Clim. Past Discuss., 6, 1077–1110, 2010
www.clim-past-discuss.net/6/1077/2010/
doi:10.5194/cpd-6-1077-2010

Climate of the Past Discussions

Climate response to freshwater perturbations in Northern or Southern Hemispheres at the last glacial inception, the last glacial maximum and the present-day

G. Philippon-Berthier[1,*], G. Ramstein[1], S. Charbit[1], and C. Ritz[2]

[1]LSCE-Orme, point courrier 129, CEA-Orme des Merisiers, 91191 Gif-Sur-Yvette Cedex, UMR 8212, France
[2]Laboratoire de Glaciologie et Géophysique de l'Environnement, UMR 5183, 38402 Saint Martin d'Hères Cedex, France
[*]now at: Center for Climatic Research, 1225 W. Dayton St., Madison, WI, USA

Received: 23 April 2010 – Accepted: 27 April 2010 – Published: 1 June 2010

Correspondence to: G. Philippon-Berthier (gwen.philippon@yahoo.fr)

Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

Freshwater inputs in North Atlantic due to huge surge of icebergs coming from ice sheets might be responsible for drastic regional and global abrupt climatic transitions. To quantify the sensitivity of climate system to these freshwater inputs, we use a model of intermediate complexity coupled to ice-sheet models for both Northern and Southern Hemispheres. We mimic the Dansgaard-Oeschger and Heinrich Events by forcing the model with appropriate freshwater perturbations. The originality of this study is to investigate with such a global model, the response of the coupled system to freshwater discharges for three different climate contexts, the Last Glacial Maximum (LGM), the Last Glacial Inception (LGI) and the present-day (PD) climates.

We first show that in all climate contexts, the North Atlantic circulation is more sensitive to freshwater flux when ice sheets are present. Secondly, the "seesaw" mechanism occurs mostly for the North Atlantic freshwater perturbation whereas it remains very weak for the Southern Ocean freshwater release. Moreover, this seesaw is generally enhanced when ice sheets are interactive. The most striking result is that the freshwater perturbation amplifies the inception of the North American ice sheet at LGI the sea-level drop associated is significantly increased and in a much better agreement with data.

1 Introduction

Many authors (Broecker, 1998; Stocker, 1998; Bender et al., 1994; Seidov and Maslin, 2001; Blunier and Brook, 2001; Clark et al., 2002; Alley et al., 2002; Knutti et al., 2004) showed that the freshwater perturbations in the Northern high latitudes due to ice caps melting like Heinrich Events (Heinrich, 1988) coming from the Northern Hemisphere ice sheets produce changes of the thermohaline circulation (THC) in the North Atlantic Ocean. More recently, Alvarez-Solas et al. (2010) suggested that the Heinrich Instability could be linked with ice shelf break-up. Independently of the exact mechanism

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

CPD

6, 1077–1110, 2010

Climate response to freshwater perturbations

G. Philippon-Berthier et al.

of Heinrich Events, the associated melt water perturbation drives or/and amplifies the decay of the meridional heat flux in the Atlantic Ocean and leads to major climate reorganizations in the Northern Hemisphere, whereas in the Southern Hemisphere, these changes are opposite phase (Stocker, 1998; Clark et al., 2002; Alley et al., 2002). This balance is thus expressed by two different patterns. The first one corresponds to an intensification of the formation of the North Atlantic Deep Water (NADW) associated with an increase of the temperature in the Northern high latitudes, while the transport of heat is slowed down in the Southern Hemisphere, and a decrease of the Antarctic surface temperature. Conversely, the second pattern shows a decrease of NADW which causes a larger heat transfer towards the South and contributes to the Antarctic surface warming and to a cooling in the Northern high latitudes. This seesaw mechanism is recorded in the ice cores and marine sediments during Dansgaard-Oeschger events (DO) (Dansgaard et al., 1982, 1993; Shakelton, 1987; Sánchez-Goñi et al., 2008; Cacho et al., 1999). The response of the climate system to freshwater perturbations has been investigated in details by models of different complexity: Ganopolski and Rahmstorf (2001) and Wang and Mysak (2006) studied the stability of the LGM and the Present-Day climates with two different models of intermediate complexity. More recently, Stouffer et al. (2007) investigated the seesaw mechanism with an atmosphere-ocean general circulation model. They applied a simple freshwater flux (1 Sv or 10^6 m^3/s during 100 yrs) and removed it under present-day conditions. Their results will be described in the discussion.

Furthermore, because the response of NADW to freshwater perturbations is non linear, a freshwater pulse may induce a total collapse or a simple strengthening depending on the hysteresis magnitude under different climatic conditions (Rahmstorf, 1995). Ganopolski and Rahmstorf (2001) and Paillard (2001) were the first authors to demonstrate behavior that the glacial climate is less stable than the present-day one through these mechanisms. Moreover, in this study, the CLIMBER climate model of intermediate complexity is forced with freshwater perturbations in the convective zones of the North Atlantic. The stability of the climate is illustrated by the hysteresis curves, which link the variations of NADW (Atlantic overturning) with the intensity of the freshwater flux. A major result pointed out in this study is dumped the shift from the "active" mode (present-day) of the thermohaline circulation (THC) to mode in glacial climate as described on Fig. 3a which is in phase space much narrow than the present day one (Fig. 3b) and therefore more sensitive to freshwater perturbations. When the North Atlantic freshwater perturbation is higher than a threshold, they obtained a break down of the THC known as "off" mode. This decrease of THC results in a cooling of the Northern Hemisphere and synchronously, in a warming of the Southern Hemisphere. Whereas our study is based on the same freshwater perturbation in North Atlantic, we performed a set of simulations with a coupled climate-ice-sheet model. Therefore, we are able to account at least for fast feedbacks associated with ice-sheet response (freshwater). Moreover, in addition to the North Atlantic Ocean perturbation derived from Ganopolski and Rahmstorf (2001), we performed simulations with a Southern Ocean freshwater perturbation because Clark et al. (2002, 2009) and Rohling et al. (2004) have provided evidences of ice-rafted debris beyond 55° S, suggesting that melt water pulse could also come from Antarctica. Accounting for the fast ice sheet response of the coupled model may be important for the following reasons: if the freshwater perturbation decreases drastically the THC, the resulting cooler climate decreases the freshwater flux from the ice sheet to the ocean which produces a negative feedback to the THC. Another scenario might also illustrate a second negative feedback: if the "negative" freshwater perturbation (i.e. increase of salinity) implies an increase of the strength of the THC; the meridional heat transport toward high latitudes is enhanced. This favors the melting of the ice sheets, which, in turn, tends to decrease the strength of NADW. This negative feedback can have their counterpart in terms of positive feedbacks. Indeed, the changes in THC will produce both variations in temperature and hydrologic cycles which will impact differently the ice sheet mass balance and the freshwater from the ice sheet. Therefore we can investigate which of these mechanisms will be the major player in the three different climates.

CPD
6, 1077–1110, 2010

In this paper, an important issue is first to understand how these freshwater perturbations impact the global climate through changes in the ocean dynamics (THC) and in the ice sheets behavior via freshwater input. Secondly, it is to explore how the different responses depend on the initial climatic context (Present-Day, PD, Last Glacial Inception, LGI, onset of the previous glacial period, Last Glacial Maximum, LGM). Our approach is based on an Earth System Model of Intermediate Complexity (Claussen et al., 2002; Ganopolski and Rahmstorf, 2001) fully coupled to high-resolution ice-sheet models for both Northern and Southern Hemispheres. We first examine the hysteresis curves: the model response to freshwater perturbation is analyzed in terms of NADW variation. Secondly, we apply the freshwater perturbation scenario developed by Ganopolski and Rahmstorf (2001) either in North Atlantic or in Southern Ocean to mimic the freshwater inputs during DO and Heinrich Events and Antarctic Maximum Events, respectively and to investigate the response of the climate system. Then, we analyze the specific role of the ice sheet and ocean dynamics with emphasis on the sensitivity of Northern Hemisphere high latitudes to the ice sheet build-up in response to North Atlantic and Southern freshwater perturbations. Finally, we analyze the impact of a North Atlantic freshwater perturbation on the extent of the ice sheets during the Last Glacial Inception (LGI).

2 Models description and coupling

The model consists of the intermediate complexity Atmosphere-Ocean-Vegetation (AOV) climate model (Petoukhov et al., 2000) coupled with two 3-D thermo-mechanical ice sheet models (Ritz et al., 1997; 2001).

