Deconstructing the Last Glacial Termination: the role of millennial and orbital-scale forcings

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Abstract

Using an Earth system model of intermediate complexity forced by continuously varying boundary conditions and a hypothetical profile of freshwater forcing, the model simulates Heinrich event 1 (H1), the Bølling warm period, the Older Dryas, the Antarctic Cold Reversal (ACR) and the Younger Dryas in close agreement with paleo-proxy data from different regions worldwide. The ACR can be simulated as the bipolar seesaw response to the AMOC recovery during the termination of H1. However, this study also demonstrates that the amplitude of the ACR can be further amplified by a rapid deglacial retreat of the Antarctic icesheets. We suggest that melting from both, the Laurentide and the Antarctic Ice sheets contributed to the sea level rise associated with Meltwater Pulse 1-A (MWP-1A). It is hypothesized that the northern hemispheric source of MWP-1A caused the Older Dryas cooling in the Northern Hemisphere, whereas the Southern Hemispheric source contributed to the ACR. The study also documents that for the majority of paleo-climate proxies considered here, the relative timing can be qual-
itatively reproduced by the transient modeling experiments. The climate model solution presented here may provide a means to further constrain dating uncertainties of some of paleo-climate proxies during the Last Glacial Termination.

Keywords: Last glacial termination, Heinrich event 1, Bølling warming, Older Dryas, Antarctic Cold Reversal, Younger Dryas, MWP-1A, Earth system model of Intermediate complexity

1. Introduction

Elucidating the driving mechanisms of glacial cycles requires both an accurate knowledge of the timing and magnitude of external forcings and an understanding of internal climate feedbacks that determine the overall climate response to these forcings. In addition, paleo-climate reconstructions that can be used to independently test the hypothesized mechanisms are essential for ascertaining our comprehension of these changes.

Paleo-modeling efforts have played an important role in quantifying the climate response to particular glacial-scale forcings. Using coupled atmosphere-ocean models, these efforts have mostly focused on particular time-slices, such as the Last Glacial Maximum (LGM, 21 ka B.P.) (Bush and Philander, 1998; Liu et al., 2000; Hewitt et al., 2001; Kitoh et al., 2001; Liu et al., 2002; Kim et al., 2003; Shin et al., 2003; Timmermann et al., 2004; Braconnot et al., 2007) or the Mid-Holocene Optimum (9-6 ka B.P.) (Joussaume et al., 1999; Otto-Bliesner, 1999; Liu et al., 2003). Results from these timeslice experiments have been compared thoroughly with existing paleo-reconstructions, such as CLIMAP-Project-Members (1981), GLAMAP (Pflaumann et al.,
and other proxies (see e.g. Shin et al. (2003) for a careful LGM model-data comparison). This modeling approach has helped to explore the spatial characteristics of past glacial climate change and understand the relative contributions of different external forcing factors (Rind, 1987; Timmermann et al., 2004; Justino et al., 2005).

However, in order to understand the physical mechanisms responsible for abrupt climate change and to identify the existence of leads, lags and phase-relationships in the climate system, it is mandatory to simulate the temporal evolution of the system in response to the time-varying forcing factors. Recently the transient modeling framework has been applied to climate models to understand the Late Quaternary climate evolution and its relation to slowly varying boundary conditions (Jackson and Broccoli, 2003; Felis et al., 2004; Lorenz and Lohmann, 2004; Tuenter et al., 2005; Renssen et al., 2005; Charbit et al., 2005; Lorenz et al., 2006; Lunt et al., 2006; Marsh et al., 2006; Timmermann et al., 2007; Timm and Timmermann, 2007; Liu et al., 2009; Timmermann et al., 2009a; Timmermann and Menviel, 2009). We recently presented a transient model solution for the Last Glacial Termination using the Earth system model of intermediate complexity LOVECLIM that was driven by changes in greenhouse gases, ice-sheet topography and albedo and orbitally-induced changes in solar radiation (Timm and Timmermann, 2007). Millennial-scale freshwater forcing was however not considered. More recently, Liu et al. (2009) performed a transient simulation for the period 22 to 15 ka B.P. using a state of the art CGCM (NCAR CCSM3) that also captured Heinrich event 1 (H1) and the transition to the Bolling warm period (Timmermann and Menviel, 2009).
The meltdown of the Laurentide and Eurasian ice-sheets started around 20 ka B.P. An acceleration occurred during H1 (17.8–15.2 ka B.P.), as massive icebergs broke off from the disintegrating ice-sheets. The resulting discharge of freshwater into the North Atlantic (Vidal et al., 1997) reduced surface density and eventually halted the formation of North Atlantic Deep Water for about 2,500 years, as suggested by the $^{231}$Pa/$^{230}$Th and ventilation age data from North Atlantic sediment cores (McManus et al., 2004; Gherardi et al., 2005; Thornalley et al., 2011) (figure 1a). At the same time, temperatures over Greenland remained low (Alley, 2000) while temperatures over Antarctica started to rise (Jouzel et al., 2007) and atmospheric CO$_2$ concentration increased from 190 to 220 ppmv (Monnin et al., 2001). The end of H1 is marked by an abrupt warming over Greenland and the North Atlantic starting around 14.7 ka B.P. Liu et al. (2009) suggested that this Bølling warming was due to the rapid resumption of the AMOC following H1, combined with the glacial-interglacial increase in atmospheric CO$_2$. Subsequent to the Bølling warming, between 14.7 and 13 ka B.P., the temperature over Greenland showed some pronounced fluctuations of about 5°C within 400-500 years. Two major cold events were identified during this period in the Norwegian Sea (Lehman and Keigwin, 1992) and in Greenland (Stuiver et al., 1995): the Older Dryas (~14 ka B.P.) and the intra-Allerød cold period (13.5–13 ka B.P.) (figure 2a).

During the same time, a 2°C surface cooling, known as the Antarctic Cold Reversal (ACR) is reconstructed for Antarctica (Petit et al., 1999; Jouzel et al., 2007). Concurrent to the ACR, sea level rose by about 20 m in about 500 years (Fairbanks, 1989; Stanford et al., 2006). The origin of this rapid
sea level rise, referred to as Meltwater Pulse 1A (MWP-1A), has been the subject of an intense debate. While Licht (2004) and Peltier (2005) propose a Northern Hemispheric origin, Kanfoush et al. (2000); Clark et al. (2002); Bassett et al. (2005) and Carlson (2009) advocate the idea of a Southern Hemispheric origin of MWP-1A. Evidence for a meltwater pulse originating from Antarctica at around 14.6 ka B.P. was found in several marine sediment cores of the Southern Ocean. Ice-Rafted Debris (IRD) layers were observed in sediment cores from the Southeast Atlantic (Kanfoush et al., 2000), the Southwest Pacific (Carter et al., 2002) and from around the Antarctic Peninsula (Noumi et al., 2007). Freshwater discharge in the Southern Ocean is also inferred from depleted oxygen values in planktic foraminifera as well as increased occurrence of marine diatoms (Chaetoceros spp) in a sediment core from the Atlantic sector of the Southern Ocean (Bianchi and Gersonde, 2004). Recent studies by Clark et al. (2002) and Bassett et al. (2005) further suggest that in order to reconcile coral-based sea level reconstructions during MWP-1A with predicted glacial isostatic adjustment, one has to invoke a significant melting of Antarctica. Bassett et al. (2005) estimated that the Antarctic ice-sheet could have contributed by up to about 15 m sea level rise during MWP-1A. The main potential source regions for this pulse in Antarctica were identified as the Weddell Sea (~9 m) and the Ross Sea (~5 m). Finally, recent modeling studies (Philippon et al., 2006; Pollard and DeConto, 2009) suggest that the accelerated melting of the Antarctic ice sheet started around 15 ka, further supporting the notion of significant deglacial freshwater forcing of the Southern Ocean during the ACR. The ACR could be explained in terms of the bipolar seesaw response to the AMOC recovery.
during the Bølling warming (Stocker, 1998), but there is also the possibility
that a near-synchronous rapid melting of the Antarctic ice sheet may have
contributed to the Southern Hemispheric cooling (Menviel et al., 2010a).

