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The effect of ocean dynamics in a coupled GCM simulation of the Last Glacial Maximum

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Abstract General circulation models (GCMs) of the climate system are powerful tools for understanding and predicting climate and climate change. The last glacial maximum (LGM) provides an extreme test of the model's ability to simulate a change of climate, and allows us to increase our understanding of mechanisms of climate change. We have used a coupled high resolution ocean–atmosphere GCM (HadCM3) to simulate the equilibrium climate at the LGM. The effect of ocean dynamics is investigated by carrying out a parallel experiment replacing the dynamic three-dimensional ocean GCM with a static thermodynamic mixed-layer ocean model. Changes to the ocean circulation, and feedbacks between the ocean, atmosphere and sea ice have an important influence on the surface response, and are discussed. The coupled model produces an intensified thermohaline circulation and an increase in the amount of heat transported northward by the Atlantic Ocean equatorward of 55°N, which is at odds with the interpretation of some proxy records. Such changes, which the thermodynamic mixed-layer ocean model cannot produce, have a large impact around the North Atlantic region, and are discussed in the study.

1 Introduction

General circulation models (GCMs) are the state-of-the-art tools used for understanding and predicting how the global and regional climate might change due to anthropogenic effects, and form an important input into the assessments of the Intergovernmental Panel on Climate Change (Houghton et al. 2001). These models fall short of reality, partly because we have an incomplete understanding of the climate system. Therefore, we must test both the models, and our understanding of climate processes and mechanisms of climate change. Past climates provide an opportunity for such tests, and offer a means of being able to evaluate the capabilities of numerical models to realistically simulate a climate change.

The last glacial maximum (LGM) has been a focus of research using numerical models and palaeoclimatic data in international projects such as the Cooperative Holocene Mapping Project (COHMAP members 1988) and the Palaeoclimate Modelling Intercomparison Project (PMIP, Joussaume and Taylor 1995) for at least the following three reasons. The LGM represents the largest global and regional climate change of recent geologic times, the main factors responsible for the different climate are believed to be well known, and there is an abundance of well-dated palaeoclimatic data available.

The large differences between the glacial and present-day climates represent perhaps an extreme challenge for a GCM's capability to realistically simulate an equilibrium climate change. Most of the earlier studies use either atmospheric GCMs (AGCMs, e.g. Gates 1976; Kutzbach and Guetter, 1986) with the LGM sea surface temperatures (SSTs) prescribed from a reconstruction provided by the Climate: Long-Range Investigation, Mapping and Prediction project (CLIMAP Project Members 1981), or atmospheric GCMs coupled to a one-layer thermodynamic model of the oceanic mixed layer and a sea ice model (e.g. Manabe and Broccoli 1985; Hewitt and Mitchell 1997). These latter models are often referred to as slab, or atmosphere-mixed layer ocean, models. However, feedbacks between the atmosphere and deep ocean may have had important influences on the climate at the LGM, which would be omitted in model simulations with simplified treatments

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of the ocean. In addition, there is a wealth of neglected palaeoceanographic data providing information about ocean dynamics and the state of the ocean below the mixed layer.

Some studies of the glacial ocean have been carried out using ocean GCMs (OGCMs) with prescribed atmospheric boundary forcing (e.g. Seidov and Haupt 1997; Bigg et al. 1998). Such studies may also have neglected important feedbacks between the atmosphere and ocean. Feedbacks between the atmosphere and deep ocean have been investigated using a low-resolution coupled ocean–atmosphere earth system model of intermediate complexity (an EMIC) to simulate the equilibrium LGM climate (Ganopolski et al. 1998), and an OGCM coupled to a simple energy-moisture balance model representing the atmosphere (Weaver et al. 1998). EMICs cannot simulate regional details, and typically parametrize either the atmosphere or ocean dynamics, or both. The zonal average ocean model cannot simulate the gyre circulations, and the heat and salt transports of the horizontal circulation are parametrized.

A three-dimensional coupled ocean–atmosphere general circulation model (OAGCM) will allow the atmosphere, ocean and sea ice to respond to glacial boundary conditions and will allow the three components to interact and feedback on each other. Until recently OAGCMs needed artificial fluxes (flux adjustments) of heat and fresh water to produce realistic and stable simulations of the present-day climate, and such models have been criticised when used for climate change simulations: these fluxes are not necessarily applicable for an altered climate, but there is no method available for recalculating them. For simulations of glacial climates a further problem may arise in regions where the sea ice expands. Since the flux adjustments are calculated for present-day ice-free surface waters, it is possible for the model to have heat flux adjustments that tend to cool the surface. In the glacial simulation, if the model produces sea ice in such regions, these adjustments would then be constantly applied to these sea-ice points, and unrealistic growth of sea ice may occur.

The only coupled OAGCM simulations of the response to glacial boundary conditions published to date are a 15-year simulation initialized from present-day conditions (Bush and Philander 1998), which was not long enough to have allowed the ocean to have adjusted to the glacial conditions, and a flux adjusted model, of more than 200 years (the initial 70 years use ocean acceleration techniques to produce an equivalent of 1680 years integration in the deep ocean), which used the present-day land–sea distribution rather than the glacial one (Kitoh et al. 2001).

In the study described here we design and carry out simulations of the climate at the LGM using a coupled high-resolution ocean–atmosphere general circulation model. The coupled model (HadCM3) developed at the Met Office's Hadley Centre does not require flux adjustments. This study attempts to simulate the quasi-equilibrium climate at the LGM with realistic topography using a non-flux adjusted fully coupled high resolution ocean–atmosphere general circulation model.

To increase the rate of cooling of the ocean the SSTs are linearly damped toward glacial values for 70 years, a technique often referred to as Haney forcing (Haney 1971). The glacial SSTs are determined from a slab model simulation of the LGM. After the "Haney forced" stage the relaxation is discontinued and the coupled model is run for a further 1000 years. Long integrations of a lower-resolution coupled ocean–atmosphere model developed at the Geophysical Fluid Dynamics Laboratory (GFDL) are used to investigate the LGM experimental design, and are included in the Appendix.

The purpose of this study is to design and describe simulations from a coupled OAGCM and a slab model of the equilibrium climate at the LGM using forcing appropriate for the LGM 21,000 years before present. By comparing the response of the slab model, with its fixed heat transports and no ocean currents, to the response of the OAGCM we can investigate the effect of the three-dimensional dynamic ocean GCM on the response. It could be argued that the climate 21,000 years before present was not at equilibrium, if considering millennial time scales, due perhaps to variability associated with ice sheet-ocean–atmosphere interactions, but such a study is beyond the scope of three-dimensional coupled ocean–atmosphere GCMs, such as the one we use, at present.

The GCMs and the LGM experimental design we employ are described in Sect. 2. The North Atlantic thermohaline circulation (NATHC) in particular undergoes substantial changes once the LGM boundary conditions are applied, as discussed in Sect. 3. Section 4 compares the OAGCM response to that of the slab model and discusses the effect of ocean dynamics on the model response. The response of the coupled model after 700 years of simulation has been briefly described by Hewitt et al. (2001a), and the findings presented here after 1000 years of simulation are consistent with those earlier results.

2 Experimental design

2.1 Coupled ocean–atmosphere GCM HadCM3

The AGCM, HadAM3, has a horizontal resolution of 2.5° by 3.75° and 19 vertical levels and is described in detail in Pope et al. (2000). The OGCM, HadOM3, is based on the GFDL "Cox" ocean model (Cox 1984). Several modifications have been made to the original GFDL ocean model (see Gordon et al. 2000). HadOM3 has a horizontal resolution of 1.25° by 1.25° and there are 20 depth levels. River runoff is included in the model using predefined river catchments, and the runoff enters the ocean at coastal outflow points. The sea ice model includes a simple thermodynamic budget and the ice thickness, concentration and snow depth are advected using the surface ocean current. Ice rheology is only crudely represented by preventing convergence of ice once the ice thickness reaches 4 m. The coupled OAGCM model, HadCM3, couples the AGCM once a day to the OGCM and the sea ice model, and does not require artificial flux adjustments. HadAM3 is integrated for a

day during which time the atmosphere–ocean fluxes are computed and then passed to the ocean/sea ice model. The ocean/sea ice model is then integrated and the ocean state (i.e. SST, sea ice fraction and depth, surface current) is passed back to the atmospheric model to enable the cycle to be repeated. The HadCM3 model, its simulation of present-day climate, and experiments with increased CO_2 concentrations are described in more detail elsewhere (Wood et al. 1999; Gordon et al. 2000).

2.2 Coupled slab ocean–atmosphere GCM HadSM3

The slab model HadSM3 couples the atmospheric GCM HadAM3 once a day to a thermodynamic mixed layer ocean model, and the sea-ice model described above. The ocean currents for the advection of ice in the sea-ice model are prescribed from values diagnosed in the HadCM2 coupled ocean–atmosphere model (Johns et al. 1997), since currents from the HadCM3 control experiment were not available at the time of developing HadSM3. The mixed layer ocean model simplifies the full ocean by representing it as a single layer (or slab) of water of constant thickness (50 m) that is assumed to be perfectly mixed, but with no horizontal or vertical heat transport. The model computes the SSTs without the expense of having to run a full ocean model. Since the model contains no ocean currents a corrective heat flux, or heat convergence as it is commonly called, is imposed, which is designed to produce realistic ocean heat transports and SSTs. In climate change experiments the heat convergence is normally unchanged in the anomaly simulation. This is effectively imposing the present-day ocean heat transports on the anomaly climate simulation, and therefore makes the assumption that the climate forcing's primary effect will not be to change the ocean heat transports. See Hewitt and Mitchell (1997) for more details.