The CLIMBER 2.3 is based on simplified representations of the atmosphere, the vegetation, ocean and sea-ice. The atmospheric resolution is 51° in longitude and 10° in latitude.

The ice-sheet models, GREMLINS, and GRISLI, deal with the dynamics of the inland ice in Northern Hemisphere and Antarctica, respectively. The GRISLI model also

CPD
6, 1077–1110, 2010

dynamically computes the position of the grounding line, which splits grounded from floating ice. This model accounts for the ice flow through the ice shelves and detects the zones where ice streams can take place. The resolutions of these models are, respectively 45 km×45 km and 40 km×40 km.

The coupling aspect has been developed by Charbit et al. (2005), Kageyama et al. (2004) and Philippon et al. (2006). Moreover, a full simulation of the last interglacial-glacial cycle has been successfully conducted with this model (Bonelli et al., 2009). Here, we only summarize the main features of the coupling procedure. The major difficulty of coupling CLIMBER to the ice sheet models lies in the difference of resolution between these models. The coupling method performs the downscaling between the annual and summer surface air temperatures and annual snowfall from the atmosphere of CLIMBER. These fields are used to compute the surface mass balance of both Northern and Southern ice sheets. The freshwater flux coming from the melting of the ice is also released to the CLIMBER oceanic model. In GRISLI, the basal melting under the ice shelves floating over the Ross and Weddell seas is parameterized. This parameterization tends to produce basal freezing in the centre of the ice shelf and melting near the grounding line and above the continental shelf as observed by Rignot and Jacobs (2002) and parameterized by Beckmann and Goosse (2003) and Philippon et al. (2006).

The altitude and the surface type of each ice sheet model grid point are returned to CLIMBER (land-ice, ice shelf, ice-free land or oceanic area) using a simple aggregation method.

Fields are exchanged every 20 years between CLIMBER and GREMLINS and CLIMBER and GRISLI. Note that the CLIMBER model we used is different from the original one (Ganopolski and Rahmstorf, 2001) because the radiative balance has been improved by using a larger number of atmospheric layers (Kageyama et al., 2004; Charbit et al., 2005). Therefore, although our results are very similar to those of Ganopolski and Rahmstorf (2001) when using CLIMBER alone, some differences in the amplitude of the climate response to the freshwater perturbation may arise.

CPD
6, 1077–1110, 2010

3 Freshwater perturbation and experimental design

To identify which feedbacks will be important when coupling the ice sheets to the climate model, we compare the responses of both CLIMBER (C) and CLIMBER-GREMLINS-GRISLI models (CGG). Both models are forced by the CO_2 atmospheric concentrations from Vostok (Petit et al., 1999), the insolation (Berger, 1978) and the sea-level variation (Imbrie et al., 1984). The description of each initial condition is given in Table 1. All the experiments we have performed are equilibrium simulations. We kept these boundary conditions, described above, fixed because our aim is to focus on the response of the climate and ice sheets to freshwater perturbations.

In the Northern Hemisphere, the initial ice sheet configurations are given by the present-day (PD and LGI simulations) and the LGM ICE-5G reconstructions (Peltier et al., 2004). For the Antarctic ice sheet, the reconstruction is obtained after a four climatic cycles simulation forced by the Vostok climate (LeMeur and Huybrechts, 1996). This four-cycles simulation is stopped at −21 kyr BP to obtain the initial state of the ice-sheet at the LGM, and is run until present-day to provide the initial ice-sheet state of the PD and LGI simulations. To study these instabilities, we forced the model with the same freshwater perturbation which were used in Ganopolski and Rahmstorf (2001) (referred hereafter as "GR01"), to mimic the Dansgaard-Oeschger (DO) (Dansgaard et al., 1982, 1993) and Heinrich Events (Heinrich, 1988). Those freshwater perturbations were introduced in the North Atlantic Ocean (Pnorth), in the 51°–76° N latitudinal belt (see Table 2).

The Pnorth perturbation is designed as follows (Fig. 1) and the results are described in Sects. 5 and 6:

1. 1000 years (yr) without any freshwater perturbation, which corresponds to the duration required for the Antarctic Bottom Water (AABW) being equilibrated with the Antarctic ice-sheet model.
2. 2 DO oscillations with minima of −0.03 and +0.03 Sverdrup (Sv), respectively. The positive forcing corresponds to a freshwater input and the negative forcing

to a freshwater output, equivalent to a salt input (commonly associated with an increased evaporation). The duration of one oscillation is 1500 yr.

3. 1 Heinrich oscillation reaching a maximum of +0.15 Sv. The duration of the Heinrich Event is 3000 yr.
4. 2 DO oscillations defined as before.

The Psouth simulations are constructed following the same experimental design, except that the freshwater perturbation, simultaneously incorporated in the Atlantic, Indian and Pacific oceans between 68° and 90° S, has an amplitude three times larger. This number is it corresponds to over 20 m of sea-level rise in 500 years equivalent discharge. To investigate the impact of Pnorth and Psouth perturbations, we also carried out "standard experiments" (STD) for PD, LGI and LGM experiments in which no freshwater perturbations are incorporated.

To explore the phase space (freshwater, NADW strength), we perform the hysteresis experiments (results described in Sect. 4) we adopted an approach similar to that used in GR01 with a step of 0.05 Sv in 1 kyr (1000 model years, Fig. 2) in the 51°–76° N Atlantic Ocean. The experimental set up is the following:

1. 2 kyr with 0 Sv perturbation.
2. 5 kyr from 0 to +0.25 Sv (noted 1).
3. 10 kyr from +0.25 to −0.25 Sv (noted 2).
4. 5 kyr from −0.25 to 0 Sv (noted 3).
5. 3 kyr with 0 Sv perturbation.

We only investigate the change in hydrology and temperature disturbing the ice sheet mass balance and not other long-term effects as isostasy.

CPD
6, 1077–1110, 2010

4 Hysteresis analysis

4.1 Comparison of CLIMBER alone results to previous studies

The hysteresis curves are displayed in Fig. 3. These diagrams are obtained in a way fully consistent with that of "GR01" using a similar North Atlantic freshwater perturbation.

We first compare our results to "GR01", for both LGM and PD climates, which are the only periods available in GR01. Moreover, we also describe our results for both Interglacial periods (PD and LGI).

Figure 3a (LGM) and 3b (PD) depict the same hysteresis shape obtained with the C model as the one given in "GR01". The width of the hysteresis diagram corresponding to the width between the two vertical branches is roughly similar in both studies (when the width is large, the climate is more stable). In the present work, it is equal to 0.08 and 0.03 Sv for PD and LGM simulations, respectively, against 0.09 and 0.01 Sv in "GR01". As previously outlined by Paillard (2001) and GR01, this shows that the PD period is more stable than the glacial one. This is further illustrated by the fact that the NADW shuts down for PD run for a freshwater perturbation equal to 0.09 Sv in our study. The same value is obtained in "GR01". In contrast, for LGM, the NADW starts to shut down completely at 0.2 Sv in our study and at 0.15 Sv in "GR01". The only difference is the initial NADW, which equals in our study to 26 and 20 Sv for PD and LGM experiments instead of 20 and 12 Sv in "GR01". These differences in amplitude could be explained by a few modifications in the CLIMBER model discussed in Sect. 2.

The larger stability of interglacial periods is also confirmed by the hysteresis width corresponding to the LGI simulation (0.06 Sv, Fig. 3c). This value is lower than that obtained for the PD period (0.09), suggesting that the LGI climate is less stable than the present-day one. A detailed discussion of changes in hydrologic cycle at LGI will be given in Sect. 6. The NADW shuts down with a freshwater perturbation equal to 0.11 Sv, which remains roughly similar to the present-day one. Note that the initial NADW at LGI is 23 Sv, which is smaller than the PD one (26 Sv).

CPD
6, 1077–1110, 2010

This result at LGI is also consistent with that of Khodri et al. (2003). These authors used the CLIMBER model at 115 ka and performed sensitivity experiments to determine the freshwater thresholds that may convert the insolation decrease at 115 ka into a maximum northern high latitudes cooling. To do this, they progressively increased the atmospheric northward freshwater transport from 0 to 0.05 Sv over the Northern seas (see Fig. 2 between 2 and 3 kyrs). This leads the interglacial climate to abruptly switch towards a glacial state when the freshwater transport is above 0.02 Sv and when THC reaches 21 Sv, which is comparable to information provided by the hystereris experiment of the present study (Fig. 3c).

