Ocean geometry and a north south resolution of four boxes, coupled to an energy balance atmosphere. Their simulations did show unforced switches in the strength of the MOC, so-called deep decoupling oscillation cycles. They propose that diffusive subsurface warming and the related sea level rise could destabilize ice shelves and ice sheets.

[6] In this paper we use a climate model of intermediate complexity to investigate potentially important oceanic feedbacks operating before and during H events in greater detail. We propose increases in global and local sea level due to changes in the Atlantic meridional overturning circulation as trigger mechanisms for the ice surges during H events, and sea level rise as well as subsurface warming in the Nordic Seas and the Labrador Sea as important feedback mechanisms that could have amplified the surging of ice from the Laurentide and other Northern Hemisphere ice sheets during these events. We also discuss a possible combination of our theory with existing hypotheses related to internal ice sheet and ice shelf instabilities.

2. Model Setup and Experiments

[7] The model used for this study is the global, coupled atmosphere-ocean-sea ice model ECBILT-CLIO version 3. The atmosphere is represented by the T21, 3-level quasi-geostrophic model ECBILT [*Opsteegh et al.*, 1998], which contains a full hydrological cycle and explicitly computes synoptic variability associated with weather patterns. The ocean model CLIO is a primitive equation, free-surface ocean general circulation model with a resolution of $3^\circ \times 3^\circ$ and twenty unevenly spaced depths layers, coupled to a thermodynamic, dynamic sea ice model [*Goosse and Fichefet*, 1999]. The coupled atmosphere-ocean model includes realistic topography and bathymetry, a simple representation of land surface processes and a bucket-style freshwater runoff scheme. It incorporates weak freshwater flux corrections. The model is freely available at the Web site http://www.knmi.nl/onderzk/CKO/ecbilt.html. The simulations presented in this paper use boundary conditions of the last glacial maximum, with adjusted topography, greenhouse gas concentrations, orbital parameters and albedo values [*Timmermann et al.*, 2004]. The LGM model results are discussed in detail by *Timmermann et al.* [2004, 2005]. In contrast to some proxy evidence the model shows only a small reduction of the MOC compared to present-day boundary conditions. However, none of the mechanisms discussed in this paper depends qualitatively on the strength or the hysteresis behavior of the MOC. Freshwater input into the whole North Atlantic between 50°N and 70°N or into the northern North Atlantic east of Greenland only (35°W to 15°E, 60°N to 75°N) is used to force changes in the Atlantic meridional overturning circulation (MOC). This freshwater input is uniformly compensated over the whole global oceans mainly to isolate the different processes in global sea level rise. Simulations without freshwater compensation show very similar dynamics of the MOC collapse, changes in regional sea level and subsurface temperatures in the North Atlantic.

[8] The results presented here are mainly based on model runs forced with a freshwater flux prescribed as a step function, with amplitude of 0.2, 0.3, 0.4 and 0.5 Sv, respectively and lasting for 1000 years, before switching back to zero (Figure 2). The model shows the typical hysteresis behavior [*Stocker and Wright*, 1991] such that there are two steady states for 0.2 Sv constant freshwater input, one with a strong meridional overturning and one

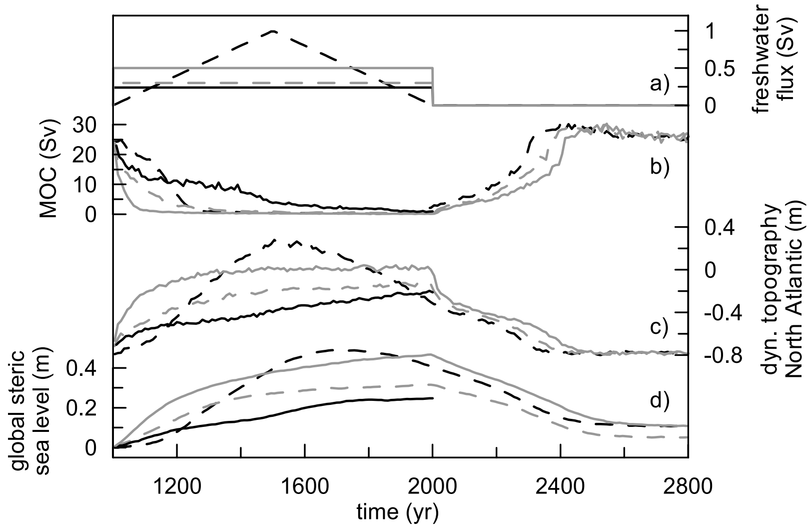

Figure 2. Transient model simulations performed with ECBILT-CLIO for four different freshwater forcings. (a) Freshwater forcing applied to the North Atlantic between 50°N and 70°N. (b) Maximum meridional overturning circulation in the North Atlantic. (c) Anomalies of dynamic topography (sea surface height) from the global mean in the North Atlantic (60°W to 20°E, 50°N to 80°N). (d) Global steric sea level change due to thermal expansion.

without. These two states are of particular interest because a comparison of the differences is free of the effect of freshwater forcing. All these model runs allow us to compare different equilibrium states, to quantify the response time of different processes to MOC changes and freshwater input, and to separate the effects caused by changes in MOC strength and freshwater flux. Additional transient simulations were calculated based on idealized equilateral triangle shaped freshwater scenarios, with a duration of 1000 years and peak amplitudes of 0.2, 0.3, 0.5, and 1.0 Sv, respectively (Figure 2).