2.3 Glacial boundary conditions

The glacial boundary conditions differ from those of the pre-industrial control simulations by modifying the land surface characteristics to include the extensive continental ice sheets, modified coastlines and topographic heights to account for a sea level lowering (Peltier 1994), lowering the atmospheric composition of CO_2 from the pre-industrial concentration of 280 ppmv to 200 ppmv, and a different pattern of insolation arising from a change to the Earth's orbit (Berger 1978).

HadCM3 uses the same land–sea distribution for both the atmosphere and ocean models. The glacial land–sea distribution is based on the OGCM bathymetry at a depth of 96 m (the nearest model level which is representative of the PMIP-inferred lowering of sea level at the LGM of about 105 m based on Peltier (1994). This land–sea distribution is modified to take into account the continental ice sheets, since some of the ice sheets occupy ocean grid-points, around high latitude continental shelves.

The ocean's glacial bathymetry is the same as in the control run except for the top seven levels (down to a depth of 96 m) which have the same land–sea distribution as described above. The ocean is not 105 m shallower at all points (as would have resulted from the sea level lowering), but wherever ocean is less than 96 m deep in the control run, this point becomes a land point, and at all other points there is no change to the depth of the ocean.

The total salt content of the ocean points that become land in the LGM simulation is evenly redistributed over all remaining ocean points so that the total global volume-weighted salinity is preserved, i.e. it is the same in the control and LGM experiments. The redistributed salt increases the volume-averaged global average salinity from 34.7 practical salinity units (PSU) to 34.9 PSU, since most of the new land points occur in the relatively fresh water of the high northern latitudes, in particular the disappearance of Hudson Bay under the Laurentide Ice Sheet and the Baltic, Barents and Kara seas under the Fennoscandian Ice Sheet. It should be noted however that this increase does not fully compensate for the amount of fresh water stored in the ice sheets, which may amount to 1 PSU. Weaver et al.(1998) state that increasing the salinity by 1 PSU made no difference to their LGM simulations. The 1 PSU addition, while changing the mean salinity did not affect the salinity gradients which are important for the ocean circulation.

2.4 HadCM3 spin-up

The time scale of response of parts of the deep ocean is so long (of the order of millennia) that it is not possible with current resources to use this high-resolution coupled model to bring the deep ocean to equilibrium with the glacial boundary conditions. However, it is important to run the experiments long enough to be close to the model's equilibrium climate state so that any residual adjustment that may occur does not dominate over the signal of the climate change. We have used a spin-up strategy designed to accelerate the cooling of the ocean. The spin-up of the HadCM3 coupled model involves three stages.

Firstly, the slab model is run for 45 years with the LGM boundary conditions prescribed. A climatology of glacial SSTs is constructed from the last 20 years of the slab model run, for use in the next stage of the HadCM3 spin-up. The coupled model does not represent the process of iceberg calving and so a water flux is needed to account for this (Lowe and Gregory 1998), determined from the amount of snow that accumulates on the ice sheets during the slab model LGM run.

Secondly, the coupled model HadCM3 is run for 70 years with the SSTs relaxed towards the LGM SSTs simulated by the slab model. The other forcing fluxes, namely the penetrative solar heat flux, the wind mixing energy, the wind stress, and the fresh water flux, are interactively calculated by the atmosphere model. The ocean model is initialized from rest, with temperature and salinity taken from the Levitus climatology for the present-day (Levitus and Boyer 1994; Levitus et al. 1994) but the salinity is increased uniformly as described previously to preserve the total volume integrated salinity due to the new land points. This configuration of the coupled model with SSTs relaxed to those of the slab model will henceforth be referred to as HadCM3H. The purpose of this second stage is to accelerate the cooling of the ocean to provide initial conditions for the next, and final, phase.

Finally, the coupled model is initialized from the end of the HadCM3H stage and has been run for 1000 years without any SST restoring, i.e. the model computes all of the surface fluxes as is done in the HadCM3 control simulation. By the end of the simulation the LGM experiment has reached a quasi-equilibrium (see Appendix for more details). Further details of the HadCM3 spin-up methodology are provided in Hewitt et al. (2001c).

3 Simulated thermohaline circulation

The HadCM3 control simulation produces a meridional overturning cell in the upper North Atlantic (the NATHC) with deep water formation at high northern latitudes (Fig. 1a). The maximum strength of the overturning in the NATHC is about 19 Sv (10^6 m^3 s^{-1}), at a depth of about 700 m around 45°N. The overturning cell penetrates down to a depth of about 2500 m producing North Atlantic deepwater (NADW), with denser Antarctic Bottom Water (AABW) intruding below the NADW throughout the rest of the depth of the ocean basin. The overturning transport cannot be calculated from observations directly. Broecker (1991) inferred that the maximum overturning is about 20 Sv based on observations of oceanic tracers, and Hall and Bryden (1982) used oceanographic measurements of currents and temperature to conclude that about 18 Sv of water

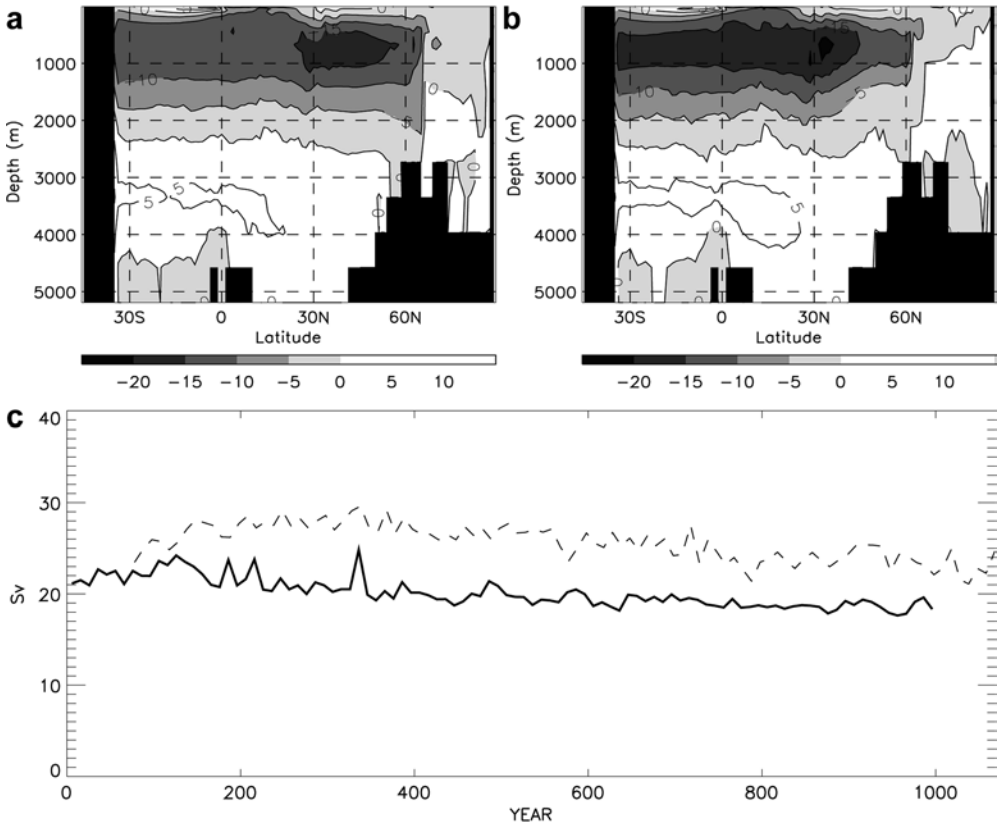

Fig. 1 North Atlantic meridional overturning stream function, in Sv (10^6 m^3 s^{-1}); **a** depth-latitude cross-section for HadCM3 control; **b** depth-latitude cross-section for final century of HadCM3 LGM experiment; **c** time series of the magnitude of the maximum decadal mean overturning in the HadCM3 control (*solid line*) and LGM simulation (*dashed line*)

is transported northward at 24°N. The model simulation is therefore consistent with oceanographic estimates. However, the vertical structure of the NATHC appears to differ to that inferred from observations, with observations suggesting that the NADW should penetrate to perhaps 4000 m (Roemmich and Wunsch 1985) with AABW deeper down.

The North Atlantic meridional overturning cell weakens at high latitudes in the LGM coupled model simulation, and shifts further south with an increase in the strength of the maximum overturning (Fig. 1b, c). The high latitude weakening and southward shift are largely due to an expansion of Arctic sea ice in the Nordic Seas at the LGM (Fig. 2), and are consistent with other modelling studies and palaeoceanographic data (Sarnthein et al. 1994; Ganopolski et al. 1998; Duplessy et al. 1988). The large glacial expansion of sea ice in the Nordic Seas provides an insulating blanket on the ocean, reducing the large loss of heat from the ocean to the cold overlying atmosphere, which weakens convection in the Nordic Seas (Fig. 3). The convection sites shift southward beyond the sea ice margin, south of Iceland, and deep convection occurs across much of the North Atlantic and into the Labrador Sea (Fig. 3b) where the sea-ice changes are not as large as in the Nordic Seas.