4.2 Impact of interactive coupling on hysteresis curves

Note that because of the lack of representation of appropriate mechanisms in the ice-sheet models, at the origin of rapid ice flow and linked to the sub-glacial hydrology, the coupled model cannot reproduce high frequency variations.

Because the climate is very cold during the LGM, the amount of meltwater coming from ice sheets is very weak and therefore the impact of ice sheets cannot be drastic. It is therefore not surprising to observe very similar pattern in the hysteresis diagrams between C and CGG. Indeed, the C LGM hysteresis width (0.03 Sv) is similar to the CGG one (0.04 Sv) (Fig. 3a).

For each interglacial period (Fig. 3b–c), the positive freshwater inputs (see label 1 in Fig. 2; until 8 kyr) do not drastically change the NADW sensitivity of the C and CGG models: both hysteresis shapes remain approximately similar to the first order. Nevertheless, a closer inspection of the diagrams shows that some differences appear. As an example, the CGG hysteresis widths compared to C ones are reduced by 0.01 (0.02) Sv in PD (LGI). In the CGG model, the NADW shuts down at a higher value (+0.01 Sv) than in the C model. The main difference lies in the fact that the CGG model is more sensitive to the salt addition (labels 2 and 3 in Fig. 2; after 8 kyr) than for the C model. The differences between the C and CGG experiments are less significant in the PD hysteresis than in the LGI one. This result was also expected because the

LGI period corresponds to the onset of the glaciation and represents therefore a more unstable context than the PD period.

In the following, we focus on the LGI period, for which the differences between C and CGG are the most significant. We analyze more precisely three time periods: 8 kyr (See Figs. 3c and 2), where the freshwater perturbation is high and NADW equals zero (no overturning); 11 kyr, where freshwater perturbation tends to zero; and 12 kyr, where freshwater perturbation equals zero.

The differences between the C and CGG responses can be linked to the surface Atlantic salt flux component (Fig. 4). At 11 kyr, the surface Atlantic density becomes higher and NADW becomes active compared to 8 kyr (Fig. 3c). The NADW recover is therefore associated with the surface Atlantic salt budget defined as: Evaporation – Precipitation – runoff – ice calving + sea-ice component. For LGI (Fig. 4), the surface Atlantic salt flux difference between CGG and C remains weak until 11 kyr. At 12 kyr (under no freshwater perturbation), the difference becomes positive and reaches 0.3 mm/day because the CGG E-P term starts to be larger than the C E-P term. This feature is linked to the reactivation of CGG NADW at 12 kyr, whereas, at this period, the C NADW remains still inactive.

For interglacial periods, mostly for LGI, there are significant differences between C and CGG simulations, due especially to an earlier restoring of NADW in the CGG simulation, which is clearly linked to hydrologic responses. This result demonstrates that the coupling with ice sheets narrows the width of the hysteresis curve and thus makes the climate more unstable than in CLIMBER alone experiment. Some consequences of this finding will be developed in more details in Sect. 6.

CPD
6, 1077–1110, 2010

5 Seesaw and dynamical response of the ocean

5.1 The Northern perturbation (Pnorth)

The aim of this part is to figure out which superimposed feedback is induced by ice sheets when climate system experiences large freshwater perturbations. In this view, we used the freshwater perturbation from "GR01" for the Dansgaard-Oeschger and Heinrich Events (as described in Sect. 3). We use the difference between the Heinrich Event (maximum of the perturbation, as referred as "H" in Fig. 1) and the stadial phase (maximum Dansgaard-Oeschger warming noted "S" in Fig. 1) because this difference shows the largest impact of the Heinrich Event on a cold climate under multiple DO events at LGM (see GR01 and NGRIP, 2004). Figure 5 shows the annual surface air temperature difference between H and S for the three climatic conditions. For all contexts, the response of the CLIMBER model alone (C) depicts a large cooling over the whole North Atlantic (Fig. 5 left), and a warming over the Southern Hemisphere. The temperature difference depicted by the LGM experiment is quite similar to that from GR01: a cooling up to 2 °C in North Atlantic and a warming about 2 °C in the Southern Ocean. Moreover, under PD and LGI conditions, the perturbation leads to more pronounced Northern Hemisphere cooling and is extended over a large part of the Northern Hemisphere. As described by Stouffer et al. (2007), using an atmosphere-ocean general circulation model, we also observe in our PD simulation that NADW decreases during the freshwater perturbation (not shown), which is associated with the drop of the northward heat transport. This feature is also depicted for LGI.

To investigate the specific role of feedbacks induced by ice sheets, we also plotted the differences between CGG and C model responses (Fig. 5 right). For the LGM and the LGI experiments, accounting for ice sheet and climate interaction leads to an increased the cooling over North Atlantic and Arctic Oceans (LGM) and over North America (LGI). The difference (H-S) shows a southward transfer of the Atlantic heat flux in the CGG experiment compared to the C one (not shown), associated with cooler temperatures high latitudes North Atlantic in the CGG experiment.

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

CPD

6, 1077–1110, 2010

Climate response to freshwater perturbations

G. Philippon-Berthier et al.

In contrast to the C experiments, the interactive ice-sheets do not produce any significant additional southern warming in LGM and LGI simulations. However, in the PD experiment, a slight additional warming is observed, mainly over East Antarctica. In summary, the pattern of temperature changes (H-S), due to the ice sheets, is enhanced compared to the one produced by CLIMBER alone: between 6 and 22% in general and up to 119% for Northern Hemisphere Pnorth temperature changes.

As expected, the feedbacks due to interactive ice sheets are more significant in the Northern Hemisphere for the LGM climate, compared to interglacial periods. These positive feedbacks are mainly due to a larger extent of the ice sheets and therefore to a higher albedo in the CGG experiment, which induce the surface cooling. Section 6 will explain that the cooler North Atlantic temperature at LGI resulting from the he Pnorth freshwater perturbation is associated with larger ice caps over North America.

5.2 The Southern perturbation (Psouth)

Figure 6 (left) depicts the difference (H-S) in the CLIMBER alone annual surface air temperature between the Heinrich Event and the stadial state. As expected, the patterns of temperature difference are reverse to those obtained with the Pnorth perturbation, and but amplitude is also lower (Fig. 5). It is important to note that whereas the freshwater perturbation is three times higher, the impact on temperature is very weak. It is possible that the amplitude of the perturbation used in this study is not large enough to produce significant temperature changes. Because the AABW differences (H-S) are also weak in all climatic contexts considered in the present work, the southern ocean temperature differences are small. The C Southern Hemisphere temperature changes (H-S) are negative (except LGM C around 60° S), whereas the north Atlantic temperature changes (H-S) are positive (Fig. 6 left).

Our PD simulations are similar to those of Stouffer et al. (2007). We also observe in our PD simulation no significant change in NADW during the southern freshwater perturbation (not shown, even if North Atlantic temperature warms up to 1 °C). In our simulation, AABW is strengthened when the freshwater forcing becomes negative (not

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |
CPD
6, 1077–1110, 2010

shown), similar to Stouffer et al. (2007), AABW switches from 5 to 22 Sv when the freshwater perturbation stops. Our PD results can also be compared to those from Weaver and al. (2003), although the climatic context is different. They used a model of intermediate complexity (UVic Earth System Climate Model) to simulate the melt-water pulse 1A dated at −14.6 ka, which led to a sea-level rise of about 20 m in a few centuries. They found that a freshwater from Antarctica explains the warming in North Atlantic, via the decline of AABW and the strengthening of NADW. Our study is consistent with this result because the cooling in Antarctica and Southern Ocean in our study is indeed associated with a decrease of AABW during the largest freshwater perturbation (H-S). The resulting warming in North Atlantic is consistent with a positive northward heat transport.

The coupled model response (CGG) to Psouth perturbation is depicted in Fig. 6 (right). Compared to Pnorth results (Fig. 5 right), the temperature pattern is reverse of what is expected. At the LGM, the ice sheets cool the surface temperature in both Northern and Southern Hemispheres except for a small part over the East Antarctica area. In the Northern high latitudes, the altitude of the ice sheets is higher in the CGG model compared to those produced by the CLIMBER smoothed topography. In other words, through the temperature-altitude relation, the ice sheets contribute to a decrease of the Northern Hemisphere surface temperature.

