Clim. Past Discuss., 7, 1363–1392, 2011
www.clim-past-discuss.net/7/1363/2011/
doi:10.5194/cpd-7-1363-2011

Climate of the Past Discussions

Systematic study of the fresh water fluxes impact on the carbon cycle

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Received: 4 April 2011 – Accepted: 15 April 2011 – Published: 26 April 2011

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

Abstract

During glacial periods, atmospheric CO_2 concentration rapidly increases and decreases by around 15 ppm at the same time as climate experiments an abrupt cooling in the North Hemisphere and warming in the South Hemisphere. Such a climate change can be triggered in models by adding fresh water fluxes (FWFs) in the North Atlantic. Yet the impact on the carbon cycle is less straightforward, and previous studies give opposite results. Because both models and added fresh water fluxes were different in these studies, it prevents any direct comparison and hinders finding an explanation for these discrepancies. In this study we use the CLIMBER-2 coupled climate carbon model to explore the impact of different additional fresh water fluxes in various conditions, including the experiments previously performed with other models. We show that the CO_2 changes caused by the fresh water flux events should be interpreted as a combination of oceanic and terrestrial processes. The initial state of the Atlantic Meridional Overturning Circulation (AMOC) prior to the addition of fresh water fluxes appears to play a crucial role. The rapid increase of CO_2 observed in ice core data can only be accounted for when the export of North Atlantic Deep Water (NADW) is relatively slow. Additionally, the terrestrial and oceanic carbon reservoirs responses are a consequence of the climate change and most importantly of the "seesaw" effect. As the latter is different in the various models it results in widely different evolution of the vegetation and oceanic carbon reservoirs. The discrepancies between the different studies can thus be explained by a combination of these factors: initial climatic and carbon cycle states, characteristics of the added fresh water flux, AMOC initial state and model "seesaw" pattern.

1 Introduction

During glacial periods, the global climate experiences rapid temperature shifts as recorded by numerous proxies from ice and sediment cores. These changes, called abrupt events (Clement and Peterson, 2008), are characterized by cool conditions in the north and simultaneous gradual warming in the south, followed by a return close to initial values (EPICA community members, 2006; Ahn and Brook, 2008; Barker et al., 2009). During these events, ice core records indicate that atmospheric CO_2 increases rapidly by around 15 ppm, and then decreases back to its initial level (Ahn and Brook, 2008).

Fresh water flux inputs in the Atlantic Ocean have been suggested as a trigger for such abrupt changes. Model simulations show that adding fresh water fluxes to the North Atlantic results in a significant Atlantic Meridional Overturning Circulation (AMOC) decrease (Stocker and Wright, 1991; Ganopolski and Rahmstorf, 2001). It leads to less heat being transported to the North Hemisphere, hence a cooling in the north and warming in the south (the so called "bipolar seesaw", Crowley, 1992; Stocker, 1998). Proxy data (presence of ice rafted debris), indicate that massive iceberg discharges happened simultaneously with some of these events (Heinrich events), which brings support to the hypothesis of fresh water input (Bond et al., 1993; Heinrich, 1988; Ruddiman, 1977). Yet the causes of the iceberg discharges are still debated (Alvarez-Solas et al., 2010).

Additionally, models studies have shown that fresh water inputs can also impact the carbon cycle. However, these studies are sparse and their results vary widely. Using an ocean model only, Marchal et al. (1998) simulated a CO_2 increase of 10–30 ppm due to the decrease of ocean solubility (caused by the warming of the Southern Hemisphere). On the other hand, simulations performed with a terrestrial biosphere model also give a CO_2 increase of approximately 6 ppm and then a decrease back to the initial value (Köhler et al., 2005) (with LGM conditions, FWF forcing of 0.3 Sv during 1000 yr). When both the ocean and terrestrial biosphere are taken into account it results in the coupling of the three main carbon reservoirs relevant for such time scales of a few thousand years, i.e. the ocean, terrestrial biosphere and atmosphere. Two climate-biogeochemical models were used for such a study. The UVic model results in a CO_2 increase of 25 ppm when forced by 0.2 Sv during 1700 yr (Schmittner and Galbraith,

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2008). The atmospheric CO_2 increase is due to the ocean which looses carbon, while the terrestrial biosphere stores more carbon. Another study gives opposite results, with the ocean taking up more carbon and the terrestrial biosphere less, resulting in a CO_2 decrease of around 13 ppm followed by an increase (Menviel et al., 2008) (with the FWF input as a triangular function increasing up to 2 Sv during 200 yr). Such different results could arise from the differences in the models themselves, the type of experiments (e.g. the type of fresh water fluxes added) or the background climate state considered (relatively cold/warm climate, slow/rapid ocean circulation...). To disentangle these effects we use a single model with different fresh water flux experiments and different climate states.

In this study we perform simulations with a climate-carbon coupled model that includes the ocean and the terrestrial biosphere. We systematically study the impact of different fresh water fluxes by modifying their amplitude, shape and duration, with three different background climate conditions. We first perform similar simulations as Schmittner and Galbraith (2008) and Menviel et al. (2008) by adding fresh water fluxes of different duration and shape in the North hemisphere in Preindustrial and Last Glacial Maximum (approximately 21 000 yr ago) conditions, which allow a direct comparison to their results. We also run additional simulations with different amplitudes of the fresh water flux to assess its impact (Fig. 1). We then explore the impact of the same fresh water fluxes in a situation with more realistic glacial CO_2 levels obtained with the "brines" mechanism previously studied (Bouttes et al., 2010, 2011). Finally we evaluate the impact of adding fresh water fluxes in the Southern Hemisphere.

2 Methods

2.1 CLIMBER-2 climate-carbon model

We use the climate-carbon coupled model CLIMBER-2 (Petoukhov et al., 2000; Ganopolski et al., 2001; Brovkin et al., 2002, 2007). CLIMBER-2 is a model of

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intermediate complexity fast enough to run numerous long simulations. The simulations are run for 20 000 yr to ensure the carbon cycle equilibrium. Running numerous simulations allows us to span various fresh water fluxes (changing the amplitude, shape and duration of the flux), background climates (modern and glacial). Hence we can both compare the results with previous studies and complete with additional ones. CLIMBER-2 includes a statistical atmosphere, a zonally averaged ocean and a terrestrial biosphere model (VECODE) (Brovkin et al., 1997).The model computes the carbon cycle both in the ocean and on land, and takes into account carbonate compensation. It has already been used for a number of studies on the links between the climate and the carbon cycle (Brovkin et al., 2002, 2007; Bouttes et al., 2010, 2011).