The stronger overturning at the LGM is due to an increase in the negative buoyancy flux of the surface waters across the mid-latitude North Atlantic and in the Labrador Sea (Fig. 4). The extremely cold conditions over the Laurentide Ice Sheet produce a strong flow of cold, dry air down the Labrador Sea into the North Atlantic where the flow joins up with the intensified mid-latitude circulation (Fig. 5b). Winds from Greenland and the expanded Arctic sea ice provide additional cold, dry polar air to the North Atlantic atmospheric circulation. An intensified Azores High then directs these winds southwards along the coast of Western Europe (Fig. 5b). The strong cold winds increase the air-sea temperature gradient in the Labrador Sea, across the mid-latitude North Atlantic, and along the coast of Western Europe. Consequently, the loss of heat from the ocean to the atmosphere is increased which produces a large negative buoyancy flux (Fig. 6a), and strong convective mixing (Fig. 3b) leading to dense deep water formation in these regions, which then drives the strong overturning cell seen in Fig. 1b, c. Changes to the hydrologic cycle also contribute to the buoyancy flux changes, but are relatively small compared to the heat flux changes described (Fig. 6). The changes to the hydrologic cycle are an increase in precipitation across much of the northern North Atlantic, but a larger increase in evaporation of water from the ocean surface due to the strong, dry winds, leading to an increase in the salinity of the surface waters.

The driving force for the intensified North Atlantic overturning cell is therefore the strong, cold low-level atmospheric winds, and a positive feedback maintains this strong overturning. The stronger overturning produces an increase in the amount of heat transported northwards in the Atlantic Ocean (Fig. 7a), causing a

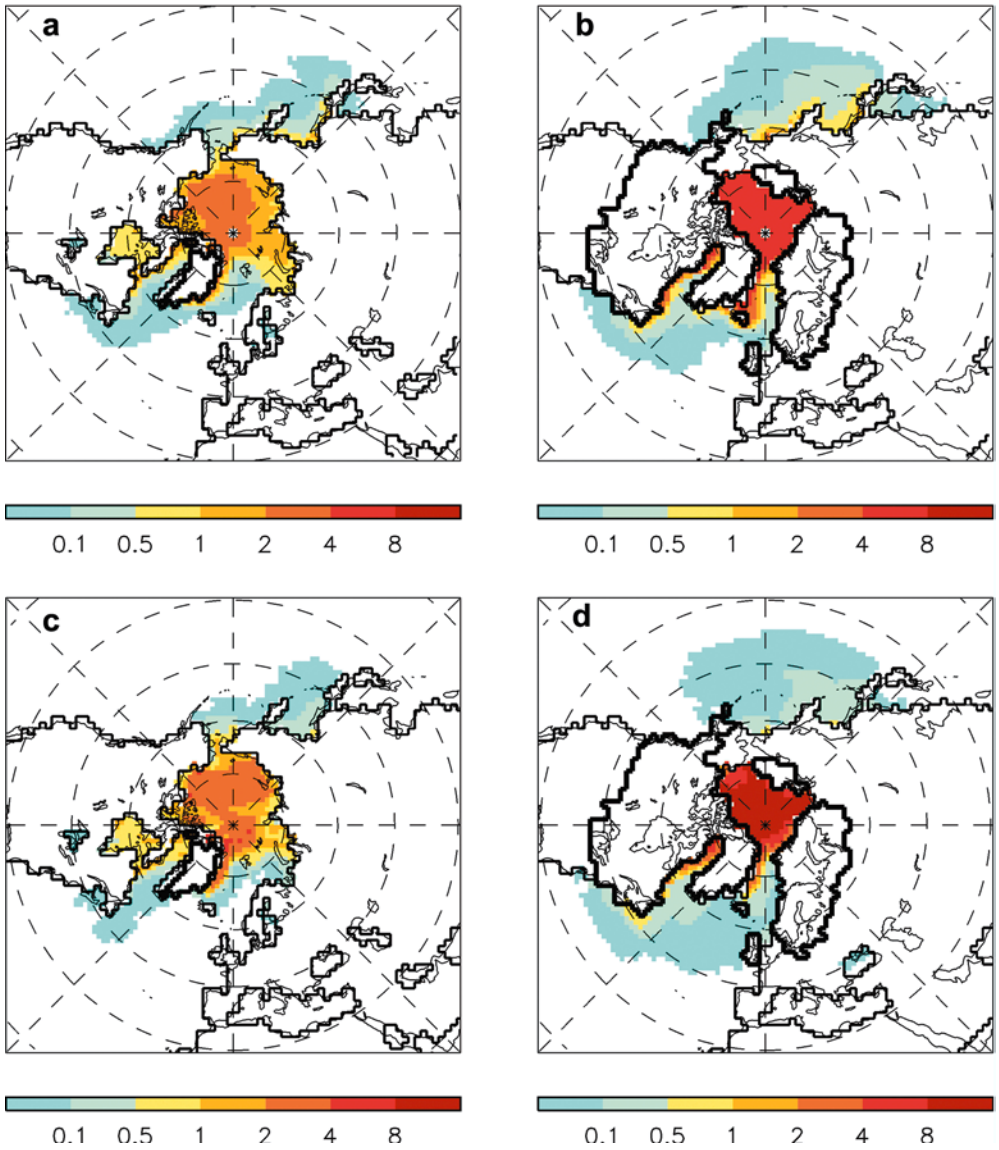

Fig. 2 Annual mean Northern Hemisphere sea ice thickness, in m. **a** HadCM3 control. **b** HadCM3 LGM. **c** HadSM3 control. **d** HadSM3 LGM

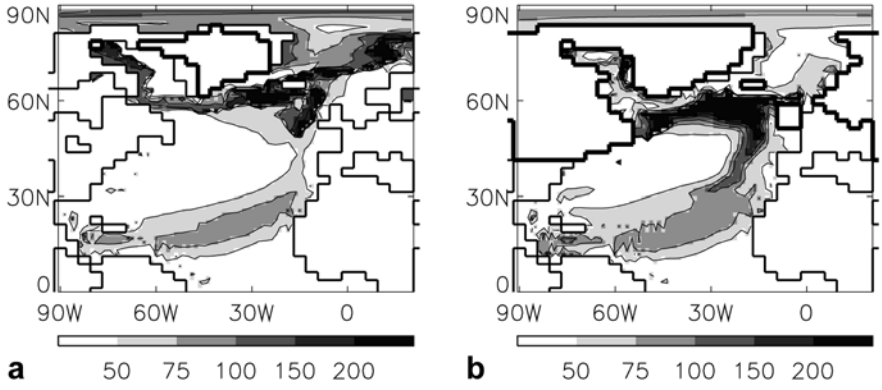

Fig. 3a, b Depth of the oceanic mixed layer in HadCM3, in m. This is the depth to which the in-situ temperature differs from the surface temperature by no more than 0.5 °C (one of the methods used by Levitus 1982). **a** Control simulation, **b** LGM simulation

warming of mid-latitude surface waters. The combination of relatively warm SSTs and cold, dry surface winds, increases both the heat loss and evaporation from the ocean to the atmosphere, which increases the salinity and hence density of the surface waters thus making the buoyancy flux more negative and strengthening the NATHC. However, there is also a negative feedback from the advection of heat. The increased northward advection of warm water from the south acts to reduce the surface density in the North Atlantic, and this acts to

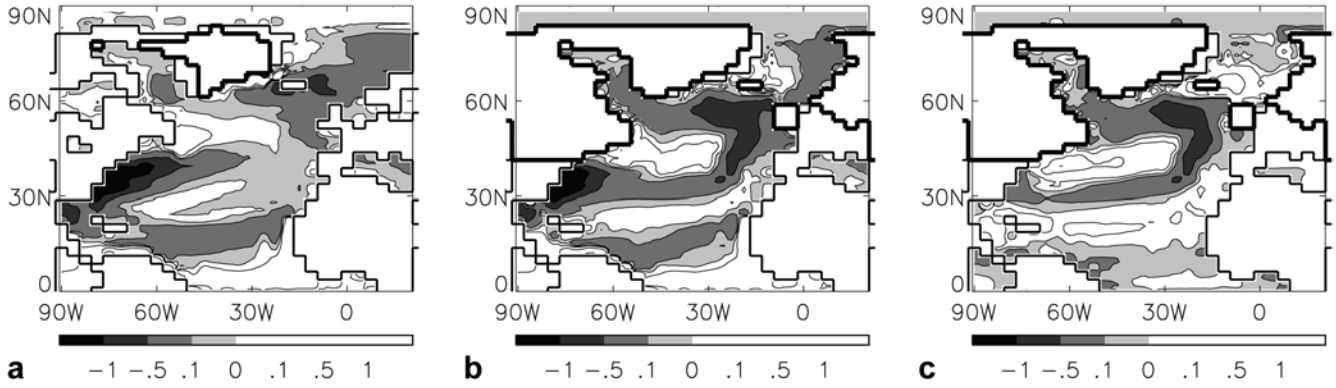

Fig. 4a–c HadCM3 Surface buoyancy flux, in 10^{-4} kg m^{-1} s^{-3}. **a** Control. **b** LGM. **c** LGM-control

weaken the NATHC and therefore the northward advection of warm water. This negative feedback is made weak by the presence of the atmosphere which absorbs some of the extra heat, and subsequently advects the heat away and radiates some to space. In addition, at low temperatures, such as those that occur in the regions of deepwater formation, the equation of state for density near the ocean surface is more sensitive to salinity than temperature, which also reduces the effect of the negative feedback. The effect of the negative feedback is to act as a brake and prevent the positive feedback from making the NATHC stronger and stronger, as discussed qualitatively already.