For the interglacial periods, both simulations show similar temperature patterns (Fig. 6 right): a cooling of the Northern high latitudes and in some parts of the Austral Ocean and West Antarctica, and a warming over East Antarctica. The cooling over Alaska is explained by an ice nucleation, which will be discussed in the next section. The impact of ice sheets tends to further cool the Northern Hemisphere (mainly at LGI) and West Antarctica and warm East Antarctica. Among the few number of studies based on a southern freshwater perturbation, Swingedouw et al. (2008) showed that the AABW formation was decreased in response to the Antarctic ice-sheet melting obtained under a 4×CO_2 scenario, which weakens the Atlantic meridional overturning circulation. Whereas our simulations are performed for different contexts, the Psouth perturbation leads to an AABW decrease and to a slight weakening of NADW (not shown), which is similar to changes of ocean circulation described in Swingedouw et al. (2008).

The seesaw balance has been pinpointed for both freshwater perturbations. Both hemispheric surface temperatures result from the overturning changes. Ice-sheets amplify the seesaw according to Pnorth, whereas in Psouth, the ice-sheet build-up over Northern Hemisphere inhibits mainly the seesaw effect

6 The impact of a freshwater perturbation on the rapid build-up of an ice sheet at LGI

Our approach in this section is to quantify, at LGI, using a climate/ice sheet coupled model how the prescribed perturbation of freshwater may contribute to enhance the ice sheet build-up.

The CLIMBER model has already been used to simulate the end of the previous interglacial and the onset of glaciation (Khodri et al., 2003; Calov et al., 2005b; Ganopolski et al., 2010). Using the CLIMBER-GREMLINS model, Kageyama et al. (2004) produced a more reliable onset with a 17 m simulated sea-level drop between −126 and −106 ka. Using a new coupling scheme accounting for the inversion of temperatures over ice sheets, Bonelli et al. (2009) recently succeeded in producing a sea-level decrease of 28 m around 110 ka BP. However, this is still underestimated compared to sea-level reconstructions (Peltier, 2004; Lambeck, 2004; Fleitout et al., 2006). Calov et al. (2005b) performed also the onset of the glaciation with CLIMBER and the ice-sheet model SICOPOLIS. They showed that the sea-level is lowered between 50 and 80 m between 125 and 110 ka but the Fennoscandian ice-sheet does not agree to the pollinic and geomorphologic data. According to Waelbroeck et al. (2002), the sea-level deduced from a compilation of coral and sediment data is dropped by around 45 m between stage 5e and 5d (115 and 110 ka, respectively). Furthermore, data provided by Landais et al. (2006), Huber et al. (2006) and Sánchez Goñi et al. (2008), demonstrate

CPD
6, 1077–1110, 2010

that, just after the onset of the glaciation (115 ka), the North high latitudes record high and abrupt frequency temperature changes.

Under no freshwater perturbation, with the CGG STD experiment, we simulate a sea-level drop of only 7.5 m for LGI. Applying the perturbation in the South does not really change the response compared to the North perturbations at LGI (not shown). In contrast, adding a freshwater perturbation in the North Atlantic, which slows down NADW, largely contributes to an additional growth of the ice sheet over the North American continent (Figs. 7 and 8). The Fig. 5 (LGI, right) shows clearly a cooling in the North America between the Heinrich and the DO event when ice-sheets are interactive. With our Pnorth perturbation, the ice-sheet build-up is enhanced and the sea-level drops by more than 16.2 m (i.e. 8.7 m more than in the STD experiment) by producing a large positive feedback on the hydrological cycle. This is mainly achieved through an enhancement of the snow accumulation in North America (Fig.8). The difference between our results and the data (Waelbroeck et al., 2002) may come from the fact the experiment is not a transient simulation.

Whereas many modeling groups performed simulations of the onset of glaciation (e.g. Khodri et al., 2003; Vettoretti and Peltier, 2004; Wang and Mysak (2002); Calov et al., 2005b), very few of them carried out transient experiments around this period. A very challenging question is to understand the rapid sea-level drop as shown by geological records. Our LGI experiment cannot reproduce the 45 m sea-level drop because it was not beyond the scope of this study to run a transient simulation consistent with available data. But when we account for a high frequency freshwater perturbation in North Atlantic, the simulated sea-level drop is about twice larger.

7 Conclusions

We performed a series of simulations of Present-Day, Last Glacial Inception and Last Glacial Maximum climates using climate/ice-sheet coupled model. With such a model, we have investigated the fast feedbacks between ice-sheets and climate. We demon-

CPD
6, 1077–1110, 2010

strate, through hysteresis curves, that the simulated climate is more unstable when ice sheets are considered as an interactive component. This study shows that the positive feedbacks, between cryosphere and climate, amplify the freshwater/salt perturbation through ocean dynamics.

For the three periods under consideration, we applied a similar perturbation than that derived from Ganopolski and Rahmstorf (2001) to mimic freshwater perturbation associated with Heinrich and DO events. First we checked that the LGM period, and with CLIMBER only, we got similar results than Ganopolski and Rahmstorf (2001). We enlarge the scope to show that, for the three climatic contexts, when the ice sheets are interactive, the surface temperature changes are enhanced in both hemispheres by the freshwater perturbation. Moreover, the far-field interactions between southern and northern hemispheres (seesaw effect) are always simulated for the three climates when the perturbation is applied in the North (Pnorth), whereas for Psouth it is not the case. Nevertheless, the amplitude of the perturbation may be too weak.

Finally, for LGI, the Pnorth experiment demonstrates that the freshwater perturbation largely enhances the ice-sheet build-up. The role of freshwater perturbation to explain a 45 m sea-level drop in less than 10 000 years at the glacial onset has therefore been shown to be a crucial process in this study. Therefore the large climate perturbations that occurred just after the inception (LGI) certainly contribute to the ice-sheet build-up and the fast and large amplitude sea-level change recorded by data. With such a tool can be applied to paleo and future climates. Recently, Bonelli et al. (2009) have shown that such cryosphere-climate model is able to simulate the last glacial-interglacial cycle. Nevertheless, this simulation only reproduces low frequency variations (orbital). To capture high frequency variations (Dansgaard-Oeschger, Heinrich Events) new studies, based on the approach developed in this work, will be necessary.

Acknowledgements. This work has been supported by PNEDC French National Programs IMPAIRS. Computing was carried out at LSCE-IPSL, supported by CEA.

We thank Amaëlle Landais and Jorge Alvarez-Solas (LSCE-IPSL), Eric De Weaver, Stephen J. Vavrus and Anders Carlson (CCR) for their fruitful comments.

The publication of this article is financed by CNRS-INSU.

References

Alley, R. B., Brook, E. J., and Anandakrishnan, S.: A northern lead in the orbital band: north-south phasing of Ice-Age events, Quaternary Sci. Rev., 21, 431–441, doi:10.1016/S0277-3791(01)00072-5, 2002.

Alvarez-Solas, J., Charbit, S., Ritz, C., Paillard, D., Ramstein, G., and Dumas, C.: Links between ocean temperature and iceberg discharge during Heinrich Events, Nature Geosciences, 3, 122–126, doi:10.1038/ngeo752, 2010.

Beckmann, A. and Goosse, H.: A parameterization of ice shelf-ocean interaction for climate models, Ocean Model., 5, 157–170, 2003.

Bender, M., Sowers, T., Dickson, M.-L., Orchardo, J., Grootes, P., Mayewski, P., and Meese, D.: Climate correlations between Greenland and Antarctica during the past 100 000 years, Nature 372, 663–666, doi:10.1038/372663a0, 2002.

Berger, A.: Long-term variations of daily insolation and quaternary climatic changes, Journal of Atmospheric Science, 35, 2362–2367, 1978.

Blunier, T. and Brook, E.: Timing of Millennial-Scale Climate Change in Antarctica and Greenland During the Last Glacial Period, Science, 291(5501), 109–112, doi:10.1126/science.291.5501.109, 2001.

Bond, G. C. and Lotti, R.: Iceberg Discharges into the North Atlantic on Millennial Time Scales During the Last Glaciation, Science, 267, 1005–1010, doi:10.1126/science.267.5200.1005, 1995.