2.2 Sinking of brines in CLIMBER-2

Brines are small pockets of very salty water released by sea ice formation. Indeed, sea ice is mainly formed of fresh water and most of the salt is rejected from the ice during its formation. In the standard version of CLIMBER-2, the flux of salt released to the ocean goes to the surface oceanic cell. The volume of the latter is quite big due to the coarse resolution, and the salt flux is diluted. Yet as brines are very dense because of their high salt content, they should rapidly sink to the deep ocean when the local conditions allows it. To avoid the dilution of such an effect the brine sink has been previously parameterized and studied in CLIMBER-2 (Bouttes et al., 2010, 2011). The relative importance of the brine mechanism is set by the parameter frac, which is the fraction of salt released by sea ice formation that sinks to the bottom of the ocean. The rest of the salt (1-frac) is mixed in the corresponding surface oceanic cell as done in the standard version. This mechanism was shown to result in a net glacial CO_2 decrease. Previous studies also showed that frac = 0.6 is a good estimation of the frac value (Bouttes et al., 2010, 2011). In this study the frac parameter is thus set to 0.6 when the sinking of brines is taken into account.

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2.3 Design of experiments

We consider three sets of additional fresh water fluxes (Fig. 1). The first one is similar to the Schmittner and Galbraith (2008) experiments in which the duration of the fresh water flux varies between 400 yr and 1700 yr, with an amplitude of 0.2 Sv (Fig. 1a). The second one is similar to the one of Menviel et al. (2008) with a linear fresh water flux increasing from 0 to 2 Sv in 100 yr then a symmetrical decrease during the following 100 yr (Fig. 1b). Finally we complete these two sets with one with varying amplitude of the FWF between 0.05 Sv and 1 Sv for a duration of 400 yr (Fig. 1c). These additional fresh water fluxes correspond to a "meter sea level equivalent" of approximately 7 to 30 m for Schmittner and Galbraith (2008), 18 m for Menviel et al. (2008) and 0.3 to 36 m with the varying amplitude experiments.

These three sets of additional fresh water fluxes are first applied in the North Atlantic (between 50° N and 67.5° N) during three background climates: the Preindustrial (PI), Last Glacial Maximum (LGM), and Last Glacial Maximum with the sinking of brines (LGM+brines). The LGM climate is obtained by modifying the orbital parameters (Berger, 1978), ice sheets (Peltier, 2004), sea level (Waelbroeck et al., 2002) and atmospheric CO_2 (Monnin et al., 2001) for the radiative code. As done in previous studies (Brovkin et al., 2002, 2007; Bouttes et al., 2010, 2011), the radiative CO_2 is imposed so as to obtain a glacial climate. This fixed radiative CO_2 is different from the geochemical CO_2 which is computed interactively by the geochemical model and discussed in this study. This allows for a better comparison with previous work and for a simpler analysis of the carbon cycle response to climate. Finally, we also assess the impact of fresh water fluxes in the Southern Ocean (between 60° S and 75° S) during glacial climate (LGM and LGM+brines) with the third set of fresh water fluxes (different amplitudes).

3 Results and discussion

3.1 Experiments with preindustrial background climate

We first perform a set of experiments with fixed preindustrial boundary conditions, so that the global climate is characteristic of the preindustrial one. Whatever the amplitude or duration of the additional freshwater flux, the AMOC is slowed down as a result of the additional fresh water (Fig. 2a, b, c). If the amount of fresh water reaches the threshold value of 0.2 Sv, the AMOC is even momentarily stopped, a situation called the "off" mode and previously studied with this model in more details (Ganopolski and Rahmstorf, 2001). The duration of the "off" mode depends both on the amplitude and duration of the fresh water flux: the longer and the bigger the flux, the longer the "off" mode lasts.

This alteration of the oceanic circulation impacts the carbon cycle. When the additional fresh water flux is too small to change the AMOC (below the threshold value of 0.2 Sv) the resulting change of CO_2 is small (less than 10 ppm, Fig. 2c and f). Above the 0.2 Sv threshold, when the AMOC shuts down, the impact on the carbon cycle is much more important. The atmospheric CO_2 first decreases for at least 1000 yr, and longer if the "off" mode still persists after 1000 yr. It starts increasing after the AMOC has started increasing again and reaches a new equilibrium with higher CO_2 value (around 300 ppm) as the AMOC also stabilizes at a higher level (around 24 Sv compared to 20 Sv initially). The existence of multiple equilibria is indeed a classical feature of some climate models (Rahmstorf, 1995) and corresponds here to a small shift in the location of convection in the North Atlantic. As we focus our study on the carbon exchanges during the transitory phase of the simulations, the final equilibrium state is here not relevant to our discussion.

The impact of additional FWF on the carbon cycle was previously studied with the LOVECLIM model in the conditions of experiment "Menviel", but with very different results compared to the ones obtained with CLIMBER-2. Indeed, Menviel et al. (2008) obtained exactly the opposite, i.e. a rapid increase of 20 ppm (in approx 100 yr) followed by a more progressive decrease of approximately 25 ppm (in 400 yr). In the similar experiment, CLIMBER-2 simulates a 10 ppm decrease in approximately 1000 yr, then an increase of around 25 ppm in 1000 yr (Fig. 2e). Both the duration and sign of the response are thus different.

These differences originate from the different evolutions of the oceanic and terrestrial biosphere carbon reservoirs. In Menviel et al. (2008) the ocean first takes up more carbon then looses it, while the vegetation looses carbon then takes it up. Because the vegetation reacts faster than the ocean, the CO_2 evolution is primarily driven by the vegetation. The ocean tends to mitigate the changes. On the contrary, in CLIMBER-2, the terrestrial biosphere first takes up carbon while the ocean looses carbon (Fig. 2h and k). Because the amplitude of the change of terrestrial biosphere is twice the one of the ocean, it dominates the CO_2 evolution, the uptake of carbon by the vegetation resulting in the CO_2 decrease.

Why is the ocean loosing carbon and the terrestrial biosphere taking up carbon? In all CLIMBER-2 simulations, ocean looses carbon while the terrestrial biosphere takes it up (Fig. 2). Because of the cessation of the AMOC, less heat is transported from the South Hemisphere to the North Hemisphere, resulting in a temperature rise in the South and a decrease in the North (Fig. 3a). At the same time, precipitation decreases in the Northern Hemisphere and increases in the Southern Hemisphere (Fig. 3b). In the ocean, colder waters increase the solubility. In the colder Northern Hemisphere, the ocean is thus more soluble and takes up more carbon, but this is mitigated by the shut-down of NADW which prevents taking that carbon to the deep ocean. In the Southern Hemisphere, AABW is not changed much (with only a 3 Sv decrease) but because of the temperature rise, the waters are less soluble and therefore take up less carbon. Overall, mainly because of the NADW shut down and the Southern Hemisphere temperature increase, the ocean looses carbon.

The terrestrial biosphere reacts to both to the temperature and precipitation changes. In the colder and dryer Northern Hemisphere, vegetation decreases (Fig. 4a). But the change of climate also tends to lower decomposition and the soils keep more carbon.