The Younger Dryas (YD) transition to colder Northern Hemispheric cli-
mates occurred around 12.9 ka B.P. (Rasmussen et al., 2006). During this
period temperatures over Greenland, the eastern North Atlantic and the
Caribbean Sea dropped by 8°C, 2°C and 4°C, respectively (figure 2). It has
been hypothesized that the YD was caused by a weakening of the AMOC
(McManus et al., 2004) due to an outburst of the late glacial Lake Agassiz
into the Arctic Ocean (Tarasov and Peltier, 2005; Murton et al., 2010). Its
effect on the Southern Hemisphere still remains somewhat elusive.

Our goal is to simulate the Last Glacial Termination using an Earth
system model of intermediate complexity, LOVECLIM, and to address the
following questions:

- What forcings are needed to reproduce the dominant climate and bio-
geochemical features of the Last Glacial Termination?
- What are the global impacts of H1, MWP-1A and the YD?
- Are climate data consistent with the notion of a Southern Hemispheric
  origin of MWP-1A?
- Is it justified to view the North Atlantic as the main driver of millennial-
scale variability during the Last Glacial Termination?
- Is the relative timing of events in paleo-proxy data from different re-
gions consistent with the relative timing in the model?
A suite of transient climate modeling experiments is conducted for the period 18-11 ka B.P. using time-varying radiative forcing due to orbitally-induced solar radiation changes and greenhouse gases. The effects of waning glacial ice-sheets on topography and albedo are captured by updating the atmospheric boundary conditions continuously with estimates of the icesheet evolution and albedo. Millennial-scale variability is produced by varying the strength of the deep water formation in the North Atlantic and in the Southern Ocean through the input of idealized freshwater pulses.

The paper is organized as follows: in section 2 we provide a description of the experimental setup and the methods employed to analyze the paleo-proxy data. Section 3 is devoted to an extensive comparison between model simulation and paleo-climate proxies covering the period 18-11 ka B.P. The paper concludes with a general discussion and a summary of the main results.

2. Model and experimental setup

2.1. The Earth-system model LOVECLIM

The model used in this study is the model of intermediate complexity LOVECLIM (Driesschaert et al., 2007; Goosse et al., 2007)\(^1\). Our experiments were conducted with version 1.1. of this model. The atmospheric component of the coupled model LOVECLIM is ECBilt (Opsteegh et al., 1998), a spectral T21, three-level model, based on quasi-geostrophic equations extended by estimates of the neglected ageostrophic terms (Lim et al., 1991) in order to close the equations at the equator. The coupling between dynamics

\(^1\)see also Goosse et al. (2010a) for the most recent LOVECLIM version 1.2
and thermodynamics is done via a linear balance equation. The sea-ice-ocean component of LOVECLIM, CLIO (Goosse et al., 1999; Goosse and Fichefet, 1999; Campin and Goosse, 1999) consists of a free-surface primitive equation model with 3°x3° resolution, coupled to a thermodynamic-dynamic sea-ice model. Coupling between atmosphere and ocean is done via the exchange of freshwater, momentum and heat fluxes. The terrestrial vegetation module of LOVECLIM, VECODE, (Brovkin et al., 1997) simulates the dynamical vegetation changes and the terrestrial carbon cycle in response to climatic conditions.

LOCH is a 3-dimensional global model of the oceanic carbon cycle (Mouchet and Francois, 1996; Menviel et al., 2008b; Goosse et al., 2010a). The prognostic state variables considered in the model are dissolved inorganic carbon (DIC), total alkalinity, phosphates (PO$_{4}^{3-}$), organic products, oxygen and silica. LOCH is fully coupled to CLIO, with the same time step. In addition to their biogeochemical transformations tracers in LOCH experience the circulation field and diffusion predicted by CLIO. LOCH computes the export production from the fate of a phytoplankton pool in the euphotic zone (0–120 m). The phytoplankton growth depends on the availability of nutrients (PO$_{4}^{3-}$) and light, with a weak temperature dependence. A grazing process together with natural mortality limit the primary producers biomass and provide the source term for the organic matter sinking to depth. The other processes described in the model include remineralization, carbonate precipitation and dissolution as well as opal production. LOCH does not include sedimentary processes but nevertheless takes into account carbonate compensation mechanisms. The atmospheric CO$_{2}$ content is updated for
each ocean time step. For our experiments, however, the atmospheric CO$_2$
was prescribed.

The coupled model has been used extensively to study different aspects of
Late Quaternary climate (Timm and Timmermann, 2007; Timm et al., 2008;
Menviel et al., 2008b,a; Timmermann et al., 2009a; Timm et al., 2010).

2.2. Experimental setup

In this study the goal is to simulate the sequence of millennial-scale events
superimposed on the slowly-varying orbital-scale background conditions dur-
ing the last deglaciation. For this we perform two transient experiments start-
ing from the background conditions of the Last Glacial Maximum (LGM)
(Timm and Timmermann, 2007; Menviel et al., 2008b). The model is then
forced for the period 21 ka to 10 ka B.P. with the time-varying evolution of so-
lar insolation (Berger, 1978), ice-sheet topography (updated every 100 years)
(Peltier, 1994), high latitude albedo and atmospheric CO$_2$. The changes in
atmospheric CO$_2$ are obtained from the EPICA Dome C ice core, Antarctica
(Monnin et al., 2001) on the EDC3 timescale (Lüthi et al., 2008). Other
greenhouse gases (CH$_4$ and N$_2$O) are kept fixed at LGM values.

In experiment DG$_{NS}$, a series of freshwater pulses is applied in both the
North Atlantic and the Southern Ocean in order to reproduce milennial-
scale events (figure 1d). The timing and amplitude of the freshwater pulses
were set so as to provide a good match with the $^{231}$Pa/$^{230}$Th data from
the North Atlantic (McManus et al., 2004) and temperature variations in
Greenland (Alley, 2000) on the GISP2 timescale, which is within dating un-
certainties of the GICC05 timescale (Rasmussen et al., 2006). Even though
a significant amount of freshwater was probably discharged into the Gulf
of Mexico during the last deglaciation (Aharon, 2003; Flower et al., 2004; Tarasov and Peltier, 2006; Sionneau et al., 2010), with potential implications for the AMOC (Roche et al., 2007), we do not add freshwater into this region. Given the uncertainties associated with the detailed freshwater routing during the Last Glacial Termination we choose a more idealized approach. To mimic H1, Older Dryas and Younger Dryas in our model, freshwater is released into a broad region in the North Atlantic centered between 55°W–10°W, 50°N–65°N and the Arctic Ocean, respectively. The amplitude of the millennial-scale freshwater forcing amounts to 0.2 Sv during the period 18 ka to 17.4 ka B.P. and 0.25 Sv between 17.4 ka B.P and 15.6 ka B.P. To specifically represent the Older Dryas we apply a freshwater pulse in the North Atlantic for the period 14.4 ka B.P. to 14 ka B.P. with a maximum input of 0.25 Sv between 14.2 and 14 ka B.P. The exact shape of the freshwater pulse is depicted in figure 1d. A background freshwater flux was also added in the North Atlantic between 14 and 13 ka B.P. to represent the melting northern hemisphere icesheets. To mimic the Younger Dryas, 0.25 Sv is applied in the Arctic Ocean (175°W–95°W, 67°N–83°N) between 13 and 12.2 ka B.P. Finally, to represent the ACR, an additional meltwater pulse is prescribed for the area enclosing the Ross and Weddell Seas (163°E–11°E, 70°S-80°S). The magnitude of this pulse is 0.15 Sv for the period 14.4 ka to 13.4 ka B.P., after which the pulse linearly decreases to 0 over 1000 years (until 12.4 ka B.P.).