Bonelli, S., Charbit, S., Kageyama, M., Woillez, M.-N., Ramstein, G., Dumas, C., and Quiquet, A.: Investigating the evolution of major Northern Hemisphere ice sheets during the last

glacial-interglacial cycle, Clim. Past, 5, 329–345, 2009, http://www.clim-past.net/5/329/2009/.
Broecker, W. S.: Paleocean circulation during the last deglaciation: A bipolar seesaw, Paleoceanography, 13, 119–121, 1998.
Cacho, I., Grimalt, J. O., Pelejero, C., Canals, M., Sierro, F. J., Flores, J. A., and Shackleton, N.: Dansgaard-Oeschger and Heinrich Event imprints in Alboran Sea paleotemperatures, Paleoceanography, 14, 698–705, doi:10.1029/1999PA900044, 1999.
Calov, R., Ganopolski, A., Claussen, M., Petoukhov, V., and Greve, R.: Transient simulation of the last glacial inception, Part I: glacial inception as a bifurcation of the climate system. Clim. Dynam., 24,545–561, doi:10.1007/s00382-005-0007-6, 2005a.
Calov, R., Ganopolski, A., Petoukhov, V., Claussen, M., Brovkin, V., and Kubatzki, C.: Transient simulation of the last glacial inception, Part II: sensitivity and feedback analysis, Clim. Dynam., 24, 563–576, 2005b.
Clark, P., Pisias, N., Stocker, T., and Weaver, A.: The role of the thermohaline circulation in abrupt climate change, Nature, 415, 853–859, doi:10.1038/415863a, 2002.
Clark, P. U., Dyke, A. S., Shakun, J. D., Carlson, A. E., Clark, J., Wohlfarth, B., Mitrovica, J. X., Hostetler, S. W., and McCabe, A. M.: The Last Glacial Maximum, Science, 325, 710–714, doi:10.1126/science.1172873, 2009.
Claussen, M., Ganopolski, A., Brovkin, V., Gerstengarbe, F.-W., and Werner, P.: Simulated Global-Scale Response of the Climate System to Dansgaard-Oeschger and Heinrich Events, Clim. Dynam., 21, 361–370, doi:10.1007/s00382-003-0336-2, 2003.
Claussen, M., Mysak, L. A., Weaver, A. J., Crucifix, M., Fichefet, T., Loutre, M.-F., Weber, S. L., Alcamo, J., Alexeev, V. A., Berger, A., Calov, R., Ganopolski, A., Goosse, H., Lohman, G., Lunkeit, F, Mokhov, I. I., Petoukhov, V., Stone, P., and Wang, Z.: Earth System Models of Intermediate Complexity: Closing the Gap in the Spectrum of Climate System Models, Clim. Dynam., 18, 579–586, doi:10.1007/s00382-001-0200-1, 2002.
Chapman, M. R. and Shackleton, N. J.: Global ice-volume fluctuations, North Atlantic ice-rafting events, and deep-ocean circulation changes between 130 and 70 ka, Geology, 27, 9, 795–798, 1999.
Charbit, S., Kageyama, M., Roche, D., Ritz, C., and Ramstein, G.: Investigating the mechanisms leading to the deglaciation of past continental Northern Hemisphere ice sheets with the CLIMBER-GREMLINS coupled model, Global Planet. Change, 48, 253–273, doi:10.1016/j.gloplacha.2005.01.002, 2005.

Dansgaard, W., Clausen, H. B., Gunderstrup, N., Hammer, C. U., Johnsen, S. J., Kristinsdottir, P. M., Rech, N.: A new Greenland ice core, Science, 218, 1273–1277, doi:10.1126/science.218.4579.1273, 1982.

Dansgaard, W., Johnsen, S. J., Clausen, H. B., Dahl-Jensen, D., Gunderstrup, N., Hammer, C. U., Hvidberg, C. S., Steffensen, J. P., Sveinbjornsdottir, A. E., Jouzel, J., and Bond, G.: Evidence for general climate instability of past climate from a 250-kyr ice core record, Nature, 364, 218–220, doi:10.1038/364218a0, 1993.

Dumas, C.: Modélisation de l'évolution de l'Antarctique depuis le dernier cycle glaciaire-interglaciaire jusqu'au futur: importance relative des différents processus physiques et rôle des données d'entrée, PhD thesis, Laboratoire de Glaciologie et Géophysique de l'Environnement, University of Joseph Fourier Grenoble, France, 183 pp., 2002.

European Project for Ice Coring in Antarctica EPICA: One-to-one coupling of glacial climate variability in Greenland and Antarctica, Nature, 444, 195–198, doi:10.1038/nature05301, 2006.

Fleitout, L. Krien, Y., and Cailletaud, G.: Post-Glacial Rebound: Lateral Viscosity Variations, Transient Creep and the Constraints on Holocene ice-sheets, EGU General Assembly, Vienna, Austria, 2–7 April 2006, EGU06-A-09224, 2006.

Gallée, H., van Ypersele, J.-P., Fichefet, T., Tricot, C., and Berger, A.: Simulation of the last glacial cycle by a coupled, sectorially averaged climate-ice sheet model, I. The climate model, J. Geophys. Res.-Atmos., 96, 13,139–13,161, 1991.

Ganopolski, A. and Rahmstorf, S.: Rapid changes of glacial climate simulated in a coupled climate model, Nature, 409, 153–158, doi:10.1038/35051500, 2001.

Ganopolski, A., Calov, R., and Claussen, M.: Simulation of the last glacial cycle with a coupled climate ice-sheet model of intermediate complexity, Clim. Past, 6, 229–244, 2010, http://www.clim-past.net/6/229/2010/.

Heinrich, H.: Origin and consequences of cyclic ice rafting in the northeast Atlantic Ocean during the past 130 000 years, Quaternary Res., 29, 143–152, 1988.

Huber C., Leuenberger, M., Spahni, R., Flückiger, J., Schwander, J., Stocker, T. F., Johnsen, S., Landais, A., and Jouzel, J.: Isotope calibrated greenland temperature record over marine isotope stage 3 and its relation to CH_4, Earth Planet. Sc. Lett., 243, 504–519, 2006.

Imbrie J., Hays, J., Martinson, D., McIntyre, A., Mix, A., Morley, J., Pisias, N., Prell, W., Shackleton, N.: The orbital theory of Pleistocene climate: support from a revised chronology of the marine δ^{18}O record, in: A. Berger, J. Imbrie, H. Hays, G. Kukla and B. Saltzman (Eds.):

Milankovitch and Climate: Understanding the Response to Astronomical Forcing, Proceedings of the NATO Advanced Research Workshop held 30 November–4 December, 1982, in Palisades, NY, 269–305, 1984.

Johnson, R. G. and Andrews, J. T.: Rapid ice-sheet growth and initiation of the last glaciation, Quaternary Res., 12, 119–134, 1979.

Kageyama, M., Charbit, S., Ritz, C., Khodri, M., and Ramstein, G.: Quantifying ice-sheet feedbacks during the last glacial inception, Geophys. Res. Lett., 31, L24203, doi:10.1029/2004GL021339, 2004.

Khodri, M., Leclainche, Y., Ramstein, G., Braconnot, P., Marti, O., and Cortijo, E.: Simulating the amplification of orbital forcing by ocean feedbacks in the last glaciation, Nature, 410, 570–574, doi:10.1038/35069044, 2001.

Khodri, M., Ramstein, G., Paillard, D., Duplessy, J. C., Kageyama, M., and Ganopolski, A.: Modelling the climate evolution from the last interglacial to the start of the last glaciation: The role of arctic ocean freshwater budget, Geophys. Res. Lett., 30, 1606, doi:10.1029/2003GL017108, 2003.

Knutti, R., Flückiger, J., Stocker, T. F., Timmermann, A.: Strong hemispheric coupling of glacial climate through freshwater discharge and ocean circulation, Nature, 430, 851–856, doi:10.1038/nature02786, 2004.

Lambeck, K.: Sea-level change through the last glacial cycle: geophysical, glaciological and palaeogeographic consequences, CR Geosci., 336 , 677-689, 2004.

Landais, A., Masson-Delmotte, V., Jouzel, J., Raynaud, D., Johnsen, S., Leuenberger, M., Schwander, J., and Minster, B.: The glacial inception recorded in the NorthGRIP Greenland ice core: information from air isotopic measurements, Clim. Dynam., 26(2–3), 273–284, doi:10.1007/ss00382-005-0063-y, 2005a.

Landais A., Jouzel, J., Masson-Delmotte, V., and Caillon, N.: Large temperature variations over rapid climatic events in greenland: a method based on air isotopic measurements, C. R. Geosci., 337, 947–956, 2005b.

