Rapid transitions and ultra-low frequency behaviour in a 40 kyr integration with a coupled climate model of intermediate complexity

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Abstract

A 40 kyr integration with the coupled atmosphere/ocean/sea-ice model of intermediate complexity ECBilt for present boundary conditions has been performed. The climate of ECBilt displays quasi-periodical behaviour with a period of approximately 13 kyr. The quasi-periodical behaviour is characterized by large changes in the overturning cell in the Southern Ocean. The southern cell fluctuates between two quasi-stationary states, with accompanying changes in the atmospheric circulation in the Southern Hemisphere. The transition between these states is rapid and resembles the polar halocline catastrophes and flushes as observed in ocean general circulation models under mixed boundary conditions. The sea-ice influence on both the surface heat and fresh water flux appears to be crucial for the existence and the prolongation of the quasi-stationary states. The atmospheric circulation of those two quasi-stationary states displays large regional differences over Antarctica, resulting in even opposite surface air temperature trends for certain locations during the transition from one state to another.

1 Introduction

Nowadays, it is more and more accepted that climate is not only slowly varying, but may also display rapid climate shifts. There is ample evidence for these shifts from paleorecords (Broecker et al. 1985; Dansgaard et al. 1993; Jouzel et al. 1995). These records also suggest that the cause of rapid climate shifts may reside in the ocean.

To study the ocean behaviour many investigators have run ocean general circulation models (OGCMs) with mixed boundary conditions (MBCs). In these studies the atmospheric temperature and precipitation minus evaporation (P-E) flux are prescribed following a fixed annual cycle. Studies with these models have revealed complex dynamical behaviour. Bryan (1986) found that an interruption of the deep-water formation, either spontaneously or by a small salt anomaly, could lead to a collapse of the thermohaline circulation (THC), a phenomenon which has become commonly known as a polar halocline catastrophe (PHC). Marotzke (1990) observed that the warming of the deep ocean in the sub polar regions, after the collapse of the THC could eventually result in an unstable stratification, which than subsequently leads to extremely strong deep-water formation and a corresponding strong thermally driven circulation, which he called a flush. The occurrence of PHCs and flushes in OGCMs has been confirmed by many other authors in different models (Wright and Stocker 1991, Weaver and Sarachik 1991, Winton and Sarachik 1993). Multiple steady states, characterized by deep-water formation in different basins, were found by Marotzke and Willebrand (1991) for a highly idealized model of the World Ocean. Multiple equilibria were also found by Hughes and Weaver (1994), Power and Kleeman (1993) and Rahmstorf (1995) for more realistic ocean bathimetry.

The possibility of oscillatory behaviour due to convective instability was first pointed out by Welander (1982) for a simple box model. Using a 2-D ocean model millennial time scale oscillations under MBCs are described by Winton and Sarachik (1993) and Winton (1995). These oscillations are characterized by two quasi-stationary states one displaying a THC and the other a PHC with rapid transitions between those states.

Mixed boundary conditions are clearly a poor man approximation of the heat and fresh water exchange of the atmosphere. To mimic the stochastic behaviour of the atmosphere a number of investigators added noise to the MBCs. These studies revealed that the addition of noise does not dramatically change the dynamical behaviour. Mikolajewicz and Maier-Reimer (1990) found that for the LSG OGCM a spatially correlated white-noise fresh-water flux generated a pronounced low-frequency signal at around 320 year. However, Pierce et al. (1995) showed that these oscillations also exist without noise for slightly different fresh water flux profiles. Very long time integrations of the order of 100 kyr with a 2D ocean circulation model forced with a stochastic white fresh-water flux have been performed by Mysak et al. (1993), demonstrating the possibility of century-scale variability. For certain regions in parameter space large oscillations of order 10 kyr were observed, characterized by rapid transitions between different quasi-stationary states of the THC.

Serious doubts concerning the relevance of MBCs for the dynamics of the coupled system have been raised by Zhang et al. (1993) and Power and Kleeman (1994). They argue that the use of MBCs with a strong restoring condition results in a too large dependency of the surface heat flux on the convective activity. This is because the atmospheric temperature is not allowed to adjust to the changing sea surface temperatures (SST). Lenderink and Haarsma (1994) showed that the occurrence of rapid changes in convective activity is tightly related to the local selfsustaining character of convection under MBCs. With weaker restoring conditions for the SST convection becomes less selfsustaining and the likelyhood for rapid changes in convective activity decreases. As a consequence Zhang et al. (1993) and Power and Kleeman (1994) state that the occurrence of PHCs is an artefact of MBCs.

From the above an apparent controversy arises. Results with MBCs do predict rapid shifts in the ocean circulation and the possibility of oscillatory behaviour on millennial time scale, in accordance with the evidence from paleo data. The physics responsible for these phenomena, however, does not seem to be correct. Attempts have been made to resolve this controversy within the context of coupled models with varying degree of complexity ranging from highly parameterized coupled box models (Nakamura et al. 1994; Marotzke and Stone 1995; Jayne and Marotzke 1999) to fully coupled GCMs. The first type of models suffer from the crude parameterizations of the atmospheric and oceanic processes, whereas the inherent complexity and computational cost of coupled GCMs prohibit detailed analyses and the investigation of longer time scales. However there are indications from these studies that indeed multiple steady states and millennial time-scale oscillations characterized by rapid changes from one state to another are not necessarily an artefact of MBCs but might occur in the climate system. Using a 2-D ocean model coupled to an atmospheric energy balance model Winton (1997) and Sakai and Peltier (1997) were able to simulate millennial time-scale oscillations which they related to the observed Dansgaard-Oeschger (D-O) cycles observed in the Greenland ice cores (Dansgaard et al. 1993). Manabe and Stouffer (1988) noted the existence of two quasi-stationary states in a coupled AOGCM. One state resembled the present climate whereas in the other state the THC has collapsed. This result is, however, criticized by Power and Kleeman (1994) because of the use of flux correction. Tziperman (1997) argued for a coupled AOGCM that a weak THC is inherently unstable, displaying large oscillations or even a collapse. Also a number of greenhouse scenario experiments with AOGCMs indicate the possibility of the weakening or even a collapse of the THC due to freshening of surface waters in the sub polar regions (Schiller et al. 1997; Wood et al. 1999).

However, although there is increasing evidence from AOGCMs that multiple steady states exist in the climate system and that small climate changes might result in large changes in the THC, to date no AOGCM has been able to simulate under constant boundary conditions millennial time-scale oscillations characterized by rapid climate changes. Until now oscillations of the THC are only observed in idealized models in diffusion dominated regimes. Here we report the results of an AO climate model of intermediate complexity (ECBilt) (Opsteegh et al. 1998). The model is sufficiently realistic that it simulates qualitatively the observed climate whereas, in contrast to state-of-the-art AOGCMs, it is still computationally cheap enough to permit long integrations. After a spin up of 4 kyr the model was integrated for another 40 kyr. During this integration the climate showed quasi-periodical behaviour with large rapid transitions between different quasi-stationary states. The residence time of the quasi-stationary states is in the order of 5 kyr, whereas the transition from one state to the other is on the decadal time scale. Analysis reveals that the influence of sea-ice on the the heat and fresh water fluxes is crucial for the occurrence of this quasi-periodical behaviour, an effect which has not been taken properly into account in experiments with MBCs.

In section 2, a short description of the coupled model is given. The results of the 40 kyr integration and an analysis of the observed rapid transitions and ultra-low frequency behaviour are presented in section 3, followed by a discussion in section 4.

2 The model

ECBilt is a spectral T21 global three level quasi-geostrophic model with simple parameterizations for the diabatic processes. The dynamical component of the atmospheric model was developed by Molteni (Marshall and Molteni 1993). The physical parameterisations are similar as in Held and Suarez (1978). As an extension to the quasi-geostrophic equations, an estimate of the neglected ageostrophic terms in the vorticity and thermodynamic equations is included as a temporally and spatially varying forcing. This forcing is computed from the diagnostically derived vertical motion field. With the inclusion of the ageostrophic terms the model simulates the Hadley circulation qualitatively correct. This results in a drastic improvement of the strength and position of the jet stream and the transient eddy activity. Despite the inclusion of these additional terms the model is two orders of magnitude faster than AGCMs. The model is realistic in the sense that it contains the minimum amount of physics that is necessary to simulate the mid-latitude planetary and synoptic-scale circulations in the atmosphere as well as its variability on various time-scales. It is coupled to a simple coarse resolution GFDL type ocean model and a thermodynamic sea-ice model. The horizontal resolution is about 5.6 times 5.6 degrees, it has 12 levels in the vertical and a flat bottom. The timestep for the atmospheric part and the sea-ice model is four hours, for the ocean it is one day. The models are synchronously coupled. For a detailed description of the atmosphere-, ocean- and sea-ice model the reader is referred to Opsteegh et al. (1998), Lenderink and Haarsma (1994, 1996) and Haarsma et al. (1996). The model runs on present generation workstations, taking 0.2 hr cpu time for the simulation of 1 yr (Power Indigo of Silicon Graphics).