The strength of the thermohaline circulation at the LGM is difficult to verify based on interpretation of proxy records. Some palaeoceanographic data supports the possibility of a stronger thermohaline circulation at the LGM (Veum et al. 1992; McCave et al. 1995) as simulated by the model (and as simulated by the coupled model study of Kitoh et al. (2001)), but other research points to a weaker circulation (Lynch Stieglitz et al. 1999). However, palaeoceanographic data and several other modelling studies seem to be more consistent in suggesting that the present-day North Atlantic deep water (NADW) was replaced with glacial North Atlantic intermediate water (GNAIW), which the HadCM3 simulation does not reproduce. Hewitt et al. (2001a) speculate that this may be a result of the model's present-day NADW (Fig. 1a) not penetrating as deeply as observations suggest, and instead occurs at depths similar to those proposed for GNAIW. Also, the gradual cooling in the deep Atlantic Ocean may eventually prevent the North Atlantic surface waters from sinking as deeply.

To summarize, whether or not a strong cell actually existed at any stage during the glacial period is still under debate, based on modelling studies and the interpretation of proxy records. However, as an aside, there is evidence for a greatly reduced North Atlantic overturning circulation (e.g. Zahn et al. 1997; McCabe and Clark 1998; Rühlemann et al. 1999) during glacial times outside of the glacial maximum period. The overturning circulation was reduced during events associated with massive discharges of fresh water into the North Atlantic, such as the Heinrich events. The fresh water is thought to have originated from sudden discharges from the continental ice sheets (see Schmittner et al. 2002, for a study of coupled ice sheet/THC interactions using an Earth System Climate model), but such a study is not part of the GCM experiments here.

4 Effect of ocean dynamics

The glacial boundary conditions produce a global average annual mean surface cooling of –3.8 K in the coupled model compared to the present-day control simulation, similar to the –3.9 K simulated by the slab model. The geographical pattern of the temperature response in the coupled model is also similar to the slab model simulation[1] (Figures 8a and b) – the largest temperature changes are over land and in the Northern Hemisphere, with maxima over the ice sheets. However there are some notable differences between the coupled model's response and the slab model's response which illustrate the effect of ocean dynamics on the model response, and raise the following questions.

1. Why does the slab model produce a localised Ice Age warming off the coast of Peru and Chile while the coupled model simulates a cooling?
2. Why does the coupled model produce a cold/warm dipole pattern of temperature response in the North Atlantic between about 30 and 45°N, while the slab model produces a general cooling pattern across the North Atlantic?
3. Why does the slab model produce a larger cooling than the coupled model over much of the mid-latitude

[1] The statistical significance of the temperature difference between the LGM and control simulations has been assessed. A two-tailed Student's *t*-test has been used based on consecutive 5-year means from the control and anomaly simulations to determine whether the multi-annual mean changes can be described as significant at and above the 95% confidence level compared to the model's internal interannual variability. The significance levels should only be regarded as approximate since meteorological variables are correlated in time and space, and the *t*-test will tend to overestimate the significance (Zwiers and von Storch, 1995).

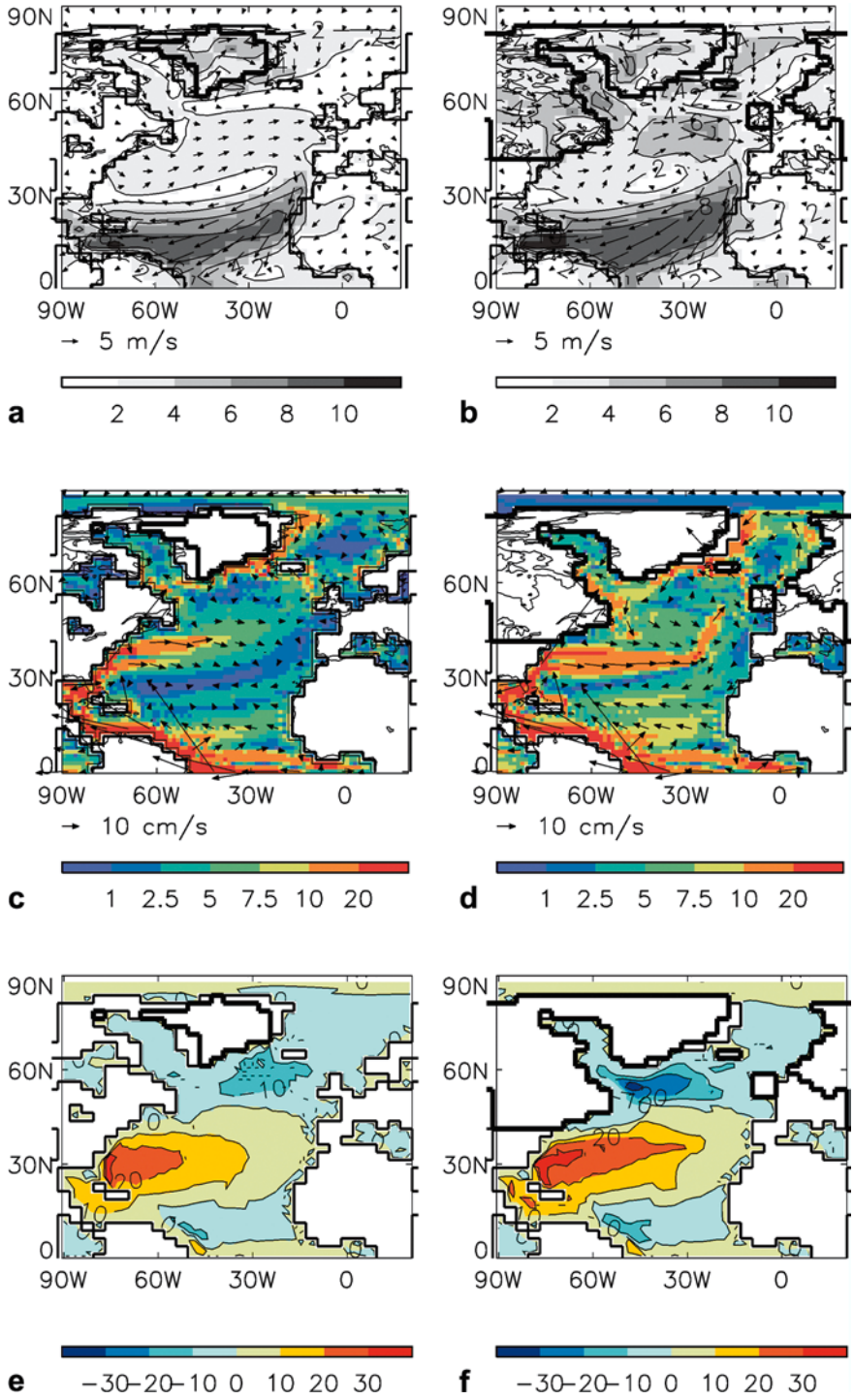

Fig. 5a–f Annual mean conditions in HadCM3 control and LGM simulations around the North Atlantic region. **a** Control surface wind, in m s^{-1}. **b** LGM surface wind, in m s^{-1}. **c** Control surface current, in cm s^{-1}. **d** LGM surface current, in cm s^{-1}. **e** Control barotropic stream function, in Sv. *Positive values* indicate clockwise flow. **f** LGM barotropic stream function, in Sv

North Atlantic (between typically 45 and 60°N)? In addition, how does the coupled model simulate *warmer* SSTs over some of the mid-latitude North Atlantic?

4. Why is the coupled model response generally larger than the slab model response in the Nordic Seas?
5. On a hemispheric scale, why does the slab model cool more than the coupled model in the Northern Hemisphere, and vice versa in the Southern Hemisphere?
6. Why does the coupled model produce a larger cooling in the equatorial East Pacific than the slab model?

The first question is relatively simple to address, and is a response peculiar, and probably spurious, to the slab model. The slab model generally produces a cooling everywhere, a thermodynamic response to the imposed glacial boundary conditions. However, the slab model produces a warming off the coast of Peru and Chile,

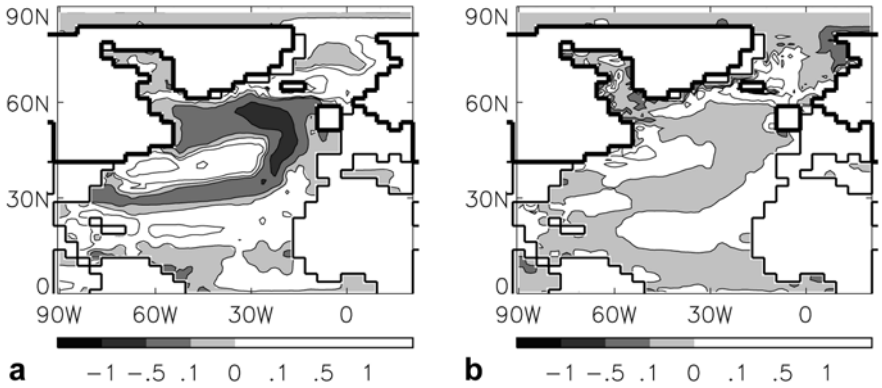

Fig. 6a, b Components of HadCM3 LGM-control surface buoyancy flux (i.e. Fig. 4c), in 10^{-4} kg m^{-1} s^{-3}. **a** Heat flux component. **b** Freshwater flux component

which the coupled model does not reproduce. The warming is due to a positive feedback associated with cloud changes. In the coupled model this feedback is weakened or removed since any surface temperature anomalies can be dissipated through oceanic advection.

The answers to the next four questions are given below, and are associated with the LGM response of the NATHC in HadCM3, which has been described in the previous section.

