Millennial-scale oscillations in the Southern Ocean in response to atmospheric CO$_2$ increase

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A B S T R A C T

A coupled climate-ice-sheet model is used to investigate the response of climate at the millennial time scale under several global warming long-term scenarios, stabilized at different levels ranging from 2 to 7 times the pre-industrial CO$_2$ level. The climate response is mainly analyzed in terms of changes in temperature, oceanic circulation, and ice-sheet behaviour. For the 4×CO$_2$ scenario, the climate response appears to be highly non-linear: abrupt transitions occur in the Southern Ocean deep water formation strength with a period of about 1200 yr. These millennial oscillations do not occur for both lower and larger CO$_2$ levels. We show that these transitions are associated with internal oscillations of the Southern Ocean, triggered by the Antarctic freshwater budget. We first analyse the oscillatory mechanism. Secondly, through a series of 420 sensitivity experiments we also explore the range of temperature and freshwater flux for which such oscillations can be triggered.

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1. Introduction

Ice core data as well as marine and continental records reveal the existence of pronounced millennial time-scale variability in the Quaternary climate system (Broecker et al., 1985; Bond et al., 1992, 1993; Sachs and Lehman, 1999; Eliot et al., 2002; Sanchez-Goni et al., 2002; Wang and Mysak, 2006; Wainer et al., 2009). As an example, the Dansgaard-Oeschger events (Dansgaard et al., 1989, 1993) are characterized by abrupt transitions occurring in a few decades, and by a period of a few thousand years. Such rapid climate variability appears to be stronger in glacial periods than during interglacials, but there is not yet a full consensus about its origin. Two types of explanation have been suggested: periodic external forcing (Keeling and Whorf, 2000; Braun et al., 2005), and internal oscillations in the climate system, for which ocean circulation is a likely candidate (Broecker et al., 1990; Clark et al., 2002; Broecker, 2006; Alvarez-Solas et al., 2010).

The main reason for the non-linear behaviour of the ocean circulation is the existence of positive feedback mechanisms. The large-scale thermohaline circulation is affected by two major positive feedbacks relying either on an advective or a convective mixing. The advective feedback is characterized by the northward advection of salty water in the Atlantic that enhances water density in the North, causing in turn a stronger thermohaline circulation (Stommel, 1961; Bryan, 1986). The convective feedback is associated with a vertical mixing which removes freshwater coming from precipitation, ice melting or calving. This process prevents the formation of a fresh light surface layer which further inhibits convection (Welander, 1982; Lenderink and Haarsma, 1994). These two mechanisms allow the existence of multiple equilibrium states of the oceanic circulation. Oceanic millennial-scale variability has been analyzed since several years by a large range of models of different complexity. The first one indicating the existence of multiple equilibria in the thermohaline circulation and illustrating the importance of the advective feedback mechanism was the Stommel’s horizontal two-box model (Stommel, 1961). The existence of self-sustained thermohaline oscillations in a vertical water column was demonstrated by Welander (1982) with a vertical two-box model. In addition to box models representing convective and advective feedbacks, Earth Models of Intermediate Complexity (EMICs) and General Circulation Models (GCMs), with more detailed dynamics, have then been used to explore the oscillatory behaviour of the ocean (Mikolajewicz and Maier-Reimer, 1990; Timmermann et al., 2003; Abshagen and Timmermann, 2004). On the other hand, it has also been suggested that an oceanic oscillatory behaviour could be excited through white-noise forcing even when the deterministic solution is completely stable (Mikolajewicz and Maier-Reimer, 1990; Canepa and Rahmestorf, 2002). Most of these studies were designed to elaborate plausible spontaneous mechanisms for the origin of Dansgaard-Oeschger (D-O) events.

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In the present climate, the deep water formation has been observed through two main processes: open-ocean deep convection and sinking of dense high salinity shelf waters. The open-ocean deep convection occurs in some regions of the Atlantic Ocean: the Labrador, Greenland, and Mediterranean Seas as well as in the Weddell Sea (Marshall and Schott, 1999). On the other hand, the rejection of salt by sea ice formation plays an important role in the formation of AABW (Nicholls et al., 2009). In this case the convection occurs in very narrow water columns, or plumes, with spatial scales <1 km (Marshall and Schott, 1999).

