NOTES AND CORRESPONDENCE

The Effect of a Large Freshwater Perturbation on the Glacial North Atlantic Ocean Using a Coupled General Circulation Model

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ABSTRACT

The commonly held view of the conditions in the North Atlantic at the last glacial maximum, based on the interpretation of proxy records, is of large-scale cooling compared to today, limited deep convection, and extensive sea ice, all associated with a southward displaced and weakened overturning thermohaline circulation (THC) in the North Atlantic. Not all studies support that view; in particular, the "strength of the overturning circulation" is contentious and is a quantity that is difficult to determine even for the present day. Quasi-equilibrium simulations with coupled climate models forced by glacial boundary conditions have produced differing results, as have inferences made from proxy records. Most studies suggest the weaker circulation, some suggest little or no change, and a few suggest a stronger circulation.

Here results are presented from a three-dimensional climate model, the Hadley Centre Coupled Model version 3 (HadCM3), of the coupled atmosphere–ocean–sea ice system suggesting, in a qualitative sense, that these diverging views could all have occurred at different times during the last glacial period, with different modes existing at different times. One mode might have been characterized by an active THC associated with moderate temperatures in the North Atlantic and a modest expanse of sea ice. The other mode, perhaps forced by large inputs of meltwater from the continental ice sheets into the northern North Atlantic, might have been characterized by a sluggish THC associated with very cold conditions around the North Atlantic and a large areal cover of sea ice. The authors' model simulation of such a mode, forced by a large input of freshwater, bears several of the characteristics of the Climate: Long-range Investigation, Mapping, and Prediction (CLIMAP) Project's reconstruction of glacial sea surface temperature and sea ice extent.

1. Introduction

The last glacial maximum (LGM), occurring about 21 000 yr ago (21 ka), represents a climate state that was substantially different from today's. There is an extensive amount of paleoclimatic data for the period and the reconstructed changes are large enough to be distinguished from present-day estimates of natural climate variability.

There is particular interest in the North Atlantic sector at the LGM where there is evidence, sometimes conflicting, of major changes to the oceanic circulation and the climate. A common view of the conditions in the North Atlantic at the LGM compared to modern conditions (based on the interpretation of proxy

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records) was of widespread and large cooling (e.g., CLIMAP Project Members 1981), limited deep convection (e.g., Duplessy et al. 1975; Boyle and Keigwin 1987) and extensive sea ice (CLIMAP Project Members 1981), all associated with a southward displaced (Duplessy et al. 1988) and weakened (McManus et al. 2004) overturning thermohaline circulation (THC).

Not all of those conditions are universally accepted. For example, some reconstructions suggest a more modest cooling compared to modern conditions (Crowley 1981; Rosell-Melé and Comes 1999), continued deep convection (Veum et al. 1992), and a fairly limited extension of Arctic sea ice [as in the Glacial Atlantic Ocean Mapping (GLAMAP) 2000 reconstruction, for example; Sarnthein et al. 2003b]. The GLAMAP reconstruction differs from the earlier landmark Climate: Long-range Investigation, Mapping, and Prediction (CLIMAP) reconstruction by using a more precise age definition of the LGM time slice, new transfer techniques to derive SSTs from microfossil assemblages, and better calibration to modern SSTs.

Perhaps the most contentious issue, however, is the "strength of the overturning circulation," a quantity that is difficult to estimate even for the present day, a point emphasized by Wunsch (2003). Similarly, the "thermohaline circulation" itself is difficult to define. Attempts have been made to deduce the strength of the THC based on proxy records, with most suggesting a weaker circulation, some suggesting no change, and one suggesting the possibility of a slightly stronger circulation (e.g., Lynch-Stieglitz et al. 1999; LeGrand and Wunsch 1995; Yu et al. 1996, respectively). Quasi-equilibrium simulations using coupled climate models forced by glacial boundary conditions also produce diverging results—strengthened overturning (Hewitt et al. 2001; Kitoh et al. 2001), little or no change (Ganopolski et al. 1998), weakened overturning (Weaver et al. 2001; Shin et al. 2003), and almost complete cessation of the overturning (Kim et al. 2003). What the models and paleodata do generally agree on is a southward displacement of the sites of deep convection and a tendency for shoaling of deepwater formation.

The discrepancies between different reconstructions and different model simulations could be explained by one or more of the following.