4.1 Sub-tropical North Atlantic response

The cold/warm dipole pattern of surface temperature response in the North Atlantic between about 30 and 45°N in the coupled model arises because, away from the coast of North America, the Gulf Stream shifts south in the LGM simulation and is replaced by a strong Labrador Current off the coast of Newfoundland (Fig. 5c, d). This produces relatively cool surface waters where the warm present-day Gulf Stream is replaced by the cold glacial Labrador Current. The glacial Gulf Stream has a more zonal flow and is stronger (which is at odds with the findings of Lynch-Stieglitz et al. 1999), giving warm surface waters where the glacial Gulf Stream flows eastwards (consistent with the findings of Crowley 1981).

The southward shift of the Gulf Stream is not associated with a shift of the line of zero Ekman pumping, as has been suggested (Keffer et al. 1988), but is in part due to the strong Labrador Current in the LGM simulation. In the LGM simulation the stronger surface winds blowing off the Laurentide and Greenland ice sheets (Fig. 5b) and the intensified Icelandic Low (Hewitt et al. 2001a) produce a strong Labrador Current and a strong subpolar gyre in the far North Atlantic (Fig. 5d, f). The larger cooling in the Labrador Sea increases the density of the water there, and this denser water flows out into the North Atlantic in the subpolar gyre forcing the Gulf Stream to flow further south, being replaced by the strong Labrador Current. The increased overturning circulation contributes to the stronger Gulf Stream current in the tropics and the more zonal flow is due to the strong subpolar gyre.

The slab model produces a similar change to the atmospheric circulation as the coupled model with strong cold surface winds, but it cannot produce changes to the Gulf Stream, or the subpolar gyre, so the dipole surface temperature pattern is not produced.

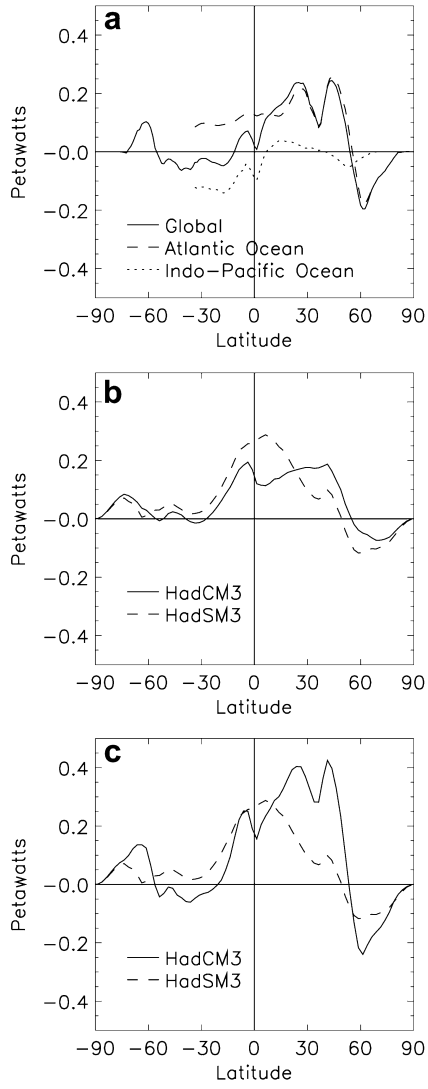

Fig. 7a–c Zonal mean northward heat transport, LGM-control. **a** Ocean heat transports in HadCM3 north of 35°S. **b** Global atmosphere heat transports in HadSM3 and HadCM3. **c** Total global heat transports in HadSM3 (i.e. atmosphere) and HadCM3 (i.e. ocean + atmosphere)

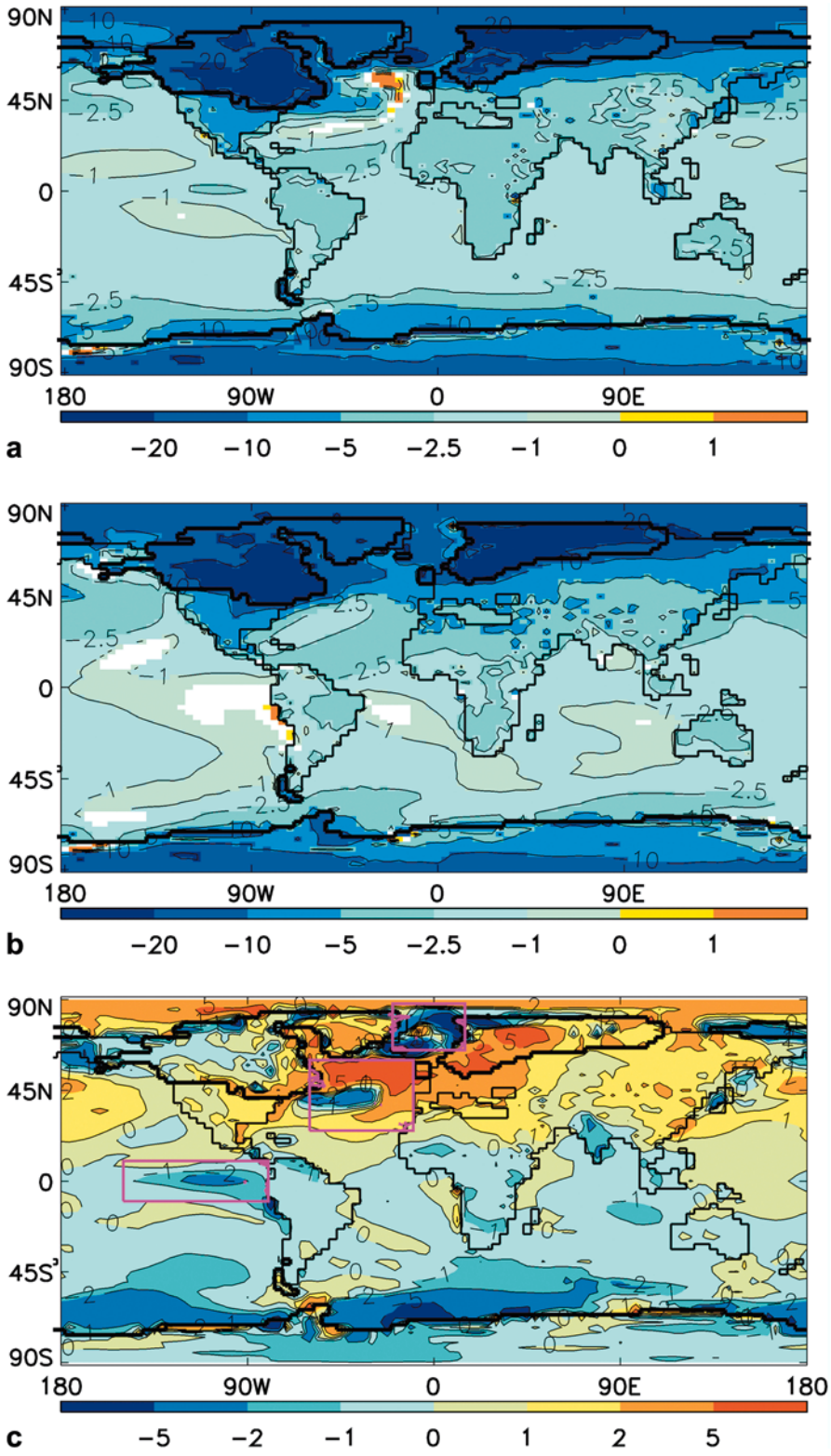

Fig. 8a–c Change in annual mean surface temperature, in °C, LGM-control. **a** HadCM3, **b** HadSM3. **c** HadCM3–HadSM3 (i.e., panel **a** – panel **b**). The *purple boxes* highlight the regions discussed in Sect. 4. The *white areas* in **a** and **b** are regions where the changes are not statistically significant based on a Student's *t*-test

4.2 Mid-latitude North Atlantic response

The slab model generally produces a larger cooling than the coupled model over much of the mid-latitude North Atlantic (between about 45 and 60°N, Fig. 8c), and there are some regions where the coupled model simulates warmer conditions at the LGM than in the control. The coupled model LGM response is relatively warm compared to the slab model response because of the increased northward heat transport from the tropics due to the stronger NATHC (Fig. 7a), and the stronger and warmer currents in the North Atlantic, in particular the

Gulf Stream and the eastern section of the subpolar gyre which bring extra heat northwards along the coast of Western Europe, westwards across the mid-latitude North Atlantic and up into the Labrador Sea and Baffin Bay.

The slab model does not simulate changes in heat transports by design. The heat transports are represented by the model's prescribed heat convergence, and this does not change between the control and LGM simulations. The slab model simply cools in response to the large atmospheric cooling, predominantly due to the presence of the continental ice sheets.

4.3 Response in the Nordic Seas

The coupled model produces a different pattern of cooling in the Nordic Seas to the slab model. Generally, the coupled model produces a larger cooling around Iceland, and in the Greenland Sea and the northeast of the Norwegian Sea, and less of a cooling than the slab model in the interior of the Nordic Seas. The relative amount of sea ice in the control simulations of different models can affect the relative magnitude of the response locally (Hewitt et al. 2001b). In regions where a model has relatively little sea ice in the control simulation, the model tends to produce a relatively large glacial cooling. While this may be a contributing factor to some of the differences between the coupled model and the slab model, it does not explain the general pattern described. For example, the coupled model has more sea ice than the slab model in the Norwegian Sea (Fig. 2a, c) but the coupled model cools more than the slab model in parts of the Norwegian Sea. The larger cooling in the coupled model around Iceland, in the Greenland Sea and the northeast of the Norwegian Sea is mainly due to the changes to the NATHC and the subsequent reduced inflow of relatively warm surface waters from the North Atlantic to the Greenland Sea. More heat is transported into the eastern Nordic Seas close to the British Isles in the coupled model LGM simulation than in the control, and this warm water limits the cooling in these regions.