Starting from a state of rest and an idealized temperature distribution, the coupled model was spun-up for 4 kyr. In order to reduce a global warming trend and to retain a North Atlantic meridional overturning circulation some tuning was performed. Details of the spin up procedure and the climate of the first 1000 year after the spin-up are discussed in Opsteegh et al. (1998). Decadal variability in the North Atlantic and the Southern ocean during this first 1000 year integration compares favourably with the observations. In the North Atlantic the model displays a mode of variability which resembles the North Atlantic Oscillation (NAO) and has a preferred time scale of about 16 years. Analysis of the Southern ocean reveals a signal similar to the Antarctic Circumpolar Wave (White and Peterson 1996). These modes are described and analysed in Selten et al. (1999) and Haarsma et al. (2000) respectively. After the spin-up the model was integrated for another 40 kyr. During this 40 kyr integration the boundary conditions, like solar irradiance, chemical composition and topography are kept constant except for the seasonal cycle in the solar irradiance.

3 Description of the ultra-low frequency variability

During the 40 kyr integration the model displays quasi-periodical behaviour. Although this is a global phenomenon it is most apparent in the Southern Ocean.

Figure 1a shows that the maximal strength of the meridional overturning in the Southern Ocean around year 3K suddenly increases from 4 Sv to 48 Sv. After this strong flush the overturning gradually decreases in about 2000 years from 30 Sv to values between 10 and 15 Sv. Around year 11K the meridional overturning is abruptly reduced to 2 Sv. A sudden increase to about 45 Sv again occurs around year 15K after which the whole cycle is repeated twice. The North Atlantic overturning, shown in Fig. 1b, also displays the quasi-periodical behaviour although the changes are much more gradual. The time series of the Southern Ocean overturning cell indicates the existence of two quasi-stationary states of about 4 and 12 Sv respectively, with rapid transitions between those states. Figure 2 shows the structure of the southern overturning cell for the two states. The accompanying thermal and salinity structures are displayed in Fig. 3 and Fig. 4 respectively. The weak southern overturning (WO) state is characterized by a large freshwater anomaly at the surface and warm deep ocean temperatures at the South Polar latitudes, whereas the strong southern overturning (SO) state displays much more saline surface waters and a vertical homogeneous temperature distribution. The convective depth of these states, as a measure of the strength of the convection, is shown in Fig. 5. It shows that deep convection mainly occurs in the Weddell, Ross and Bellingshausen sea and that during the WO state it is much reduced compared to the SO state.

The rapid transition between a state with deep convection and strong overturning to a state with shallow convection and weak overturning is commonly known as a polar halocline catastrophe (PHC). A PHC starts with an interruption of the convection. A fresh surface layer then readily develops because of the absence of the supply of saline water from the deeper layers by convection. The formation of the freshwater cap at the surface strongly stabilizes the stratification. This positive local feedback has been investigated by numerous investigators (e.g. Lenderink and Haarsma 1994).

Seeking the cause of the interruption of the deep convection and the transition from the SO to the WO phase we noticed that prior to the interruption, the deep ocean gradually cools in the areas of convection, thereby reducing the temperature difference between the surface layer and the deep ocean to almost zero (Fig. 6). This reduces the driving force of the convection (upward thermal buoyancy flux). The convection stops when the upward thermal buoyancy flux no longer compensates stratification due to surface freshening (downward buoyancy flux) as pointed out by Winton (1997). The cooling of the deep ocean is most prominent in the Ross sea, which is the location where the convection is first interrupted. The deep ocean cools because the poleward horizontal diffusion and advection of heat in the deep ocean no longer compensate the heat loss due to convection and vertical advection in the Southern Ocean. Figure 1a shows that there is a gradual increase of the meridional overturning in the Southern Ocean before the onset of a PHC. This is caused by the increasing surface salinity in the convecting areas, shown in Fig. 6, which is due to the advection of more saline water from lower latitudes by the southern meridional cell. This is an advective positive feedback which was first pointed out by Stommel (1961). The enhancement of the southern meridional overturning increases the downward vertical advection of cold water to the deep ocean close to Antarctica which results in a cooling of the deep ocean and eventually causes the shut down of convection due to unsufficient upward thermal buoyancy flux. After the shut down of the convection in the Ross sea the convection in the Weddell sea also stops about 50 years later. The interruption of the convection in the Weddell sea is due to the freshening of the surface waters by the advection of fresh water from the Ross sea by the Antarctic circumpolar current when the halocline in the Ross sea has developed. The advective positive feeback which enhances the southern meridional overturning and thereby cooling of the deep ocean due to the downward advection of cold surface waters is crucial for the transition from the SO to the WO phase.

During the WO phase there is a gradual increase of the deep ocean temperatures as shown in Fig. 7. These temperatures rise because there is no longer a supply of cold surface waters by convection. A budget analysis (results not shown) reveals that the rate of increase of the deep temperatures in the Antarctic seas is set by diffusive processes. The rapid rise just after the shut down of the convection is caused by horizontal diffusion from the midlatitudes into the Antarctic region. The slower rise during the second half of the WO period is determined by the downward vertical diffusion from the tropical and midlatitude surface layers toward the deep ocean, which is subsequently diffused horizontally to the deep Antarctic seas. During this period not only the Antarctic seas, but the entire deep ocean displays a gradual increase of temperature. The rise of deep ocean temperatures eventually destabilizes the deep Antarctic ocean, causing the re-onset of convection and the termination of the WO phase. Because of the build up of a large heat reservoir in the deep ocean during the WO phase, the convection is initially strong, resulting in a flush with a very strong southern overturning of about 40 Sv (Fig. 1b).

A schematic picture of the different processes involved in the oscillation is depicted in Fig. 8. In this figure also the positive effect of the increased sea-ice cover during the WO state on the forming of the polar halocline is sketched. This effect will be discussed more in detail in section 3.1. The mechanism of the oscillation shown in Fig. 8, apart from the role of sea-ice, is similar to the one described by Winton and Sarachik (1993) and Winton (1995, 1997), which they called deep-decoupling oscillations. Essential for this type of oscillations to occur is that during the coupled phase with a strong overturning the deep ocean heat budget displays a small imbalance. As discussed above the small imbalance results in a cooling of the deep ocean until the upward buoyancy flux vanishes, resulting in a halocline catastrophe. As stated by Winton and Sarachik (1993) this imbalance occurs when there is an incompatibility between the effect of deep overturning on salt and heat transports and the deep heat budget. Sensitivity analyses performed Winton and Sarachik (1993) and Winton (1995, 1997) revealed that the deep-decoupling oscillations disappeared for larger vertical diffusivity and higher atmospheric temperatures. In addition increasing windstress and horizontal diffusivity also will stabilize these oscillations. Because of the sensitive dependence on atmospheric temperatures Winton (1997) suggested that deep-decoupling oscillations might explain the observed Dansgaard-Oeschger (D-O) cycles in the Greenland ice cores (Dansgaard et al. 1993), which only occured during glacial periods. Deep-decoupling oscillations occur both in two- and three-dimensional ocean models, but over a much broader range of forcing in three dimensional models. The nonlinearity of the equation of state appeared not to be essential (Winton and Sarachik 1993) in contrast to the oscillations studied by Cai and Chu (1997). Winton and Sarachik also show that deep-decoupling oscillations are very different from the loop-type oscillations (Welander 1986) in which a salinity anomaly can be traced as it passes through the deep overturning circulation. These type of oscillations were also simulated in their experiments but had a much smaller amplitude and time scale.

A quasi-periodical oscillation in the Southern Hemisphere with a much shorter time scale of about 320 years was observed by Mikolajewicz and Maier-Reimer (1990) in the Hamburg LSG OGCM with realistic bottom topography. In this experiment the LSG model was forced with MBCs plus an additional white noise forcing in the freshwater fluxes. Pierce et al. (1995) (Hereafter PBM) made a detailed analysis of this oscillation. The mechanism described in PBM appears to be qualitatively similar to the mechanism operating in ECBilt and the deep-decoupling oscillations. In addition they showed that this oscillation also occurs without added noise for slightly different freshwater profiles. The outstanding difference between the oscillation of PBM and ECBilt is the factor of 10 difference in time scale. This could have several causes. In the first place the absence of bottom topography in our model strongly underestimates the barotropic transport in the Antarctic region (Cai 1994). Therefore in order to obtain a realistic barotropic mass transport through Drake Passage, the boundary value of the streamfunction Ψ at Antarctica was set at 100 Sv (Opsteegh et al. 1998). This procedure, however, does not entirely correct for the underestimated meridional mass and heat transport, resulting in slower temperature trends in the Antarctic. Secondly the ocean circulation in PBM is forced with MBCs implying the absence of a change in E-P fluxes due to changes in the atmospheric circulation and sea-ice coverage. In PBM a thermodynamic sea-ice model is employed, but in that study the sea-ice model only affects the freshwater flux due to brine rejection and melting and not due to changes in E-P. Below in section 3.1 we will argue that the E-P changes, mainly resulting from changes in sea-ice coverage, are an important cause for the ultra long time scale in ECBilt.

3.1 Sea-ice changes and Atmospheric response

The transition between an SO and a WO state is accompanied by large changes in sea-ice coverage in the Weddell and Ross sea as shown in Fig. 9ab: From almost ice free conditions during the SO state to complete ice-coverage during the WO state. The total sea-ice coverage in the Antarctic ocean is halved from $8 \times 10^6\ km^2$ during the WO state to $4 \times 10^6\ km^2$ during the SO state. However even in the the WO state the sea-ice coverage is considerably smaller than the observed sea-ice coverage of the present climate which in winter is about $16 \times 10^6\ km^2$ (Ropelewski 1989). This is due to the considerably warmer climate of ECBilt in the the midlatitude and polar regions. Close to the Antarctic the SST during the WO state are about 6K warmer than observed (Haarsma et al. 2000).