1) Incorrect boundary conditions for the LGM climate applied to some or all of the models. For example, the modeling studies cited above used the ICE-4G reconstruction of the LGM ice sheets, which has recently been revised (Peltier 2004) in light of new constraints. In addition, none of the modeling studies cited included changes to atmospheric aerosols or vegetation, both of which could have been very different at the LGM, or allowed for any variability in the ice sheet volume. In the case of aerosols, we do not know the direct and indirect effects of aerosols for present day very well, let alone at the LGM.
2) Incorrect interpretation of some of the proxy records. For example, the southward shift of the sites of deep convection and shoaling of deep water formation could lead to problems of interpretation of some records. There could be problems with the dating of some proxy records, problems identifying the LGM in some records, or problems interpreting some proxy records, particularly when trying to quantify the strength of the ocean's THC, perhaps from proxy indicators of water mass properties.
3) Inability of some or all of the models to simulate the THC at the LGM. Climate models have shortcomings, partly due to uncertainties in how to represent some subgrid-scale physical processes in the climate system, and partly due to potentially missing processes and feedbacks. It has previously been reported that the sensitivity of the THC in coupled ocean–atmosphere general circulation models (OAGCMs) to climate change is highly variable (Cubasch et al. 2001), for reasons that are not yet completely understood.
4) Simple theoretical models have suggested the possibility that more than one mode of the THC exists for a given climate forcing (e.g., Welander 1986). It is possible (but difficult to assess) that different initial conditions or experimental designs could lead to different modes of the THC in three-dimensional coupled climate models.

Since the LGM reconstructions differ, the second possibility listed above raises the question of whether some of the reconstructions are wrong. For example, Wunsch (2003) questions the conclusions reached by some studies of paleodata, especially the likelihood of a reduced circulation given windier conditions during glacial periods.

However, a further consideration is that there is evidence of considerable *variability* in the ocean circulation in the North Atlantic during glacial times (e.g., Keigwin and Jones 1994), associated with Dansgaard-Oeschger cycles and Heinrich events, for example. In particular, Heinrich event 2 (H2) and Heinrich event 1 (H1) occurred at either side of the LGM, with H2 finishing about 20 100 ^{14}C yr ago (20.1 ^{14}C ka), equivalent to about 23 700 calendar years ago (23.7 ka), and H1 commencing about 15.0 ^{14}C ka, equivalent to about 17.6 ka (Bard 1998; Grousset et al. 2001). In the case of H2, it is possible that the THC had not fully recovered in the time considered for the LGM in some proxy records. In addition, there is evidence of a meltwater pulse earlier than H1 (Yokoyama et al. 2000; Clark et al. 2004), at about 19 ka (approximately equivalent to 17.0 ^{14}C ka). The interval considered for the LGM varies between different reconstructions. For example, the CLIMAP reconstruction (CLIMAP Project Members 1981) centers the LGM at 18.0 ^{14}C ka (equivalent to about 21.1 ka). The GLAMAP 2000 reconstruction (Sarnthein et al. 2003a) has two definitions of the LGM time slice, one being 15–18 ^{14}C ka (18–21.5 ka) for North Atlantic sediment records and the other being 16–19.5 ^{14}C ka (19–23 ka) for western equatorial and South Atlantic sediment records.

This raises the possibility that different proxy records could be providing information on different ocean events or states (modes). One such state could be characterized by weak overturning circulation (weak relative to the general state throughout the glacial period) in the North Atlantic, relatively cold SSTs, and extensive sea ice cover such as may have occurred at either side of the LGM during H2 and H1 (Bard et al. 2000), with ocean cores showing that the North Atlantic during H1 was cooler than during the LGM (Bond et al. 1993; Rasmussen et al. 1996). A different state could have relatively strong overturning, relatively warm SSTs, and reduced sea ice. The meltwater pulse at 19 ka, which could have influenced some records in both CLIMAP and GLAMAP, has supporting evidence of reduced overturning in the North Atlantic (Clark et al. 2004). Several studies with coupled climate models also support the possibility of considerable variability in the strength and nature of the THC in response to freshwater input (Manabe and Stouffer 1997; Ganopolski and Rahmstorf 2001; Rind et al. 2001; Renssen et al. 2002; Claussen et al. 2003).

Here we extend the work of Hewitt et al. (2001), who used the Hadley Centre Coupled Model version 3 (HadCM3) coupled OAGCM forced with boundary conditions appropriate for the LGM, using the specifications of the first phase of the Paleoclimate Modelling Intercomparison Project (PMIP1; available online at http://www-lsce.cea.fr/pmip). The model's simulation in the region of the northern North Atlantic is at odds with several reconstructions. The model produces anomalously warm LGM surface conditions over parts of the North Atlantic relative to the CLIMAP reconstruction (CLIMAP Project Members 1981) and a slightly intensified THC, at odds with the findings of Lynch-Stieglitz et al. (1999) or McManus et al. (2004).