Usual convective schemes on EMICs and GCMS are based either on the application of very high mixing coefficients when the water column becomes unstably stratified (as in CLIMBER-2) or on actual mixing of the water column. This results in a mixing of the whole grid-cell area, and therefore on larger spatial scales than observed. In spite of these limitations, deep water formation occurs under these conditions, the large-scale oceanic circulations are reproduced in CLIMBER-2 and a reasonable Atlantic overturning circulation can be attained for present-day conditions (Rahmstorf and Ganopolski, 1999; Petoukhov et al., 2000), but also for the Last Glacial Maximum (LGM) (Ganopolski and Rahmstorf, 2001) and for the last interglacial period (Khodri et al., 2003).

Nevertheless, the more stable present-day interglacial period is being currently perturbed by anthropogenic emissions of greenhouse gases. Theoretical considerations about ocean dynamics and glacial variability of the thermohaline circulation lead several authors to argue that the climatic stability of the Northern hemisphere high latitudes could change in response to the increase of greenhouse gases in the atmosphere, due to the warming of oceanic surface layer and freshwater input from Greenland melting (Broecker, 1997; Rahmstorf and Ganopolski, 1999; Rahmstorf, 2003; Swingedouw et al., 2007; Charbit et al., 2008). However, this context does not favour the occurrence of millennial oscillations in North Atlantic Ocean. As noted by Rahmstorf (2003): “D-O events are not present in the Holocene possibly because the North Atlantic Ocean circulation is not close to a threshold in a warm climate”. Therefore, the question arises as to whether anthropogenic warming may drastically impact the climate stability of the Southern hemisphere: Could this warming trigger an oscillatory behaviour of the Southern Ocean? Could this warming destabilize the Antarctic ice sheet leading to periodic ice discharges into the Southern Ocean, similarly to Heinrich events in the Northern hemisphere?

In order to answer these questions, we use in this paper the CLIMBER-2 (CLIMate-BiosphEre) Earth System Model (Petoukhov et al., 2000) coupled to Northern and Southern ice-sheet models (Ritz et al., 1997, 2001). Such a numerical tool allows the investigation of the long-term behaviour of Northern and Southern high latitudes in response to climate warming as well as the feedbacks between atmosphere, ocean and polar ice sheets. Due to massive freshwater discharges coming from the potential destabilization of the ice sheets and to the associated changes in oceanic circulation, the climate response to anthropogenic warming may be highly non linear. Therefore, in the present study, we explore the potential occurrence of rapid climate variability in the future within a large spectrum of perturbations (in terms of CO2 and ice-sheets response) and we investigate the mechanisms that induce a new oscillatory behaviour.