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The warmer and more humid conditions of the Southern Hemisphere lead to an increase of the vegetation (Fig. 4b). Because there is more vegetation going to the soils, the latter increase as well. Overall, the vegetation first decreases rapidly then increases while the soils content increase, resulting in a small decrease then increase of the terrestrial biosphere, followed by a decrease when the climate returns towards its initial condition. Hence the atmospheric CO_2 increase is mostly driven by the soil carbon changes from the "seesaw" response.

Why is the carbon-cycle response different in LOVECLIM? Although the additional fresh water flux is the same for both models, the response of the AMOC is different: the AMOC increases as soon as the FWF stops in LOVECLIM (i.e. after 200 yr), whereas in CLIMBER-2 the "off" mode persits for a few more hundred years. The AMOC reaches its new stable level around 600 yr after the beginning of the FWF in LOVECLIM, but after around 1000 yr in CLIMBER-2. The timing of circulation and climate changes is thus different. It is more rapid in LOVECLIM compared to CLIMBER-2, but also different in amplitude. In particular, the "seesaw" effect is relatively small in LOVECLIM resulting in very different vegetation response. On the other hand the change of NADW is more important and faster. This could lead to a greater role of the upper part of the ocean through solubility changes. Because the absolute change of temperature seems of greater amplitude in the Northern Hemisphere, the increased solubility due to the cooling could dominate and explain the increased uptake of carbon from the ocean. A complete explanation could only be obtained by running additional simulations with LOVECLIM.

3.2 Experiments with LGM background climate

In the following we explore the response of the model to additional fresh water fluxes in the context of LGM climate. Two backgrounds conditions are considered: a standard one (LGM) that can be compared to previous studies (Menviel et al., 2008; Schmittner and Galbraith, 2008) and an original one taking into account the sinking of brines rejected by sea ice formation around Antarctica (LGM+brines).

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3.2.1 Experiments with the standard LGM background climate

We now explore the impact of fresh water fluxes on the carbon cycle with a LGM background climate. Compared to the PI results, the simulations with the LGM background climate present one major difference (Fig. 5). The carbon cycle equilibrates at a different level than the initial one due to the AMOC evolution during the PI, while the final level is the same as the initial one with the LGM climate, because the AMOC stabilizes at its initial value. Other than that, the evolution is roughly similar, although the initial values are different due to the different background conditions. This confirms that the final equilibrium state that was found in the PI experiments is not relevant to the transient fluxes of carbon during the FWF perturbation. Consequently, the general evolution of CO_2 is a relatively small decrease, then a bigger increase and a decrease to the initial level.

Two ocean-atmosphere-vegetation models have already studied the carbon cycle response to fresh water fluxes during the LGM. With the LOVECLIM model (Menviel et al., 2008), it again results in an oceanic carbon uptake and loss of carbon from the terrestrial biosphere, but the result is slightly different. Because the ocean takes up carbon more rapidly, it results in a decrease of CO_2 during the first 400 yr, then a small increase. On the other hand, the UVic model (Schmittner and Galbraith, 2008) gives opposite results: the ocean looses carbon while the terrestrial biosphere takes up carbon. It leads to a CO_2 increase in the UVic model and a CO_2 decrease in LOVECLIM, then a return towards near initial conditions.

The CLIMBER-2 results are more similar to the UVic ones, with a loss of carbon from the ocean and an uptake from the terrestrial biosphere. The result is however a bit different, with first a small decrease of CO_2. It then increases as in Schmittner and Galbraith (2008). The CO_2 evolution depends on the tight interplay of ocean and terrestrial biosphere. Because the terrestrial biosphere reacts faster than the ocean its response at the beginning of the fresh water flux addition is important for the CO_2 evolution.

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Why do CLIMBER-2 and Schmittner and Galbraith (2008) seem to be in better agreement, while Menviel et al. (2008) give opposite results? Such discrepancies seem to arise from the different climatic response to the AMOC decrease. In particular, as recently reviewed (Kageyama et al., 2010), the LOVECLIM model exhibits a much smaller climate change compared to the Uvic model. The "seesaw" pattern is more pronounced in the Uvic results, with cooling in the Northern Hemisphere and warming in the Southern Hemisphere. With the LOVECLIM model the warming in the Southern Hemisphere is comparatively very small or not seen. Similarly, in the Uvic model the addition of fresh water fluxes leads to a dryer Northern Hemisphere and wetter Southern Hemisphere, whereas it globally becomes dryer everywhere in the LOVECLIM model. Hence it results in a very different behavior of the carbon cycle both in the terrestrial biosphere and ocean. In the Uvic model, as for CLIMBER-2, the terrestrial biosphere carbon content increases mostly because it becomes warmer and wetter in the Southern Hemisphere leading to a tree cover increase. In LOVECLIM, the climate becomes cooler and dryer in most of the world which triggers a global decrease of tree cover and a loss of carbon from the vegetation. Because it is globally cooler everywhere with LOVECLIM the ocean can store more carbon thanks to an increase in solubility. But in Uvic and CLIMBER-2 the warmer South Ocean leads to a loss of oceanic carbon. The carbon cycle evolution is thus tightly driven by the climate change. As the latter is highly model dependent, the modification of the carbon cycle vary significantly between the different models.

3.2.2 Experiments with LGM background climate and brines

In the previous experiments, the background climate was the LGM, but the simulated CO_2 level was not coherent with the glacial CO_2 level inferred from ice core data (around 190 ppm during the LGM, between 190 ppm and 220 ppm from −65 000 yr to the LGM, compared to approximately 257 ppm in CLIMBER-2). Previous studies have shown that it is possible to simulate the glacial carbon cycle in better agreement with the data by including a brine mechanism (Bouttes et al., 2010, 2011). With the

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following, the rejection of salt during sea ice formation around Antarctica leads to a sinking of very saline and dense water to the deep ocean. It results in a more stratified deep ocean that contains more carbon. We use this LGM with brines climate background to explore the evolution of the carbon cycle when an additional FWF is put in the Atlantic.

The addition of FWF still leads to an uptake of carbon by the terrestrial biosphere and a release of carbon by the ocean. However the amplitude of the terrestrial biosphere increase is smaller, while the oceanic one is similar (Fig. 6). Hence there is no more CO_2 decrease at the beginning of the experiment: CO_2 increases then decreases. This CO_2 evolution is in better agreement with proxy data (Ahn and Brook, 2008).