In this study we describe an idealized experiment with the same model (HadCM3) to investigate particular events that have been recorded during glacial times, namely due to massive inputs of freshwater into the Labrador Sea and North Atlantic. The purpose of this study is to yield insights into some of the issues surrounding data from the North Atlantic for the LGM and their interpretation; in particular, we explore some aspects of the first and second possibilities listed above (associated with boundary conditions for the model and proxy records). Our study is intended to be illustrative rather than definitive. This is essentially a sensitivity study designed to explore some of the mechanisms that could be responsible for the different responses. Since, doubtless, there are still issues with the prescribed boundary conditions, the forcings used, the model, the experimental design deployed, and the paleodata themselves, this study should not just be seen as an exercise in model validation, although it does have some implications for model–data comparisons. It seeks to provide a framework for investigating the model's response to freshwater inputs during the LGM and to increase our understanding of the paleorecord.

Section 2 describes the model we use and the design of the experiments. The experiments are idealized and represent a large freshwater event during a glacial period, of which Heinrich events H2 and H1 (which occurred a few thousand years on either side of the LGM) are extreme examples. Section 3 describes the model results, concentrating on the response of the THC to adding a massive amount of freshwater, and consequently describing the effect of the THC changes on the climate, particularly around the North Atlantic.

Here, for the first time, we present evidence from a three-dimensional coupled ocean–atmosphere climate model that the differing reconstructions from proxy records in the North Atlantic could all have occurred at different times during or around the last glacial maximum period, with the very cold conditions and very weak THC occurring during periods of large meltwater inflow to the northern North Atlantic. It is therefore possible that different model results and different paleoclimatic reconstructions are relevant to different climate states during the glacial period.

2. Model description and experimental design

The model that we use, HadCM3 (Gordon et al. 2000), is a coupled atmosphere–ocean–sea ice general circulation model used extensively for climate studies. The atmospheric component of HadCM3 has a horizontal resolution of 2.5° by 3.75° and 19 vertical levels. The ocean model has a horizontal resolution of 1.25° by 1.25° and 20 depth levels. The sea ice model includes a simple representation of sea ice thermodynamics and dynamics. The model does not use flux adjustments

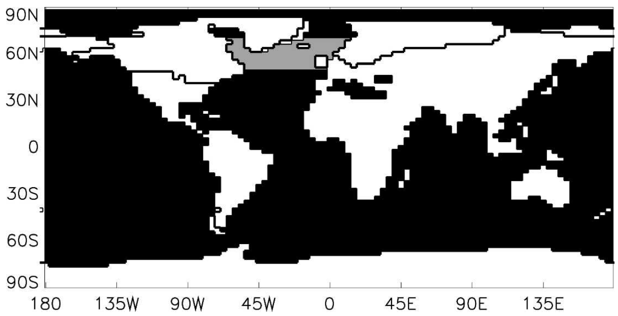

FIG. 1. A map showing the region over which the freshwater forcing is applied (in gray), the LGM land–sea distribution, and the boundaries of the prescribed continental ice sheets (thick black lines).

(Manabe and Stouffer 1988) and has a stable multimillennial climate simulation for present day. The HadCM3 model, its simulation of present-day climate, experiments with increased CO_2 concentrations, and experiments with glacial boundary conditions are described in more detail elsewhere (Wood et al. 1999; Gordon et al. 2000; Hewitt et al. 2003). HadCM3 produces a realistic simulation of the modern climate, comparable with other state-of-the-art OAGCMs.

To investigate the effect of freshwater perturbations on the glacial climate, we perform two integrations of the model. A 1000-yr-long simulation of the climate at the last glacial maximum has already been carried out using HadCM3 with PMIP1 LGM boundary conditions (Hewitt et al. 2003). We have started our two new integrations from the end of Hewitt et al.'s 1000-yr-long simulation. The baseline experiment is a 500-yr-long continuation of the LGM experiment. This integration will hereafter be referred to as experiment *LGM*. A second integration has been performed, identical to *LGM* but with 1 Sv (1 Sv $\equiv 10^6$ m^3 s^{-1}) of freshwater (FW) continuously added over the surface of the North Atlantic Ocean between 50° and 70°N (Fig. 1) for 100 yr. The freshwater flux does not alter the in situ ocean temperature and for numerical reasons is, in fact, a negative salinity flux. This experiment will hereafter be referred to as experiment *LGM_FW*. By comparing the *LGM_FW* experiment to the *LGM* experiment, we can see what effect a high-latitude freshening of the North Atlantic and Nordic Seas has on the glacial climate. We show comparisons between the 100-yr average from the *LGM* experiment and an average over years 51–100 from the *LGM_FW* freshwater experiment.