2. Model description

The model of intermediate complexity (Petoukhov et al., 2000) is based on simplified representations of atmosphere, vegetation, ocean and sea-ice. The atmosphere component is based on a statistical-dynamical approach and has a coarse spatial resolution of 10° in latitude and approximately 51° in longitude. This module is designed to resolve large-scale processes (~1000 km), whereas statistical characteristics of the synoptic variability are parameterized as diffusion terms. The radiation scheme accounts for water vapour, CO2 and the computed cloud cover (stratiform and cumulus). The ocean model is composed of three zonally averaged basins (Atlantic, Indian and Pacific) connected around Antarctica. The equations are solved on a 2.5° latitudinal grid with 20 vertical layers. The mean zonal effect of the horizontal wind-driven gyres is taken into account through imposed advective transport at key latitudes such as those covered by the Southern Ocean. The sea ice component predicts the ice fraction and thickness for each grid cell and includes a simple treatment of advection and diffusion of sea ice. In the model hierarchy, CLIMBER-2 is placed between energy balance models and General Circulation Models (GCMS) (Claussen et al., 2002). It describes a large set of processes and feedbacks in the climate system and favourably compares to GCMS for present-day and glacial climates (Petoukhov et al., 2000; Ganopolski et al., 2001), but has a much faster computational time due to its low spatial resolution and simplified governing equations. Moreover, this model has been revealed to be very useful to understand very important features of the glacial oceanic millennial variability (Ganopolski and Rahmstorf, 2001; Ganopolski et al., 2001; Ganopolski and Rahmstorf, 2002; Roche et al., 2004; Braun et al., 2005). The Antarctic ice sheet model, GRISLI (Ritz et al., 2001), is a 3-D ice-sheet model (40 x 40 km). It predicts the evolution of the geometry of the Antarctic Ice Sheet (AIS) and accounts for the thermomechanical coupling between temperature and velocity fields. It deals with both inland and floating ice and explicitly computes the migration of the grounding line. The Northern hemisphere ice sheet model, GREMLINS (45 km x 45 km) is developed in the same way than GRISLI, except that it only deals with inland ice (Ritz et al., 1997). These models, as the other ice-sheet models (Huybrechts, 1990; Greve et al., 1998; Huybrechts and de Wolde, 1999; Marshall et al., 2002) do not successfully reproduce small scale processes, such as the acceleration of the ice flow from outlet glaciers, which has recently been shown to be an important process for the acceleration of the Greenland melting (Rignot and Kanagaratnam, 2006). However, they include a representation of the main mechanisms responsible for slower and large-scale processes and can reasonably be used to simulate the behaviour of polar ice sheets over the next thousands years. The coupling method between GREMLINS and CLIMBER-2 is described in (Charbit et al., 2005) and (Kageyama et al., 2004): the mean annual and summer (June–July–August) surface air temperatures and annual snowfall computed by CLIMBER-2 are given to GREMLINS through downscaling techniques to compute the surface mass balance. In turn, the altitude and the nature (land ice, ice free land or ocean) of each ice-sheet model grid point are returned to CLIMBER-2. The freshwater fluxes resulting from ice-sheet melting is released into the ocean. The coupling method between CLIMBER-2 and GRISLI is based on the same procedure, but the oceanic temperatures at 500 m-depth are also used to compute the basal melting under the ice shelves (Philippin et al., 2006). This fully coupled climate-cryosphere model (referred to hereafter as CLIMBER-IS) has been used to reproduce the last glacial inception (Kageyama et al., 2004), the last deglaciation (Charbit et al., 2005), the future deglaciation of the Greenland ice sheet (Charbit et al., 2008) and the Antarctic contribution to the sea-level rise through last glacial termination (Philippin et al., 2006).

3. Experimental set-up

CLIMBER-2 is forced by insolation and a set of different anthropogenic atmospheric CO2 scenarios starting from the pre-industrial level (i.e. 280 ppm, between 0 and 1860 AD) and stabilized at the following levels: 560, 840, 1120, 1400 and 1960 ppm, that is 2, 3, 4, 5 and 7 times the pre-industrial level. This highest value was estimated to approximately correspond to the total available fossil fuel supplies (Archer et al., 1997). Another experiment (control run)
forced with a constant CO₂ concentration fixed to 280 ppm has also been carried out. The CO₂ scenarios are obtained with an increasing rate of 0.21% per year from 1860 to 2000 AD, and of 1% per year after 2000 AD. Once the above values are reached, the atmospheric CO₂ concentrations are stabilized. The duration of the simulations is 10,000 model years. The initial Antarctic ice-sheet topography is obtained after a 200,000 years integration of GRISLI to obtain a vertical temperature profile in the ice consistent with the history of the ice sheet. For Greenland, we start from the present-day observed topography. The initial climatic state is obtained by a 10,000 years integration of CLIMBER alone.

4. Results

4.1. General results

The simulated mean global surface air temperature starts to increase as soon as the anthropogenic CO₂ increases. This increase is referred to hereafter as the anthropogenic perturbation (Fig. 1a). In all simulations, the mean temperature continues to rise after the maximum CO₂ concentration is reached, during the stabilization phase, until year 3000 approximately. In the following, the period spanning from 3000 AD until the end of the simulation will be called “the post-perturbation equilibrium”. The maximum warming ranges from 3 (2×CO₂) to 7.5 °C (7×CO₂). During the anthropogenic perturbation, both Southern and Northern deep water formations decrease as a response to a reduction of the surface water density caused by warming and freshwater input from Antarctica and Greenland (Fig. 1b, c). This mechanism is enhanced with the increase of the CO₂ perturbation. Nevertheless, when the maximum atmospheric CO₂ is reached at the beginning of the post-perturbation period, the Antarctic Bottom Water (AABW) and the North Atlantic Deep Water (NADW) formations tend to recover the equilibrium. However, this recovery takes place very differently between the Southern and the Northern Hemispheres and also between the different CO₂ experiments. The AABW post-perturbation signal is characterized by two possible states: the pre-industrial value (~17 Sv) for the control, 2×CO₂ and 3×CO₂ experiments and an enhanced value (~21 Sv) for 5×CO₂ (with still some abrupt excursions towards a reduced value) and 7×CO₂ simulations. On the other hand, the 4×CO₂ AABW signal seems to oscillate between these two different states (Fig. 1b). The behaviour of the NADW formation is simpler: for all the simulations, the post-perturbation values, ranging from 19 to 22 Sv are smaller than the control one (23 Sv), with the lowest value corresponding to the highest CO₂ perturbation (7×CO₂) (Fig. 1c).