To better evaluate the impact of Southern hemispheric meltwater pulses, another experiment similar to \( DG_{NS} \) (\( DG_N \)) was conducted. In \( DG_N \), no meltwater pulse is added in the Southern Ocean between 14.4 ka and 12.4 ka.
The results of experiment $\text{DG}_{NS}$ are compared to a variety of paleo-proxy records (figure 3 and table 1), which represent variations in oceanic circulation, temperature, precipitation and biogeochemical variables in different regions of the world. To obtain a more representative comparison of model outputs with paleo-proxy records, when possible, composite timeseries were generated. These composite timeseries were produced by first interpolating each paleo-proxy data onto an equidistant grid in time using bicubic splines and then averaging the different paleo-proxy records. In the model-paleo data comparison conducted here, we allowed for some dating uncertainties in the proxy records. Optimal match between the local proxy timeseries and the simulated physical variable was sometimes achieved by shifting the proxy timeseries by a few hundred years (black arrows in timeseries figures, table 2). The shift that we had to apply to obtain a good match between model solution and paleo-data was smaller than the respective published absolute dating uncertainties. The time series shifting depends on the timing of the freshwater forcing sequences prescribed in the transient simulations.

### 3. The deglacial history in the Northern Hemisphere

#### 3.1. Heinrich event 1

During H1, and as a result of the chosen freshwater forcing the simulated AMOC collapses (figure 1a), in agreement with the $^{231}\text{Pa}/^{230}\text{Th}$ data.

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$^2$Unlike some of the experiments described in Timm and Timmermann (2007), experiments $\text{DG}_{NS}$ and $\text{DG}_N$ do not employ any orbital acceleration (Lorenz and Lohmann, 2004).
from the Bermuda rise core OCE326-GGC5 (McManus et al., 2004). Figure 1a also displays the $^{231}\text{Pa}/^{230}\text{Th}$ data from the Iberian margin core SU81-18 (Gherardi et al., 2005) and the ventilation age as inferred from South Iceland rise sediment cores (Thornalley et al., 2011). Even if all the records show a weakened AMOC at some time during H1, there are some substantial differences among them, which will further be discussed in section 5. The AMOC weakening reduces the poleward Atlantic heat transport at 40°N by 0.5 PW (not shown), which leads to an enhanced Northern Hemispheric cooling relative to the LGM (figure 2 and figure 4 left), in accordance with temperature paleo-proxy data from Greenland (Alley, 2000) (figure 2a), the North Atlantic (Bard et al., 2000; Calvo et al., 2001; Waelbroeck et al., 2001; Bard, 2002; Martrat et al., 2004, 2007) (figure 2c), the tropical North Atlantic (Lea et al., 2003; Niedermeyer et al., 2009) (figure 2d), the Mediterranean Sea (Cacho et al., 1999, 2001), Europe (Ivy-Ochs et al., 2006), Asia (Wang et al., 2002; Evans et al., 2003) and Africa (Powers et al., 2005). The enhanced meridional temperature gradient in the North Atlantic induces a strengthening of the northeasterly trades and a southward shift of the Intertropical Convergence Zone (ITCZ) (Timmermann et al., 2005) (figure 4, middle). In agreement with paleoproxies, drier conditions prevailed during H1 in Europe (Peyron et al., 2005), the northern part of Africa (Peck et al., 2004; Lamb et al., 2007; Williams et al., 2006; Niedermeyer et al., 2009), the Middle East (Bar-Matthews et al., 2003; Vaks et al., 2003; Shakun et al., 2007; Fleitmann et al., 2009) and the northern part of South America. As shown in figure 5a, we simulate dry conditions in the Cariaco basin as soon as the AMOC weakens. The reflectance record from the Cariaco sediments (Peterson et al., 2000)
indicates relatively dry conditions between 18 and 16.2 ka B.P., however the drying becomes more intense at 16.2 ka B.P. The reason behind this feature remains unexplained. In the Gulf of Guinea, the southward ITCZ shift leads to drier conditions in agreement with Ba/Ca data indicating greater surface salinity during H1 in that area (Weldeab et al., 2007). The AMOC shut down also leads to a reduced ratio of summer/winter precipitation over China in agreement with δ¹⁸O records from Chinese caves (Wang et al., 2001; Dykoski et al., 2005) (figure 5c). Furthermore, our model solution supports the notion of a reduced summer monsoon over the Arabian Sea (Ivanochko et al., 2005) (figure 5d).

As a result of the simulated southward shift of the ITCZ (figure 4, middle), the reduced Atlantic-Pacific freshwater export (not shown) and the reduced Asian summer monsoon (figure 5c), North Pacific salinity increases at the onset of H1, as described in Okazaki et al. (2010). North Pacific waters become dense enough to generate North Pacific Deep Water (NPDW) formation in the Bering Sea and the Gulf of Alaska. Subsequently a deep Pacific Meridional Overturning Cell (PMOC) establishes. Maximum meridional transports in this cell attain values of up to 20 Sv. Southward spreading of well-ventilated NPDW is in accordance with a compilation of radiocarbon data obtained from marine sediment cores that suggest a drop of eastern North Pacific ventilation ages by ~1,000 years (figure 1b) (Okazaki et al., 2010). As can be seen in figure 4 left, the formation of NPDW leads to a reorganization of surface currents in the Indo-Pacific: the Indonesian Throughflow transport decreases (~50%) and the warm and saline water of the subtropical Pacific are diverted into the northern North Pacific, supplying the sinking
branch of the PMOC (Chikamoto et al., 2010; Menviel et al., 2010b; Okazaki et al., 2010). A slowdown of the Indonesian Throughflow during Heinrich events has also been inferred from a surface and upper thermocline temperature record in the Timor Sea (Zuraida et al., 2009), providing support to our analysis. The formation of NPDW leads to an increase in poleward heat transport in the North Pacific of 0.5 PW at 40°N. In experiment $DG_{NS}$, the enhanced transport of warm equatorial waters to the North Pacific combined with greater insolation and atmospheric CO$_2$ lead to a slight warming of the ocean surface in the North Pacific during H1 (16–18 ka B.P.) (figure 4, left). In agreement with SST paleoproxies, a warming is observed in the northeastern Pacific where warm waters converge (Sarnthein et al., 2006), while a relatively small area around Japan cools (Harada et al., 2006, 2010) (figure 4, left). At the beginning of H1, the ITCZ shifts southward in the Pacific, however as NPDW formation enhances, the ITCZ shifts back to its initial position in the Pacific basin (figure 4, middle).

In the North Pacific, the formation of NPDW leads to an increase in the dissolved oxygen content (by up to 3 ml/L) in the upper 2500 m, in agreement with benthic foraminiferal assemblages in marine sediment cores from the Northwest Pacific (Shibahara et al., 2007). The well-oxygenated waters flow equatorwards from the North Pacific. The main branch flows as a deep western boundary current but well ventilated waters are also advected along the California basin. The dissolved oxygen content along the coast of California increases by about 50% down to about 2000 m depth. This is in qualitative agreement with redox sensitive metal content, $\delta^{15}$N and total organic carbon records from marine sediment core ODP 893A (Kennett and
Ingram, 1995; Emmer and Thunell, 2000; Ivanochko and Pedersen, 2004) as well as with \( \delta^{15}N \) record from marine sediment core JPC-56 (Pride et al., 1999) (figure 1c). Despite the fact that changes in \( \delta^{15}N \) can have numerous causes (Somes et al., 2010), the redox sensitive metal content and organic carbon records from the California basin support our hypothesis of greater ventilation during H1.

Due to the colder conditions and increased stratification, simulated marine export production is low in an area spanning the North Atlantic, the Atlantic coast of North and Central America as well as the Caribbean Sea, in agreement with paleo-proxy reconstructions (Peterson et al., 2000; Vink et al., 2001) (figure 4, right). Due to the stronger upwelling, marine export production significantly increases (+25%) in the Equatorial Atlantic (Northwest African coast and Brazilian coast) in agreement with paleo-proxy records from the Brazilian coast (Jennerjahn et al., 2004).