The region where the freshwater is applied has been chosen as an approximate and idealized representation of where anomalously fresh surface water may have existed during periods of massive discharges of icebergs that were released periodically from the ice sheets that surrounded the North Atlantic during glacial times, such as could have occurred during Heinrich events or during the meltwater pulse 19 000 yr ago. The large area chosen accounts for the possibility of the source regions being either the Laurentide Ice Sheet or the European ice sheets, an issue that is currently under debate (e.g., Grousset et al. 2001).

Since the LGM, about 50×10^6 km^3 of ice has melted from the land-based ice sheets, raising the sea level by about 120 m (Church et al. 2001). At the termination of the LGM, a rapid decrease in ice volume of about 10% occurred within a few hundred years (Yokoyama et al. 2000). Clark et al. (2004) comment that this rapid decrease in ice volume would have produced a freshwater flux of between 0.25 and 2 Sv over 100–500 yr, but their data from the Irish Sea indicates that the meltwater event at 19 ka may have been considerably shorter in duration than 500 yr. The freshwater forcing of 1 Sv for 100 yr used in the *LGM_FW* experiment is comparable to such a rapid decrease, although the area over which this freshwater is added to the upper ocean is probably somewhat larger than in reality. This is necessary since the ocean model cannot handle very large water fluxes over a small area because of the assumption of a rigid ocean surface used in the design of the ocean model.

3. Results

a. *Effect of freshwater perturbation on the THC*

The 500-yr-long continuation of the *LGM* simulation does not differ noticeably from the first 1000 yr that have been described elsewhere (Hewitt et al. 2003) and so will not be described in detail here. The effect of the freshwater perturbation is seen by comparing the *LGM_FW* simulation to the *LGM* simulation.

Once the freshwater is applied over the northern North Atlantic (Fig. 1), the overturning circulation rapidly weakens (Fig. 2). After about 50 yr, the maximum strength of the overturning is less than 8 Sv, significantly weaker than the 24 Sv in the *LGM* experiment and with smaller interannual and interdecadal variability. The overturning does not weaken much more after 50 model years in the *LGM_FW* simulation. Since most of the adjustment has occurred over the first 50 yr, the following results are averaged over the final 50 yr of the *LGM_FW* simulation.

In the *LGM* experiment, there is a fairly vigorous extensive cell in the upper 2500 m of the North Atlantic (Fig. 2b) with deep convection across the northern North Atlantic (Fig. 3b), as reported in Hewitt et al. (2003). While the depth extent of this cell could be seen to be consistent with proxy records suggesting Glacial North Atlantic Intermediate Water (GNAIW) forma-