[9] We have used the Bern2.5D model in addition to the ECBILT-CLIO model to explore the magnitude of the steric sea level rise due to a MOC collapse. The Bern2.5D model is also a coupled climate model of intermediate complexity, but much simpler and thus more efficient. It consists of a zonally averaged dynamical ocean, an energy moisture balance atmosphere and a thermodynamic sea ice component [*Stocker et al.*, 1992] and has been used to study the stability of the MOC and the climatic effects related to changes in the latter [*Stocker and Schmittner*, 1997; *Knutti and Stocker*, 2002]. The results from the Bern2.5D model, which we show here, use present-day boundary conditions. The inclusion of three different mixing parameterizations allows studying the uncertainty of the MOC response to how subgrid-scale mixing processes are represented in these typically coarse-resolution models. The ocean mixing parameterizations considered are horizontal/vertical diffusion (HOR), isopycnal/diapycnal diffusion (ISO) and isopycnal diffusion with the Gent/McWilliams (GM) parameterization [*Gent et al.*, 1995] and are identical to those used in previous studies with the same model [*Knutti and Stocker*, 2000; *Knutti et al.*, 2000].

[10] Freshwater input into the North Atlantic is used in both models to achieve a change in the strength of the MOC. There are, however, several other mechanisms discussed in section 3.1 that probably can induce changes in the MOC.

3. Heinrich Event Mechanism Based on Oceanic Changes

[11] While many of the trigger mechanisms proposed for H events involve ice sheet internal [*MacAyeal*, 1993] or ice sheet related mechanisms [*Alley et al.*, 2006], we propose an oceanic trigger for these events, presuming the ice sheets are close to a threshold and sensitive to external perturbations [*Bond et al.*, 1993; *MacAyeal*, 1993]. Evidence that there is a strong coupling between the ocean, ice shelves and ice sheets does come from present-day measurements made in Antarctica. Changes in ocean subsurface temperatures and sea level do affect ice shelves and perturbations are propagated hundreds of kilometers inland [*Bamber et al.*, 2000; *Rignot and Jacobs*, 2002; *Bindschadler et al.*, 2003; *Payne et al.*, 2004]. The breakup of ice shelves buttressing large amounts of inland ice leads to an acceleration of the ice stream velocities and to increased glacier surges [*De Angelis and Skvarca*, 2003; *Rignot et al.*, 2004; *Scambos et al.*, 2004; *Dupont and Alley*, 2005].

[12] Figure 3 gives an overview of the trigger and feedback mechanisms that are involved in our Heinrich event hypothesis. The mechanisms involve changes in subsurface temperature as well as in sea level. The oceanic temperature patterns are strongly affected by the ocean currents. Strong changes in the MOC for example do affect sea surface temperature [*Stocker et al.*, 1992; *Mikolajewicz and Maier-Reimer*, 1994; *Ganopolski and Rahmstorf*, 2001; *Vellinga and Wood*, 2002; *Schmittner et al.*, 2003], subsurface and deep ocean temperatures [*Knutti and Stocker*, 2000; *Knutti et al.*, 2004]. Sea level varies globally and/or locally by (1) adding volume (i.e., melting of continental ice

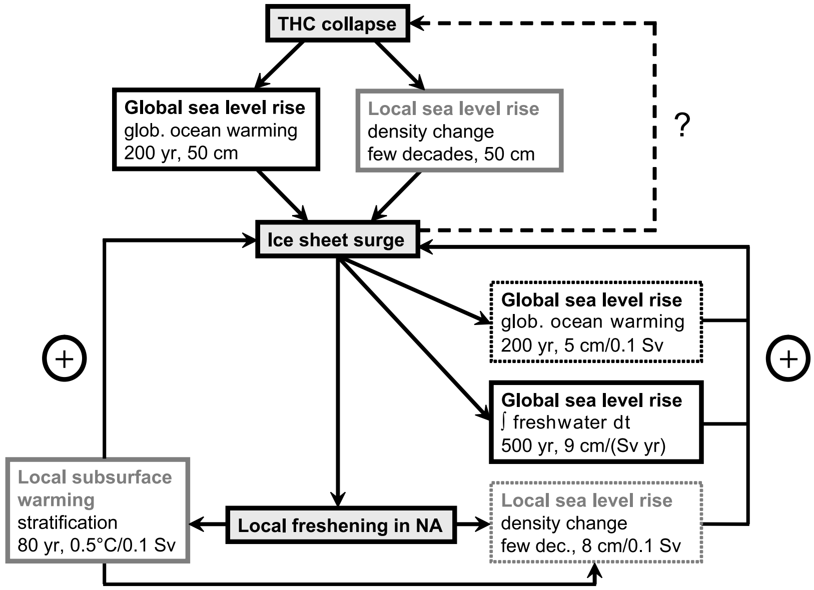

Figure 3. (top) Trigger and (bottom) positive feedback mechanisms that are responsible for the large ice surges from Northern Hemisphere ice sheets during Heinrich events according to our hypothesis. Each box represents a single mechanism and contains its cause, its characteristic response timescale, and the order of magnitude of the sea level rise or the subsurface warming. The timescales were deduced from the ECBILT-CLIO model except for the eustatic sea level rise, which is determined by the duration of ice calving. We have used near-equilibrium runs with collapsed meridional overturning circulation (MOC) but different strength of freshwater forcing to quantify the magnitude of the effects that are given per 0.1 Sv. Mechanisms that are of minor importance are shown in dotted boxes. The dashed arrow in the upper part indicates the possibility that a freshwater anomaly caused by small meltwater or iceberg discharge from a Northern Hemisphere ice sheet could have been the origin of the MOC collapse. Other mechanisms that could have caused the MOC collapse are discussed in the text. Plus signs on the two bottom loops indicate that the net feedback loops are positive.

sheets), (2) changing the water volume with constant mass through heating/cooling, and (3) changing locally the density distribution through changes in ocean currents or the freshwater budget. Figure 2 gives an overview of the sea level changes in response to freshwater input into the North Atlantic and a MOC shutdown as predicted by the ECBILT-CLIO model.