What can explain these changes? The initial level of NADW export is lower in the LGM+brines simulations compared to the LGM ones (15 Sv compared to 20 Sv in the LGM experiments). When the fresh water flux is added, the decrease of NADW export is smaller, and the induced climate change is slightly smaller (Fig. 7). The changes of terrestrial biosphere are thus less important (Fig. 8) and the corresponding uptake of carbon smaller. Indeed, the ultimate driver of the CO_2 change appears to be both the timing and amplitude of the NADW decrease. As shown on Fig. 9, the maximum CO_2 increase depends on the integral of NADW below its initial value. It thus appears necessary to have a shut down of NADW for a sufficient duration to obtain a CO_2 increase similar to the data (20 ppm). In the model, this requires a substantial addition of fresh water (0.2 Sv during 1700 yr). It shall be noted, however, that the amount of freshwater flux to be added in one model to shutdown the AMOC is very much model dependent. The CO_2 amplitude change during AMOC shutdown thus gives a measure of the length of disruption of the AMOC. Or, more importantly, if we have a period of time in the past with disrupted AMOC and measures of the CO_2 for the same period, we may have an estimation of the rate of AMOC before the shutdown using the CO_2 change as a measure for the rate.

It thus appears that the strength of the NADW prior to the event plays a role because it impacts the amplitude of the climate change. The latter directly impacts the evolution

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of the terrestrial biosphere. In LOVECLIM, the NADW initial level is very high compared to UVic and CLIMBER-2: 30 Sv compared to 20 Sv in CLIMBER-2 at LGM, 15 Sv for LGMb and only 13 Sv in UVic. The initial value of NADW can partly explain the differences in the amplitude of the vegetation response, but not the opposite sign between Menviel and the others which arises from the different "seesaw" patterns obtained.

3.3 FWF in the Southern Hemisphere: similar or different impacts?

We finally perform the same set of experiments as the "Amplitude" ensemble (Fig. 1c), but we now add the FWF in the Southern Hemisphere. We consider two background climate states: the Last Glacial Maximum (LGM) and Last Glacial Maximum with the sinking of brines (LGM+brines). Adding fresh water flux in the Southern Hemisphere impacts mainly the Antarctic Bottom Water (AABW) export whose intensity diminishes (Fig. 10a and b). The NADW export is also slowed down (Fig. 10c and d), but none of them shuts down.

The CO_2 evolution is very different from the one obtained with the FWF in the Northern Hemisphere, showing a decrease then an increase (Fig. 10e and f). This is due to the ocean and terrestrial biosphere reservoirs whose evolutions are the opposite as the ones with the additional FWF in the Northern Hemisphere (Fig. 10g, h, i and j). Indeed, temperature and precipitation now decrease in both hemispheres (Fig. 11), with greater amplitudes in the Southern Hemisphere. The colder and dryer conditions result in a decrease of carbon in the vegetation.

In the ocean, the temperature drop makes the water more soluble. As NADW nor AABW are "off", the surface water which contains more carbon can be sent to the deep ocean, thus the observed ocean carbon uptake. Because the amplitude of the ocean change is more important than the vegetation one, the ocean evolution prevails and it results in a CO_2 decrease. Hence adding fresh water fluxes in the Southern Hemisphere cannot explain the CO_2 increase during Heinrich events. However, it could play a role during the deglaciation, as the data show a momentary stop in the CO_2 increase during the Bolling-Allerod (Ahn and Brook, 2008).

3.4 Comparison with data: discussion

The results are improved both in amplitude and timing for the CO_2 evolution when the sinking of brines is taken into account during the LGM. Both the amplitude and timing match better the CO_2 change as recorded in ice cores (Fig. 12) (Ahn and Brook, 2008). This could be due to the initial state of NADW, whose strength is lower with the brines compared to the simulation without. With the sinking of brines and the induced stratification, carbon is stored in the abyssal glacial ocean (Bouttes et al., 2010, 2011), this different carbon cycle state could explain part of the change, as the ocean contains more carbon and can thus be more sensitive to changes of NADW with respect to loosing or taking carbon. Such an impact of the sinking of brines on the state of the glacial ocean remains to be tested with 3-dimensional models.

A slightly slower NADW is in agreement with data indicating that the intensity of NADW during the LGM is comparable or lower than today, but not shut down nor accelerated (Lynch-Stieglitz et al., 2007; McManus et al., 2004). Furthermore, a more realistic representation of the brines mechanism would depend on climate, with brines deep sinking being reduced when the Southern Ocean temperatures are rising. This additional feedback would likely amplify the carbon oceanic response and consequently the atmospheric CO_2 increase.

Finally, the fresh water flux in the Southern Ocean leads to an opposite CO_2 evolution marked by a decrease followed by an increase. Such an impact of fresh water fluxes from Antarctica could explain the CO_2 plateau observed during the deglaciation, at the time of the Bolling Allerod (around 14 000 to 12 300 yr BP) (Monnin et al., 2001). It corresponds to the period when the Antarctic ice melting and retreat becomes more significant (Clark et al., 2009; Mackintosh et al., 2011).

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4 Conclusions

In conclusion, the response of the carbon cycle strongly depends on a close interplay between the ocean and vegetation responses. The results are very model dependent, especially because the models considered all have a different LGM state and a different "seesaw" pattern in response to the AMOC devrease. In particular the NADW strength differs among the model. The simulation including the sinking of brines not only is in better agreement with LGM proxy data as previously studied (Bouttes et al., 2010, 2011), but it also results in a better response of the carbon cycle to the addition of fresh water fluxes compared to proxy data. It thus supports the crucial role of brines in setting the glacial ocean state and climate. This mechanism and its effects in such experiments should be tested with other models, especially 3-D ocean models.

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

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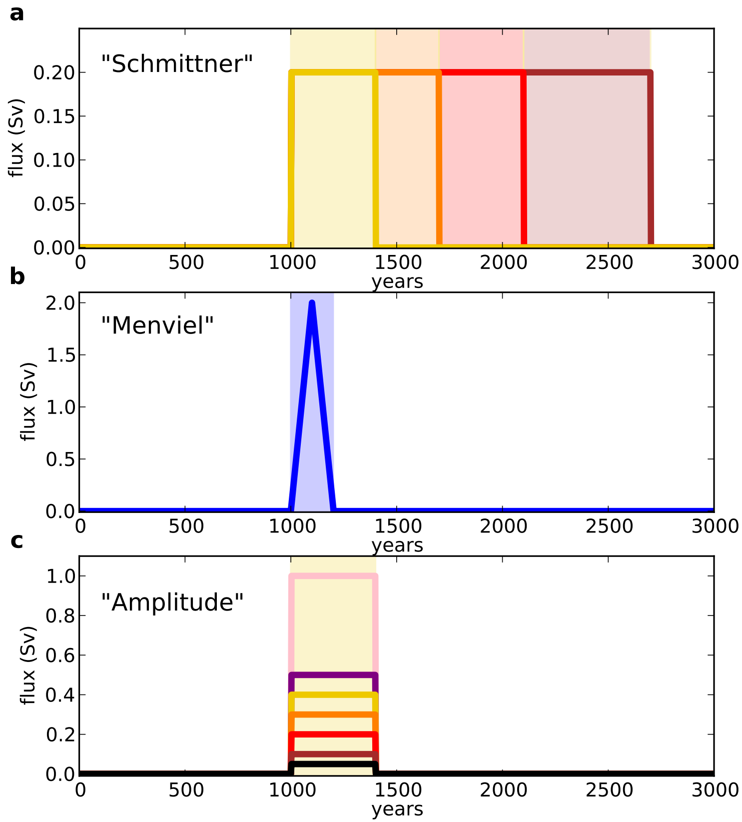

Fig. 1. Evolution of the fresh water flux (Sv) added in the three types of experiments: **(a)** for the "Schmittner" experiments based on Schmittner and Galbraith (2008), **(b)** for the "Menviel" experiments based on Menviel et al. (2008) and **(c)** for the "Amplitude" experiments.