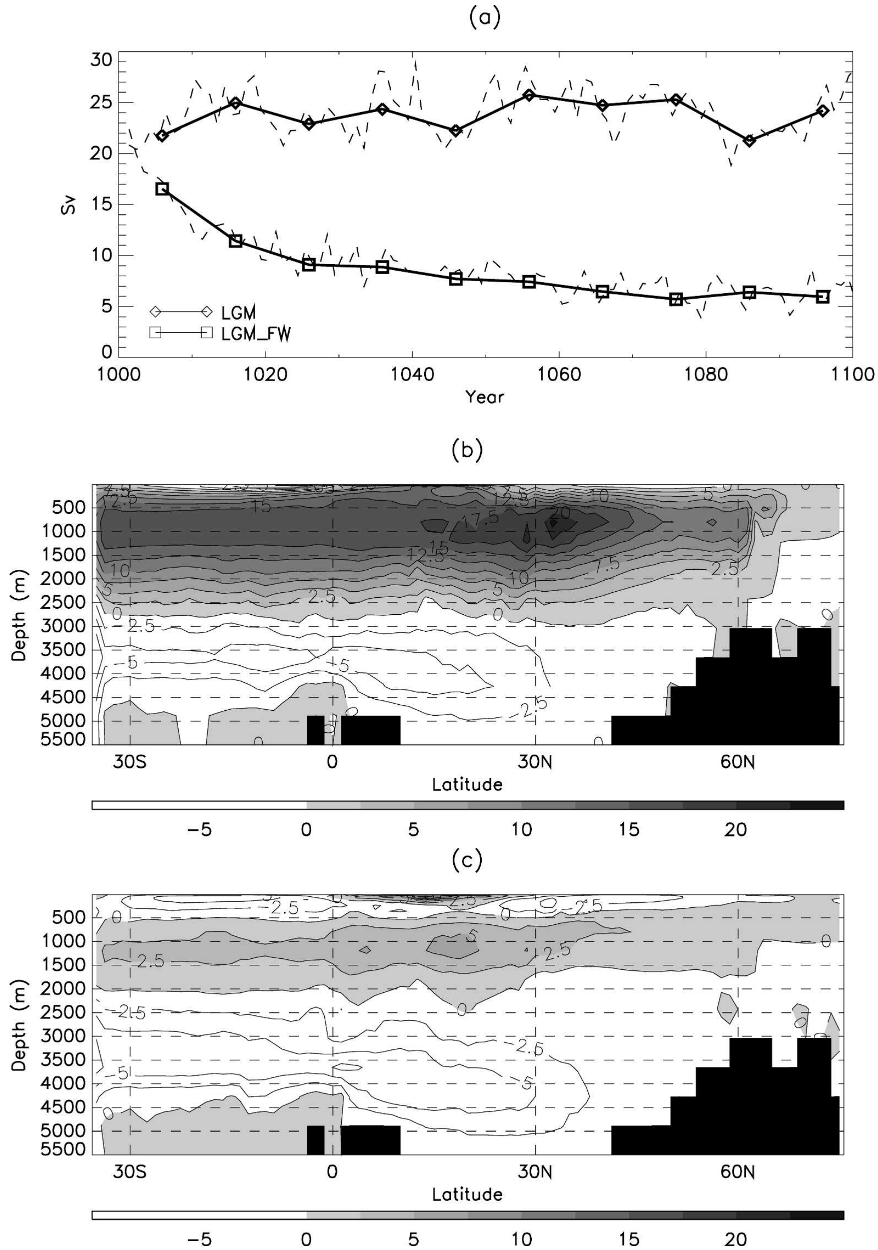

FIG. 2. The Atlantic meridional overturning streamfunction (in Sv). The (a) time series of maximum strength, based on decadal (solid lines) and annual averages (dashed lines) from the *LGM* (diamonds) and *LGM_FW* (squares), (b) *LGM* during model years 1001–1100, and (c) *LGM_FW* during model years 51–100.

tion, the study by Hewitt et al. (2003) points out that the North Atlantic Deep Water formation in the simulation of the modern climate does not extend as deep as observations suggest, and as a result, the cell is not shallower at the LGM compared to modern times, in disagreement with the proxy records. However, it is the changes between the *LGM* and *LGM_FW* simulations that we are interested in here, so the changes described below use the *LGM* simulation as the control, unless stated otherwise.

The structure of the *LGM_FW* circulation is dramatically altered by the freshwater forcing (Fig. 2c). In the *LGM_FW* experiment, the cell becomes much weaker and is much less extensive both vertically (shoaling by between 500 and 1000 m) and latitudinally, with very little deep convection in mid- and high latitudes. These

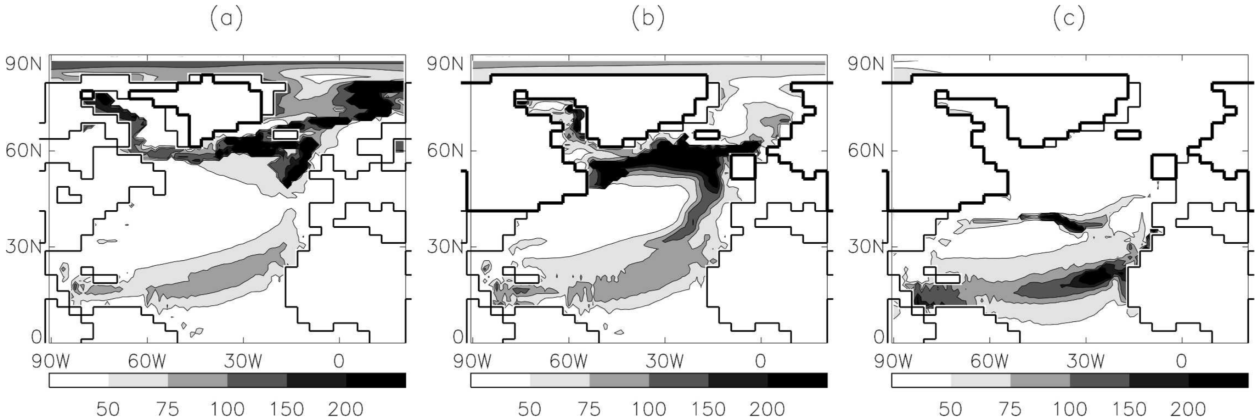

FIG. 3. The annual mean mixed-layer depth (in m) for the (a) modern simulation during model years 1001–1100, (b) *LGM* during model years 1001–1100, and (c) *LGM_FW* during model years 51–100.

changes are brought about as follows: the salinity and density of the surface waters are reduced in the region where the freshwater perturbation is applied, and these fresher, less-dense waters are gradually advected over much of the upper few hundred meters of the North Atlantic Ocean. The reduction in surface density makes the water column more stable, and convection is dramatically weakened, inhibiting deep mixing.