The large changes in sea-ice coverage between a SO and a WO state have a strong impact on the heat and fresh water balance. The latent and sensible heat fluxes are strongly reduced over the sea-ice covered area as illustrated in Fig. 10ab, resulting in a net heatflux reduction between 50 and 100 Wm^{-2}. As shown by Zhang et al. (1993) and Power and Kleeman (1994) a strong reduction in surface heatflux, which further enhances the stability of the water column, is an essential prerequisite for the occurrence of a PHC.

Closer inspection of the timing of sea-ice formation shows that sea-ice formation starts after the convection has stopped and the SST has been cooled to freezing level. Before the onset of the interruption of the convection the SST in the convective areas are above the freezing level of sea-water (-2^0C) as shown in Fig. 6. The SAT are however well below -2^0C, due to the cold air coming from the Antarctic continent. The large temperature difference between SST and SAT results in surface heatfluxes in the order of 100 Wm^{-2}. As soon as the convection stops the heat source from the deep ocean is cut off and the upper ocean is cooled by the cold Antarctic air to freezing level resulting in the formation of sea-ice. Inspection of the time series of the heatflux and sea-ice coverage (not shown) reveals that the increase in sea-ice coverage at the transition from the SO state to the WO state occurs one or two years later than the reduction of surface heat flux. Sea-ice formation is important for the development of a PHC, but it is not the trigger. The trigger is the interruption of convection due to insufficient upward thermal buoyancy flux.

In response to the reduced surface heatflux also the atmospheric circulation significantly changes over Antarctica. Figure 11a-d shows the difference between the WO and SO state for SST, surface air temperature (SAT), 800 hPa geopotential height (Φ_{800}) and precipitation respectively. The largest differences in SST occur in the Ross and Weddell sea with maximum values of about 4 ^{0}C. Due to the insulating effect of sea-ice, which strongly reduces the atmosphere-ocean heatflux, the response in SAT over the ice covered areas is much larger with maxima of about 8 ^{0}C. The circumpolar vortex (Fig. 11c) is increased and modified. The modification is responsible for the occurrence of localized areas of warming around 100 ^{0}W and 60 ^{0}E due to the increased advection of warm air from the north. The changes in the atmospheric circulation also alter the precipitation pattern up to 50 cm/year. However changes in the P-E flux are dominated by sea-ice induced changes in evaporation. The net effect is a 30 % increase during the sea-ice covered WO state in the P-E flux averaged over the areas of deep water formation from 30 to 40 cm/year. This enhances the freshwater cap.

In order to test the importance of this increase in the freshwater flux for the prolongation of the WO state we did the following experiment. We averaged, starting from year 19K during the SO state, over 100 year the daily freshwater fluxes retaining the annual cycle. These fresh water fluxes were then used instead of the actual fresh water fluxes in a 1000 yr coupled integration starting from year 11K just after the onset of the WO state. We will call this experiment FFW. Figure 12 (thin line) shows that already after about a decade the southern overturning increases, with a small flush, to a steady value of about 7 Sv. Closer inspection reveals that convection in the Ross Sea has started, whereas the Weddell sea is still ice covered without convection. Although this experiment suggests that the change in the freshwater fluxes is crucial for the maintenance of the WO state a caveat is that in experiment FFW the freshwater flux, apart from the annual cycle, is fixed in time. Hence we cannot discard the possibility that the presence of interannual fluctuations is essential for the maintenance of the WO state. To check this we performed a control experiment in which we repeated the FFW experiment, but now using the freshwater flux averaged over 100 year during the WO state instead of the SO state. During this control experiment the model remained in the WO state, confirming that it is the change in the freshwater fluxes and not the stochastic fluctuations which is crucial for the maintenance of the WO state. Also in this short experiment the deep ocean temperatures are rising, which similarly as in the control run would ultimately result in a flush.

In order to see whether the intermediate overturning (IO) situation, with only convection in the Ross sea, is an artefact of the imposed freshwater fluxes, we continued experiment FFW for another 1000 years but now fully coupled. During this integration the model remained in the IO state, which thus appears to be another (quasi-)stationary state. In contrast to the WO state the deep ocean temperature does not show an upward trend. At present we cannot answer the question whether the IO state is really stationary or eventually becomes unstable. Only another ultra long integration, which has not been carried out yet, can answer this question.

Repeating experiment FFW but starting from year 12K, 1000 years after the onset of the WO state, in which there has already been a significant increase of the heat storage in the deep ocean, does show a rapid switch from the WO to the SO state (Fig. 12, thick line). This explains why the transition from the WO to the IO state is never observed during the 40 kyr integration. The prolongation of the WO state due to the effect of sea-ice enables the building up of the heat storage in the deep ocean. This inhibits the transition from the WO to the IO state.

In another experiment, a transition to the IO state could also be accomplished from the SO state by imposing a freshwater perturbation in the surface layers of the Weddell sea. This freshwater perturbation consisted of replacing the salinity of the upper 80 m by a hundred year mean salinity during the WO state.

One might speculate about the existence of another (quasi-)stationary state which is the mirror of the IO state: convection in the Weddell sea and no convection in the Ross sea. To test this we performed a similar freshwater perturbation experiment as described above but now for the Ross sea. The result is that after the shut down of the convection and the increase of the sea-ice coverage in the Ross sea the convection also stopped in the Weddell sea after about 50 years. This is caused by the advection of fresh water from the Ross sea to the Weddell sea by the Antarctic circumpolar current. This counter acts the immediate atmospheric response which shows a decrease in SAT over the Weddell sea due to changes in the atmospheric circulation. The mirror of the IO can thus not be triggered from the SO state: A freshwater perturbation in the Ross sea enforces a transition from the SO to the WO state. This explains why during the 40 kyr run a transition from an SO to an IO state never happened, because as we discussed before the interruption of the convection always started in the Ross sea. Also starting from the WO state the mirror of the IO state could not be triggered: Salt perturbations in the Weddell sea indeed initiated convection there but this convection soon died out due to the advection of fresh water from the Ross sea. We therefore conclude that the mirror of the IO state does not exist.

A natural question is why in the WO state when the Ross and Weddell sea are covered with sea-ice, the convection does not move to other locations outside the sea-ice covered region as often observed in numerical experiments in the North Atlantic (Rahmstorf 1994). Closer inspection of Fig. 5 reveals that no new convecting areas are emerging, but that the convective activity in the Bellingshausen sea, located between 70 ^{0}W and 100 ^{0}W has enhanced. However, this enhanced convection is not strong enough to prevent a continously warming of the deep ocean. This is due to the fact that this convection area is located at the border of Antarctica with higher SST's (about 3 ^{0}C), than in the Ross and Weddell sea (about 1 ^{0}C), which are intensively cooled by by continental winds from the surrounding Antarctica. From this we conclude that the relatively warm climate of ECBilt around Antarctica prevents the formation of cold bottom water in other locations than the Ross and Weddell sea.

In summary we conclude that the sea-ice changes are crucial for the ultra-long quasi-periodical behaviour in the 40 kyr run. The influence of sea-ice on the dynamics of the ocean circulation has received relatively little attention. Using mixed boundary conditions Lenderink and Haarsma (1996) stressed the fact that sea-ice enhances the self sustaining character of convection. Lohman and Gerdes (1998) and Jayne and Marotzke (1999) studied the sensitivity of the sea-ice effects in a coupled model. These studies confirm the importance of the effect of sea-ice on the ocean circulation. However, they also show that the sign of this effect sensitively depends on the atmospheric heat and moisture transports which in their models had to be parameterized. Our results indicate that for the Southern Ocean the sign of the sea-ice effect is positive. After the interruption of the deep convection due to unsufficient upward thermal buoyancy flux, caused by the cooling of the deep ocean, the sea-ice facilitates the occurrence of a PHC due to its insulating effect and strongly enhances the period of the oscillations, because of the increase in the freshwater flux. The absence of bottom topography probably also enhances the time scale. The magnitude of this effect could, however, not be addressed within the context of this study.

3.2 Northern Hemisphere response

Variations in the North Atlantic (NA) overturning (Fig. 1a), are closely connected to those of the southern overturning cell. The NA overturning shows a rapid decrease after the transition from the SO to the WO state in the Southern Ocean. This is due to the freshening of the surface waters in the Southern Ocean after the shut-down of the convection. These relatively fresh surface waters are advected on a decadal time scale into the North Atlantic where the convection is reduced, resulting in a weaker NA overturning. The subsequent gradual increase of the NA overturning after this rapid decrease is due to the continously rising of deep ocean temperatures in the Southern Ocean during the WO state and the subsequent slow diffusive spreading of this heat into the North Atlantic and other oceans. This enhances the convection in the North Atlantic which strengthens the NA overturning. After the re-onset of the SO state the NA overturning is, after a short enhancement as a result of the salinity pulse from the Southern Ocean, again slowed down due to the cooling of the deep oceans.

Figure 13 shows the difference in the North Atlantic meridional volume transport and zonal mean temperature between periods of strong and weak NA overturning. The maximum change in the meridional volume transport of about 5 Sv occurs around 10^o N. The poleward branch of this anomalous overturning cell displays an upward tilt. As a consequence the zonal mean temperature anomaly shows a minimum of 0.2 ^{o}C around 500 m depth at 40^o N due to the advection of cold water from below which pushes the thermocline closer to the surface. The temperature rise of about 1 ^{o}C related to the increase of the meridional overturning is concentrated at the intermediate levels around 2000 m. A significant change in the zonal mean surface temperature is only notable poleward of 70^o N, where intensified convection mixes warm water from below to the surface.