3.2. Bolling Warming, Older Dryas and Intra-Allerød cold period

At the end of H1 (~15.2 ka B.P.), the simulated AMOC recovers quickly regaining a strength of about 27 Sv around 14.6 ka B.P (figure 1a). As can be seen in figures 2 and 6 (left), the temperatures in Greenland and in the North Atlantic increase abruptly during the recovery phase of the AMOC (15 to 14.6 ka B.P.), until reaching a maximum during the Bolling Warm Period (BWP). As already discussed by Liu et al. (2009), the rapid transition to the BWP can be explained by a combination of a rapid resumption of the AMOC following H1, an overshooting effect of the AMOC, as well as by the contemporaneous increase in atmospheric CO\(_2\) concentrations (Köhler et al., 2010). The amplitude of the simulated increase in spring temperature over
Greenland between 15.2 ka B.P. and 14.5 ka B.P. amounts to about 12°C, which is similar to the 13°C temperature difference estimated from GISP2 ice core for the same time period (Alley, 2000) (figure 2a). Off the Iberian margin the simulated SST and alkenone-based SST (Bard et al., 2000; Bard, 2002; Martrat et al., 2004, 2007) both increase between H1 and BWP by about 3–4°C (figure 2c). In the Cariaco basin however, the simulated 1.5°C SST increase during the winter season is only half of that reconstructed from Mg/Ca data in core PL07-39PC (Lea et al., 2003) (figure 2d). The resumption of the AMOC leads to a reorganization of surface ocean currents and to the decline in the formation of NPDW.

Such an intense and rapid warming in the North Atlantic at about 15 ka B.P. is likely to have consequences on the massive ice sheets that are still present over North America and northern Europe. During this period melting of the Laurentide and Fennoscandian ice sheets accelerated (Peltier, 1994; Thornalley et al., 2010; Gregoire, 2010) thus adding freshwater into the North Atlantic. In response to this freshwater perturbation the simulated AMOC weakens from 27 Sv at the peak of the Bølling warming down to about 12 Sv during the Older Dryas (∼14 ka B.P.) and levels off at about 18 Sv during the Allerød (between 13.8 and 13 ka B.P.) (figure 1a). These variations in simulated AMOC strength are in overall agreement with the ventilation age history recorded in the South Iceland rise (Thornalley et al., 2011). The significant freshwater-induced reduction in AMOC strength during the Older Dryas is not observed in the $^{231}$Pa/$^{230}$Th records (McManus et al., 2004; Gherardi et al., 2005). Higher resolution $^{231}$Pa/$^{230}$Th records for the period 14.2 and 13.5 ka B.P would be needed to test this assertion. Due to the
intermittent AMOC weakening at 14 ka B.P., modeled air temperatures over Greenland drop by about 6°C, summer air temperature decreases by 1.8°C in the Netherlands, SST decrease by about 1.2°C off the Iberian margin and by about 0.7°C in the Cariaco basin (figure 2 and 7, left). This simulated cooling is in very good agreement with temperature estimates from Greenland ice core (Alley, 2000) and chironomid-inferred July temperatures from the Netherlands (Heiri et al., 2007). However the cooling in the Cariaco basin SST reconstruction (Lea et al., 2003) is lower than in the model results and no significant cooling is observed in marine sediment cores off the Iberian margin (Bard et al., 2000; Bard, 2002; Martrat et al., 2004, 2007).

Over northern North America, the waning of the ice sheet and the resulting changes in height induce a local surface warming due to the lapse-rate effect, even during the Older Dryas. At about 13.8 ka B.P. the simulated temperatures in the North and Equatorial Atlantic recover to levels equivalent to the Bølling warming, in agreement with SST proxies from that region.

Greenland ice cores also record another millennial-scale cold event, known as the Intra-Allerød cold period (IACP, 13.5-13 ka B.P., figure 2a) (Stuiver et al., 1995), which is not reproduced in our transient modeling experiments. This millennial-scale cold event was also observed in the Norwegian Sea (Lehman and Keigwin, 1992), on the South Iceland Rise (Thornalley et al., 2010) as well as in terrestrial records from Northern Europe (von Grafenstein et al., 1999; Heiri et al., 2007) (figure 2b) and North America (Yu and Eicher, 2001). In addition ventilation age anomalies indicate decreased ventilation south of Iceland during that period (Thornalley et al., 2011). However no significant decrease in North Atlantic SST was reconstructed for that time
period according to higher resolution SST proxies (Cacho et al., 1999; Bard et al., 2000; Cacho et al., 2001; Calvo et al., 2001; Bard, 2002; Lea et al., 2003; Martrat et al., 2004, 2007). Given these discrepancies between the proxy-based spatial extent of the IACP and the model simulation it thus remains unresolved whether large-scale freshwater forcing was the sole cause for these variations.

During the Bølling warm period, the ITCZ swings back to the north, leading to wetter conditions over Europe (Peyron et al., 2005), the northern part of South America (Peterson et al., 2000), North Africa (Peck et al., 2004; Williams et al., 2006; Weldeab et al., 2007; Niedermeyer et al., 2009), the middle East (Bar-Matthews et al., 2003; Vaks et al., 2003; Ivanochko et al., 2005; Shakun et al., 2007; Fleitmann et al., 2009), China (Wang et al., 2001; Dykoski et al., 2005) and Tibet (Kramer et al., 2010) (figures 6, middle and 5).

Due to the intense warming and retreat of the sea-ice edge in the North Atlantic the marine export production more than doubles in this region (figure 6, right). It also increases considerably in the Caribbean Sea (Peterson et al., 2000; Vink et al., 2001). Marine export production however decreases in the North Pacific (-5%) due to the reduction of the NPDW formation and in the Equatorial Atlantic (-25%) due to the reduction in the strength of the northeasterly trades (Jennerjahn et al., 2004).

3.3. Younger Dryas

In two recent studies (Tarasov and Peltier, 2005; Murton et al., 2010) it was suggested that a glacial meltwater pulse into the Arctic Ocean caused the YD cold period. The shut down of the AMOC at 13 ka B.P. (figure 1a)
leads to a cooling of about 10°C over Greenland, 2°C in the Netherlands, 3°C off the Iberian margin and 1°C in the Cariaco basin (figure 2). The simulated relative strength of the overturning in the North Atlantic is weaker than relative changes recorded by the $^{231}$Pa/$^{230}$Th record of the Bermuda rise and Iberian margin sediment cores (McManus et al., 2004; Gherardi et al., 2005). Ventilation age anomalies from the South Iceland rise indicate an interruption of the ventilation in the second part of the YD (Thornalley et al., 2011). Over Greenland, the amplitude of the simulated cooling during the YD is similar to the one estimated from the GISP2 borehole (Cuffey and Clow, 1997; Alley, 2000), but the simulated temperature at 13 ka B.P. (before the YD) is about 5°C warmer than the one estimated in GISP2 (figure 2a). This 5°C offset from the observations lasts until the end of the experiment $DG_{NS}$. Part of this temperature discrepancy between the model and the GISP2 record could be explained by an elevation change at GISP2. Using the relationship between height and $\delta^{18}O$ (Vinther et al., 2009), a 200 m increase in the GISP2 icesheet would lead to a 1.2 permil decrease in $\delta^{18}O$. Assuming a slope of 2.8°C per 1 permil for the deglaciation, the recorded temperatures would be $\sim$3.3°C lower. On top of that the adiabatic lapse rate (0.98°C/100m) has to be taken into account. A 5°C temperature offset could thus be explained by a 180-200 m increase in the icesheet height at GISP2. The simulated cooling off the Iberian margin is overestimated compared to the estimated SST anomaly (Bard et al., 2000; Bard, 2002; Martrat et al., 2004, 2007), while it is in agreement with a reconstructed 3°C cooling in the Western Mediterranean Sea (Cacho et al., 1999, 2001). In the Cariaco basin, however, the simulated cooling is underestimated. The abrupt YD cooling
is also observed in paleorecords from continental Europe (von Grafenstein et al., 1999; Heiri et al., 2005; Heiri and Millet, 2005; Peyron et al., 2005; Genty et al., 2006).