LeMeur, E. and Huybrechts, P.: A comparison of different ways of dealing with isostasy: Examples from modelling the Antarctic ice sheet during the last glacial cycle, Ann. Glaciol., 23, 309–317, 1996.

Marchal, O., Stocker, T. F., and Joos, F.: A latitude-depth, circulation-biogeochemical ocean model for paleoclimate studies: model development and sensitivities, Tellus, 50B, 290–316, 1998.

Masson-Delmotte, V., Jouzel, J., Landais, A., Stievenard, M., Johnsen, S. J., White, J. W. C., Sveinbjornsdottir, A., and Fuhrer, K.: GRIP Deuterium excess reveals rapid and orbital-scale changes of Greenland moisture origin, Science, 309, 118–121, doi:10.1126/science.1108575, 2005.

North Greenland Ice Core Project NGRIP: High-resolution record of Northern Hemisphere climate extending into the last interglacial period, Nature, 431, 147–151, doi:10.1038/nature02805, 2004.

Paillard, D.: Glacial Hiccups, Nature, 409, 441–442, doi:10.1038/35051691, 2001.

Peltier, W. R.: Global Glacial Isostasy and the Surface of the Ice-Age Earth: The ICE-5G (VM2) Model and GRACE, Annu. Rev. Earth Pl. Sc., 32, 111–149, doi:10.1146/annurev.earth.32.082503.144359, 2004.

Petoukhov, V., Ganopolski, A., Brovkin, V., Claussen, M., Eliseev, A., Kubatzki, C., and Rahmstorf, S.: CLIMBER-2: a climate system model of intermediate complexity: part I. Model description and performance for present climate, Clim. Dynam., 16, 1–17, 2000.

Petit, J. R., Jouzel, J., Raynaud, D., Barkov, N. I., Barnola, J. M., Basile, I., Bender, M., Chappellaz, J., Davis, M., Delaygue, G., Delmotte, M., Kotlyakov, V.M., Legrand, M., Lipenkov, V. Y., Lorius, C., Pepin, L., Ritz, C., Saltzman, E., and Stievenard, M.: Climate and atmospheric history of the past 420 000 years from the Vostok ice core, Antarctica, Nature, 399, 429–436, doi:10.1038/20859, 1999.

Philippon, G., Ramstein, G., Charbit, C., Kageyama, M., Ritz, C., and Dumas, C.: Evolution of the Antarctic ice sheet throughout the last deglaciation: A study with a new coupled climate-north and south hemisphere ice sheet model, Earth Planet. Sc. Lett., 248, 750–758, doi:10.1016/j.epsl.2006.06.017, 2006.

Rahmstorf, S.: Bifurcations of the Atlantic thermohaline circulation in response to changes in the hydrological cycle, Nature, 378, 145–149, 1995.

Rignot, E. and Jacobs, S. S.: Rapid bottom melting widespread near Antarctic ice sheet grounding lines, Science, 296, 2020–2023, doi:10.1126/science.1070942, 2002.

Ritz, C., Fabre, A., and Letréguilly, A.: Sensitivity of a Greenland ice sheet model to ice flow and ablation parameters: consequences for the evolution through the last climatic cycle, Clim. Dynam., 13, 11–24, 1997.

Ritz, C., Rommelaere, V., Dumas, C.: Modeling the evolution of Antarctic ice sheet over the last 420 000 years: implications for altitude changes in the Vostok region, J. Geophys. Res.-Atmos., 106(D23), 31943–31964, 2001.

CPD
6, 1077–1110, 2010

Climate response to freshwater perturbations

G. Philippon-Berthier et al.

Rohling, E. J., Marsh, R., Wells, N. C., Siddall, M., and Edwards, N. R.: Similar meltwater contributions to glacial sea level changes from Antarctic and northern ice sheets, Nature, 430, 1016–1021, doi:10.1038/nature02859, 2004.

Sánchez Goñi, M. F., Landais, A., Fletcher, W. J., Naughton, F., Desprat, S., and Duprat, J.: Contrasting impacts of Dansgaard-Oeschger events over a western European latitudinal transect modulated by orbital parameters, Quaternary Sci. Rev., 27, 1136–1151, doi:10.1016/j.quascirev.2008.03.003, 2008.

Seidov, D., Maslin, M.: Atlantic Ocean piracy and the bipolar climate see-saw during Heinrich and Dansgaard-Oeschger events, J. Quaternary Sci., 16, 321–328, doi:10.1002/jqs.595, 2001.

Severinghaus, J. P. and Brook, E. J.: Abrupt Climate Change at the End of the Last Glacial Period Inferred from Trapped Air in Polar Ice, Science, 286, 930–934, doi:10.1126/science.286.5441.930, 1999.

Shackleton, N.J.: Oxygen isotopes, ice volume and sea level, Quaternary Sci. Rev., 6, 183–190, doi:10.1016/0277-3791(87)90003-5, 1987.

Stocker, T. F., Wright, D. G., and Mysak, L. A.: A zonally averaged, coupled ocean-atmosphere model for paleo-climate studies, J. Climate, 5, 773–797, 1992.

Stocker, T.F.: The seesaw effect, Science, 282, 61–62, doi:10.1126/science.282.5386.61, 1998.

Stouffer, R.J., Seidov, D., and Haupt, B. J.: Climate Response to External sources of Freshwater: North Atlantic versus the Southern Ocean, J. Climate, 20, 436–448, 2007.

Swingedouw, D., Fichefet, T., Huybrechts, P., Goosse, H., Driesschaert, E., and Loutre, M.-F.: Antarctic ice-sheet melting provides negative feedbacks on future climate warming, Geophys. Res. Lett., 35, L17705, doi:10.1029/2008GL034410, 2008.

Vettoretti, G. and Peltier, W. R.: The Sensitivity of Glacial Inception to Orbital and Greenhouse Gas Climate Forcing, Quaternary Science Reviews, 23, 499–519, doi:10.1016/j.quascirev.2003.08.008, 2004.

Waelbroeck, C., Labeyrie, L., Michel, E., Duplessy, J. C., McManus, J. F., Lambeck, K., Balbon, E., and Labracherie, M.: Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records, Quaternary Sci. Rev., 21, 295–305, 2002.

Wang, Z. and Mysak, L. A.: Glacial abrupt climate changes and Dansgaard-Oeschger oscillations in a coupled climate model, Paleoceanography, 21, PA2001, doi:10.1029/2005PA001238, 2006.

Wang, Z. and Mysak, L. A.: A simple coupled Atmosphere-Ocean-Sea Ice-Land Surface Model for climate and paleoclimate studies, J. Climate, 13, 1150–1172, 2000.
Weaver, A. J., Saenko, O. A., Clark, P. U., and Mitrovica, J. X.: Meltwater Pulse 1A from Antarctica as a trigger of the Bølling-Allerød warm interval, Science, 299, 1709–1713, doi:10.1126/science.1081002, 2003.

CPD
6, 1077–1110, 2010

Table 1. Boundary conditions.

	Insolation	Atmospheric CO_2	Sea-level variation
PD	Pre-industrial	280 ppm	0.0 m
LGM	−21 ka	190 ppm	−127.5 m
LGI	−115 ka	267 ppm	−20.0 m

CPD
6, 1077–1110, 2010

Table 2. Acronyms for the experiments.