The changes in the Northern Hemisphere climate related to these localized changes in SST (not shown) are small. Only very close to the area of deep water formation in the northern part of the GIN sea significant changes in SST and SAT of about 1 ^{o}C are observed. Due to the very small area were these changes occur the resulting changes in Φ_{800} are weak, less than 0.2 dm outside that area.

4 Discussion

A 40 kyr integration with ECBilt reveals a quasi-periodical behaviour with a period of about 13 kyr. The largest changes occur in the Southern Hemisphere. The mechanism of this quasi-periodical behaviour appears to be similar as the deep-decoupling oscillations described by Winton and Sarachik (1993) and Winton (1997). An important difference is the crucial role of the sea-ice on the surface heat and freshwater fluxes. Both the reduction of the surface heatflux and the enhancemeant of the P-E flux enhance the stability of the water column thereby prolonging the duration of the WO state.

This study is the first to report the occurrence of rapid climate changes in a coupled climate model in which the atmospheric circulation, the ocean circulation and the sea-ice coverage are computed interactively without the use of fluxcorrection techniques. Our results confirm the results of earlier studies with OGCMs with some noteworthy differences. The strong reduction in heat flux during a PHC under MBCs without sea-ice is due to a large relaxation constant for the heatflux and the fact that the atmospheric temperature cannot adjust to the SST. This is not realistic (Zhang et al. 1993; Power and Kleeman 1994). In our study the reduction in heatflux is caused by the increase in sea-ice coverage. In addition the increase in P-E during a PHC due to the reduction of the evaporation over sea-ice has not been taken into account in experiments with MBCs which did use a sea-ice model like the study of PBM.

In analyzing the deep-decoupling oscillations Winton (1997) demonstrated that a perequisite for these oscillations to occur is sufficient cold deep ocean temperatures. If the deep ocean temperatures are too warm, the upward thermal buoyancy flux will always be sufficient to balance the downward buoyancy flux due to surface freshening. In this case convection will be sustained and a PHC will never occur. Winton (1997) used this argument to defend the hypothesis that deep-decoupling oscillations are a possible explanation for the D-O oscillations, which are only observed during the glacial periods and not during the warmer interglacial periods. Using the same argument we state that due to the warm deep ocean temperatures in the North Atlantic in ECbilt, which are in the order of 5 ^0C, deep-decoupling oscillations are not observed there. Only in the Weddell and the Ross sea the deep ocean temperatures, in the order of 1 ^0C, are sufficiently cold for generating a PHC. After the generation of a PHC the gradual rise of of the deep ocean temperatures due to the advection and diffusion of heat from lower latitudes eventually destabilizes the vertical stratification and cause the re-onset of convection. Only in a very cold climate we expect a PHC as a stable solution of the ocean circulation.

The period of the oscillation in ECBilt, which is in the order of 10000 year, differs widely from the periods of two other oscillations mentioned in this paper, the oscillation described by PBM with a period of about 300 years and the deep-decoupling oscillations of Winton and Sarachik (1993) and Winton (1997), which are in the order of about 5000 years, although the basic mechanism in those three oscillations appears to be the same. We hypothesize that these large differences in time scale are due to the different sizes of the basins involved and the different feedback mechanisms which are operating. The short time scale in PBM is probably due to the small basin and the absence of the E-P feedback of sea-ice, whereas the long time scales in Winton (1997) and ECBilt are caused by the large size of the basin and the positive feeback of sea-ice respectively. The absence of bottom topography in ECBilt might also affect the period of the oscillations although at present we do not know the sign and the magnitude of this effect.

Due to the simplifications made in the ocean model, like the absence of bottom topography and sea-ice dynamics, the coarse resolution and the crude parameterisation of convection, care should be taken in the extrapolation of these results to the observed past and present climate. In addition changes in atmospheric chemical composition, ice sheets, land-sea distribution and orbital parameters have not been taken into account. Below we will discuss some of these model assumptions and their relevance for the simulated oscillations.

Because sea-ice changes are crucial for the rapid changes between a WO and a SO state the absence of sea-ice dynamics should be taken into account in the interpretation of these results. Present day Antarctic bottom water (AABW) formation depends sensitively on the northwards advection of sea-ice due to the windstress. The induced northwards freshwater transport away from the Antarctic coast by the sea-ice drift results in saltier and thus denser water close to the Antarctic coast and therefore contributes to the occurrence of deep convection and the formation of AABW. This mechanism might prevent the occurence of the WO state. On the other hand the enhanced deep convection might increase the southern overturning cell and the cooling of the deep ocean and thereby the possibility of a transition from the SO to the WO state. At present we cannot answer this, future experiments with sea-ice dynamics included should clarify this point.

Due to the coarse resolution of the ocean model the diffusion is overestimated with respect to advective processes especially at deeper levels. As discussed the slow warming of the deep Antarctic ocean during the WO state is largely controlled by diffusion in the model. The warming and the resulting transition to the SO state might therefore be due to the strong diffusion in the model. Whether the advection would take over the role of the diffusion at higher resolution we don't know. However as discussed before inclusion of bottom topography would enhance the meridional mass and accompanying heat transport. In the absence of bottom topography the simulation of the southern overturning cell is affected because the model is not able to create an east-west pressure difference along the latitude of Drake passage. The southern cell is confined to poleward latitudes of Drake passage and is generated by pressure differences due to the zonal asymmetry of Antarctica.

In order to simulate a realistic Antarctic circumpolar current the barotropic mass transport was fixed at 100 Sv as described before. This, however, violates the dynamic balance between the density differences and mass flow. This imbalance could be a source for oscillatory behaviour, although earlier simulations with a flat bottom and a prescribed Drake passage transport using mixed boundary conditions (Marotzke and Willebrand, 1991) do not suggest this.

The above discussion shows that the simulated oscillation is clearly affected by the model assumptions. Despite this, we are confident that the effect of sea-ice changes on the heat and freshwater balance are qualitatively correct and that as a result PHC's and multiple steady states can occur in coupled climate models. Whether oscillations as observed in ECBilt will also occur in more realistic climate models is a question we cannot answer within the present context. However, recent paleoclimatological research of ice-cores does suggest the occurrence of rapid climate changes and millennial-scale variability in the Antarctic region (Jouzel et al. 1995; Steig et al. 1998, 2000; Kanfoush et al. 2000), which supports our results.

A present controversy is the timing between climate changes in Antarctica and the North Atlantic during the Dansgaard-Oescher events (Stocker 1998). The Vostok and Byrd core indicate an asynchronous behaviour of those regions (Blunier et al. 1997) during the Younger Dryas, with the Antarctic cooling preceding 1.8 kyr. In contrast with this, recent analyses of the Taylor Dome core (Steig et al. 1998) does suggest that they behaved synchronously. Our results suggest that this discrepancy could be due to the different locations of those Antarctic ice cores. Fig. 11 shows that the Byrd and Vostok core are situated in the region where SAT is affected by the atmospheric circulation in contrast to the Taylor Dome core which is situated close to the Ross Sea. Due to these different locations the temperature response, as shown in Fig. 11b, is opposite during the transition from the WO to the SO state or vice versa. This explanation of the differences between the Vostok and Byrd Cores and the Taylor Dome core, would lead to the conclusion that the deepwater formation was simultaneously depressed in the North Atlantic and Southern Ocean during the Younger Dryas.

However there remain still problems with the interpretation of the Antarctic ice cores due to the differences in time scales between the different cores. The Byrd and Vostok cores display a gradual cooling during the Antarctic cold reversal, whereas the Taylor Dome shows a much more rapid warming especially at the end of the last glacial maximum (Steig et al. 1998). The characteristics of the Taylor Dome are therefore more comparable with those of the Greenland ice cores which are also characterized by abrubt changes, although the changes in the Taylor Dome core are not yet as pronounced and rapid as in the Greenland ice cores. The time series of SAT in ECBilt at the locations of the ice cores does show abrubt changes at the transitions between the two quasi-stationary states. This is in accordance with Taylor Dome but in contrast to what is observed in Byrd and Vostok. As discussed in section 3.2 the temperature changes over Greenland simulated by ECBilt are rather small about 1 ^oC, much smaller than the temperature changes indicated by the Greenland ice cores. In addition these changes are gradual and not abrubt as observed. We therefore do not claim that the oscillations in ECBilt are an explanation for the observed paleorecords. The main message of the results of ECBilt is that climate changes in Antarctic are not necessarely of uniform sign. A result which, we believe, is not specific to ECBilt. The difference in sign between the response of Taylor Dome and the two other ice cores, Vostok and Byrd, is caused by the excitation of planetary waves with small zonal wave number, mainly 2 and 3. This is due to the location of the Ross and Weddell sea, which are the two main centers of diabatic heating. The NCEP reanalyses (Kalnay et al. 1996) reveal that the first EOF of 850-hPa geopotential height over Antarctica is dominated by wavenumber 3 with troughs located over the Ross and Weddell sea (Haarsma et al. 2000), indicating that this a preferred mode of variability of the atmosphere that can easily be excited.