4.2. The 4×CO₂ simulation

The 4×CO₂ run displays an interesting feature: the global temperature signal exhibits oscillations between warmer and cooler states (Fig. 1a). These oscillations are more clearly depicted in the Southern Hemisphere (Fig. 2a), than in the Northern Hemisphere (Fig. 2c), with an amplitude of ~0.7 °C. Following the CO₂ increase, the intensity of AABW formation (Fig. 2a) decreases from ~16 to ~10 Sv. When the CO₂ concentration is stabilized, the AABW progressively recovers until year 3000. Then, oscillations between two equilibrium states appear, shifting in a few decades from ~15 Sv to ~23 Sv. These oscillations are correlated with the atmospheric temperature response. NADW decreases from its pre-industrial value (23 Sv) to 15 Sv at the end of anthropogenic CO₂ perturbation, and then, asymptotically recovers towards a lower value around 20 Sv. The Antarctic oscillatory behavior is also reflected in the NADW signal with an amplitude of about one order of magnitude lower than that of the AABW signal.

The Antarctic ice sheet (AIS) loses mass from the beginning of the anthropogenic perturbation until the end of the simulation with a larger melting rate between 2000 and 3500 AD (Fig. 2c). At the end of the simulation, the ice mass lost is about 7.6% compared to the pre-industrial state, and the ice volume tends to further decrease as shown in Fig. 2. The smaller Greenland ice sheet reacts more dramatically and has melted completely around 4500 AD (Fig. 2c). Whereas the total melting of the Greenland ice sheet is of uppermost relevance from a climatic perspective (Charbit et al., 2008), the freshwater provided into the North Atlantic Ocean and corresponding to an ice volume loss of ~3. 1015 m³ does not excite any oscillatory behaviour. By contrast, the release of meltwater coming from the Antarctic ice sheet, (equivalent to an ice volume variation of ~2. 1015 m³) seems to be able to disturb the dynamical stability of the Southern Ocean.

5. Low frequency oscillations in the Southern Ocean

5.1. Description of the oscillations

A spectral analysis of the Southern hemisphere mean surface air temperature simulated in the 4×CO₂ experiment for the time period corresponding to the post-perturbation equilibrium (from 3000 to the end of the simulation, Fig. 3) reveals the existence of both fundamental and secondary components, having periods of respectively ~1180 and 450 years. The latter component has a weight about two orders of magnitude lower than the former one and will not be discussed in the following. More interesting are the unexpected millennial time-scale oscillations (i.e. ~1180 years) called hereafter low frequency oscillations (LFO). The amplitude of these oscillations is ~0.7 °C for the whole southern hemisphere. These abrupt temperature variations occur in a few decades and are mainly located over the
convection zones of the Southern Ocean. Over the Pacific sector the amplitude can be as high as 3 °C, and exceeds 4 °C in some locations of the Atlantic Ocean (Figs. 4a, b). This suggests that the temperature oscillations are due to oceanic processes. Moreover during a transition from a cool to a warmer state (resp. warm to cooler), the surface layer oscillates between low and high values. Potential density differences between subsurface (750 m-depth) and surface layers from 60° to 75°S for the 3×CO2, 4×CO2 and 7×CO2 experiments. Low values of this gradient correspond to a highly mixed water column, while higher values correspond to more stratified situations. After 3000 AD, the 3×CO2 simulation depicts the largest gradient, similar to its pre-industrial value, while the 7×CO2 experiment produces the smallest one. This suggests that, during the post-perturbation period, the vertical mixing regime in the 5×CO2 is similar to the pre-industrial situation, while for the 7×CO2 simulation convection is strongly enhanced. As expected, in the 4×CO2 experiment, the behaviour is rather different, and the potential density gradient oscillates between low and high values. This is consistent with a periodical switch between a more stratified mode and a more convective one.