C	CLIMBER model
CGG	model fully coupled = CLIMBER+GREMLINS+GRISLI
STD	simulation CGG without any perturbations
Pnorth	simulation with freshwater perturbation in the North Atlantic Ocean
Psouth	simulation with freshwater perturbation in the Antarctic Ocean

CPD
6, 1077–1110, 2010

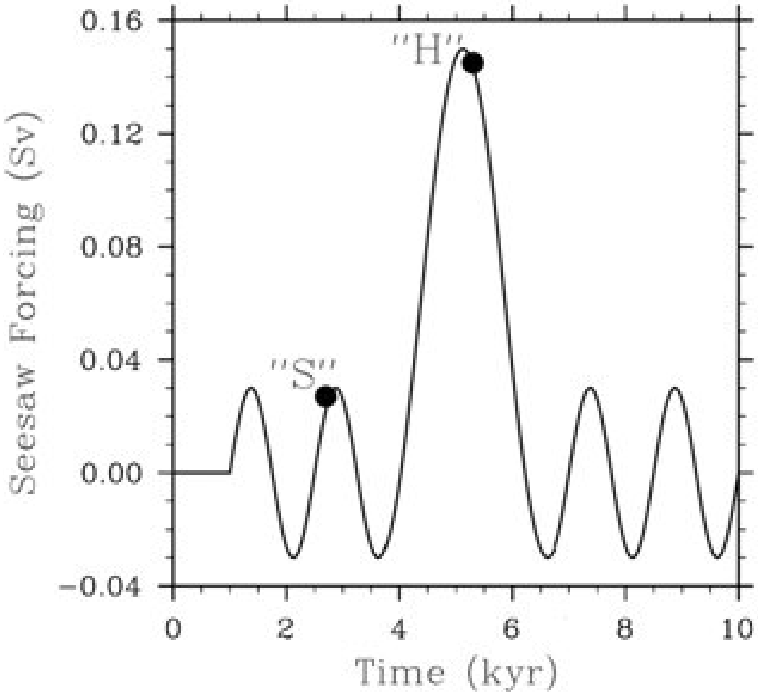

Fig. 1. Pnorth freshwater forcing (in Sv) along the time of the simulation. The stadial event, noted "S" is defined as a cold DO at 2.7 kyr; the Heinrich Event, noted "H" is defined when the freshwater perturbation is the highest, at 5.3 kyr. This scenario is derived from Ganopolski and Rahmstorf (2001). Psouth is three times higher in amplitude.

CPD
6, 1077–1110, 2010

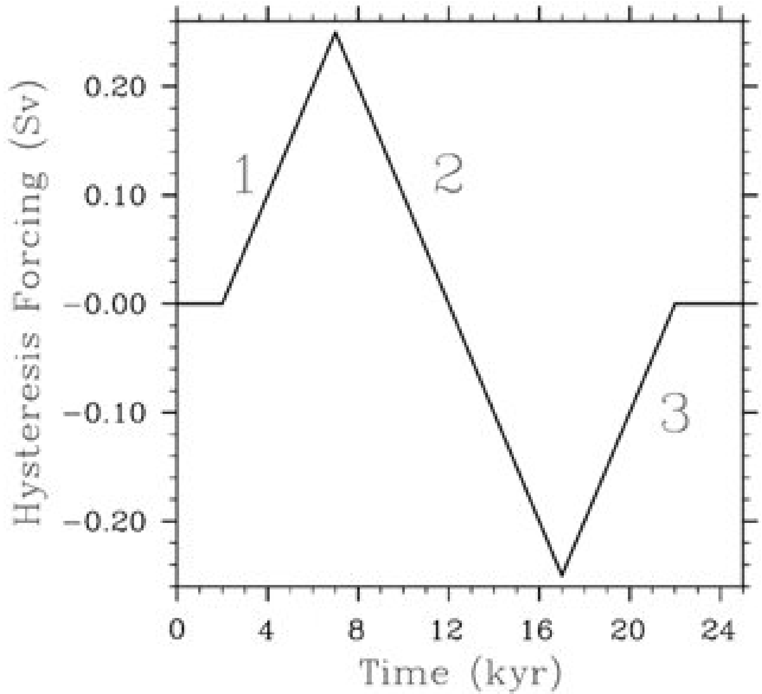

Fig. 2. Freshwater forcing (in Sv) for the hysteresis experiments.

CPD
6, 1077–1110, 2010

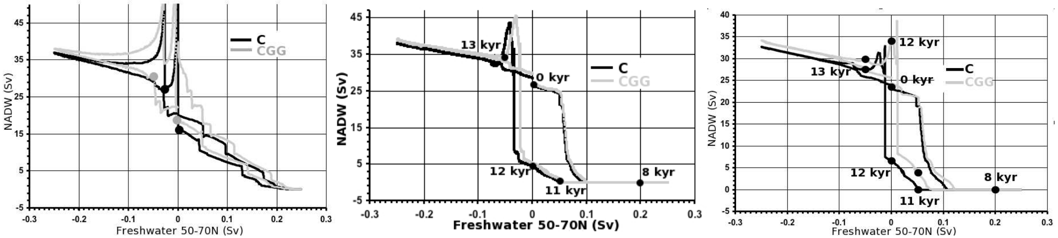

Fig. 3. Hysteresis diagrams for the Atlantic thermohaline circulation for the North Atlantic freshwater perturbation for the CLIMBER-GREMLINS-GRISLI experiment (grey curve), the CLIMBER experiment (black curve). A/ LGM. B/ PD. C/ LGI.

CPD
6, 1077–1110, 2010

Climate response to freshwater perturbations

G. Philippon-Berthier et al.

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◂◂ | ▸▸
◂ | ▸
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

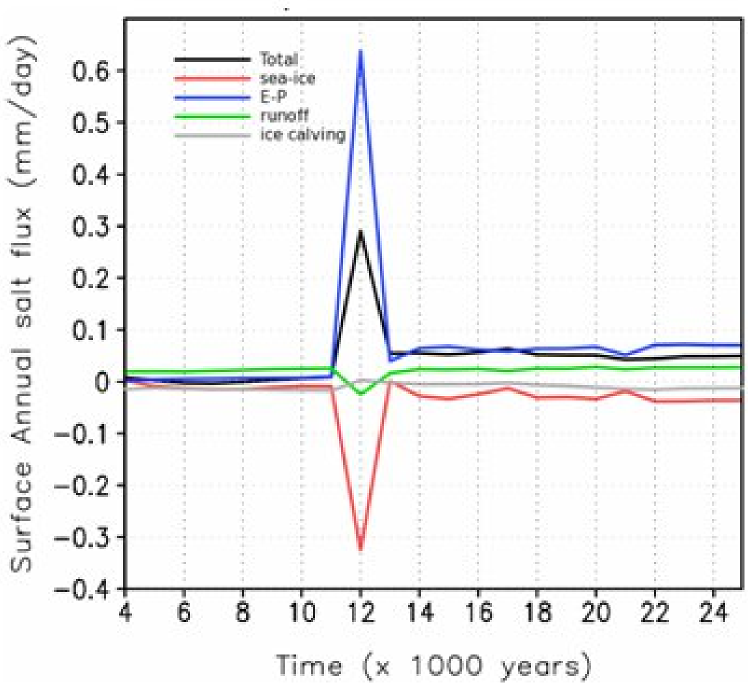

Fig. 4. LGI Surface Atlantic annual salt flux difference (mm/day) between CGG and C (CGG-C) at 60° N, and along 4 kyr to 25 kyr. The surface annual salt flux is equals to: (evaporation – precipitation, blue curve) – runoff-ice calving + sea-ice component. When the values are positive, the ocean gains salt and the atmosphere loses salt; for negative values, the ocean wins freshwater and the atmosphere loose freshwater.

CPD
6, 1077–1110, 2010

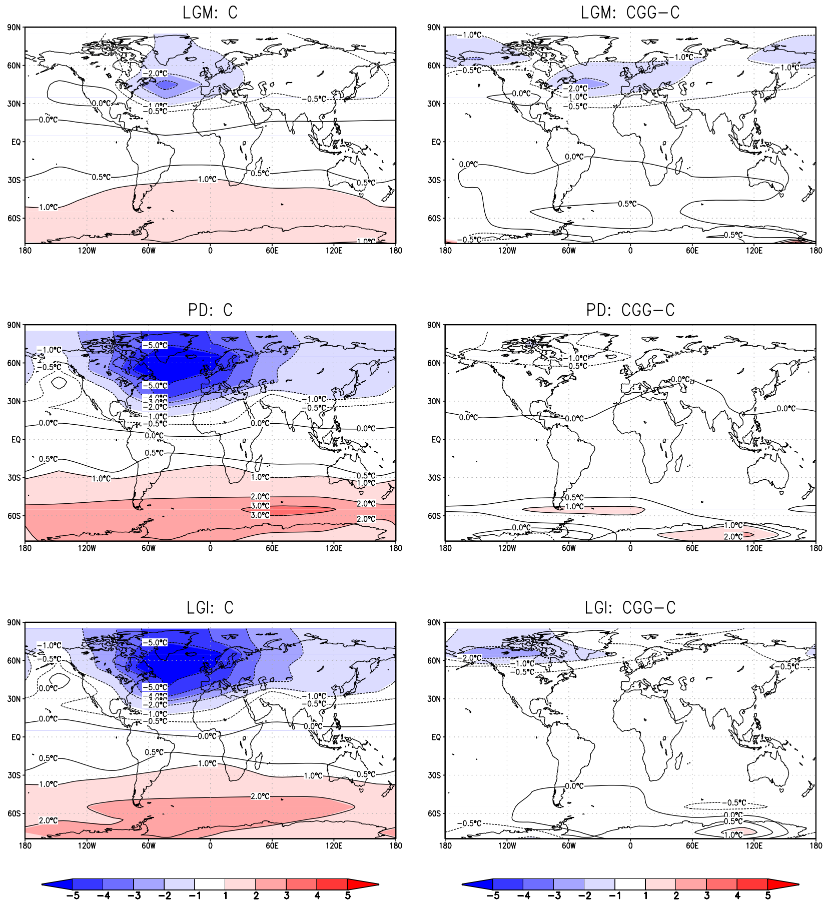

Fig. 5. Pnorth Surface annual air temperature (°C) difference between the Heinrich Event (time = 5.3 kyr) minus the stadial phase (time = 2.7 kyr) for LGM (top), PD (middle) and LGI (bottom). In left, the figures show the difference (H-stadial) from the C experiment and in right the difference (H-stadial) between the CGG and C experiments.