Another consequence of this study concerns the frequently observed trends in deep ocean temperatures in 'state-of-the-art' coupled AOGCMs (Von Storch 1994; Manabe and Stouffer 1996; Tett et al. 1997; Cai and Gordon 1999). This trend is usually interpreted as a signal that the AOGCM is still not in equilibrium. The temperature fluctuations in the deep ocean in ECBilt are in the order of 2 ^oC for a period of 5 kyr (Fig. 7), resulting in a 'trend' of about 0.04 ^oC/century. This falls within the range of observed trends in AOGCMs (Von Storch 1994; Manabe and Stouffer 1996; Tett et al. 1997). Therefore our results suggest that the observed deep ocean temperature trends in AOGCMs are not necessarily a reflection of the fact that the AOGCM is not yet in equilibrium, but that they could be part of an ultra low-frequency oscillation.

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5 Figure Captions

Figure 1: Extrema of the streamfunction of the meridional overturning [Sv]. (a) Southern ocean (minimum); (b) Atlantic basin (maximum).

Figure 2: Time mean Southern Ocean meridional overturning [Sv] for the periods 12-14 kyr (WO) (a) and 18-24 kyr (SO) (b).

Figure 3: Time mean zonal mean temperature [^{o}C] for the periods 12-14 kyr (WO) (a) and 18-24 kyr (SO) (b).

Figure 4: As Fig. 3 but now for salinity [Psu] and upper 1000 m only.

Figure 5: Time mean convective depth [m] for the periods 12-14 kyr (WO) (a) and 18-24 kyr (SO) (b).

Figure 6: Surface temperature (thick solid) [^oC], deep ocean temperature at 3750m [^oC] (thin solid) and surface salinity [PSU] (dotted) in the Ross sea for the period 10-20 kyr. A rapid transition from the SO the WO state takes place around year 11 k and the transition back to the SO state around year 15 k.

Figure 7: Mean Southern Ocean (south of 45 ^{o}) temperatures [^{o}C] at 3750 m depth.

Figure 8: Schematic picture of the mechanism of the oscillations in ECBilt.

Figure 9: Sea-ice coverage (dark shaded) over the Antarctic ocean during the WO (12-14 kyr) (a) and SO (18-24 kyr) (b) phase.

Figure 10: Latent (a) and sensible heatflux (b) [Contour interval: 20 Wm^{-2}] during the WO (12-14 kyr) and SO (18-24 kyr) phase.

Figure 11: Difference between WO (12-14 kyr) and SO (18-24 kyr) phase of SST (a), SAT (b) [Contour interval: 1 ^{o}C], Φ_{800} (c) [Contour interval: 1 dm] and precipitation (d) [Contour intervals: -20 -10 -5 5 10 20 cm/year, values less than -20 cm/year are denoted by light grey shading and values more than 20 cm/year by dark grey shading]. **B** is station Byrd, **V** Vostok and **T** Taylor Dome.

Figure 12: Maximum of Southern ocean meridional overturning [Sv]. thin line: experiment FFW; thick line: As experiment FFW, but starting 1000 years later.

Figure 13: Difference in North Atlantic meridional volume transport [Sv] (a) and zonal mean temperature [^{o}C] (b) between the periods 15.00-15.25 kyr (strong NA overturning) and 11.80-12.05 kyr (weak NA overturning).

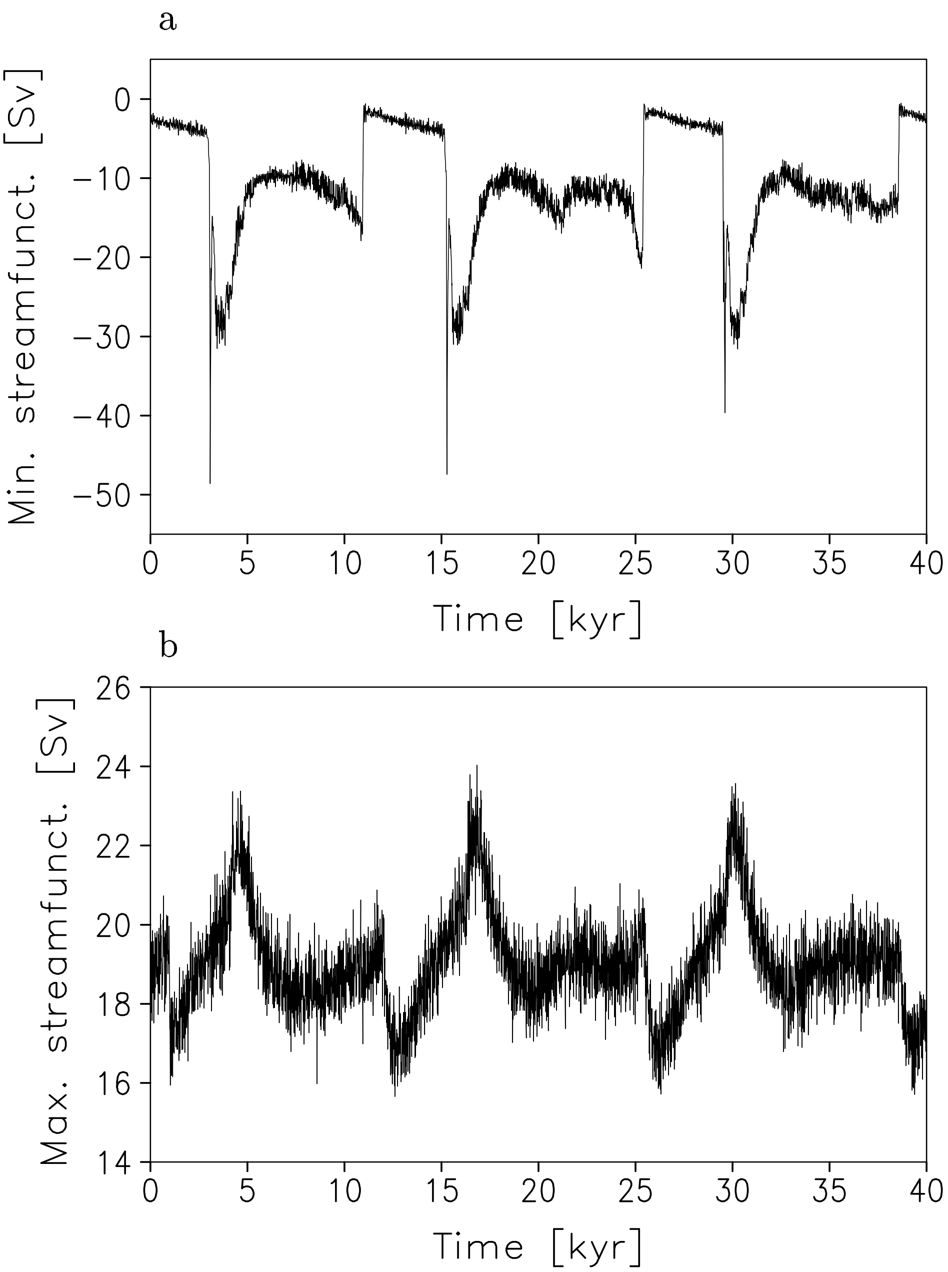

a
Min. streamfunct. [Sv]
0
−10
−20
−30
−40
−50
0
5
10
15
20
25
30
35
40
Time [kyr]
b
Max. streamfunct. [Sv]
26
24
22
20
18
16
14
0
5
10
15
20
25
30
35
40
Time [kyr]

Figure 1

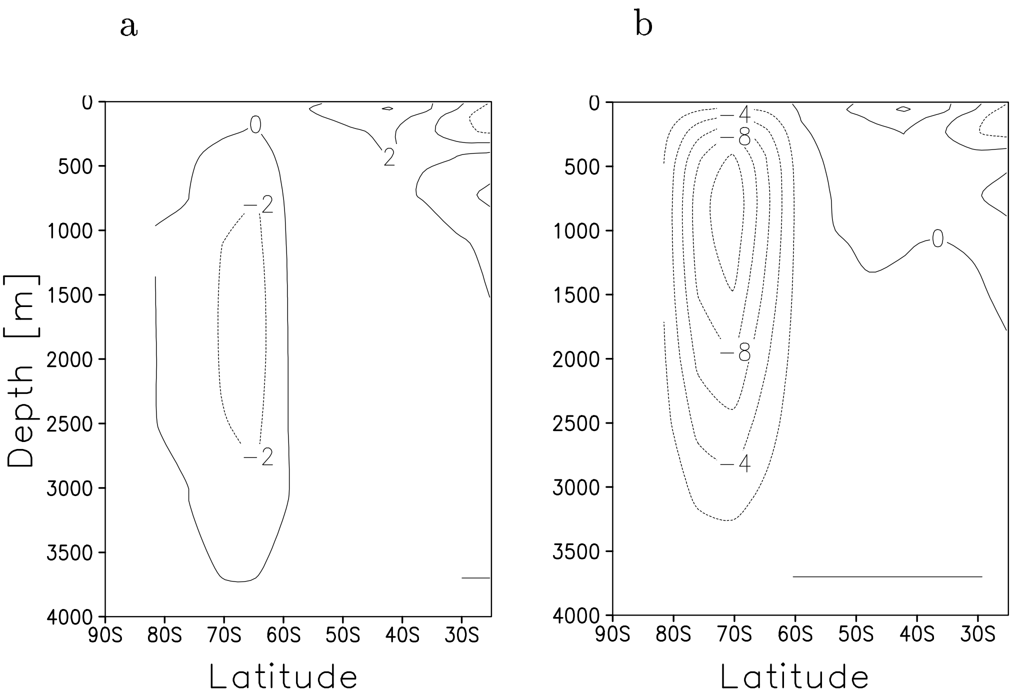

a
b
0
500
1000
1500
2000
2500
3000
3500
4000
Depth [m]
90S 80S 70S 60S 50S 40S 30S
Latitude
0
−2
−2
2
−4
−8
−8
−4
0