[13] The mechanisms and feedbacks related to sea level and oceanic subsurface temperature changes will be discussed in detail in the following sections of this paper. They operate under a broad range of boundary conditions and we therefore do not attempt to model one particular H event, but illustrate potential interactions of the ocean with ice shelves and ice sheets during all H events. Isostatic effects and changes in the dynamic topography of sea level due to gravitational effects caused by the presence of the ice sheets are not discussed here. They are probably small at the ice sheet margins and have a longer timescale than the processes considered here. If anything, they would have reinforced the vulnerability of the ice sheets to ocean perturbations toward the end of a Bond cycle when the ice sheets have reached the intermittent largest extent.

3.1. The Trigger

[14] The trigger mechanism for H events in the hypothesis discussed here is a collapse of the MOC, which induces a sea level rise in the North Atlantic because of two independent mechanisms. First, we find that the global reorganization of the oceanic density structure causes a change in the dynamic topography of sea level in the North Atlantic, consistent with *Levermann et al.* [2005]. The deviation of sea surface elevation from the global mean is defined as dynamic topography. It is determined, among other factors, by the oceanic density structure. A weakening of the MOC reduces the advection of dense, salty water into the North Atlantic, therefore reducing the local density and increasing sea level locally. The rise in the dynamic topography related to a MOC shutdown in the ECBILT-CLIO model is about 0.6 m in the Nordic Seas (35°W to 20°E, 65°N to 80°N) and in the Labrador Sea (90°W to 50°W, 50°N to 75°N). Figure 4 shows a map of the change in dynamic topography related to the shutdown of the MOC. The adjustment timescale of this process in relation to the MOC change is determined by wave propagation mechanisms and is in the order of a few decades (Figure 2c) [*Huang et al.*, 2000; *Cessi et al.*, 2004; *Timmermann et al.*, 2005].

[15] The second effect is caused by a temporary strong oceanic heat uptake in response to a MOC shutdown, which, in turn, leads to an additional global sea level rise (the steric sea level rise) due to the thermal expansion of the ocean water [*Knutti and Stocker*, 2000]. The characteristic timescale of the steric sea level rise is related to diffusive

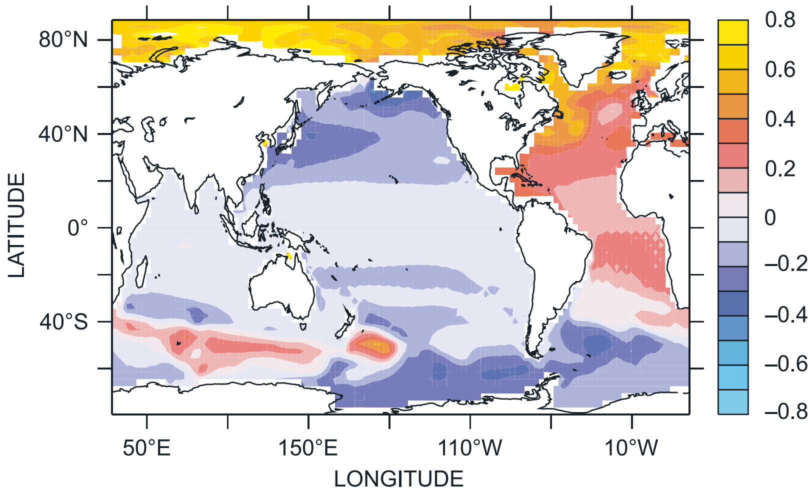

Figure 4. ECBILT-CLIO changes in the dynamic topography (in m) due to a MOC collapse, calculated as difference between two near-equilibrium states with active and collapsed MOC but with identical freshwater forcing. The regional sea level rise in the Nordic Seas and the Labrador Sea due to the MOC collapse is in the order of 0.6 m.

ocean heat uptake and predicted by the ECBILT-CLIO model to be in the order of 200 years (Figure 2d). Figure 5 shows the equilibrium strength of the MOC and the global equilibrium steric sea level rise for different constant freshwater flux scenarios and two different climate models, the ECBILT-CLIO and the Bern2.5D model. Both models predict a global steric sea level rise in the order of 0.3 to 0.5 m for a MOC collapse, depending on how ocean mixing is parameterized.

[16] The resulting total sea level rise in the North Atlantic Ocean, in the order of 1 m within a short timescale, might be too small to destabilize the large ice shelves and ice streams of the Laurentide ice sheet but it has the potential to destabilize smaller marine-based ice streams around the Nordic Seas assuming that they are more sensitive to small sea level perturbations than the large ones. This would result in ice surging into the northern North Atlantic, which in turn ignites strong positive feedback mechanisms that lead to further ice surges from the Northern Hemisphere ice sheets. These feedbacks will be discussed in section 3.2.