The main physics of such oscillations, based on temperature and salinity considerations, were previously suggested by (Welander, 1982) with a vertical 2-box model. Following (Welander, 1982), this behaviour can also be illustrated in the CLIMBER-2 model. Fig. 6 displays salinity–temperature diagrams for the Southern Ocean between surface and subsurface by the salinity and temperature gradients. Three different time periods can be identified:

1. The pre-industrial equilibrium: It provides identical (∆T, ∆S) values for all the simulations (see green circle in Fig. 6a–c), with a vertical 2-box model. Following (Welander, 1982), this behaviour can also be illustrated in the CLIMBER-2 model.

Fig. 2. a) Mean annual Southern hemisphere surface air temperature (black line) and intensity of the Antarctic Bottom Water formation (in absolute values) (blue line) for the 4×CO2 scenario; b) Mean annual Northern hemisphere surface air temperature (black line) and intensity of the North Atlantic Deep Water formation (in Sv) (red line) for the 4×CO2 scenario; c) Antarctic (blue line) and Greenland ice volumes (red line) for the 4×CO2 scenario; d) Freshwater flux coming from the Antarctic ice sheet to the Atlantic Ocean. The solid line corresponds to the standard 4×CO2 scenario freshwater flux and the dashed line corresponds to the averaged freshwater flux that has been used to force a CLIMBER-alone experiment (Section 5.2). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
stabilized, the surface warming slows down and the melting rate of ice sheets decreases leading in turn to an increase of the surface salinity. This process favours a progressive decrease of the temperature and salinity gradients. Therefore, in these three experiments, when the stratification process is no longer sustained, the deep water formation progressively recovers towards its pre-industrial level and the convection increases. The AABW circulation reaches therefore a new equilibrium state. In the 3×CO2 simulation, the ocean has not sufficiently been shifted from the initial conditions to allow the occurrence of a new different equilibrium. Consequently, the resulting equilibrium state is characterized by a stable regime very similar to the pre-industrial one, though slightly more stratified (Fig. 6a). By contrast, in the 7×CO2 experiment, the warming undergone by the ocean during the anthropogenic perturbation is large enough to reach the deep layers. This acts to sustain a convective vertical mixing. Indeed, during the post-perturbation period, a new reorganization of water properties leads the ocean to a new stable equilibrium state corresponding to a highly convective mode (Fig. 6c). In between, the equilibrium reached in the 4×CO2 simulation is unstable and the Southern Ocean oscillates between a more stratified and a more convective modes.

5.2. Mechanisms at the origin of low frequency oscillations

To investigate whether the low frequency oscillations (LFOs) stem from an intrinsic behaviour of the ocean or from any climate-ice sheet feedback, a set of additional experiments has been carried out by unplugging one or two ice-sheet components (i.e. Northern and/or Southern ice-sheet models). It appears that the LFOs occur only when the CLIMBER-2 model is coupled to GRISLI (not shown). Therefore the mechanism responsible for their occurrence is likely to be related to the freshwater flux coming from the AIS and released to the Southern Ocean. To test this assumption, we used the Antarctic freshwater flux simulated in the coupled CLIMBER-IS 4×CO2 experiment to force the CLIMBER-2 model unplugged to the ice sheet components (CLIMBER-alone). The newly simulated AABW and Southern Hemisphere temperature signals are also characterized by the occurrence of LFOs (not shown), with a period of ~1170 yr, very similar to the period of the oscillations simulated in the coupled 4×CO2 experiment (~1180 yr). Moreover, in the coupled and CLIMBER-alone simulations, these oscillations have the same amplitude (~0.7 °C). This confirms

Fig. 4. a) Surface air temperature anomaly (in °C) for the 4×CO2 scenario between an atmospheric warm state (4.2 kyr) and an atmospheric cool state (4.0 kyr). b) Surface air temperature anomaly (in °C) for the 4×CO2 scenario between an atmospheric cool state (4.9 kyr) and an atmospheric warm state (4.8 kyr). c) Latitude–depth diagram of the oceanic temperature anomaly (in °C) for the 4×CO2 scenario between an atmospheric cool state (4.9 kyr) and an atmospheric warm state (4.8 kyr) in the Atlantic Ocean. d) Latitude–depth diagram of the oceanic temperature anomaly (in °C) for the 4×CO2 scenario between an atmospheric cool state (4.9 kyr) and an atmospheric warm state (4.8 kyr) in the Atlantic Ocean.