Figure 2

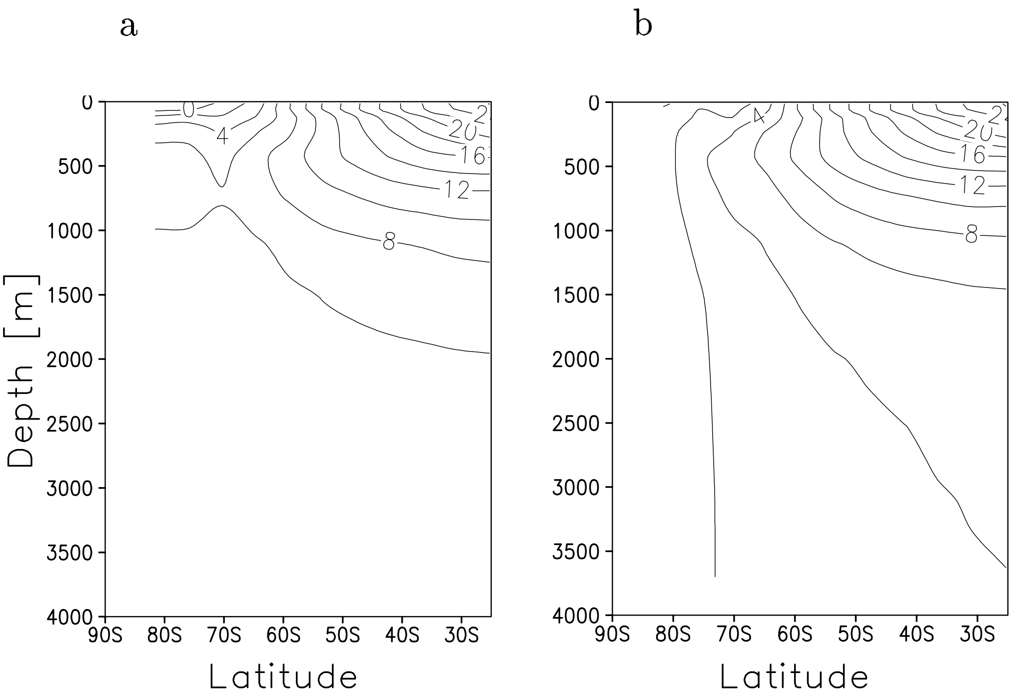

a
b
Depth [m]
0
500
1000
1500
2000
2500
3000
3500
4000
90S
80S
70S
60S
50S
40S
30S
Latitude
0
4
8
12
16
20
24

Figure 3

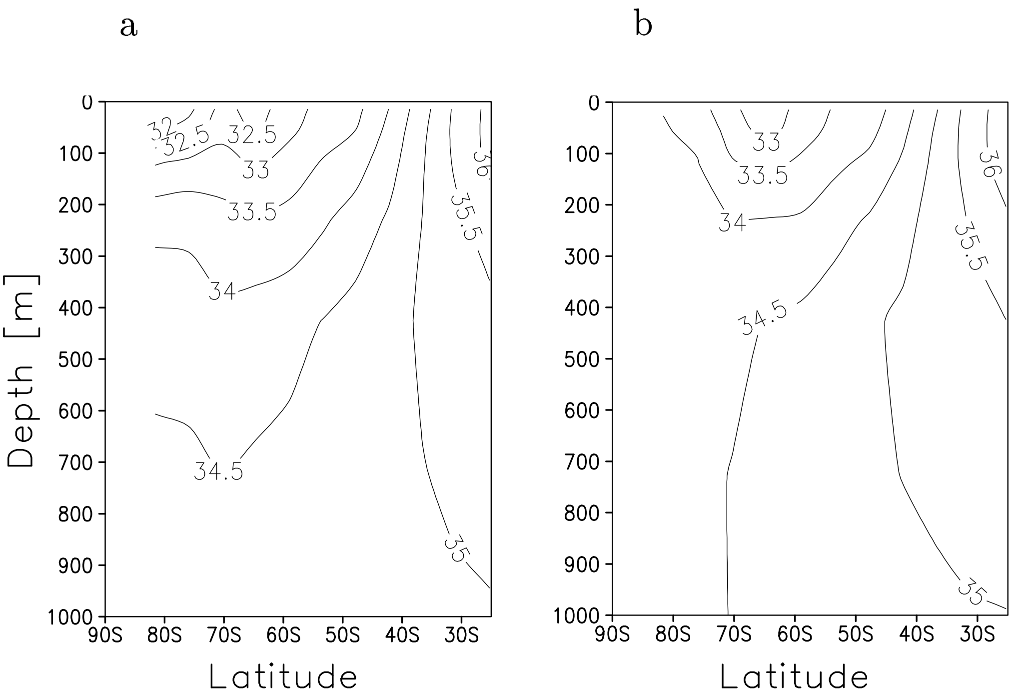

a
b
Depth [m]
0
100
200
300
400
500
600
700
800
900
1000
90S
80S
70S
60S
50S
40S
30S
Latitude
32
32.5
33
33.5
34
34.5
35
35.5
36

Figure 4

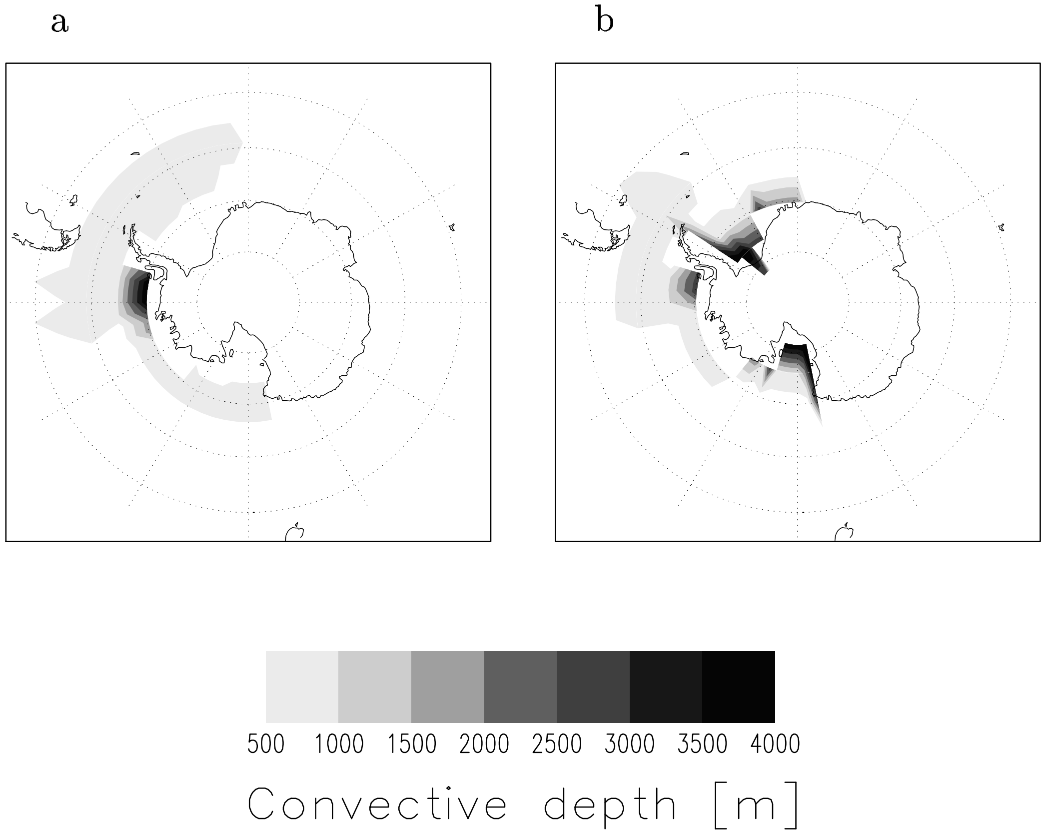


Figure 5

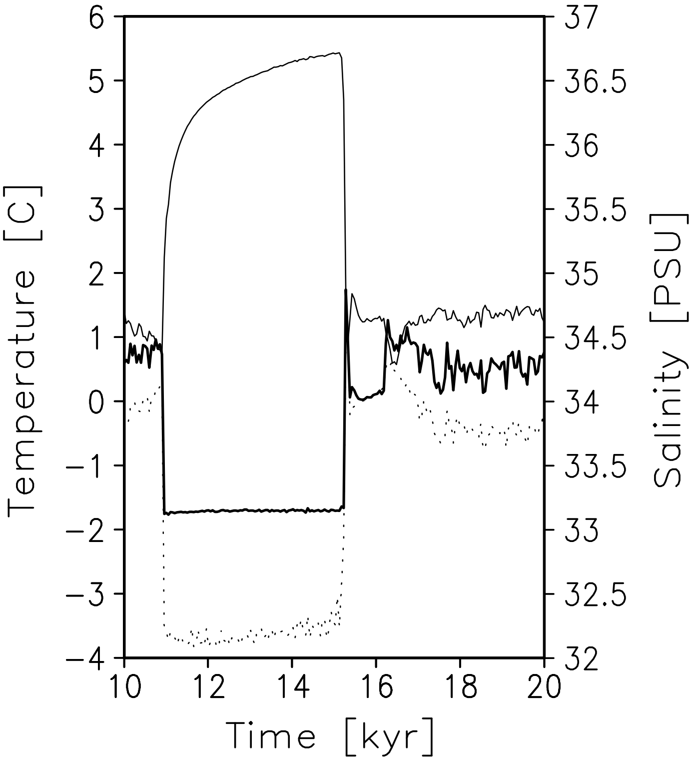

Temperature [C]
Salinity [PSU]
Time [kyr]
6
5
4
3
2
1
0
−1
−2
−3
−4
37
36.5
36
35.5
35
34.5
34
33.5
33
32.5
32
10
12
14
16
18
20