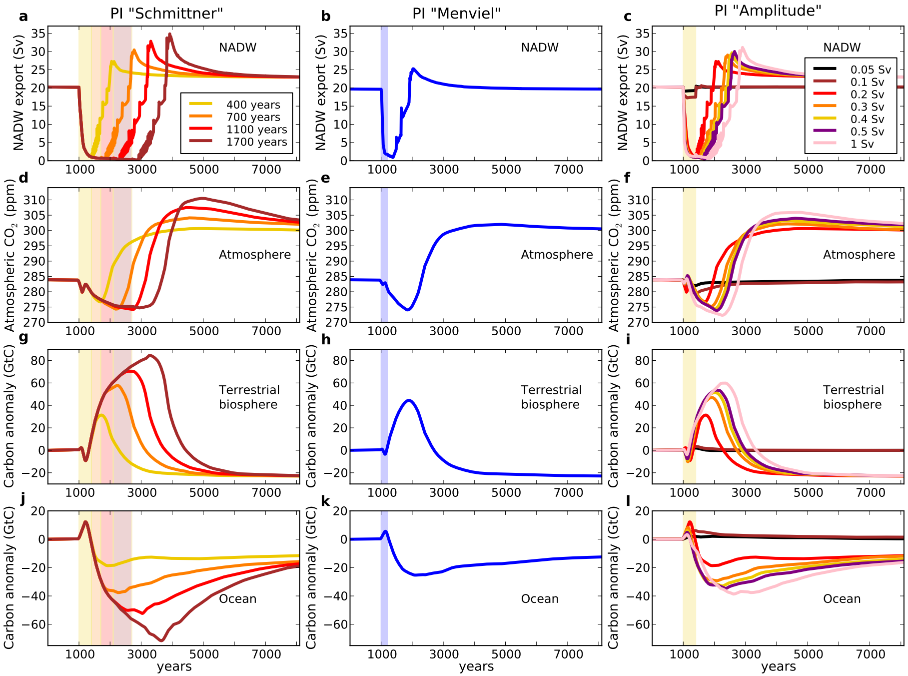

Fig. 2. Evolution of **(a, b, c)** NADW export (Sv) (maximum of the AMOC), **(d, e, f)** atmospheric CO_2 (ppm), **(g, h, i)** carbon content anomaly in the terrestrial biosphere (GtC) and **(j, k, l)** carbon content anomaly in the ocean (GtC) during the simulations with the Preindutrial (PI) background climate state. The fresh water flux is added in the North Atlantic and vary as described in Fig. 1 for the three types of experiments: **(a, d, g, i)** "Schmittner", **(b, e, h, k)** "Menviel" and **(c, f, i, l)** "Amplitude".

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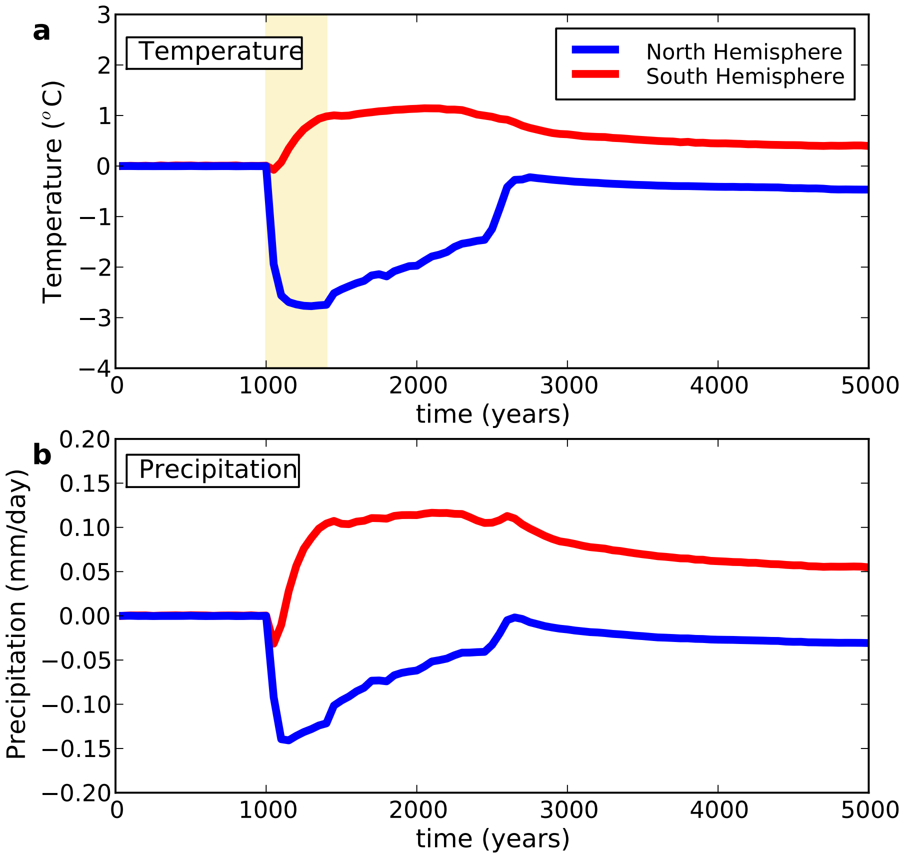

Fig. 3. Evolution of **(a)** air temperature (°C) and **(b)** precipitation (mm day^{-1}) in the North (blue) and South (red) Hemispheres for the simulation during the Preindustrial with a fresh water flux added in the North Atlantic of 0.5 Sv during 400 yr.