Similar to the H1 climate response, the anomalous meridional SST gradient during the YD in the tropical Atlantic leads to a southward shift of the ITCZ. In agreement with paleoproxies, drier conditions are simulated over the northern part of South America (Maslin and Burns, 2000; Peterson et al., 2000; Haug et al., 2001), Europe (Goslar et al., 1999; Kerschner et al., 2000; Peyron et al., 2005), Northern Africa (Peck et al., 2004; Williams et al., 2006; Lamb et al., 2007; Weldeab et al., 2007; Niedermeyer et al., 2009) and the Middle East (Bar-Matthews et al., 2003; Vaks et al., 2003; Ivanochko et al., 2005; Shakun et al., 2007; Fleitmann et al., 2009). Reduced summer precipitation is also simulated over China (Wang et al., 2001; Dykoski et al., 2005) (figure 5).

The weakening of the AMOC during the YD leads to the formation of NPDW. As a result, younger ventilation ages are recorded in marine sediment cores from the North Pacific (figure 1b) (Okazaki et al., 2010) and dissolved oxygen content increases along the coast of California (Kennett and Ingram, 1995; Pride et al., 1999; Emmer and Thunell, 2000; Ivanochko and Pedersen, 2004) (figure 1c).

4. The deglaciation history in the Southern Hemisphere

4.1. Heinrich event 1

As a result of the simulated AMOC shut down during H1, the AABW cell strengthens, attaining values of up to 12 Sv between 17.6 and 14.8 ka
B.P. (figure 8a). This intensification of AABW formation is in qualitative agreement with Neodymium data from the South Atlantic core RC11-83 (Piotrowski et al., 2004). Lower $\epsilon_{Nd}$ values indicate the predominance of NADW at the site of core RC11-83 while higher values indicate the predominance of AABW. Accompanying the NADW decrease and the AABW increase, the model simulates an intensification of the poleward heat transport at 30°S by about 0.6 PW. As depicted in figure 8, temperatures at high southern latitudes increase almost linearly from 17.6 to 14.8 ka B.P. This warming is a combination of different factors: rising atmospheric CO$_2$ concentrations (Timmermann et al., 2009b), increase of austral spring mean insolation changes affecting the sea-ice extent around Antarctica, and the bipolar seesaw effect leading to a warming of the Southern Hemisphere. Over Antarctica, the simulated air temperature during austral spring follows quite closely the temperature derived from EPICA Dome C ice core $\delta$D record (Jouzel et al., 2007). Both show an increase of about 4°C between 18 and 15 ka B.P. At that time, the derived temperature from the EPICA Dome C ice core suddenly increases by about 2°C in 200 years (figure 8b). This temperature spike is not simulated by our model. As austral spring insolation is an important pacemaker of the glacial termination in the Southern Hemisphere (Timmermann et al., 2009b), simulated austral spring temperatures were used for the comparison with Antarctic records.

In the Atlantic and Pacific sectors of the Southern Ocean (figure 8c), the simulated annual mean SST increases by 2.7°C between 18 and 15 ka B.P., which is in agreement with the reconstructed SST increase from a composite of marine sediment cores that we obtained by averaging linearly interpolated
alkenone SST reconstructions for cores MD97-2120 (Pahnke et al., 2003), ODP1233 (Kaiser et al., 2005), TN057-13PC4 (Nielsen, 2004; Anderson et al., 2009) and TN057-21 (Barker et al., 2009). At the same time period, the simulated SST increases by 1.9°C in the South Pacific (32°S-42°S, figure 8d). A composite of SST reconstructions from marine sediment cores MD03-2611 (Calvo et al., 2007) and H214 (Samson et al., 2005) indicates a larger increase of the order of 3°C, but the overall timing matches well with the transient simulation.

The cooling of the North Atlantic during H1 in combination with the warming of the Southern Hemisphere lead to a southward shift of the ITCZ and generally wetter conditions over the Southern Hemisphere (figure 4, middle), in agreement with paleoproxies from the Bolivian (Baker et al., 2001a,b; Blard et al., 2009) and Peruvian (Thompson et al., 1995) Altiplano (figure 9a-b) as well as with δ18O records from South African stalagmites (Holmgren et al., 2003) (figure 9d). The South American Summer Monsoon is also quite intense during H1, as suggested by the δ18O recorded in stalagmites from Botuverá cave (Brazil) (Cruz et al., 2005; Wang et al., 2007) and our modeling results (figure 9c).

Due to the warmer conditions during H1 and retreating sea-ice in the Southern Ocean, simulated marine export production increases in these areas by about 15% (figures 10a and 4, right), in close correspondence with the data of Anderson et al. (2009) and Sachs and Anderson (2005).

4.2. Antarctic Cold Reversal

Simulated surface temperatures at high southern latitudes reach a local maximum around 14.8 ka B.P (figure 8b-c). The rapid strengthening of the
AMOC towards the end of H1 leads to a weakening of AABW formation starting 14.8 ka B.P. (figure 8a). The warmer conditions in the Southern Ocean as well as higher sea level around 14.6 ka B.P. (Bindschadler et al., 2003; Peltier, 2005), may have helped to destabilize the Antarctic ice shelves and ice sheets (Flückiger et al., 2006), thus adding freshwater to the Southern Ocean. Proxy-evidence for a meltwater pulse originating from Antarctica at around 14.6 ka B.P. was found in several marine sediment cores of the Southern Ocean and was discussed in section 1. Adding freshwater to the Southern Ocean in our simulation $DG_{NS}$ further weakens the formation of AABW and slows down the Southern Ocean overturning cell to 5 Sv until about 13 ka B.P, in qualitative agreement with Neodymium data from the South Atlantic (Piotrowski et al., 2004) (figure 8a). In contrast, when no freshwater is added to the Southern Ocean (experiment $DG_N$), the AABW stays moderately strong (~12 Sv). The latter scenario does not agree well with the Neodymium data.

Reduced AABW formation in experiment $DG_{NS}$ leads to a decrease in southward heat transport at 30°S of about 0.2 PW. As a result, temperatures decrease at high and mid Southern latitudes (figure 7, left). Austral spring temperature averaged over Antarctica are about 0.8°C lower (figure 8b), which compares reasonably well with the cooling recorded in the $\delta$D data from EPICA Dome C (Jouzel et al., 2007) and Vostok ice cores (Petit et al., 1999) and qualitatively with the $\delta^{18}$O from Taylor Dome ice core (Steig et al., 1998). While the EPICA Dome C and Vostok ice core records had to be shifted by 600 years, the $\delta^{18}$O record from Talos Dome ice core based on the new GICC05 timescale fits well with our simulation without any time
shift. Over the Pacific and Atlantic section of the Southern Ocean (40°S-55°S), the simulated cooling of 1.4°C is in agreement with a reconstructed 1.2°C cooling in the Southern ocean SST index derived in this work from various data sets (Barker et al., 2009; Nielsen, 2004; Anderson et al., 2009; Pahnke et al., 2003; Kaiser et al., 2005) (figure 8c). Finally in the South Pacific (32°S-42°S), a simulated SST decrease of 0.9°C compares well with a reconstructed SST decrease of 1.2°C (Calvo et al., 2007; Samson et al., 2005) (figure 8d). Alkenone-based SST estimates from marine sediment core GIK 17748-2 (32.45°S,72.02°W) (not shown) also suggest a similar cooling of 1°C during the ACR (Kim et al., 2002).

A part of the cooling in the Southern Hemisphere prior to the addition of freshwater is caused by the reduction of AABW formation at 14.8 ka B.P. When no freshwater is added to the Southern Ocean (experiment DG₅), the cooling amounts to 0.3°C over Antarctica and 0.6°C in the Southern Ocean. The cooling at high Southern latitudes in experiment DG₅S is more pronounced over the ocean than over Antarctica (figure 11). In experiment DG₅ the cooling is reduced significantly compared to DG₅S, particularly over Antarctica and the Atlantic and the Eastern Pacific side of the Southern Ocean.

During the ACR the temperature decrease in the Southern Hemisphere leads to relatively drier conditions over most of the Southern Hemisphere (figure 7, middle). This is in good qualitative agreement with paleoproxies that document decreasing precipitation over the Bolivian and Peruvian Altiplano (Baker et al., 2001a,b; Thompson et al., 1995) and a reduction of summer rainfall over South Africa (Holmgren et al., 2003) (figure 9).
model also simulates a weakening of the South American Summer Monsoon over Brazil during the ACR in agreement with the δ¹⁸O record from Botuverá cave (Brazil) (Wang et al., 2007).