Figure 6

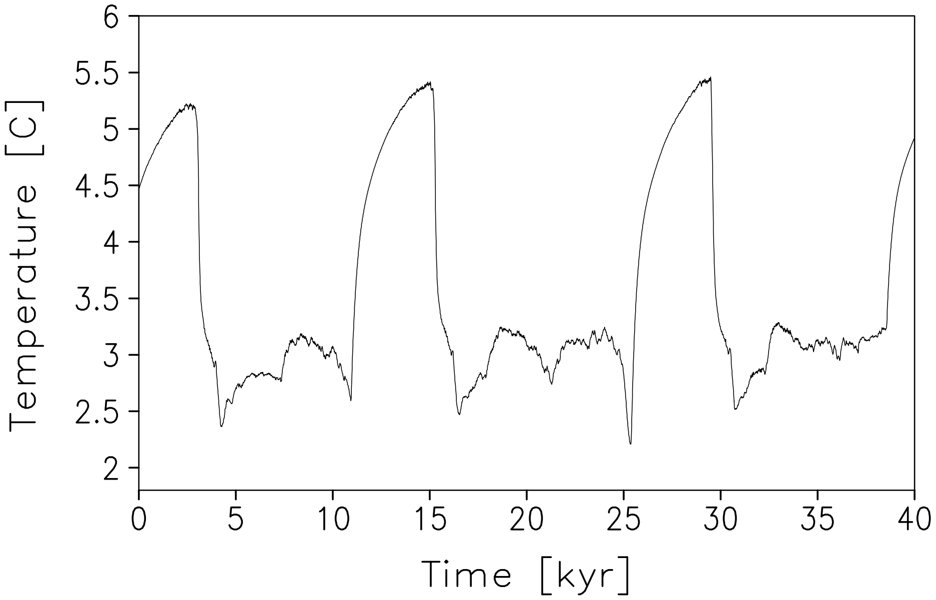


Figure 7

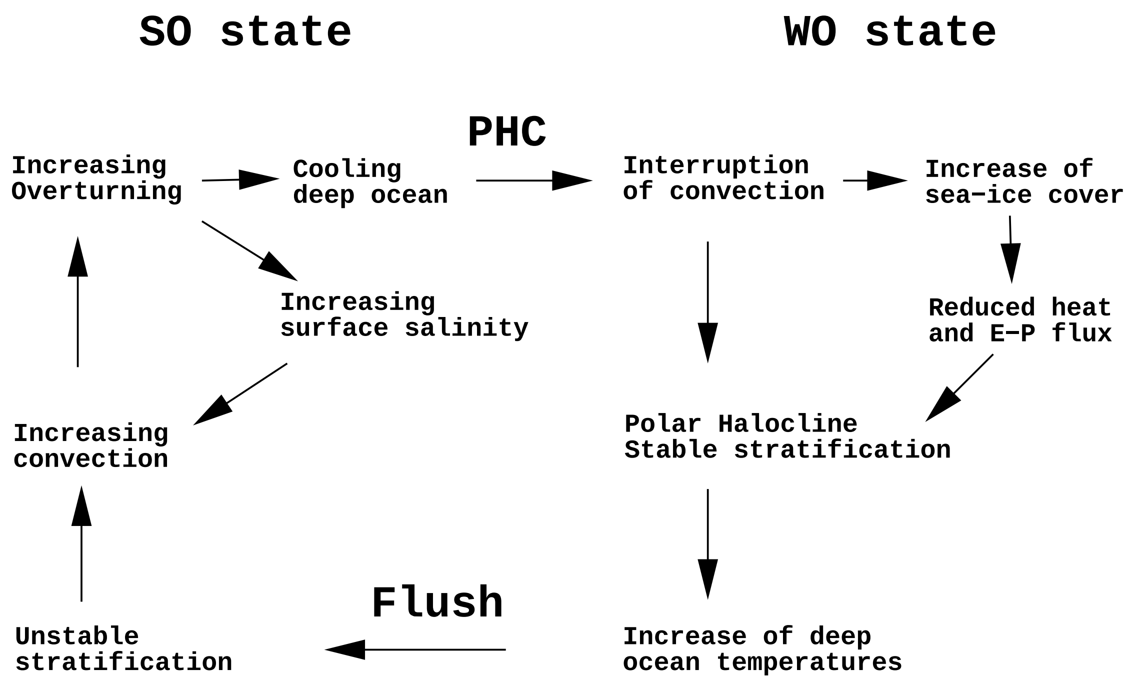

SO state
WO state
PHC
Increasing Overturning
Cooling deep ocean
Interruption of convection
Increase of sea−ice cover
Increasing surface salinity
Reduced heat and E−P flux
Increasing convection
Polar Halocline Stable stratification
Flush
Unstable stratification
Increase of deep ocean temperatures

Figure 8

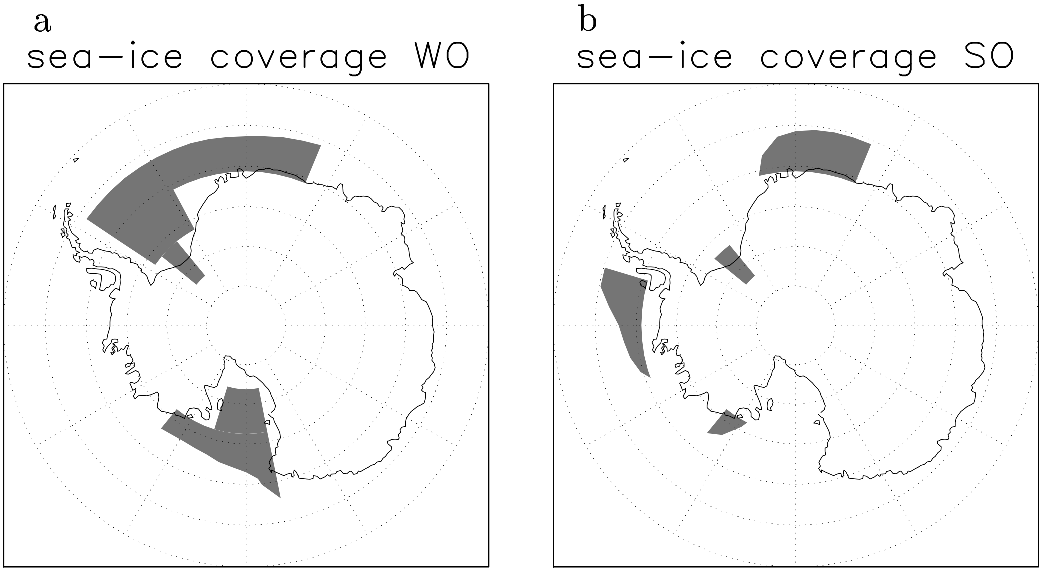
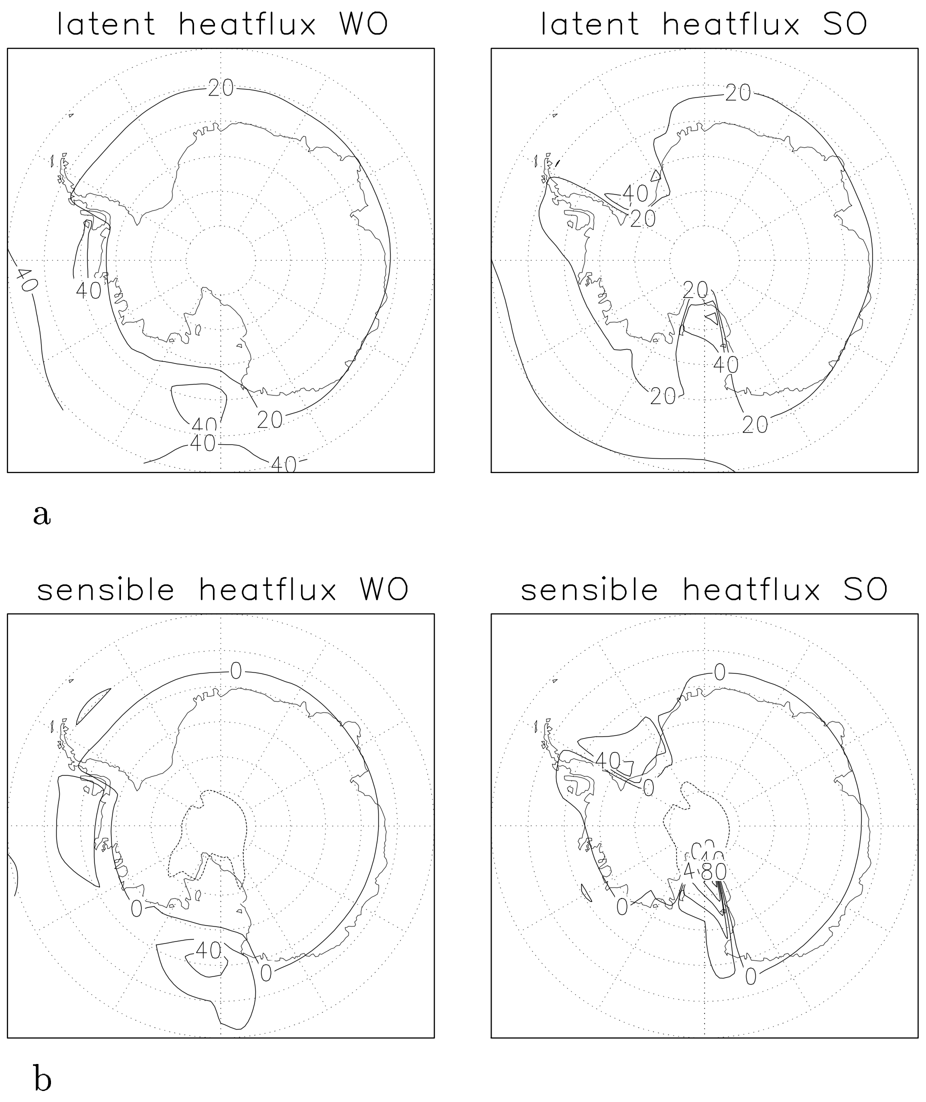