[17] Our trigger mechanism starts with the assumption of a MOC collapse. Marine data and model experiments indeed indicate that H events are accompanied by a complete shut down of the MOC [*Shackleton et al.*, 2000; *Ganopolski and Rahmstorf*, 2001; *Sarnthein et al.*, 2001; *Elliot et al.*, 2002; *Schmittner et al.*, 2002; *Hemming*, 2004; *Knutti et al.*, 2004; *McManus et al.*, 2004]. The exact timing of the MOC collapse, however, is still under debate. Nevertheless, there are indications that the MOC did weaken substantially before the large ice calving [*Zahn et al.*, 1997; *McManus et al.*, 2004; I. R. Hall et al., Accelerated draw-down of meridional overturning in the late-glacial Atlantic triggered by transient pre-H event freshwater perturbation, submitted to *Geophysical Research Letters*, 2006, hereinafter referred to as Hall et al., submitted manuscript, 2006] during H events. The question whether freshwater input into the North Atlantic or different mechanisms are responsible for the drastic change in the Atlantic meridional overturning strength remains unsolved. Some models indicate that the glacial MOC was close to its instability threshold [*Ganopolski and Rahmstorf*, 2001; *Schmittner et al.*, 2002], which implies that a small perturbation, for instance in the

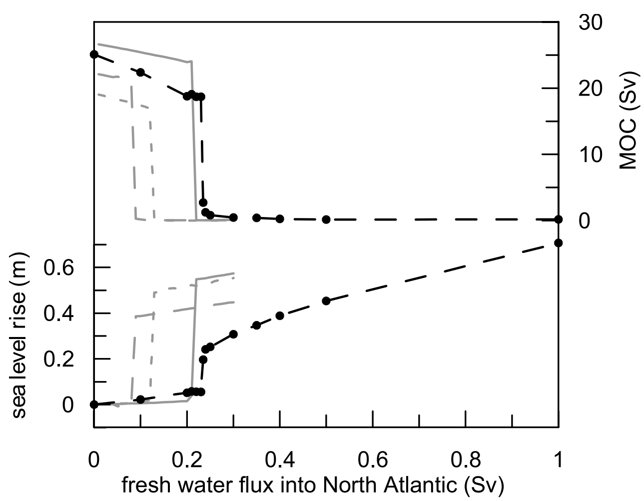

Figure 5. Maximum meridional overturning circulation (top) and global steric sea level rise (bottom) for near-equilibrium model simulations forced with different constant freshwater fluxes into the North Atlantic. Shown are the results from the ECBILT-CLIO model (black dots and dashed line) and from the Bern2.5D model with different mixing parameterizations: isopycnal mixing (shaded solid line), horizontal mixing (shaded dashed line), Gent/McWilliams mixing parameterization (shaded dotted line). For the ECBILT-CLIO model, 14 states were calculated (black dots). Changes are given at year 800 to 1000 after the freshwater perturbation starts and has not completely reached equilibrium (see Figure 2d). The Bern2.5D model runs were performed every 0.01 Sv and represent equilibrium changes after 5000 years.

freshwater budget of the North Atlantic, is able to weaken or even to shut down the MOC. Several other mechanisms have been suggested as potential triggers for a shift, weakening or shutdown of the circulation, including relaxation from an unstable/metastable state [*Ganopolski and Rahmstorf*, 2002; *Schmittner et al.*, 2002; *Schulz et al.*, 2002], oscillatory behavior [*Aeberhardt et al.*, 2000], noise induced transitions [*Knutti and Stocker*, 2002], transitions induced by a very weak forcing [*Ganopolski and Rahmstorf*, 2001], a combination of forcing and noise [*Ganopolski and Rahmstorf*, 2002], slow changing boundary conditions [*Wood et al.*, 1999] or spontaneous transitions due to internal variability [*Sakai and Peltier*, 1997; *Hall and Stouffer*, 2001; *Goosse et al.*, 2002; *Schaeffer et al.*, 2002].

[18] On the basis of the discussion above it is conceivable that the MOC collapses at the end of the last D-O event of a Bond cycle without a large or even with no input of freshwater into the North Atlantic. Only the MOC collapse itself then triggers the onset of significant freshwater input into the North Atlantic related to the H event. Even though the MOC shutdown in the model runs presented here is achieved by a rather large freshwater input into the North Atlantic (Figures 2a and 2b), we emphasize that the cause for the MOC shutdown is not part of the hypothesis presented here. Our model results can shed light on the impacts of a MOC collapse as described above, whereas they cannot solve the questions of the trigger of the MOC collapse itself.

[19] If the MOC was only reduced at the beginning of the H stadial but not fully collapsed, the same qualitative trigger would operate, but with smaller amplitude. The model suggests that both the local and global steric part of the sea level rise is approximately linear for a reduction of the MOC to about 5 Sv (consistent with *Levermann et al.* [2005]), with a substantial nonlinear increase below 5 Sv. Thus for example, a reduction of the MOC from 25 to 10 Sv would increase sea level in the North Atlantic by about 0.4 m.

3.2. Amplifying Feedback Mechanisms

[20] As soon as the sea level rise due to the MOC collapse leads to the first ice discharge into the northern North Atlantic, two strong positive feedback loops come into play that amplify possible ice surges and induce additional ice shelves to collapse and ice sheets to calve (Figure 3). The two feedback loops are related to an ongoing sea level rise and a subsurface warming in the Labrador Sea as well as in the Nordic Seas. Both feedback loops have the potential to further destabilize ice shelves and ice sheets in the North Atlantic region and will be discussed in this section.