In the simulation of the modern climate, there is deep mixing of several hundred meters in the Nordic Seas (Fig. 3, where the mixed-layer depth is calculated as the depth to which the in situ temperature is within 0.5 K of the SST), driven by a large heat flux in winter from the relatively warm ocean to the relatively cold atmosphere, cooling the surface waters and thus making them denser. In the *LGM* experiment, the expansion of sea ice across the Nordic Seas (Fig. 4b) provides an insulating blanket on the ocean, reducing the large loss of heat from the ocean to the cold overlying atmosphere, which inhibits the deep mixing there. The region of deep mixing shifts southward beyond the sea ice margin in the *LGM* experiment to the high-latitude North Atlantic and Labrador Sea, at around 60°N. In the *LGM_FW* run, however, there is no deep mixing anywhere north of about 40°N. The deep mixing in the *LGM_FW* experiment is inhibited by changes to the density of the upper ocean from both freshwater (as discussed above) and heat fluxes. The heat fluxes are reduced because thick sea ice subsequently expands across the North Atlantic (Fig. 4c; discussed below), which, as in the *LGM* experiment, acts as an insulating blanket on the ocean.

The consequence of the above changes on the THC in the *LGM_FW* experiment is that the reduced vertical mixing and lower surface density weaken the supply of dense water to the deep ocean in the North Atlantic, which reduces the deep ocean density in the North Atlantic. This, in turn, weakens the deep ocean meridional density gradient, which weakens the THC (Stommel 1961).

While the effect on the THC of adding freshwater to the ocean has been studied for the modern ocean using OAGCMs (Manabe and Stouffer 1997) and for the modern and glacial oceans using earth system models of intermediate complexity (EMICs; Ganopolski and Rahmstorf 2001; Schmittner et al. 2002), there have been no studies for the glacial ocean using an OAGCM. Below we discuss the impact on the surface temperature and sea ice during glacial times, which has not been shown before.

b. *Effect of freshwater perturbation on climate*

The freshwater perturbation has a dramatic effect on the temperature across much of the earth's surface (Fig. 5c). There is widespread cooling across most of the Northern Hemisphere in the *LGM_FW* experiment compared to the *LGM* experiment, and a smaller but comparably widespread warming across most of the Southern Hemisphere. This hemispheric seesaw effect has been observed in proxy records, for example, in marine isotope stage 3 between approximately 60 and 30 ka (Voelker 2002) and in Greenland and Antarctic ice cores for the past 90 000 yr (Blunier and Brook 2001). The largest cooling in the model is seen in and around the North Atlantic region, with cooling exceeding −20°C compared to the *LGM* simulation over the high-latitude North Atlantic, and cooling of about −10°C across much of central and southern Greenland and the northern North Atlantic and Labrador Sea. The large-scale pattern of temperature change simulated in these glacial experiments is similar to that reported by Vellinga and Wood (2002), who studied the