Due to the cooling at high southern latitudes around ~14 ka B.P., the simulated summer sea-ice edge advances to about 53-57°S in agreement with paleoproxies (Bianchi and Gersonde, 2004) (contour in figure 7, right). The extended sea-ice coverage induces a decrease in export production in the Southern Ocean due to light limitation. Both the simulated export production and opal production in the Southern Ocean significantly decrease during the ACR (figure 10 and figure 7, right), in agreement with paleoproxies from the Southern Ocean. Marine sediment core TN057-13 (Anderson et al., 2009) exhibits a decline in opal flux during the ACR, which corresponds to a decrease in diatom-productivity. In addition, a decrease in alkenone content indicating a decrease in marine productivity during the ACR is recorded in marine sediment core MD97-2120 (Sachs and Anderson, 2005). When no freshwater pulse is added to the Southern Ocean (experiment DGₙₛ), the simulated anomalies in marine export and opal production obtained are much smaller and last for only 1,000 years instead of 1,700 years for DGₙₛ (figure 10).

4.3. Younger Dryas

The shut down of the AMOC and the associated strengthening of AABW formation lead to an enhanced poleward heat transport into the Southern Ocean. Surface temperatures thus increase rapidly in the Southern Hemisphere. In conjunction with an increase of radiative forcing and changes in obliquity the simulated austral spring air temperature increases by 2.6°C
between 14 and 12 ka B.P., which is somewhat less than for the temperature reconstructions from the EPICA Dome C and Vostok ice cores (Jouzel et al., 2007; Petit et al., 1999) (figure 8b). Both the simulated and reconstructed SST in the Southern Ocean increase by 2°C. However the simulated SST increases more abruptly than the reconstructed one (figure 8c). In the South Pacific, the simulated SST increases by 1.6°C compared to a reconstructed SST increase of about 3°C (figure 8d). In addition, the simulated SST decreases by 1°C at 11.5 ka B.P., in contrast to the paleo-proxy indices used here. Due to the warmer conditions at high southern latitudes and reduced sea-ice coverage, the export and opal productions also increase strongly in agreement with paleoproxies data (Sachs and Anderson, 2005; Anderson et al., 2009) (figure 10).

In agreement with paleo-records, wetter conditions prevail over the Bolivian Altiplano (figure 9a) (Baker et al., 2001a,b). Wetter conditions are also simulated over the Peruvian Altiplano (figure 9b), while the Huascaran δ18O records show only a slight increase in precipitation (Thompson et al., 1995). Our model simulates a strengthening of the South American Summer Monsoon over Brazil, in agreement with the δ18O record from Botuverá cave (Cruz et al., 2005; Wang et al., 2007) (figure 9c).

5. Discussion and Conclusion

The transient deglaciation experiments performed in this study successfully capture the dominant climate and some biochemical features of the last deglaciation in both hemispheres. This good correspondence between model results and paleo proxy data suggests that most of the millennial-
scale variability recorded during the last deglaciation can be explained by
freshwater-induced variations in the deep water formation strength in the
North Atlantic, Southern Ocean and North Pacific. Based on our modeling
results it is justified to conclude that the North Atlantic is the key driver of
millennial-scale variability during the Last Glacial Termination.

In this study freshwater was mostly added in a broad area of the North
Atlantic, not taking into account differences that would arise from changes in
the freshwater routing around the North Atlantic (e.g. through the Norwe-
gian Sea, Labrador Sea, Gulf of Mexico) (Aharon, 2003; Tarasov and Peltier,
2006; Sionneau et al., 2010). Therefore, we did not differentiate between the
North Atlantic deep water formation sites, i.e. the Labrador Sea and the
Greenland Iceland and Norwegian (GIN) Sea. As seen in figure 1a, paleo-
proxy records from different North Atlantic locations suggest slightly different
scenarios. The Bermuda rise $^{231}$Pa/$^{230}$Th data (McManus et al., 2004) and
the South Iceland rise ventilation age record (Thornalley et al., 2011) indi-
cate a weakened AMOC already at 18 ka B.P., whereas the Iberian margin
$^{231}$Pa/$^{230}$Th data shows a weakened AMOC starting 16 ka B.P. (Gherardi
et al., 2005). Gherardi et al. (2005) suggested that the Labrador Sea NADW
formation site might have weakened before the GIN site, which is actually
not supported by the data presented in Thornalley et al. (2011). In the
Northern Hemisphere, the Cariaco basin reflectance record (Peterson et al.,
2000) is the only data set indicating a late (~16.2 ka B.P.) initiation of H1.
On the other hand, the South Iceland rise data suggest a greater ventilation
between 16 and 15 ka B.P., which is not seen in other North Atlantic circu-
lation and temperature records. Even if a strong weakening of the NADW
formation most likely occurred during H1, the impact of variations in the
North Atlantic convection sites needs to be further investigated. Whether a
more detailed knowledge of the freshwater routing into the North Atlantic
would help to reconcile the conflicting timing in paleo-proxy data during H1
remains to be shown.

A shut down of the AMOC during H1 leads to cold conditions over the
North Atlantic while the Southern Hemisphere starts to warm. A related
reduction of Atlantic-Pacific moisture transport leads to an increase of North
Pacific salinity and eventually triggers deep water Formation in the Bering
Sea. This leads to the establishment of a deep Pacific Meridional Overturning
Circulation during H1 in agreement with paleo-proxy records (Okazaki et al.,
2010). The closing of the Bering Strait is necessary for the formation of
NPDW (Okumura et al., 2009). Further North Pacific paleo-SST records
and modeling studies would be needed to better constrain the associated
SST patterns in the North Pacific during H1.

The rapid and strong resumption of the AMOC at about 15 ka B.P.
leads to the Bolling warming in the North Atlantic and initiates the ACR at
high Southern latitudes. The ACR can be simulated by a weakening of the
AABW formation, either as a bipolar seesaw response to the strengthened
AMOC or as a result of a freshwater input in the Southern Ocean. The
knowledge of AABW dynamics is too limited at the moment to conclude
whether the bipolar seesaw is sufficient to weaken the AABW or to estimate
the amount of freshwater necessary to weaken the AABW for about 2,000
years. In agreement with paleoproxies (Sachs and Anderson, 2005; Anderson
et al., 2009), our model successfully simulates a decrease in export and opal
production in the Southern Ocean during the ACR due to the sea-ice edge advance and colder conditions. It should be however noted that our model does not include iron limitation. Changes in export production therefore do not take into account any possible changes in iron supply to the euphotic zone as well as non-Redfield processes.

Concurrently in the North Hemisphere, the Older Dryas (14.4–14 ka B.P) is simulated by a substantial weakening of the AMOC. Our simulation thus suggests that melting from both, Antarctica and the Northern Hemispheric ice sheets contributed to MWP-1A at ~14.4 ka B.P. The exact timing, shape and location of the different freshwater pulses contributing to MWP-1A still need to be better constrained to allow for a more quantitative comparisons between climate model simulations and paleo-proxy data. The model results suggest that climate sensitivity to southern ocean freshwater input is largest in the high southern latitudes. Future model-proxy comparisons could thus improve the estimates of the Antarctic meltwater contribution to MWP-1A.

The Younger Dryas is represented by a freshwater input into the Arctic Ocean following the hypothesis of Tarasov and Peltier (2005) and Murton et al. (2010). The induced shut down of the AMOC captures most but not all of the features of the YD in the Northern Hemisphere. $^{231}\text{Pa}/^{230}\text{Th}$ data of McManus et al. (2004) suggests that only a weakening and not a complete shut down of the AMOC occurred during the YD. Our model is not able to simulate an intermediate strength of the AMOC. In addition, the amplitude of the cooling observed in GISP2 (Alley, 2000), in the North Atlantic (Cacho et al., 1999; Bard et al., 2000; Cacho et al., 2001; Calvo et al., 2001; Waelbroeck et al., 2001; Bard, 2002; Martrat et al., 2004, 2007) and in the Cariaco
basin (Lea et al., 2003) can only be obtained in our model as the consequence of a shut down of the AMOC. In the Southern Hemisphere, temperatures increase in phase with the strengthening of AABW formation, starting around 13 ka B.P. We therefore find that, similarly to H1, temperature and precipitation variations between the North Atlantic and the Southern Hemisphere are anti-correlated during the YD, in conformance with the bipolar seesaw mechanism.