[21] Global and local sea level rise progresses after the MOC collapse due to three mechanisms. The largest and dominant contribution to sea level rise originates from the surging and melting ice itself that leads to a global eustatic sea level rise. Its evolution over an H event corresponds to the integral of the total freshwater release from ice sheets into the global ocean. Sea level reconstructions indicate that the total sea level rise along with H events is between 10 and 30 m [*Yokoyama et al.*, 2001; *Chappell*, 2002; *Siddall et al.*, 2003], most of which is due to eustatic sea level rise. Eustatic sea level rise therefore is a positive feedback mechanism for ice surges during H events. It was suggested that part of the eustatic sea level rise during H events is caused by ice surges in Antarctica [*Rohling et al.*, 2004]. This does not contradict our H event hypothesis as long as part of the global freshwater input is released into the North Atlantic and the transient evolution of the MOC shows a strong reduction or a shut down during the H stadial. Another but much smaller feedback mechanism is the local sea level rise in the North Atlantic due to density changes in the water column related to the freshwater input. In the equilibrium model runs this effect is in the order of 8 cm per 0.1 Sv of constant freshwater input into the North Atlantic (Figures 2a and 2c), the characteristic adjustment timescale of this process is in the order of decades. Note that this sea level rise is only seen as long as the freshwater flux is maintained. A small global steric sea level rise due to global ocean heat uptake is also observed in response to additional freshwater input into the North Atlantic after the MOC collapse. Our equilibrium runs for different amounts of freshwater input into the North Atlantic (Figures 2a and 2d and Figure 5) show that the effect is in the order of 5 cm per 0.1 Sv of constant freshwater input. The characteristic timescale of this process is related to diffusive ocean heat uptake and is about 200 years. From these model simulations we conclude that the effects on local and global sea level due to local freshening and changes in the global ocean heat content respectively are small compared to the eustatic component.

[22] An additional positive feedback loop beside the sea level rise is the subsurface warming in the North Atlantic in response to the freshwater input. While the MOC collapse itself leads to a vigorous cooling of several degrees at the surface as well as below the surface of the North Atlantic, additional freshwater input after the MOC collapse leads to almost no change in the North Atlantic SST but to a significant warming of the subsurface temperatures below 100 m north of 40°N. *Shaffer et al.* [2004] used a low-resolution, zonally averaged climate model with idealized Atlantic and Southern Ocean geometry and found a slow warming of the whole Atlantic basin above 2500 m depth after a MOC shutdown through downward diffusion of heat at low latitudes. The climate model (including an ocean general circulation model) used here allows studying the subsurface warming in a more realistic setup and in greater detail. The subsurface warming in our model is due to increased stratification caused directly by freshwater discharge, thus has a short response timescale, and is limited to the North Atlantic (north of 40°N). While there is warming in both studies, the processes described here are clearly different in nature. Evidence for a subsurface warming in the North Atlantic after freshwater experiments was also simulated but not discussed in earlier general circulation model runs with the ECHAM3/LSG model [*Schiller et al.*, 1997] (see *Stocker and Johnsen* [2003] and Figure 1b).

[23] We focus on the temperature changes in the depth range between 100 and 500 m, because many of the marine-based ice sheets in the Northern Hemisphere have reached to these depths [*Sejrup et al.*, 2003] and were therefore exposed to temperature changes in these depths. Subsurface