Fig. 5. Potential density gradient (kg/m3) for the 3×CO2 (yellow), 4×CO2 (black) and 7×CO2 (purple) scenarios as a function of time in kyr. High values of this gradient correspond to highly stratified states. Note that the 3×CO2 and 7×CO2 evolutions depict the extreme expected values (strong and reduced gradient respectively) of the oscillatory 4×CO2 case. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
that the oscillatory behaviour is triggered by the meltwater coming from Antarctica.

However, the freshwater flux used in this experiment is also characterized by an oscillatory behaviour (Fig. 2d, solid line) due to the feedbacks between LFOs in Southern Ocean and the mass balance of the Antarctic ice sheet. Indeed, during the warm phase of LFOs, the larger oceanic and surface air temperatures lead to an increase of basal melting under the ice-shelves and to an enhanced ablation around the coastal zones of the Atlantic sector, increasing the freshwater flux. During the cooler phase of LFO the process is reversed with reduced basal melting and ablation, and a decrease of freshwater input into the ocean. The succession of these positive feedbacks results in an oscillatory behaviour (Fig. 2d, dashed line) of the Antarctic ice sheet. The oscillatory behaviour of the ocean induces in turn a cyclic instability of the surface air temperature around the convection zones (>4°C) that is propagated until mid-latitudes, ~30°S (more than 1°C).

The schematics of Fig. 7 describes the processes involved in this millennial variability and explains the persistence of the oscillation even in presence of constant FWFs: during episodes of strong convection, sea-ice cover decreases enhancing thereby the ocean-to-atmosphere heat flux. The deep ocean is progressively cooled until the water column becomes stably stratified and the convection cannot be maintained. This new situation is accompanied by an abrupt surface cooling and by an increase of sea-ice cover. As convection is reduced during this phase, the vertical column is poorly mixed implying the occurrence of a halocline. The heat stored in the column progressively warms the deep ocean leading eventually to an unstable stratification and a new episode of strong convection.

5.3. Sensitivity experiments

Among the experiments considered in the present work, millennial oscillations only occur under the 4×CO₂ scenario. This suggests that LFOs may occur for specific values of either surface air temperature or the magnitude of the freshwater flux. Moreover, the driving of CLIMBER-alone forced by a constant freshwater flux (Section 5.2) also demonstrates that a change in the magnitude of the freshwater forcing may induce changes in both amplitude and period of the resulting low frequency oscillations. To explore the range of temperature and freshwater flux conditions allowing the LFOs occurrence a set of 420 sensitivity experiments has been carried out.

To achieve this goal, 30 additional CO₂ scenarios have been constructed in the same way as the previous ones (2×CO₂ to 7×CO₂) with levels stabilized from 600 to 2000 ppm. These CO₂ scenarios have been used to force the coupled CLIMBER-IS model to obtain a range of FWF and southern hemisphere temperature values (averaged over the period 4800–10,000 AD). Each averaged FWF value has then been multiplied by 14 constant factors ranging from 0.05 to 2.0. As a result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result, we obtain 420 constant through time FWF values used to force the CLIMBER-alone model, exactly in the same way as the experiment result.

To answer this question, we carried out another CLIMBER-alone experiment forced by a new freshwater flux constant through time (i.e. non-oscillating). This new flux has been obtained by averaging the FWF simulated in the CLIMBER-IS 4×CO₂ experiment over the period spanning between 4800 and 10,000 model years within the post-perturbation period (Fig. 2d, dashed line). In this experiment, the LFOs are still present in the 4×CO₂ run with a smaller fundamental period of ~880 years and with roughly the same amplitude (~0.6°C). Therefore, this new experiment demonstrates that the millennial time-scale oscillations do not result from an external oscillatory forcing but rather stem from an intrinsic behaviour of the Southern Ocean triggered by a constant freshwater flux released at high latitudes and coming from the destabilisation or the melting of the Antarctic ice sheet. The oscillatory behaviour of the ocean induces in turn a cyclic instability of the surface air temperature around the convection zones (>4°C) that is propagated until mid-latitudes, ~30°S (more than 1°C).
(Fig. 8a). The period of oscillations progressively decreases with the increase of temperature, until the complete disappearance of the millennial component. The maximum amplitude of the oscillations is located at the centre of the parameter phase space and the amplitude is more and more attenuated when approaching the bounds of this region (Fig. 8b).