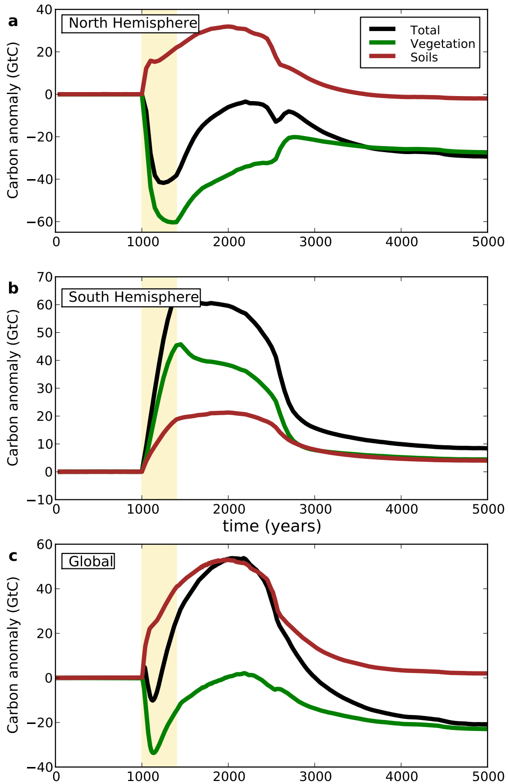

Fig. 4. Evolution of the carbon content anomaly in the terrestrial biosphere (GtC) **(a)** in the North Hemisphere, **(b)** in the South Hemisphere and **(c)** total of the two hemispheres. The total carbon in the terrestrial biosphere (black) is decomposed in its two subreservoirs: the vegetation (green) and soils (brown). In the simulation the fresh water flux added in the North Atlantic has an amplitude of 0.5 Sv during 400 yr.

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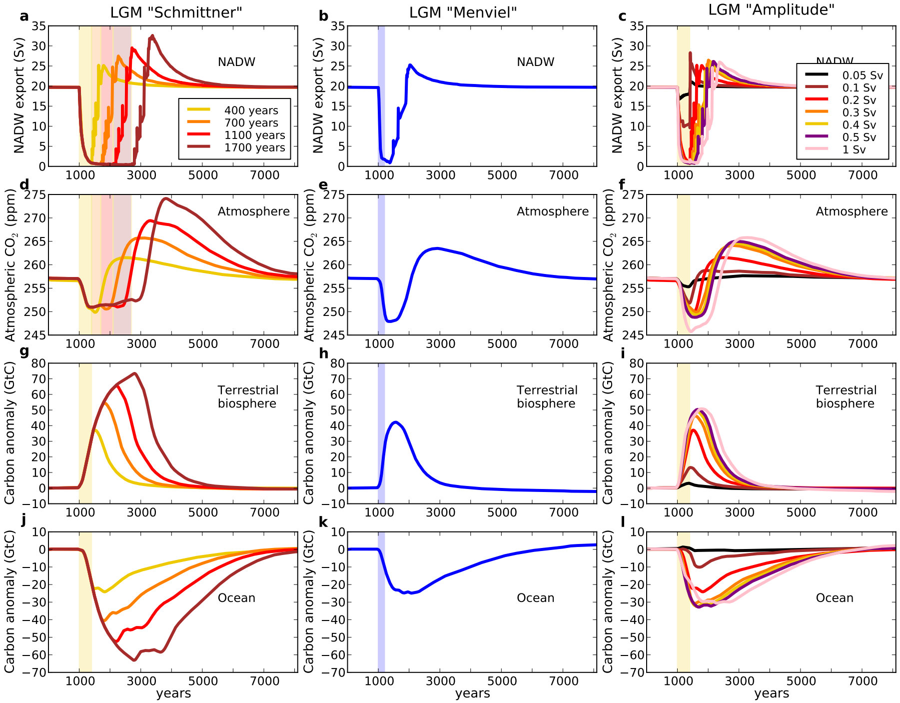

Fig. 5. Evolution of **(a, b, c)** NADW export (Sv) (maximum of the AMOC), **(d, e, f)** atmospheric CO_2 (ppm), **(g, h, i)** carbon content anomaly in the terrestrial biosphere (GtC) and **(j, k, l)** carbon content anomaly in the ocean (GtC) during the simulations with the Last Glacial Maximum (LGM) background climate state. The fresh water flux is added in the North Atlantic and vary as described in Fig. 1 for the three types of experiments: **(a, d, g, i)** “Schmittner”, **(b, e, h, k)** “Menviel” and **(c, f, i, l)** “Amplitude”.

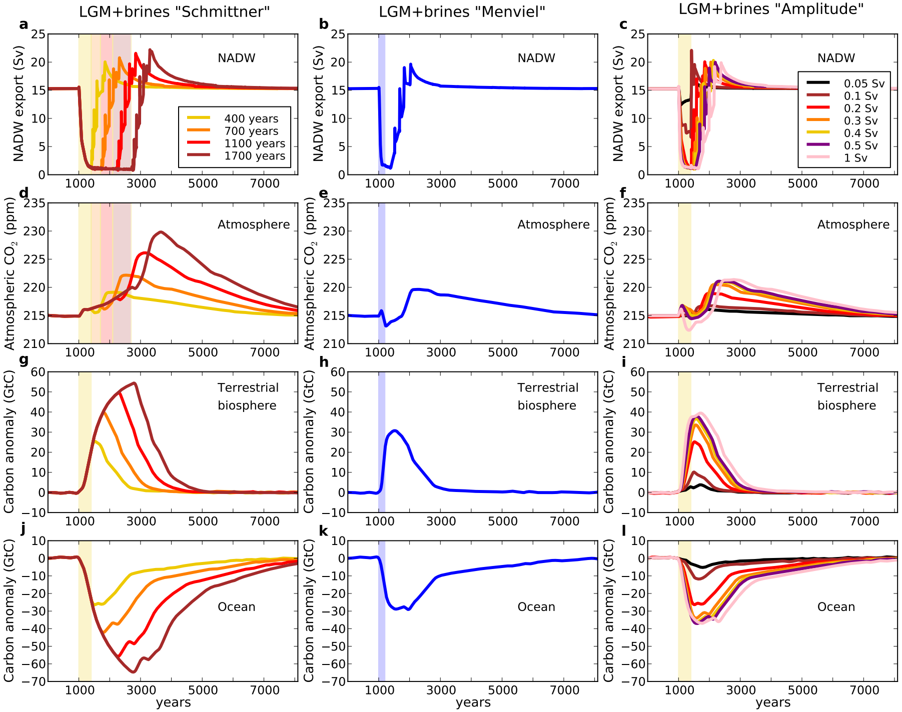

Fig. 6. Evolution of **(a, b, c)** NADW export (Sv) (maximum of the AMOC), **(d, e, f)** atmospheric CO_2 (ppm), **(g, h, i)** carbon content anomaly in the terrestrial biosphere (GtC) and **(j, k, l)** carbon content anomaly in the ocean (GtC) during the simulations with the Last Glacial Maximum background climate state and taking into account the sinking of brines (LGM+brines). The fresh water flux is added in the North Atlantic and vary as described in Fig. 1 for the three types of experiments: **(a, d, g, i)** "Schmittner", **(b, e, h, k)** "Menviel" and **(c, f, i, l)** "Amplitude".