Since the slab model does not allow the heat transports, or ocean currents, to change it cannot produce the reduced inflow from the North Atlantic to the western Nordic Seas nor the increased heat transport to parts of the eastern Nordic Seas. The coupled model cooling is further enhanced by the positive sea ice albedo feedback, as the region cools, more sea ice forms producing a larger cooling. The coupled model produces thicker glacial sea ice than the slab model in western and northern regions of the Nordic Seas (Fig. 2b, d) and slightly thinner sea ice in the eastern Nordic Seas.

The sea ice simulated by both the coupled model and the slab model is less extensive than that suggested by the CLIMAP project (CLIMAP Project Members 1981), but is in good agreement with more recent reconstructions (de Vernal et al. 2000). In particular, the model simulated seasonally ice-free waters in the Nordic Seas are supported by data (Hebbeln and Wefer 1997; Lassen et al. 1999). Hebbeln and Wefer (1997) state that a poleward oceanic heat transport kept the Norwegian Sea seasonally ice free as far north as 80°N, consistent with the coupled model and slab model results (Fig. 2). Sediments containing ice-rafted detritus imply that there was a northward current in the Norwegian Sea (Hebbeln et al. 1994), and this would suggest an anticlockwise circulation in the Nordic Seas (Hebbeln and Wefer 1997; Lassen et al. 1999), a situation supported by the coupled model results (Fig. 5d).

4.4 Northern Hemisphere versus Southern Hemisphere response

In the Northern Hemisphere the increased glacial northward ocean heat transport in the coupled model, as reported, offsets the cooling caused by the LGM boundary conditions. The slab model, with its fixed ocean heat transports, can only simulate changes to the amount of heat transported polewards by the atmosphere.

The large glacial cooling over the Northern Hemisphere continental ice sheets (Fig. 8) increases the meridional atmospheric temperature gradient between the mid-latitudes and the equatorial regions. The increased temperature gradient at the LGM increases the amount of heat transported northwards by the atmosphere up to the mid-latitudes in both the coupled model and slab model simulations (Fig. 7b).

In the coupled model the additional increase in the amount of heat transported northward by the ocean in the LGM simulation (Fig. 7a, c) means that on a hemispheric scale the Northern Hemisphere does not cool as much in the coupled model LGM experiment as it does in the slab model LGM experiment. The warming effect of the increased northward heat transport in the coupled model is largest in the Atlantic region because this is where the ocean heat transports increase (Fig. 7). By comparison, the amount of heat transported northward in the Northern Hemisphere by the Indo-Pacific Ocean hardly changes. The large-scale zonal circulations in the atmosphere enable it to transport additional heat relatively rapidly to other regions, and so the coupled model also cools less than the slab model over much of the northern continents, the Arctic Ocean and the Pacific Ocean.

As a consequence of more heat being transported northwards, from both the low latitudes and the South Atlantic Ocean, the Southern Hemisphere subsequently cools more in the coupled model than in the slab model. The greater cooling in the coupled model in the Southern Hemisphere leads to further cooling in the Southern Ocean due to the positive sea ice albedo feedback. Consequently, the coupled model produces much larger increases to the thickness of the sea ice at the LGM than the slab model around Antarctica (Fig. 9b, d).

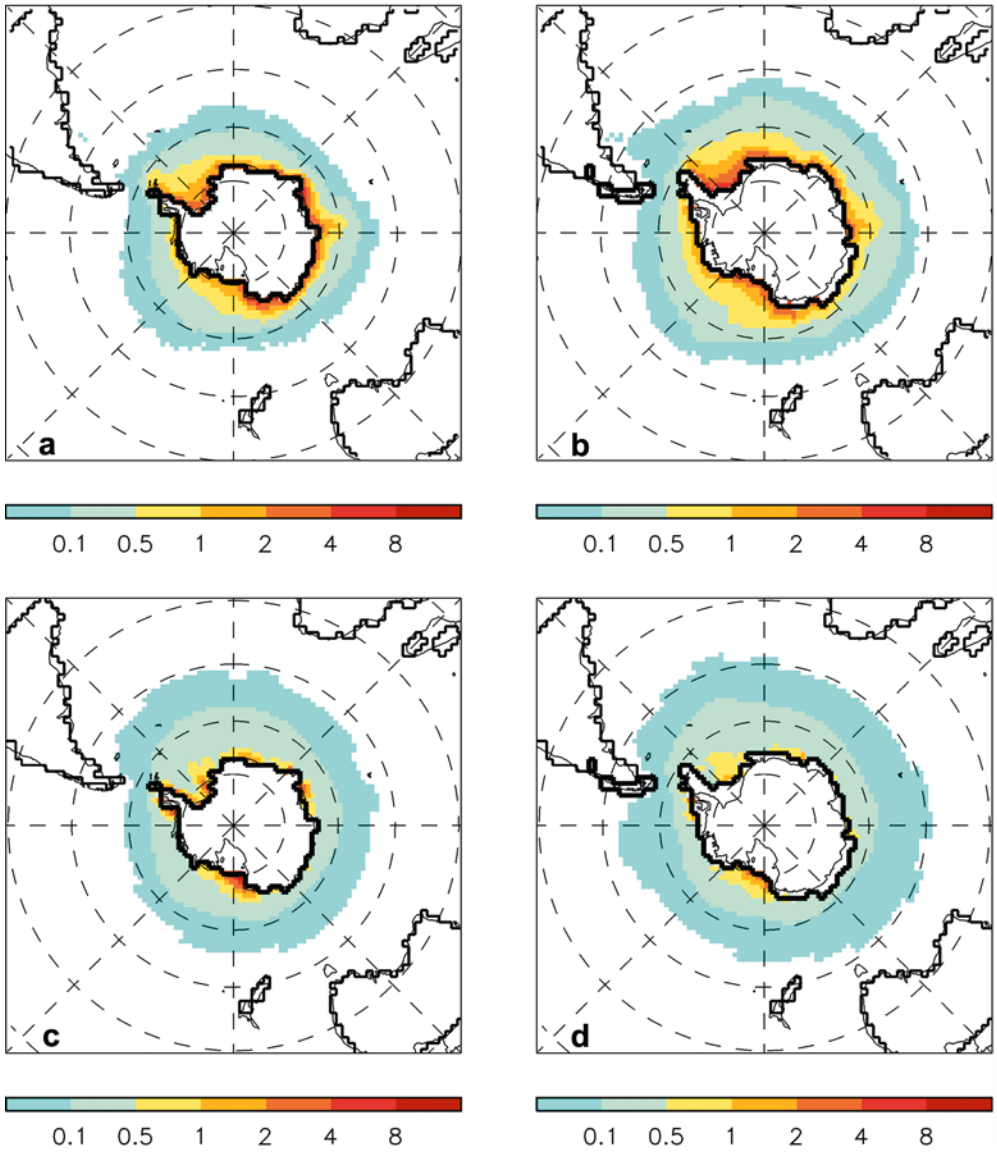

Fig. 9a–d Annual mean Southern Hemisphere sea ice thickness, in m. **a** HadCM3 control. **b** HadCM3 LGM. **c** HadSM3 control. **d** HadSM3 LGM

4.5 Cooling in equatorial East Pacific

The control simulation has a monsoon-type trough over southwest North America (Fig. 10). In the LGM simulation the extremely cold conditions over the continental ice sheets produces large-scale subsidence and anticyclonic conditions (as reported by COHMAP members 1988), particularly over the Laurentide Ice Sheet, with a ridge of high pressure extending southwards towards Central America. This ridge weakens the trough over southwest North America producing a more easterly flow at the LGM in the tropical east Pacific around Central America. The stronger easterly component to the winds produces stronger divergence along the equator and greater upwelling of cold water. The coupled model produces a larger cooling in the equatorial east Pacific than the slab model (Fig. 8c). Stronger equatorial upwelling has been supported by palaeoclimatic data (Pisias and Rea 1988; Sarnthein et al. 1988), and the hypothesis to explain the increased upwelling has been stronger trade winds in response to the increased thermal gradient. However, as Lautenschlager and Herterich (1990) point out, reconstructions such as CLIMAP produce a decrease of the meridional SST gradient in the tropical Pacific at the LGM, which decreases the strength of the Pacific Hadley circulation and the trade winds in their AGCM simulations using CLIMAP glacial SSTs. Lautenschlager and Herterich (1990) realised that stronger near-coastal easterly winds could be the mechanism producing stronger upwelling. In the HadCM3 coupled model LGM experiment it is indeed a stronger easterly component of the trade wind flow in the East Pacific, and not necessarily just stronger trade winds, that increases the upwelling.