Fig. 7. Scheme describing the mechanisms at work to shift the Southern Ocean from the warm to the cold mode. Based on figures 8 and 6 in Haarsma et al. (2001) and Meissner et al. (2007) respectively.

Fig. 8. Spectral analysis of the Southern Ocean variability. More than 420 additional CLIMBER-alone simulations have been carried out to explore the range of temperature and FWF values allowing the occurrence of LFOs (see text). For each run, a MTM analysis (as in Fig. 3) of the Southern Hemisphere temperature signal has been computed for the post-perturbation period. Each colour point corresponds to a single simulation. a) The colour scale is associated with the power of the main component of the MTM spectrum and has been interpreted in terms of amplitude in temperature change during each oscillation; b) The main MTM spectral frequency has been extracted and interpreted in terms of periodicities (i.e. the colour code is associated with the oscillations period). Black circles noted A, B, C, D, E and F correspond to CTRL, 2×CO₂, 3×CO₂, 4×CO₂, 5×CO₂ and 7×CO₂ CLIMBER-IS simulations respectively. Solid and dashed gray lines represent the 99%-significance and 95%-significance iso-lines on the MTM spectra: areas within the solid and dashed gray lines gather simulations in which the main spectrum periodicities are significant above 99% and 95% respectively. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
In other words, a larger warming favours the system instability. Moreover if the FWF becomes greater than 10 mSv, oscillations can be produced with amplitudes smaller than 0.4 °C. For FWF values greater than 12 mSv the period becomes smaller than 100 years.

On the other hand, the CLIMBER-2 model does not produce weather noise. As mentioned in the introduction, it has been shown (Mikolajewicz and Maier-Reimer, 1990; Ganopolski and Rahmstorf, 2002) that a quasi-oscillatory behaviour of the ocean can be attained through white-noise forcing by a mechanism of stochastic resonance. To determine whether a similar behaviour can be observed in the future Southern Ocean, we carried out several additional simulations around the oscillatory parameter space in which we added white-noise to the mean FWFs. Following the methodology described in Ganopolski and Rahmstorf (2002), the standard deviation of the white noise corresponds to the 25% of the present-day net freshwater flux to the Southern Ocean. Nevertheless, in contrast with the Dansgaard-Oeschger-type oscillations, the noise-forcing scenario does not have any significant effect on the future Southern Ocean oscillatory phase space shown here. In the case of glacial North-Atlantic, adding a FWF noise can be very efficient to create an abrupt transition provided that the ocean was close to the bifurcation between two stable states (one stable and one metastable). In contrast, for the future Southern Ocean, there is not a real bifurcation between two states, but a first stable state (corresponding to the less convective regime in CTRL, 2×CO₂ and 3×CO₂), a second stable state (corresponding to a more convective regime in 5×CO₂ and 7×CO₂) and a large region in between both states (around 4×CO₂) where the ocean shows periodical solutions and therefore oscillations arise. In other words, because the Southern Ocean oscillations are self-sustained, the main requirement for their occurrence is the phase space location rather than the forcing provided by the noise itself. This explains why our phase space illustrated in Fig. 8 does not show any substantial change through white-noise forcing.