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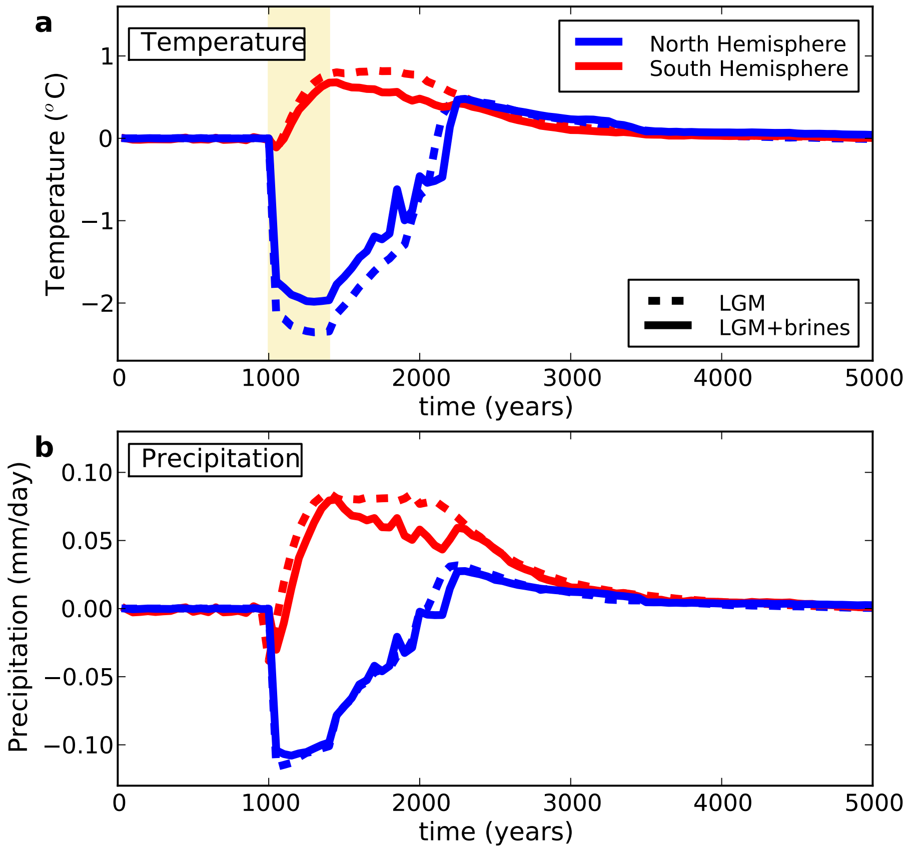

Fig. 7. Evolution of **(a)** air temperature (°C) and **(b)** recipitation (mm day^{-1}) in the North (blue) and South (red) Hemispheres for the simulation with a fresh water flux added in the North Atlantic of 0.5 Sv during 400 yr. Two background climate states are considered: the Last Glacial Maximum (LGM, dashed lines) and the Last Glacial Maximum with the sinking of brines (LGM+brines, solid lines).

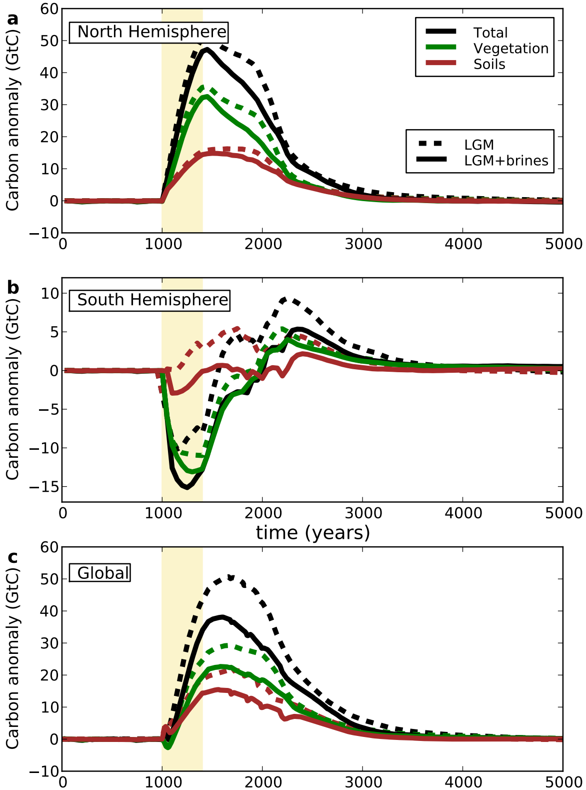

Fig. 8. Evolution of the carbon content anomaly in the terrestrial biosphere (GtC) **(a)** in the North Hemisphere, **(b)** in the South Hemisphere and **(c)** total of the two hemispheres. The total carbon in the terretrial biosphere (black) is decomposed in its two subreservoirs: the vegetation (green) and soils (brown). In the simulation the fresh water flux added in the North Atlantic has an amplitude of 0.5 Sv during 400 yr. Two background climate states are considered: the Last Glacial Maximum (LGM, dashed lines) and the Last Glacial Maximum with the sinking of brines (LGM+brines, solid lines).

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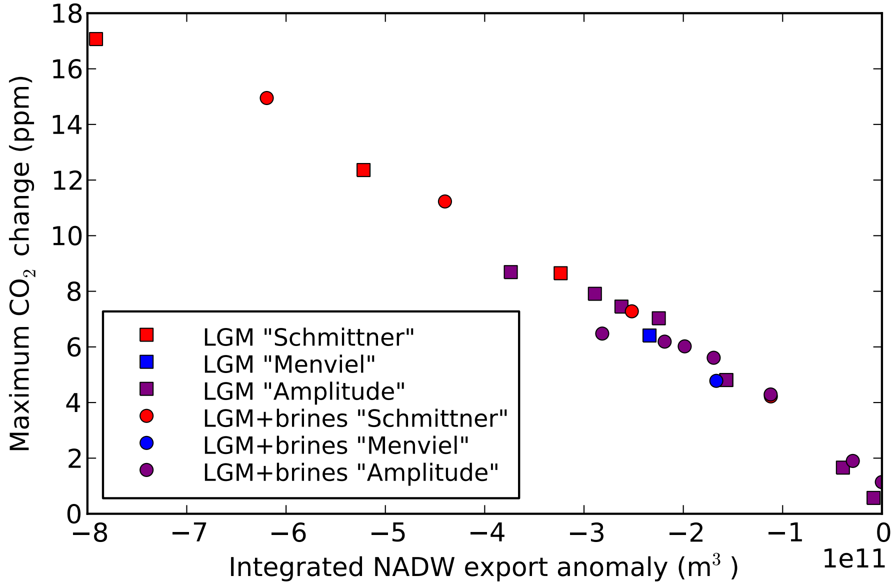

Fig. 9. Maximum of the CO_2 change (as a difference from the initial value) (ppm) as a function of the integrated export of NADW anomaly (taken as the maximum of the AMOC, as a difference from the initial value).