The slab model produces a similar change to the atmospheric circulation as the coupled model with a ridge extending over southwest North America that weakens

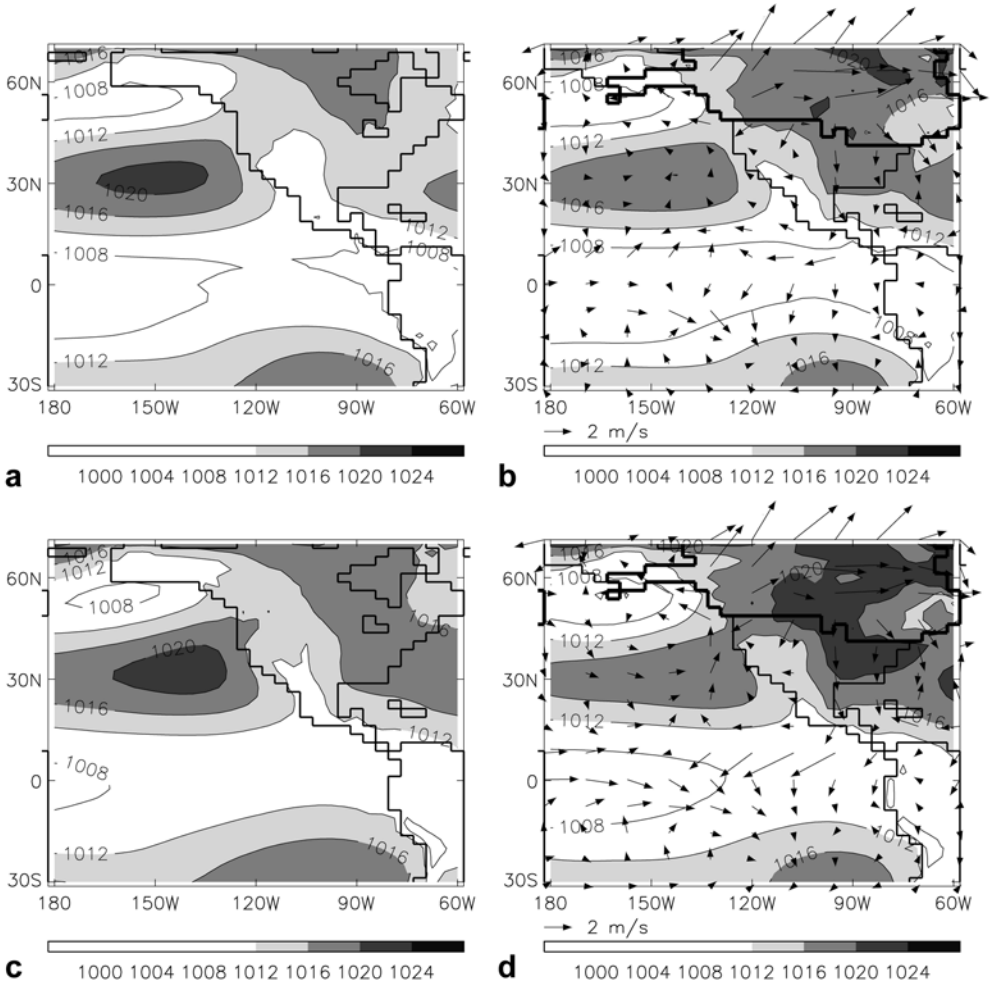

Fig. 10a–d Annual mean pressure at mean sea level, in mb. **a** HadCM3 control. **b** HadCM3 LGM. **c** HadSM3 control. **d** HadSM3 LGM. The *arrows* in **b** and **d** depict the vector difference between LGM and control low-level winds in HadCM3 and HadSM3 respectively

the monsoon-type trough producing stronger easterly flow in the equatorial East Pacific. However, since the slab model does not simulate ocean dynamics, it cannot simulate changes to upwelling and does not produce the enhanced cooling in the East Pacific. Therefore, even though the atmospheric response of the slab model is similar to that of the coupled model in this region, the effect of ocean dynamics is to produce a different local response in the coupled model.

5 Summary and conclusions

We have carried out a 1000-year long experiment to simulate the climate at the LGM using the coupled ocean–atmosphere GCM HadCM3. A brief evaluation of the coupled model LGM response after 700 years has been made by Hewitt et al. (2001a). Their study showed that changes in the ocean circulation, namely the intensified thermohaline circulation and increased northward ocean heat transport in the Atlantic equatorward of 55°N, can have an important influence on the surface response. We have carried out additional experiments using an atmospheric model coupled to a thermodynamic-only 50 m deep ocean model to investigate the effect of ocean dynamics and the deep ocean on the response of the GCM to LGM boundary conditions.

The coupled model is initialized from present-day conditions. The LGM boundary conditions are applied instantaneously, and cause the climate system to cool. The GFDL low-resolution coupled model suggests that parts of the deep ocean respond on time scales of longer than 1000 years, so we have devised a strategy to accelerate the cooling of the HadCM3 coupled model. In the first stage of the experiment the SSTs are relaxed towards the slab model LGM SSTs for 70 years to accelerate the cooling towards a glacial state.

In the second stage of the experiment the relaxation is discontinued and the surface and mixed layer respond fairly quickly, i.e. within a few decades, to the glacial boundary conditions, and have reached a quasi-equilibrium. The deep ocean is still slowly adjusting, even after 1000 years, but the trends in the deep ocean are very small by then. The GFDL low-resolution coupled model equilibrium experiments suggest that the SST relaxation used in the first stage of the LGM experiment produced an enhanced cooling that would take several hundred years longer without this initial stage, representing a considerable saving in computer time,

although still a relatively small fraction of the total time required to reach equilibrium.

The LGM forcing produces a more vigorous NATHC (than in the control), and feedbacks maintain this stronger overturning cell throughout the simulation. The magnitude of the strength of the NATHC at the LGM is difficult to verify based on interpretation of proxy records, with some studies supporting the possibility of a stronger thermohaline circulation at the LGM, and others suggesting a weaker circulation. However, there does seem to be more consistent evidence for a shoaling of deep water production, which the HadCM3 simulation does not reproduce, and a greatly reduced NATHC during glacial times outside of the glacial maximum period, which the HadCM3 study does not attempt to reproduce.

The LGM surface temperature response of the coupled model is generally similar to that of the slab model over large spatial scales. Regionally, however, there are some notable differences between the responses of the coupled model and the slab model. The influence of ocean dynamics in the coupled model response has been investigated since the slab model does not explicitly represent ocean dynamics and assumes no changes in the ocean heat transports.

The coupled model produces enhanced equatorial upwelling in the East Pacific, consistent with some palaeoclimatic reconstructions, due to an increase to the eastward component of the northeast trade winds, leading to a larger cooling in that region than the slab model. As an aside, the localized cooling in the equatorial East Pacific in the HadCM3 LGM simulation may have implications for El Nino/La Nina-type variability, and this could be an interesting topic for future work.

The Gulf Stream in the interior of the North Atlantic shifts south in the coupled model simulation, producing relatively cold conditions where the LGM Labrador Current replaces the present-day Gulf Stream. The stronger LGM overturning cell in the tropics and the stronger subpolar gyre produce a stronger glacial Gulf Stream which transports more heat and produces the relatively warm conditions where the glacial Gulf Stream flows. Again, the slab model cannot simulate changes to the Gulf Stream or the Labrador Current.

The increased northward heat transport in the Atlantic Ocean in the coupled model produces a smaller cooling than the slab model over the mid-latitude North Atlantic, and even produces some regions of warming. The slab model cannot simulate changes to the oceanic heat transports, and cools everywhere in response to the large atmospheric cooling.

The high-latitude weakening and southward shift of the NATHC results in a reduced inflow of warm surface waters to the Greenland Sea. The slab model cannot produce such changes to the heat transports and so the coupled model produces a larger cooling across much of the Nordic Seas than the slab model. The increased northward heat transport in the Northern Hemisphere in the coupled model reduces the cooling in the coupled model, compared to the slab model. However, the increased northward heat transport in the coupled model in the Northern Hemisphere occurs partly at the expense of the Southern Hemisphere which cools more in the coupled model than in the slab model.

On a final note, it will be interesting to compare the results presented here with other OAGCM studies to assess which features of the LGM simulation may be model-dependent. Glacial reconstructions of vegetation, ice sheets, and atmospheric aerosols are somewhat controversial and so other modelling studies may be useful to investigate the effect of altering such boundary conditions.

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Appendix A

A.1 Results from the spin-up

The transient adjustment of the model to the glacial forcing does not necessarily have any physical implications for how climate may have changed in reality since the boundary conditions are implemented instantaneously in these experiments, and so only a limited description of the results is included. In the real world such large changes involving greenhouse gases occur over many decades, and changes involving ice sheets and orbital forcing occur over many millennia.

During the HadCM3H stage the relaxation towards the slab model's SSTs causes the model to cool. The HadCM3H stage was deliberately not run to equilibrium because the coupled model was not expected to produce the same equilibrium SSTs as the slab model, mainly because the slab model neglects both changes in the ocean dynamics and changes in the response of the ocean below the mixed layer.

The surface and mixed layer of the ocean are partly isolated from the deep ocean by the presence of the permanent thermocline. As a result of this isolation the surface and mixed layer respond fairly quickly, i.e. within a few decades, to the glacial boundary conditions, and have reached a quasi-equilibrium by the end of the simulation. At abyssal depths there is a residual small rate of cooling (0.1 °C/century) of a similar order of magnitude to that in the control (0.05 °C/century).