6. Discussion

As mentioned by Winton (1997), the occurrence of oscillations in the vertical mixing regime requires sufficiently cold deep ocean temperatures. On the contrary, a too warm deep ocean sustains a permanent convection state, as observed in our 5×CO₂ and 7×CO₂ experiments. According to (Winton 1997) and (Haarsma et al., 2001) such a mechanism could play a crucial role to trigger D-O events in a glacial climate and could explain the absence of ocean millennial variability during the Holocene. In the same way, within the context of a future climate, the high latitudes of the Southern Ocean could be the only ones to experience sufficiently cold temperatures to allow the occurrence of LFOs, in line with our findings. Indeed, several previous studies (e.g. Stocker and Wright, 1991; Mysak et al., 1993; Winton and Sarachik, 1993; Hughes and Weaver, 1994; Weaver and Hughes, 1994; Sakai and Peltier, 1997; Winton, 1997; Marotzke and Scott, 1999), based on ocean models of various complexities, have been devoted to the dynamical behavior of the ocean and raised the possibility for the ocean to fall into an oscillatory behavior at the millennial time-scale. Most of these studies were mainly focused on the North Atlantic. To our knowledge, only two modeling works suggest the existence of millennial variability in the Southern Ocean (Haarsma et al., 2001; Meissner et al., 2007). In these works a quasi-stable state (with no or very little deep water formation) is periodically disturbed by episodes of strong convection, called "flushes", whereas, in our work, convection is always active, and the oscillations are characterized by transitions between a state of pronounced stratification and a state where convection is enhanced by more than 10 Sv. Despite this difference, the physical mechanisms at the origin of the oscillatory behavior and pointed out in those works are fundamentally the same than those involved in the present study. Nevertheless, these studies are different from ours in many respects. Firstly, we account here for the transient effects of different imposed CO₂ evolving scenarios. Haarsma et al. (2001) analyze an ultra-low (13 kyr) frequency behavior of the ocean under constant boundary conditions. Meissner et al. (2007) focused on the same issue, also under fixed equilibrium conditions, but liberating one crucial component of the system as the carbon cycle is. This allowed them to account for superimposed CO₂ degassing (sequestration) from the ocean which amplifies the warming (cooling) as well as they depict a threshold behavior (~440 ppm of atmospheric CO₂) for oscillations. On the other hand, our study is also the first to consider the fully coupled effects between ocean, atmosphere and ice sheets. In particular, the dynamics of the Antarctic ice sheet has been shown to play a crucial role for the millennial variability through the freshwater released to the Southern Ocean. Therefore, thanks to the coupling between CLIMBER and the 3D ice-sheet models (GREMLINS and GRISLI), the dependence of the oscillations on more realistic freshwater values is accounted for in the present study. This latter aspect combined with the transient character of our simulations and with the major efficiency in terms of computational costs of our coupled model, allowed us to explore and constraint the phase space of the Southern Ocean in which an oscillatory behavior appears. Even if the precise window of such oscillations will depend on the intrinsic characteristics of the different models, this study offers a highly useful guide for future studies dealing with the Southern Ocean behavior under warmer climates.

It is important to note that we do not observe dramatic purges (i.e. a future analogue of glacial Heinrich events) of ice from Antarctica. The response to both CO₂ perturbation and LFOs themselves do not induce rapid or strong periodical variations of the ice volume. However, recent studies indicate a strong acceleration of outlet glaciers after the catastrophic collapse of the Larsen B ice shelf in the Antarctic Peninsula (Hulbe and Fahnestock, 2004; Rignot et al., 2004; Hulbe et al., 2008; Rignot et al., 2008). This suggests that our ice-sheet model likely underestimates the fast ice flow episodes triggered by the ocean and atmosphere surface warming around Antarctica.

7. Conclusion

The coupled atmosphere-ocean-vegetation-ice-sheet model (CLIMBER-GREMLINS-GRISLI) has been used to study through numerous sensitivity tests the possible occurrence of millennial variability in the climate system of the next 8000 years. The 4×CO₂ CLIMBER-IS experiment shows the presence of instabilities in the Southern Hemisphere temperature signal, associated to a periodical shift between two different vertical regimes in the deep water formation areas of the Southern Ocean: a more convective one (with a strong mixing of water properties) and a more diffusive one (with stratified layers). We demonstrate that these oscillations do not result from the action of any oscillatory forcing. A constant freshwater flux is sufficient to generate such instabilities, providing that the system is placed in the range of forcing values (freshwater flux and surface air temperature) that enables such periodical solutions.

The present study confirms that the Southern Ocean warming and the increase of the freshwater flux from Antarctica favors the occurrence of rapid climate variability. Differently than in glacial periods, in a window of boundary conditions accessible by the anthropogenic perturbation, millennial-scale oscillations are triggered in the Southern Ocean. These oscillations that could occur in the next thousand of years have a huge impact on regions around Antarctica in terms of ocean stratification, sea ice cover and temperature changes. Moreover, strong changes in the atmosphere-to-ocean CO₂ flux could also be observed through changes in deep water formation. Finally, this new abrupt low frequency oscillations could significantly perturb the ice flow dynamics of the grounded Antarctic ice sheet on the next millennia.

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