Figure 12 shows the time series of simulated changes in terrestrial primary production in different latitudinal bands compared to the atmospheric CH$_4$ as recorded in GISP2 ice core (Brook et al., 2000). Wetlands are a main source of atmospheric CH$_4$ and it has been estimated that tropical region wetlands ($20^\circ$N–$30^\circ$S) were responsible for about 60% of the total wetland emissions, while the high northern latitudes wetlands ($45^\circ$N–$70^\circ$N) were responsible for about 38% (Bartlett and Harris, 1993). Due to the cold and dry conditions prevailing in most of the Northern Hemisphere during H1, terrestrial primary production slightly decreases in the latitudinal band 0–$20^\circ$N (–10%). Wetter and warmer conditions in the Southern Hemisphere lead to greater terrestrial primary production in the latitudinal band 0–$30^\circ$S (+20%), which could induce the measured atmospheric CH$_4$ increase. Overall, the simulated terrestrial carbon stock during H1 increases by about 100 GtC. The abrupt increase in atmospheric CH$_4$ at 15 ka B.P. corresponds to the Bølling warming and the associated warmer and wetter conditions in the Northern Hemisphere. The terrestrial primary productivity increases by ~20% in the latitudinal band 0–$20^\circ$N and 35% over $45^\circ$N–$70^\circ$N (figure 12). As a result of the drier conditions prevailing in the Southern Hemisphere
during the ACR, the terrestrial primary production slightly decreases (~5%) in the latitudinal band 0–30°S. Overall, the terrestrial carbon stock increases by about 260 GtC between 15 and 13 ka B.P. Finally, the drier conditions over the Northern Hemisphere during the YD induce a decrease in terrestrial primary production by 60% in the latitudinal band 0-20°N and by 30% in the latitudinal band 45-70°N. The terrestrial carbon stock decreases by only 60 GtC during the YD after which it increases by almost 300 GtC.

In this study we suggest a sequence of changes in the strength of deep water formation in the North Atlantic, the Southern Ocean and the North Pacific, which gives a reasonable representation of the reconstructed millennial-scale climatic variability during the Last Glacial Termination. To a large extent, we find that the relative timing of events observed in paleo-proxy data from different regions is consistent with the relative timing in the model. Ensemble-simulations with some form of data-assimilation (Goosse et al., 2010b; Jackson et al., 2010) could help to put an improved constraint on the freshwater forcing time series and, subsequently, the phase relations among proxy time series. A more detailed study of the climatic influence of the different deep water formation sites in the North Atlantic (similar to (Roche et al., 2010)) would also help to improve the understanding of the deglacial climatic changes.

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Figure 1: (a) Time series of the maximum of the meridional stream function (Sv) in the North Atlantic for experiment $DG_{NS}$ (green) compared to $^{231}Pa/^{230}Th$ data (black) from core OCE326-GGC5 (McManus et al., 2004) (filled circles) and from core SU81-18 (Gherardi et al., 2005) (empty circles) as well as ventilation age data (grey) from cores RAPID 10-1P, 15-4P and 17-5P (Thornalley et al., 2011). The time series of McManus et al. (2004) was shifted by 200 years backward to better fit the GISP2 data (Alley, 2000). H1 stands for Heinrich event 1, BWP for Bolling Warm Period, OD for Older Dryas and YD for Younger Dryas. (b) Time series of the maximum overturning strength in the North Pacific for experiment $DG_{NS}$ (green) compared to the ventilation ages (yrs) as recorded in marine sediment cores from the Western North Pacific (black). The ventilation ages curve is the smoothed spline interpolation of averaged benthic–planktic foraminifera ages and projection ages in the Western North Pacific (Okazaki et al., 2010). (c) Time series of dissolved oxygen content (ml/L) averaged over the California basin (135°W-114°W, 25°N-40°N) for experiment $DG_{NS}$ (green) compared to a composite of two $\delta^{15}N$ record (black) from the California Basin (marine sediment cores JPC-56 (Pride et al., 1999) and ODP 893A (Emmer and Thunell, 2000)) and compared to the Mo/Al data from core ODP 893A (grey circles) (Ivanochko and Pedersen, 2004). (d) Freshwater input (Sv) in the North Atlantic (black) and in the Southern Ocean (green) as applied in experiment $DG_{NS}$. Time series are 21 year running means for the model outputs.
Figure 2: (a) Time series of spring season (MAM) air temperature anomalies (°C) averaged over Greenland (47°W-0°, 60°N-80°N) for experiment DG_{NS} (green) compared to the temperature reconstruction from Greenland ice core GISP2 (black) (Alley, 2000). IACP stands for Intra-Allerod Cold Period. (b) Summer (JJA) air temperature anomalies (°C) averaged over the Netherlands (0°E-10°E, 45°N-55°N) for experiment DG_{NS} (green) compared to chironomid-inferred July air temperature anomalies (°C) from Hijkermeer, the Netherlands (Heiri et al., 2007). (c) SST anomalies (°C) averaged over the Iberian margin (11°W-8°W, 37°N-40°N) for experiment DG_{NS} (green) compared to a reconstructed temperature index (black) obtained by averaging different alkenone-based SST reconstructions from cores SU81-18 (Bard et al., 2000), MD95-2042 (Bard, 2002), ODP-161-977A (Martrat et al., 2004) and MD01-2444 (Martrat et al., 2007). The time axis of the alkenone-based SST reconstructions was shifted forward by 200 years to better fit the model results and the other northern hemispheric paleoproxies. (d) Winter season (DJF) SST anomalies (°C) averaged over the Cariaco basin (55°W-52°W, 10°N-14°N) for experiment DG_{NS} (green) compared to Mg/Ca SST reconstruction (°C, black) from core PL07-39PC (Lea et al., 2003). As marine productivity is the highest during the winter season, this season was chosen to best represent temperature variations. Anomalies are with respect to pre-industrial time. Time series are 21 year running means for the model outputs.
Figure 3: Map showing the locations of the paleo-proxy records used to compare model results. Dark blue stars represent paleo-proxy records of oceanic circulation (McManus et al., 2004; Piotrowski et al., 2004; Gherardi et al., 2005; Okazaki et al., 2010; Thornalley et al., 2011). Light blue stars represent terrestrial and marine temperature records (Petit et al., 1999; Alley, 2000; Bard et al., 2000; Lea et al., 2003; Pahnke et al., 2003; Nielsen, 2004; Kaiser et al., 2005; Samson et al., 2005; Calvo et al., 2007; Heiri et al., 2007; Jouzel et al., 2007; Martrat et al., 2007; Barker et al., 2009; Stenni et al., 2011). Orange stars show the location of paleo-precipitation records (Thompson et al., 1995; Peterson et al., 2000; Baker et al., 2001a,b; Wang et al., 2001; Holmgren et al., 2003; Ivanochko et al., 2005; Wang et al., 2007; Weldeab et al., 2007). Red stars represent biogeochemical paleo-proxy records (Pride et al., 1999; Emmer and Thunell, 2000; Sachs and Anderson, 2005; Ivanochko and Pedersen, 2004; Anderson et al., 2009)
Figure 4: (left) Annual SST anomalies (°C, shaded) and current anomalies (m/s, vectors) averaged over 0-200 m for H1–LGM conditions (16–18 ka B.P.), (middle) Annual precipitation anomalies (cm/yr, shaded) and annual wind stress anomalies (Pa, vector), (right) Annual export production anomalies (gC/m²/yr) for H1–LGM and overlaid 0.1 m annual sea-ice contour for H1.
Figure 5: (a) Time series of JJA precipitation anomalies (cm/yr) averaged over the Cariaco basin (55°W-52°W, 10°N-14°N) for experiment DG_NS (green) compared with the reflectance (% black) measured in a marine sediment core from the Cariaco basin (ODP 1002) (Peterson et al., 2000). Lower reflectance has been associated with increased marine productivity due to greater riverine input. (b) Time series of JJA precipitation anomalies (cm/yr) averaged over western Equatorial Africa (15°W-15°E, 4°N-20°N) for experiment DG_NS (green) compared with SSS reconstruction (black) from the Gulf of Guinea core MD03-2707 (Weldeab et al., 2007). (c) Time series of JJA/DJF precipitation averaged over China (114°E-124°E, 28°N-35°N) for experiment DG_NS (green) compared to δ¹⁸O (permil, black) as recorded in stalagmites of the Hulu cave (China) (Wang et al., 2001). (d) Time series of JJA precipitation anomalies over the Arabian Sea (45°E-65°E, 5°N-15°N) for experiment DG_NS (green) compared to δ¹⁵N (permil, black) as recorded in marine sediment core 905 from the Arabian Sea (Ivanochko et al., 2005). Anomalies are with respect to pre-industrial time. Time series are 51 year running means for the model outputs.
Figure 6: (left) Annual SST anomalies (°C, shaded) and current anomalies (m/s, vectors) averaged over 0-200 m for Bolling warming–H1 conditions (14.6–16 ka B.P.), (middle) Annual precipitation anomalies (cm/yr, shaded) and annual wind stress anomalies (Pa, vector), (right) Annual export production anomalies (gC/m²/yr) for BWP–H1 and overlaid 0.1 m annual sea-ice contour for BWP.