The atmosphere, with its relatively low heat capacity, responds quickly to changes in SST. However, the changes deeper in the ocean influence the SSTs over long time scales through advection and mixing. For example, the SSTs in the tropical East Pacific cool by a degree or two over the first four centuries of the HadCM3 LGM simulation (Fig. 11), after which time the cooling becomes negligible. Therefore, the response of the overlying atmosphere, for example the surface winds, may also adjust over the first few centuries of the simulation. One must therefore be cautious in performing and interpreting the atmospheric response in experiments of only one or two centuries if there are large drifts in the deep ocean. We therefore describe our simulation as having reached a "quasi-equilibrium" because the deep ocean is still slowly adjusting, even after 1000 years. It is possible that on longer time scales the atmosphere and mixed layer of the ocean may not have reached their final equilibrium state, but it should be noted that the trends in the deep ocean are very small after 1000 years. The GFDL low-resolution coupled model equilibrium experiments suggest that the

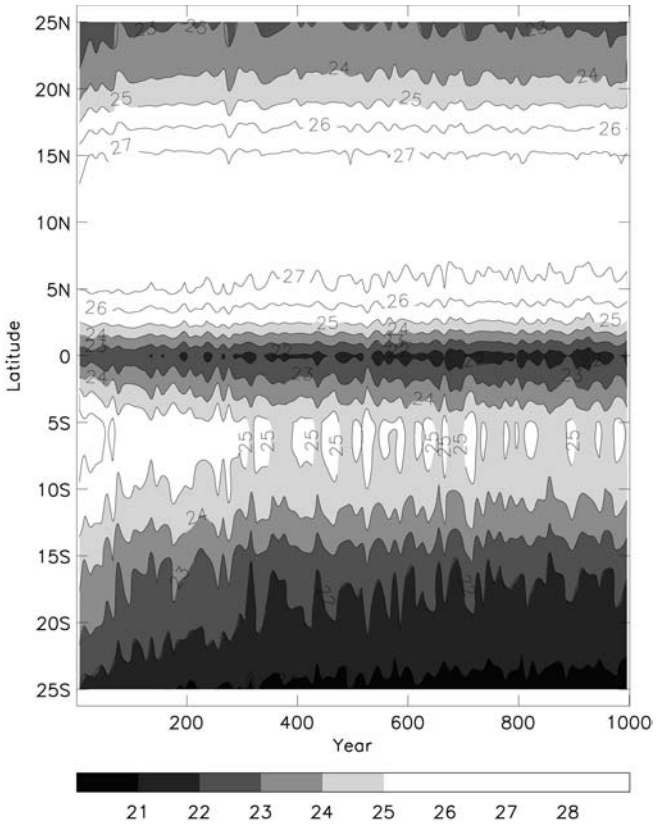

Fig. 11 Time versus latitude plot of SST from the HadCM3 LGM run averaged between 90°W and 120°W for the latitude belt 25°S to 25°N. Contours every °C

SST relaxation used in the first stage of the LGM experiment produced an enhanced cooling that would take several hundred years longer without this initial stage, representing a considerable saving in computer time, although still a relatively small fraction of the total time required to reach equilibrium (see A.2).

A.2 Estimation of the time scales of response of a cooling of the coupled ocean–atmosphere system

Since we do not have a global three-dimensional data set of glacial conditions from which to initialize the model, we use the present-day climate state. It is not possible to run the high-resolution model HadCM3 for the thousands of years needed to bring the ocean–atmosphere system to full equilibrium. However, we have made use of long simulations performed using a low-resolution coupled OAGCM developed at GFDL to estimate the time scales of response of the coupled system.

GFDL low-resolution coupled model and experimental design

The GFDL coupled model is only briefly described here. For more details see Manabe et al. (1991) and the references therein. The AGCM represents the predicted variables using spherical harmonics with the associated grid point values at a horizontal resolution of 4.5° by 7.5°, and 9 vertical levels (Gordon and Stern 1982). Insolation varies seasonally, but not diurnally. The OGCM has a horizontal resolution of 4.5° by 3.75°, and 12 depth levels. The sea-ice model includes a simple thermodynamic budget and the ice moves with the ocean currents. The atmospheric, ocean and sea-ice components exchange fluxes of heat, water and momentum

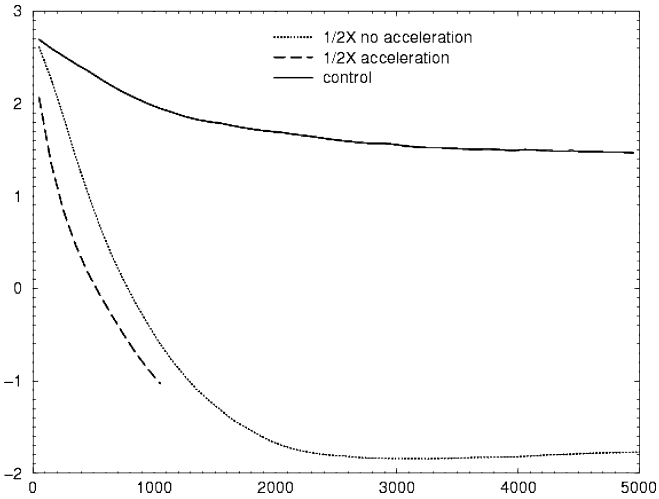

Fig. 12 Time series of global average ocean temperature at mid-depth (2228 m) from the GFDL runs

once per day. To prevent large climate drifts, the fluxes of heat and water are adjusted using values which vary seasonally, but not on longer time scales. To prevent the unrealistically large growth of sea-ice thickness due to problems with negative heat flux adjustments as noted earlier, the sea-ice thickness is limited to 10 m. Whenever the thickness exceeds this value, the thickness is reset to 10 m and the corresponding flux of heat is put into the surface heat budget.

The model is initialized from a long control integration. The CO_2 concentration is reduced at a rate of 1% per year until the concentration is one half its original value (referred to as "1/2X no acceleration" in Fig. 12), and then the CO_2 concentration is held fixed at this value for more than 4000 model years, when the model climate reaches a statistical equilibrium with the changed radiative forcing.

A further experiment has been carried out employing the experimental design strategy used in the HadCM3 LGM spin-up. The GFDL model is forced with atmospheric CO_2 concentrations halved. The model is run for 70 years with the SSTs restored to those determined from a slab model simulation with CO_2 concentrations halved. The model is then run for a further 1000 years without any restoring. This run is referred to as "1/2X acceleration" in Fig. 12.

The global average radiative forcing on the climate system of halving CO_2 concentrations is similar to, although smaller than, the radiative forcing due to imposing LGM boundary conditions (Hewitt and Mitchell 1997). However, the regional pattern of radiative forcing is very different. The reduction of CO_2 produces a fairly homogeneous horizontal spatial pattern, while the presence of large continental ice sheets at the LGM produces a large localized forcing, concentrated over the mid-and high latitudes of the Northern Hemisphere (Hewitt and Mitchell 1997). We have not explored the implications of this in our analysis later.

GFDL low-resolution coupled model results

The halving of CO_2 concentrations causes the model to cool. The rate of cooling varies at different depths and latitudes in the ocean. The slowest response occurs at about 2000 m depth in the Northern Hemisphere ocean, and not the deep ocean (Fig. 13). The long time scales occur in both the North Pacific and the North Atlantic oceans.

Figure 12 shows how the global average ocean temperature evolves in time at these mid-depths in the ocean. At mid-depths the ocean gradually cools by about 1 K over the initial 2000 years of the GFDL control simulation (Fig. 12), with a negligible cooling

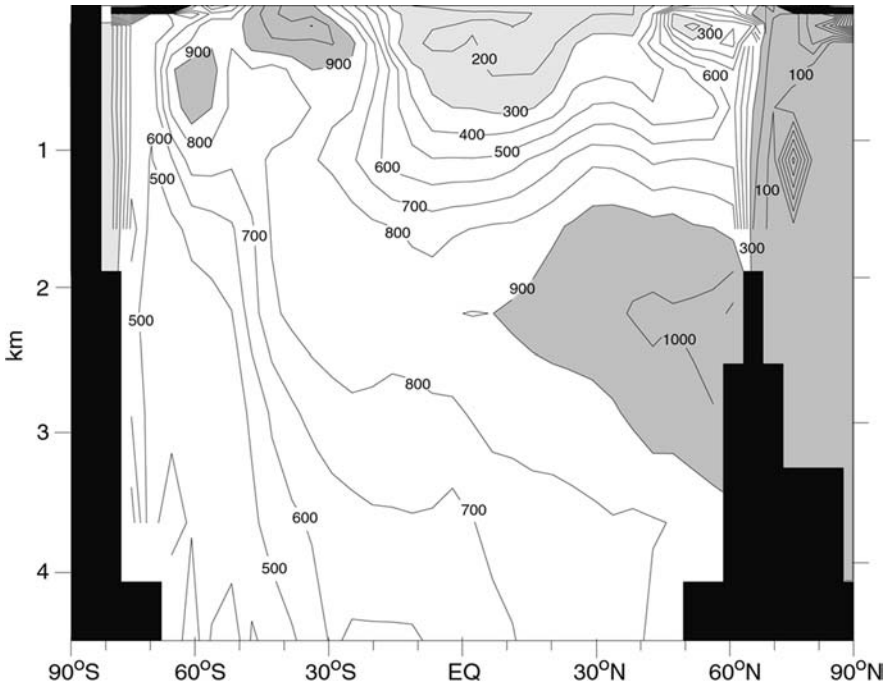

Fig. 13 The time (years) to reach 70% of the total, equilibrium response from the beginning of the integration (the first occurrence). The fraction of the total response is computed by dividing the response at any given time by the total response at each grid location. The zonal mean, latitude versus depth response time scales are shown

thereafter, indicative that the simulation has reached equilibrium. It also takes about 2000 years for the model to reach its equilibrium state (Fig. 12) once CO_2 concentrations are halved.

Figure 12 suggests that without the initial HadCM3H stage, the GFDL coupled model would take about 300–500 years longer to reach the same mid-depth ocean cooling as the experiment with the initial SST restoring term. Since the LGM experiment uses a different climate forcing and a different model (with different parameterization of vertical mixing which could have an impact on the time scale of response) we cannot quantify the acceleration the HadCM3H stage produces for the LGM experiment. However, since the largest cooling occurs in the mid- and high-latitudes of the Northern Hemisphere in the LGM experiment, the regions where the GFDL coupled model takes the longest to cool, it is likely that the HadCM3H LGM experiment has a similar, if not larger, acceleration towards equilibrium.

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