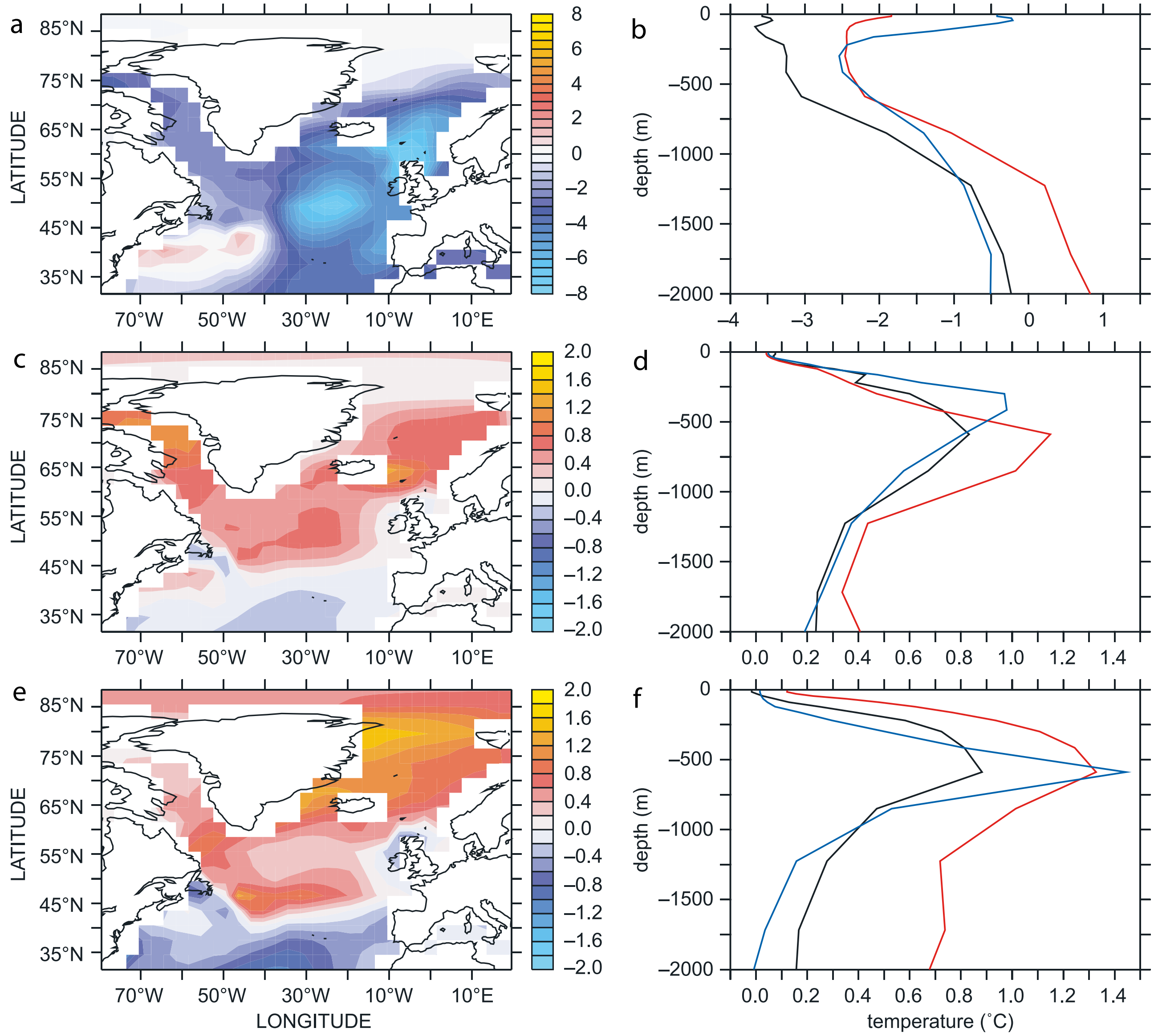

Figure 6. ECBILT-CLIO model results for (left) average subsurface temperature change (°C) in the depth interval 100 to 500 m and (right) temperature change profiles for the northern North Atlantic (60°W to 20°E, 50°N to 80°N, black), for the Nordic Seas (red), and the Labrador Sea (blue). (a and b) Difference between two near-equilibrium states with active and collapsed MOC but with identical freshwater forcing (0.2 Sv) showing the widespread cooling of the subsurface temperatures in the North Atlantic related to the MOC collapse. (c and d) Difference between two near-equilibrium states with collapsed MOC forced with freshwater inputs of 0.4 and 0.3 Sv, differing by 0.1 Sv. Freshwater forcing is applied to the North Atlantic between 50°N and 70°N. (e and f) Same as Figures 6c and 6d but with freshwater forcing applied to the northern North Atlantic east of Greenland. Figures 6c–6f show the subsurface warming in the North Atlantic related to freshwater input after the MOC collapse. Note the different temperature scales of Figures 6a and 6b versus Figures 6c–6f.

ocean temperature changes at the front of an ice shelf and in the area of the grounding line have the potential to destabilize the ice shelf, which in turn leads to an acceleration of the ice flow of the ice stream [*Rignot and Jacobs*, 2002; *Payne et al.*, 2004]. Figures 6a and 6b show the average cooling in this depth range and the temperature change versus depth, respectively, because of the MOC collapse only. Figures 6a and 6b represent the difference between two model runs with the same freshwater forcing of 0.2 Sv, but with different states of the MOC (“on” state versus “off” state) due to the hysteresis behavior as described in the model section. The average temperature between 100 and 500 m in the Nordic Seas and the Labrador Sea decreases by 2.4°C and 1.7°C, respectively. The average temperature in the same depth over the northern North Atlantic (60°W to 20°E, 50°N to 80°N) shows a decrease of 3.5°C. Additional freshwater input into the North Atlantic after the MOC collapse leads to a widespread subsurface warming. Figures 6c and 6d show the difference between two equilibrium model runs, both with a collapsed MOC but with a difference in the freshwater input into the North Atlantic of 0.1 Sv (freshwater forcing of 0.4 and 0.3 Sv, respectively). The average subsurface warming between 100 and 500 m depth derived from such equilibrium runs is in the order of 0.5°C per 0.1 Sv with peak warmings around and even over 1°C at about 350 to 850 m depth in some parts of the North Atlantic (Figure 6d). The response time of the subsurface warming to the freshwater forcing is in the order of 80 years. The subsurface warming results from a reduced ocean heat loss to the atmosphere due to the stronger stratification of the upper ocean, which is caused by the freshwater lens.

[24] The subsurface warming can also be seen in model runs with freshwater input into the northern North Atlantic east of Greenland only (Figures 6e and 6f). The average equilibrium warming between 100 and 500 m depth in the Nordic Seas is larger in this case (0.9°C per 0.1 Sv) with a peak warming of 1.3°C around 500 m depth. It results again from a reduction in the ocean to atmosphere heat flux because of the freshwater lens. Subsurface warming can even be seen outside the region where the freshwater forcing is applied. The warmed water masses of the Nordic Seas get advected into the Labrador Sea, leading to an average equilibrium subsurface warming (100 to 500 m depth) of 0.5°C per 0.1 Sv and a peak warming of over 1.4°C at a depth of about 600 m in the Labrador Sea. The subsurface warming is also propagated with the subpolar gyre. Ice surges into the Nordic Seas at the beginning of H events therefore could induce a local subsurface warming and have a far field impact on subsurface temperatures in the Labrador Sea. This in turn could lead to the destabilization of additional ice shelves and ice streams not only around the Nordic Seas but also around the Labrador Sea and therefore could help to trigger the large ice surges from the Laurentide ice sheet through the Hudson Strait into the Labrador Sea [*Bond and Lotti*, 1995; *Hemming*, 2004].