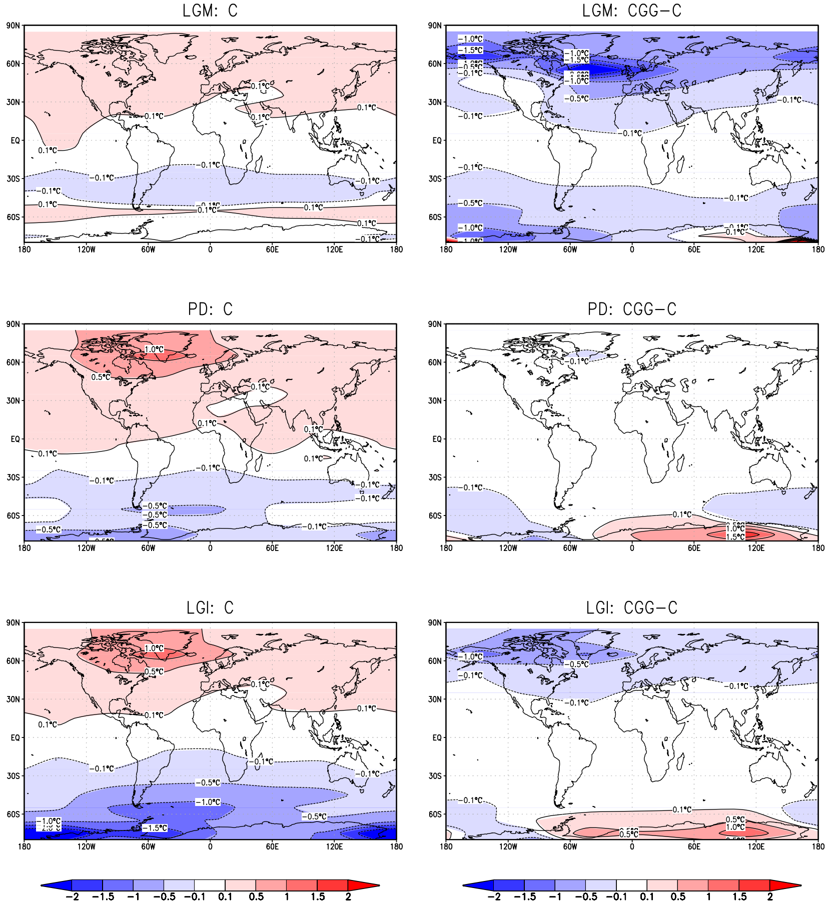

Fig. 6. Same as Fig. 5 but only for the Psouth perturbation. Note the color scale is different.

CPD
6, 1077–1110, 2010

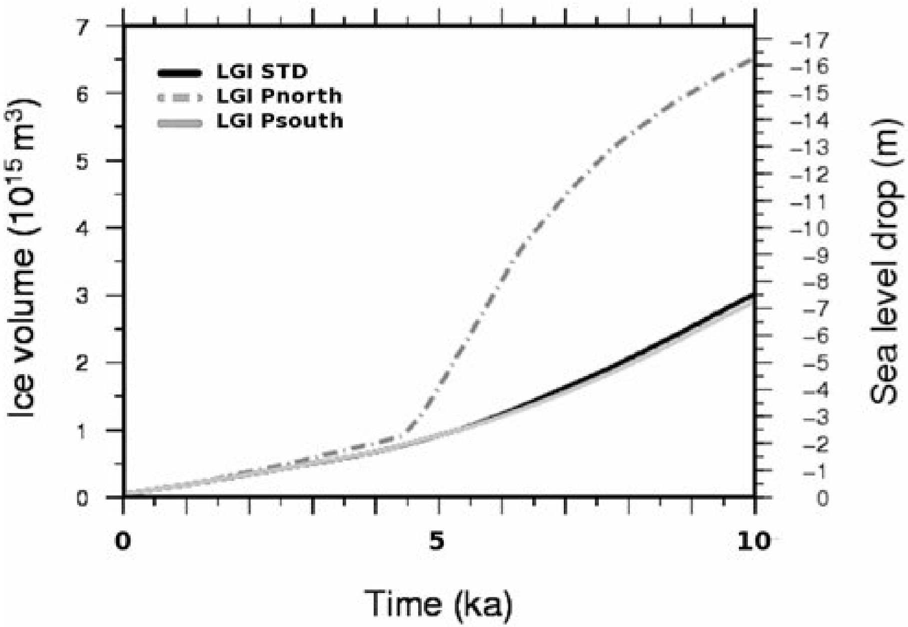

Fig. 7. Evolution of the ice volume (left) and sea-level drop corresponding (right) for North America for the LGI simulation. The black curve refers to the STD simulation at 115 ka, the dark gray dashed line, is for the Pnorth simulation and the light gray line for the Psouth simulation.

CPD
6, 1077–1110, 2010

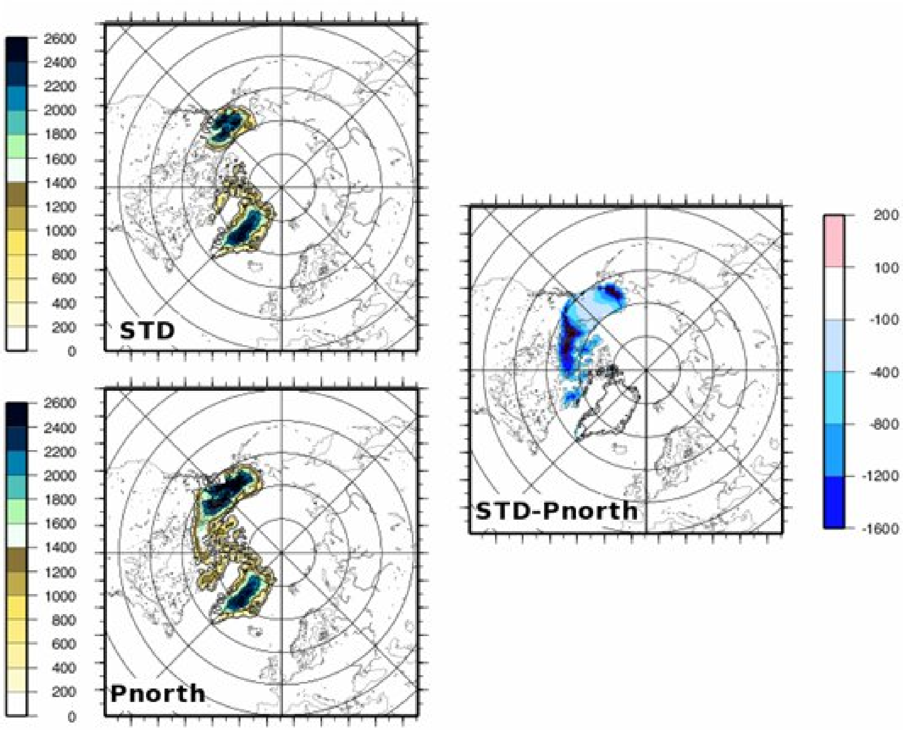

Fig. 8. Ice thickness (in meters) simulated by the STD and Pnorth LGI experiments The Difference (STD-Pnorth) represents the STD ice thickness minus Pnorth ice thickness.