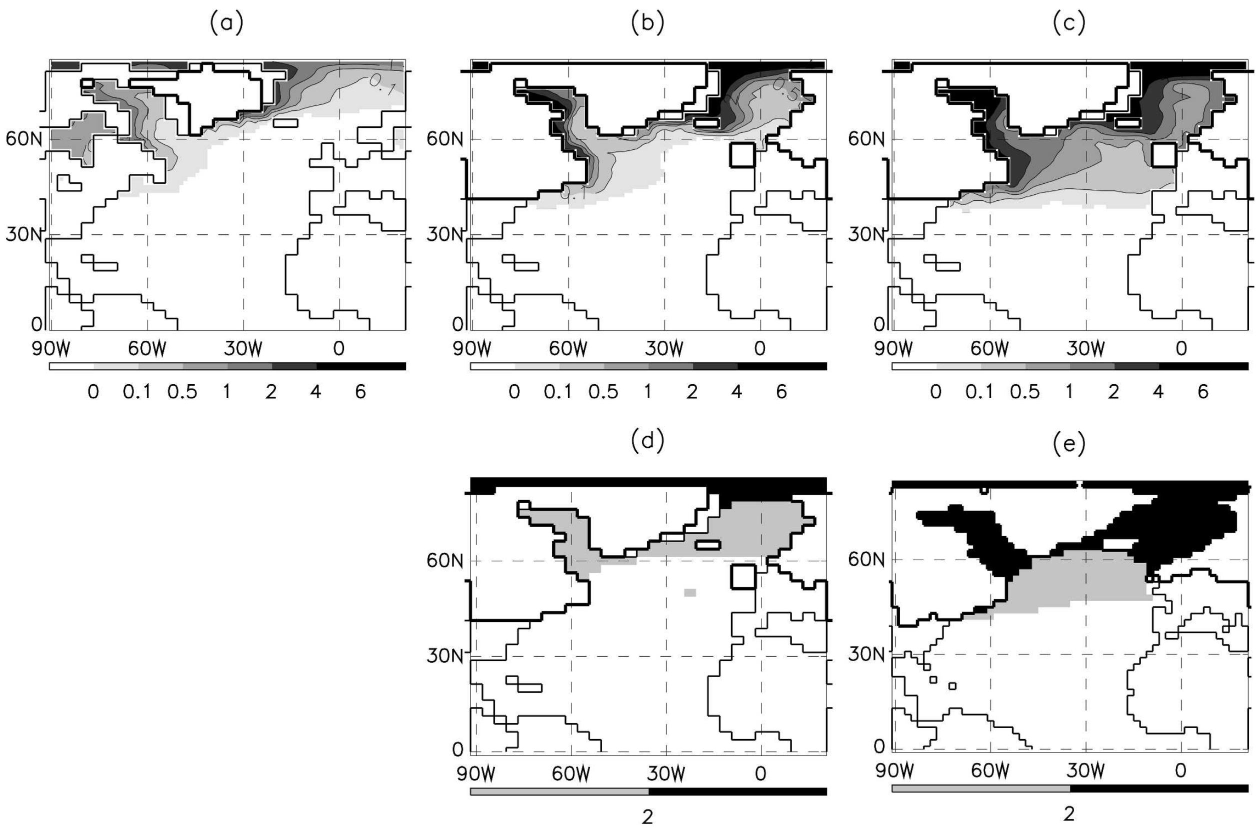

FIG. 4. The modeled annual mean sea ice depth (in m) for the (a) modern simulation, (b) *LGM*, and (c) *LGM_FW*. The approximate maximum (light gray) and minimum (dark gray) sea ice extents from (d) GLAMAP 2000 and (e) CLIMAP.

impact on the modern climate of turning off the THC using HadCM3 and, similar to the response seen in the Climate and Biosphere Model (CLIMBER-2) EMIC (Ganopolski and Rahmstorf 2001), who applied a slowly increasing (to 0.15 Sv) freshwater forcing over about 1500 yr between 50° and 80°N in the Atlantic.

The large-scale pattern of the surface temperature changes can be attributed to the weakened THC. Hewitt et al. (2001) describe how the strengthened THC in the *LGM* simulation (compared to the simulation of the modern climate) increases the amount of heat transported northward in the Atlantic Ocean (Fig. 6). The warming from the increased ocean heat transport is large enough for the LGM ocean surface to be locally warmer than in the modern climate simulation in the Gulf Stream region (Fig. 7a). This warm Gulf Stream restricts the cooling in the Labrador and Nordic Seas.

In the *LGM_FW* experiment, however, the weakened THC reduces the amount of heat transported northward in the Atlantic Ocean (Fig. 6). This leads to widespread cooling across the North Atlantic Ocean (Fig. 7b). The pattern of large cooling over the northern North Atlantic and more modest cooling and even small warming in the tropical Atlantic in the *LGM_FW* experiment is more similar to that seen in the CLIMAP reconstruction than in the GLAMAP reconstruction (Figs. 7c,d), while the pattern seen in the *LGM* experiment is more similar to the GLAMAP reconstruction.

The sea ice coverage in the *LGM* experiment also bears similarities to the GLAMAP 2000 reconstruction (cf. Figs. 4b,d), with permanent sea ice restricted to the high-latitude Arctic and the coastal waters of Greenland and North America, and seasonal sea ice coverage on the Labrador and Nordic Seas and high-latitude North Atlantic. The more extensive sea ice coverage in the *LGM_FW* experiment is approaching that seen in the CLIMAP reconstruction (cf. Figs. 4c,e), with permanent sea ice covering more of the Labrador and Nordic Seas and seasonal sea ice covering the North Atlantic equatorward to about 40°N.

The enhanced North Atlantic cooling in the *LGM_FW* experiment also affects other Northern Hemisphere regions (Fig. 5c).The large-scale zonal circulation in the atmosphere transports the cold air rapidly to other regions; such a teleconnection is seen in