[25] The subsurface warming due to freshwater input into the North Atlantic or the northern North Atlantic east of Greenland does not counterbalance the cooling caused by the MOC collapse in most North Atlantic regions except for model runs with quite a large freshwater input. This finding, however, does not exclude subsurface warming as an important mechanism for destabilizing ice shelves and marine-based ice streams. Proxy data suggest that a substantial MOC weakening or even collapse happens up to several centuries before the H event takes place [*Zahn et al.*, 1997; *McManus et al.*, 2004; Hall et al., submitted manuscript, 2006]. The ice shelves and ice streams adapt to the colder subsurface temperatures after the MOC weakening and advance into the sea [*Hulbe*, 1997; *Alley et al.*, 2006]. The new state of the ice is then probably more vulnerable to perturbations and a subsurface warming at that time could lead to the disintegration of the shelves and ice streams [*Hulbe*, 1997; *Rignot and Jacobs*, 2002; *Payne et al.*, 2004; *Alley et al.*, 2006].

[26] The proxy evidence for subsurface warming during H events currently remains controversial. Variations in benthic $\delta^{18}O$ data can be interpreted as a change in subsurface temperature as well as in deepwater formation by brine rejection and are often interpreted as the latter [*Dokken and Jansen*, 1999; *van Kreveld et al.*, 2000]. However, there are indications that the low benthic $\delta^{18}O$ signals associated with H events in the northern North Atlantic represent a warming of the deep water masses rather than deepwater formation by brine rejection [*Bauch and Bauch*, 2001; *Moros et al.*, 2002; *Rasmussen and Thomsen*, 2004].

3.3. Non-Heinrich Stadials

[27] Non-Heinrich stadials are characterized by ice calving around the Nordic Seas only [*Bond and Lotti*, 1995; *Elliot et al.*, 1998; *Dokken and Jansen*, 1999; *van Kreveld et al.*, 2000], by a weakening of the MOC (in contrast to a complete shut down during H event stadials) [*Ganopolski and Rahmstorf*, 2001; *Elliot et al.*, 2002] and a much smaller sea level rise than during H events [*Shackleton et al.*, 2000; *Siddall et al.*, 2003]. Are the mechanisms proposed in this paper able to explain these findings? Our model simulations suggest that the dynamic topography in the North Atlantic and the global steric sea level increase close to linearly for a reduction of the MOC to about 5 Sv, with a substantial larger, nonlinear increase below 5 Sv. Overall, the sea level rise in the North Atlantic due to a MOC weakening is smaller than due to a complete MOC collapse but could possibly still be large enough to trigger ice calving of one or more ice sheets around the Nordic Seas and initiate at least some of the described feedback mechanisms. The eustatic sea level rise, however, is small during non-Heinrich stadials [*Shackleton et al.*, 2000; *Siddall et al.*, 2003]. The resulting perturbations on the Laurentide ice sheet remain too small to induce large ice surges from the Laurentide ice sheet, either because the subsurface warming and the sea level rise are too small or because of the larger stability of the Laurentide at these times than during H event stadials at the end of a Bond cycle [*Bond et al.*, 1993; *MacAyeal*, 1993]. We conclude that our hypothesis does not contradict the findings from paleodata characterizing non-Heinrich stadials, but further work is necessary to better understand the behavior of the climate system during non-Heinrich stadials.

4. Summary and Conclusion

[28] We have discussed processes associated with local and global sea level rise as well as subsurface temperature changes by which the ocean can interact with ice shelves and marine-based ice streams. In Figure 7 we show a possible sequence of events during a Heinrich stadial to illustrate how the suggested processes potentially link different components of the climate system (the following numbers in the text refer to Figure 7). The sequence starts with (1) a vigorous reduction of the MOC or a MOC collapse at the end of the last D-O event of a Bond cycle, initiating the cold stadial phase. This causes (2) global

(steric) and local (dynamic) sea level rise in the North Atlantic, which triggers (3) the destabilization of a few ice shelves and ice streams grounded on the seafloor, and causes some ice streams to surge. The freshwater input into the northern North Atlantic from the first destabilized ice sheets induces (4) additional (mainly eustatic) sea level rise as well as (5) a subsurface warming in the North Atlantic. Both effects act as strong positive feedback mechanisms and have the potential to destabilize additional ice shelves and ice streams. The stress on the ice shelves and marine-based ice streams from the Laurentide increase over time and finally (6) and (7) trigger the large surge from the Laurentide ice sheet. The feedback loops break down and (8) the ice surges come to an end when the ice sheets reach a new equilibrium and the amount of ice calving decreases [*MacAyeal*, 1993]. Subsequently, (9) the MOC resumes because of the strongly reduced freshwater input and the temperature in the Northern Hemisphere increases abruptly [*Mikolajewicz and Maier-Reimer*, 1994; *Manabe and Stouffer*, 1997; *Marchal et al.*, 1998; *Schmittner et al.*, 2002], marking the start of the first large D-O event of a Bond cycle. Sea level starts falling slowly as the ice sheets are starting to grow again.