Figure 9



latent heatflux WO
latent heatflux SO
a
sensible heatflux WO
sensible heatflux SO
b

Figure 10

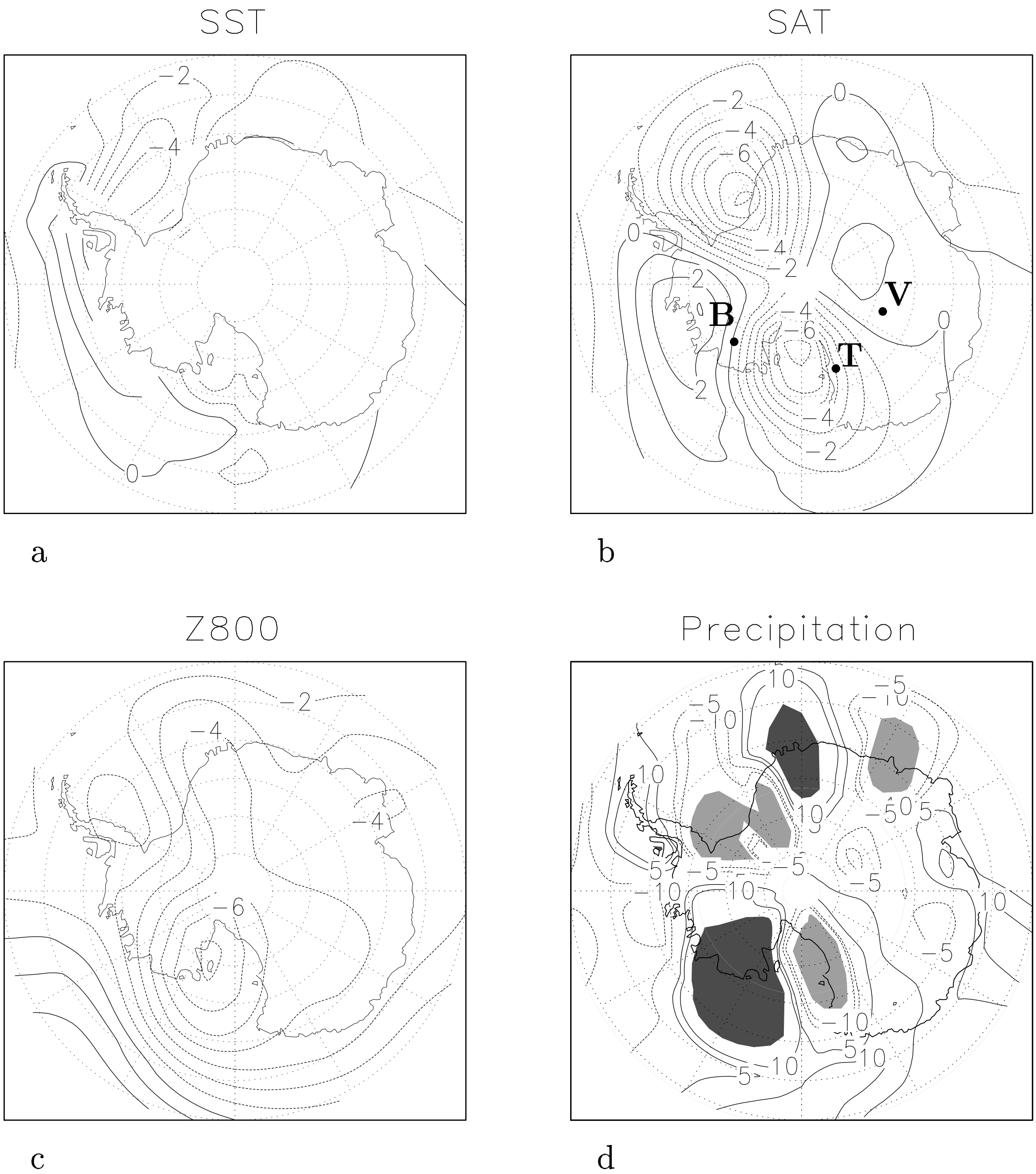


Figure 11

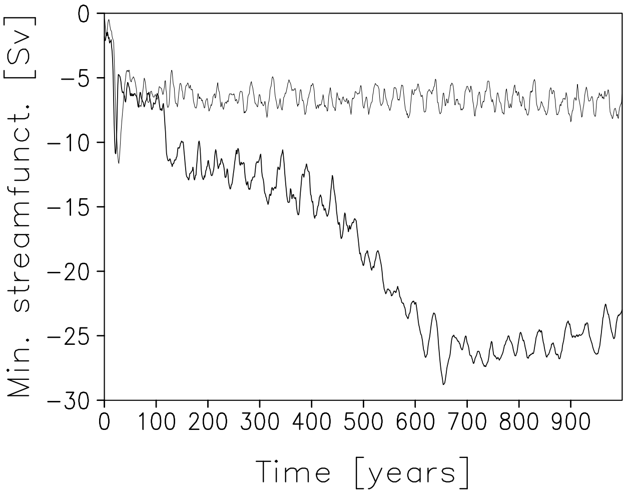
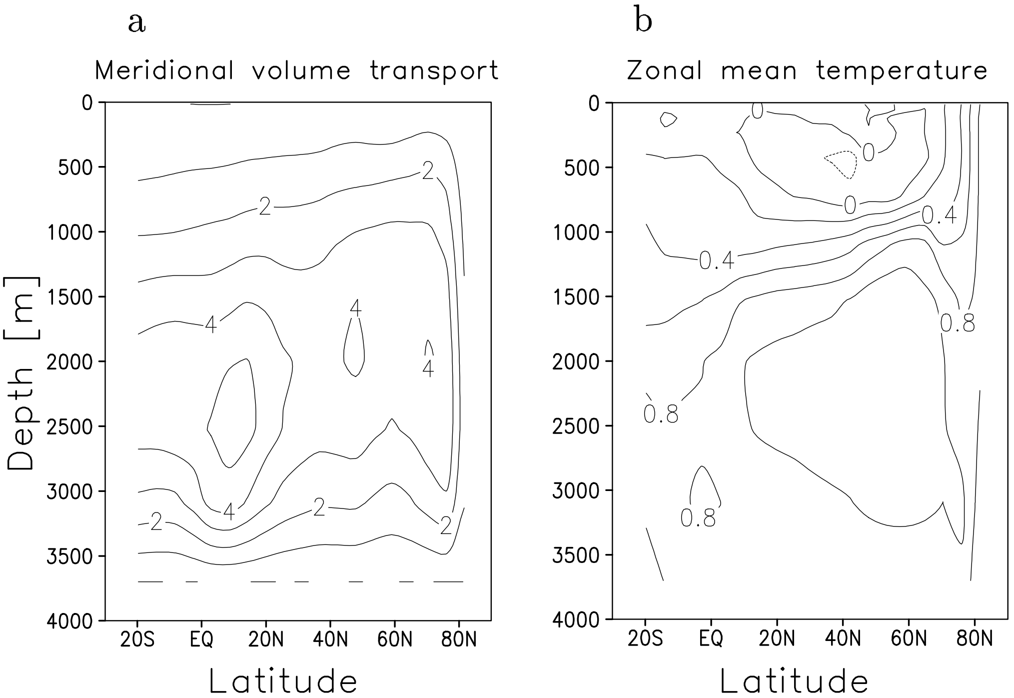

Min. streamfunct. [Sv]
0
−5
−10
−15
−20
−25
−30
0
100
200
300
400
500
600
700
800
900
Time [years]

Figure 12



a
Meridional volume transport
Depth [m]
Latitude
0
500
1000
1500
2000
2500
3000
3500
4000
20S
EQ
20N
40N
60N
80N
2
4
b
Zonal mean temperature
Depth [m]
Latitude
0
0.4
0.8

Figure 13