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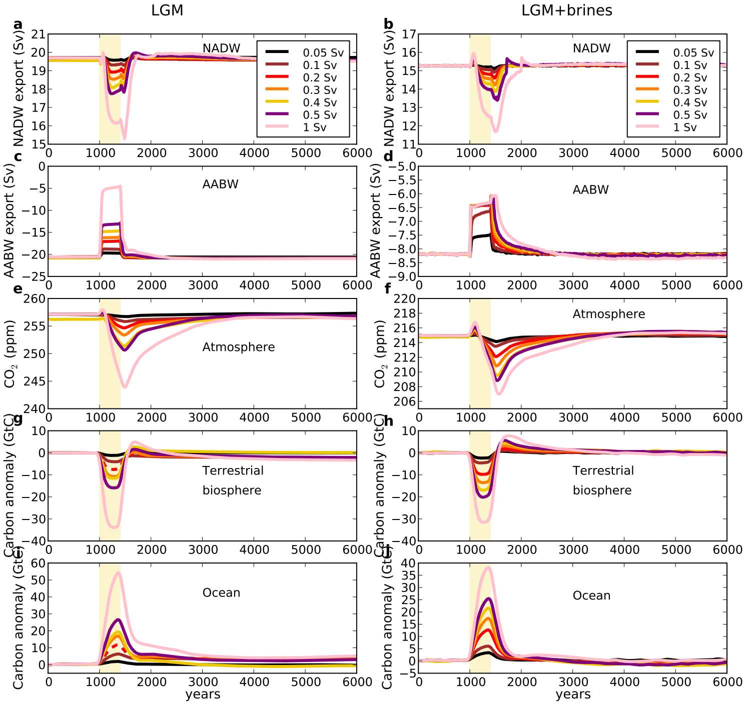

Fig. 10. Evolution of: **(a, b)** NADW export (Sv) (maximum of the AMOC), **(c, d)** AABW export (Sv) (minimum of the AMOC, the sign indicates that the transport is from the South to the North), **(e, f)** atmospheric CO_2 (ppm), **(g, h)** carbon content anomaly in the terrestrial biosphere (GtC) and **(i, j)** carbon content anomaly in the ocean (GtC). The simulations are performed with two background climate states: **(a, c, e, g, i)** during the Last Glacial Maximum (LGM) and **(b, d, f, h, j)** during the Last Glacial Maximum with the sinking of brines (LGM+brines). The fresh water fluxes are added in the South Atlantic and follow the "Amplitude" experiments (Fig. 1c).

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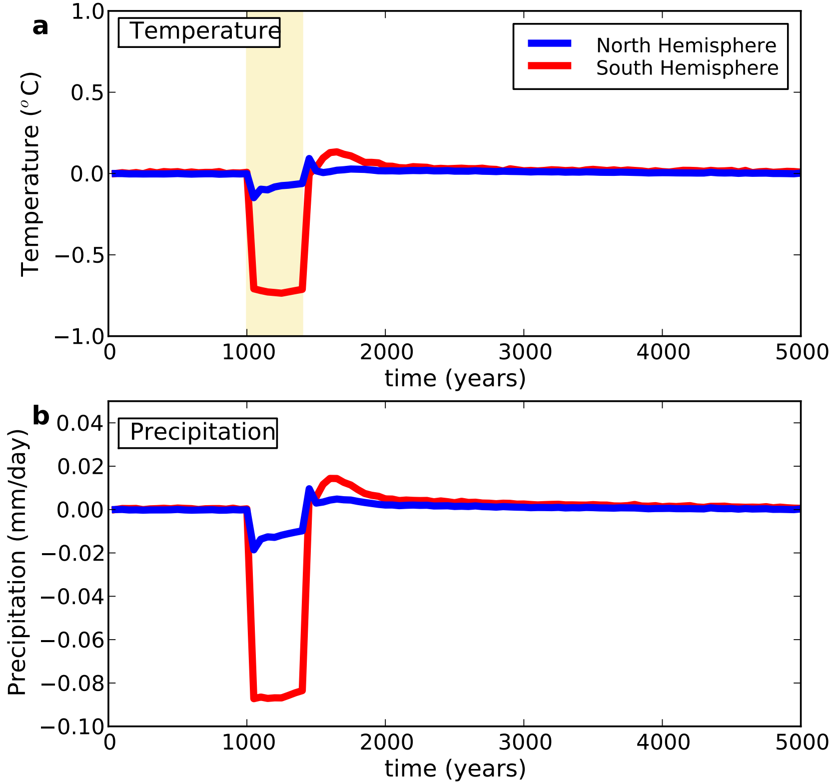

Fig. 11. Evolution of **(a)** air temperature (°C) and **(b)** precipitation (mm day^{-1}) in the North (blue) and South (red) Hemispheres for the simulation during the Preindustrial with a fresh water flux added in the South Atlantic of 0.5 Sv during 400 yr.

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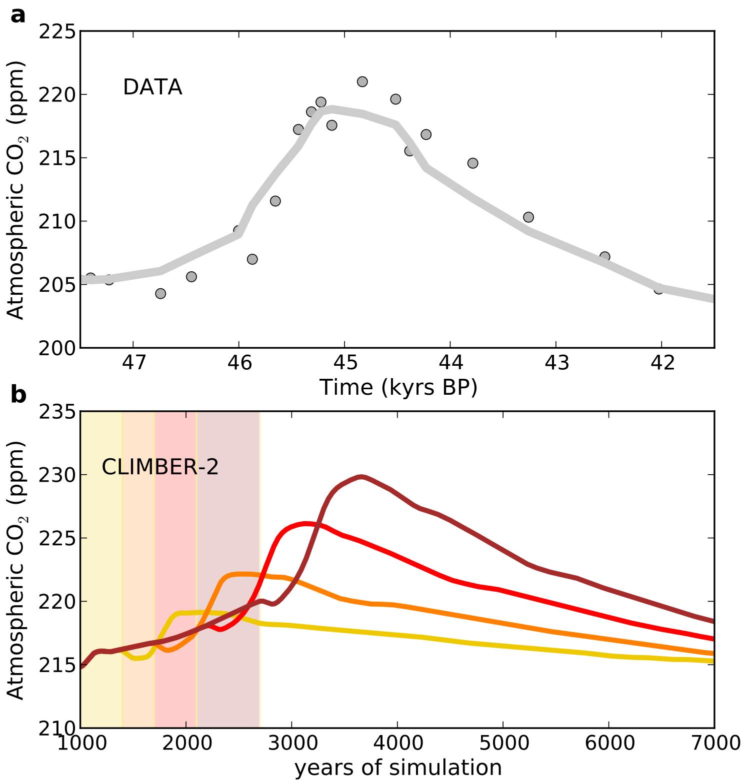

Fig. 12. Evolution of the atmospheric CO_2 (ppm) **(a)** from ice core data for the H5 event (the dots are the data and the solid line the smoothed signal) (Ahn and Brook, 2008) and **(b)** as simulated by the CLIMBER-2 model with additional fresh water fluxes in the North Atlantic following the "Schmittner" experiments (Fig. 1a).

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