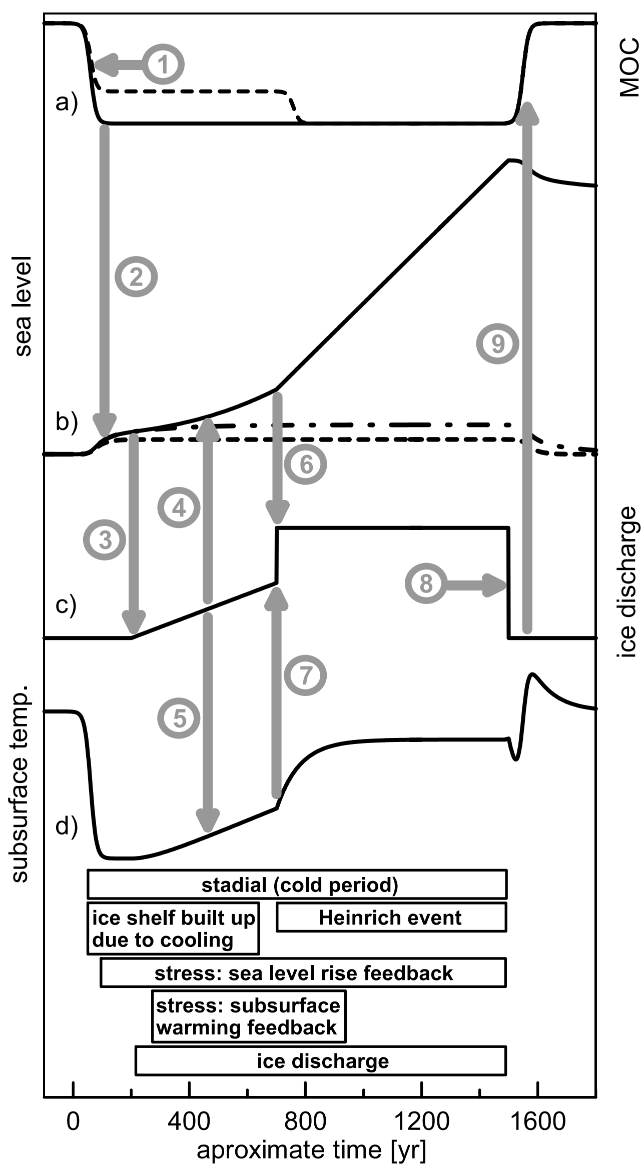

Figure 7. Sketch of a Heinrich event stadial according to our hypothesis. (a) Two possible pathways of the North Atlantic meridional overturning circulation, showing a full collapse at the end of the last Dansgaard-Oeschger event of a Bond cycle (solid line) and a substantial decrease followed by a full collapse later on (dashed line). (b) Sea level anomalies due to changes of the dynamic topography in the North Atlantic (dashed line), North Atlantic dynamic topography plus global steric sea level (dash-dotted line), North Atlantic dynamic topography plus global steric and eustatic sea level (solid line). (c) Ice discharge of the Northern Hemisphere ice sheets. (d) Subsurface temperature in the Nordic Seas and the Labrador Sea. The boxes at the bottom of Figure 7 indicate time periods of the stadial cold period, of ice shelf built up due to the cold ocean surface and subsurface temperatures, of increasing stress on the ice shelves and marine-based ice streams due to sea level rise and subsurface warming, of ice discharge in the Northern Hemisphere and of the Heinrich event itself with ice calving from the Laurentide. Shaded arrows indicate triggers and positive feedback mechanisms, which are part of our hypothesis. The step by step description of events in our hypothesis (corresponding to numbers 1 to 9) is found in the text in section 4.

[29] Our hypothesis is not in contrast, but rather complements other hypotheses related to ice sheet instabilities and surge mechanisms [*Hulbe*, 1997; *Moros et al.*, 2002; *Hulbe et al.*, 2004; *Alley et al.*, 2006]. A combination with the theory that ice sheets around the Nordic Seas have triggered H events [*van Kreveld et al.*, 2000] is for example conceivable, too. Ice surges caused by ice sheet internal processes [*MacAyeal*, 1993] of one or more ice sheets around the Nordic Seas could be responsible for the MOC weakening and subsequent shut down, which in turn initiates the trigger and feedback mechanisms discussed here. The same could be true for early ice surges in the Lancaster Strait and Baffin Island region [*Papa et al.*, 2006]. We conclude that a complete theory or model of the interaction between atmosphere, ocean, sea ice, ice shelves and ice sheets during glacial climate variations will have to incorporate regional surface temperature, ocean subsurface temperature and sea level changes as driving forces for the ice shelves and ice sheets.

[30] This work is based on model simulations only and is a plausible scenario given today's knowledge about H events from paleodata and modeling efforts. However, it will be important to test the hypothesis with additional and more detailed paleodata that will arise in the future. The following catalogue gives four suggestions for how to test the model predictions of our hypothesis. (1) Of high interest is the exact timing of the MOC collapse and the start of the main freshwater input into the North Atlantic. With our hypothesis we expect the MOC collapse (or at least a strong MOC weakening) to precede the large meltwater input during H events. (2) To get a better understanding of the evolution of the freshwater flux over time an accurately

dated high-resolution sea level record over the last glacial period is of great interest and will help to shed additional light on the nature of H events. Such a record would be the best proxy for the timing and the amount of freshwater release into the global ocean, although it would still not resolve the contributions from the different ice sheets. Nevertheless, it will be an interesting test whether or not the sea level record is in contradiction to our theory. (3) Another central question is whether a subsurface warming as our model predicts for the North Atlantic can be confirmed with paleodata for the time period of H events after the severe cooling related to the MOC collapse. (4) Further, our mechanism for H events requires a strong direct coupling of the ocean and the Northern Hemisphere ice sheets. This is only the case for marine-based ice streams and ice shelves. The reconstruction of their abundance and extent as well as their importance to buttress ice streams is therefore of great interest to test our hypothesis.

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J. Flückiger and J. W. C. White, Institute of Arctic and Alpine Research, University of Colorado, Campus Box 450, Boulder, CO 80309, USA. (flueckiger@colorado.edu)

R. Knutti, Climate and Global Dynamics, National Center for Atmospheric Research, P.O. Box 3000, Boulder, CO 80307, USA.