Figure 7: (left) Annual SST anomalies (°C, shaded) and current anomalies (m/s, vectors) averaged over 0-200 m for ACR–Bolling warming conditions (14–14.6 ka B.P.), (middle) Annual precipitation anomalies (cm/yr, shaded) and annual wind stress anomalies (Pa, vector), (right) Annual export production anomalies (gC/m²/yr) for ACR–BWP and overlaid 0.1 m annual sea-ice contour for ACR.
Figure 8: (a) Time series of the maximum overturning strength in the bottom cell of the Southern Ocean (Sv) for experiment DG\textsubscript{NS} (green) and DG\textsubscript{N} (cyan) compared to $\varepsilon_{\text{Nd}}$ (black) as recorded in marine sediment core RC11-83 (Piotrowski et al., 2004). Lower $\varepsilon_{\text{Nd}}$ values indicate the predominance of NADW at the site of core RC11-83 while higher values indicate the predominance of AABW. The time series of Piotrowski et al. (2004) has been shifted by 500 years forward. (b) Time series of austral spring (SON) air temperature anomalies (°C) averaged over 70°S-90°S for experiments DG\textsubscript{NS} (green) and DG\textsubscript{N} (cyan) compared to temperature reconstructions (°C) from EPICA Dome C ice core (black stars) on the EDC(3) time scale (Jouzel et al., 2007) and from Vostok ice core (black thin line) (Petit et al., 1999) as well as compared to the $\delta^{18}O$ record from Talos Dome ice core (grey) (Stenni et al., 2011). The time axis of the EPICA Dome C and Vostok temperature reconstructions were shifted backward by 600 years to better fit the model results and the northern hemispheric paleoproxies. Simulated austral spring temperatures were used as suggested by the study of Timmermann et al. (2009b). (c) Time series of annual SST anomalies (°C) averaged over the Pacific and Atlantic part of the Southern Ocean (170°E-10°E, 40°S-55°S) for experiments DG\textsubscript{NS} (green) and DG\textsubscript{N} (cyan) compared to a composite of SST reconstructions (°C, black) from marine sediment cores MD97-2120 (Pahmeke et al., 2003), ODP1233 (Kaiser et al., 2005), TN057-13PC4 (Anderson et al., 2009; Nielsen, 2004) and TN057-21 (Barker et al., 2009). (d) Time series of annual SST anomalies (°C) averaged over the Southwest Pacific (130°E-180°E, 32°S-42°S) for experiment DG\textsubscript{NS} (green) compared to a composite of SST reconstructions (°C, black) from marine sediment cores MD03-2611 (Calvo et al., 2007) and H214 (Samson et al., 2005). Anomalies are with respect to pre-industrial time. Time series are 21 year running means for the model outputs.
Figure 9: (a) Time series of annual precipitation anomalies (cm/yr) averaged over Bolivia (71°W-62°W, 15°S-25°S) for experiment DGNS (green) compared to % of freshwater diatoms (black x) in Lake Titicaca (Bolivia) (Baker et al., 2001b) and to a natural gamma ray profile (c.p.s., grey circles) from the Salar de Uyuni (Bolivia) (Baker et al., 2001a). The time axis of the Lake Titicaca and the Salar de Uyuni records were shifted forward by 400 years. (b) Time series of DJF precipitation anomalies (cm/yr) averaged over Peru (85°W-70°W, 5°S-15°S) for experiment DGNS (green) compared to δ18O record (permil, black) from Huascaran ice core (Peru) (Thompson et al., 1995). (c) Time series of DJF/JJA precipitation averaged over Brazil (60°W-44°W, 20°S-32°S) for experiment DGNS (green) compared to δ18O record (permil, black) from stalagmites of Botuverá cave (Brazil) (Wang et al., 2007). (d) Time series of JJA/DJF precipitation anomalies averaged over South Africa (25°E-35°E, 20°S-30°S) for experiment DGNS (green) compared to δ18O record (permil, black) from stalagmites of the Makapansgat Valley (South Africa) (Holmgren et al., 2003). Anomalies are with respect to pre-industrial time. Time series are 51 year running means for the model outputs.
Figure 10: (a) Time series of opal production (mol/m$^2$/yr) in the Southern Ocean ($52^\circ$S-$64^\circ$S) for experiments $\text{DG}_{NS}$ (green) and $\text{DG}_N$ (cyan) compared to the opal flux record (g/cm$^2$/kyr, black) from core TN057-13 (Anderson et al., 2009). The time axis of the opal flux was shifted backward by 200 years. (b) Time series of marine export production (gC/m$^2$/yr) in the Pacific side of the Southern Ocean ($150^\circ$E-$190^\circ$E, $42^\circ$S-$49^\circ$S) for experiments $\text{DG}_{NS}$ (green) and $\text{DG}_N$ (cyan) compared to the alkenone content record (ng/g, black) from core MD97-2120 (Sachs and Anderson, 2005). The alkenone record is shown for the period 18 to 10.4 ka B.P.

Figure 11: Austral spring air temperature anomalies (14–14.8 ka B.P., °C) over Antarctica and Southern Ocean for (left) experiment $\text{DG}_{NS}$ and (right) experiment $\text{DG}_N$
Figure 12: Time series of annual terrestrial primary production anomalies (GtC/yr) averaged over the latitudinal band 0–30°S (cyan), 0–20°N (red), 45°N–70°N (magenta). The green line represents the sum of the cyan, red and magenta lines. The black line represents the time series of the atmospheric CH$_4$ (ppb) as recorded in GISP2 ice core (Brook et al., 2000).
### Table 1: Paleo-proxy records used to compare the model outputs of this study.

<table>
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<tr>
<th>Location</th>
<th>Core</th>
<th>Lat.</th>
<th>Long.</th>
<th>Elev (m)</th>
<th>Proxy</th>
<th>Reference</th>
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<td>Ice core δ¹⁸O, borehole</td>
<td>(Alley, 2000)</td>
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<tr>
<td>Greenland</td>
<td>GISF2</td>
<td>72.6°N</td>
<td>38.5°W</td>
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<td>Hjornemoer</td>
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<td>Chironomids</td>
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Table 2: Temporal shift applied to paleo-proxy records used to compare the model outputs of this study. A positive (negative) shift means that the paleo data had to be shifted forward (backward) in time to match the model simulation.

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<th>Location</th>
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<th>Reference</th>
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<td>500</td>
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<td>(-) 200</td>
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<td>Antarctica</td>
<td>EPICA Dome C</td>
<td>Ice core δD</td>
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