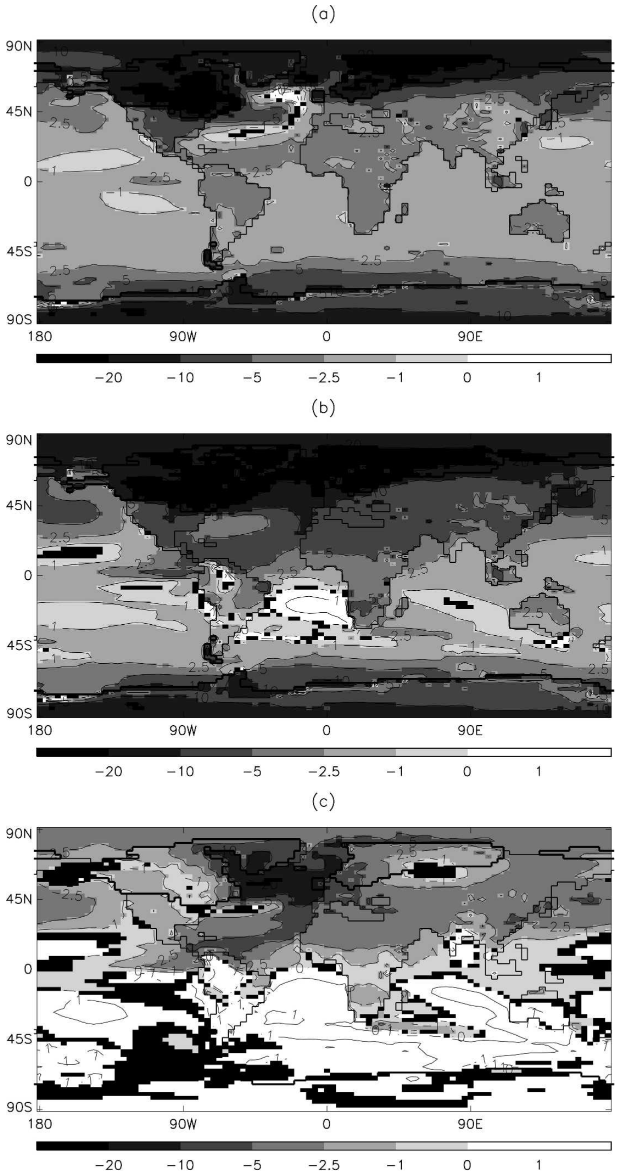

FIG. 5. The annual mean surface temperature changes (in K) for the (a) *LGM* minus modern simulation, (b) *LGM_FW* minus modern simulation, and (c) *LGM_FW* minus *LGM*. The changes that are not statistically significant based on a Student's *t* test are masked out in black.

continental records where the Atlantic climate system extended its influence at least as far as the central Mediterranean region (Allen et al. 1999). As a consequence of less heat being transported northward in the North Atlantic Ocean in the *LGM_FW* experiment, more heat remains in the South Atlantic Ocean (Fig. 6), which in combination with the large-scale zonal transport by the atmosphere, produces the widespread warming (compared to the *LGM* simulation) across the Southern Hemisphere.

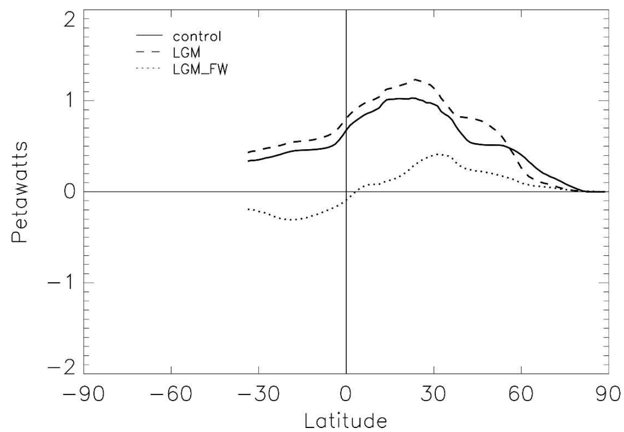

FIG. 6. The annual mean northward ocean heat transports in the Atlantic Ocean.

4. Discussion

Three-dimensional coupled ocean–atmosphere climate models are now being used to simulate the equilibrium climate at the last glacial maximum, and comparisons are being made between those model simulations and reconstructions of climate parameters from proxy records. However, we should be realistic with our expectations when making such comparisons. The models are not perfect, the climate reconstructions are not perfect, and even a perfect model will not produce an accurate simulation if the boundary conditions (or climate forcing), be they either constant or varying, are not properly representative of the climate being considered.

We have carried out experiments with a three-dimensional coupled ocean–atmosphere climate model (HadCM3) to investigate the effect of a massive freshwater event on the glacial climate. We compare two model simulations. One simulation, *LGM*, has boundary conditions appropriate for the last glacial maximum (extensive continental ice sheets, modified coastlines and topographic heights, glacial CO_2 concentrations, and a change to the earth's orbit following the PMIP1 specifications); the other, *LGM_FW*, has the same glacial boundary conditions as *LGM* but has 1 Sv of freshwater continuously added to the North Atlantic between 50° and 70°N. The effect of adding freshwater to the ocean and investigating the response of the THC has been shown in work with earth system models of intermediate complexity for modern times and LGM (Ganopolski and Rahmstorf 2001; Schmittner et al. 2002) and with OAGCMs for modern times (Manabe and Stouffer 1997), but here we discuss the impact on the surface temperature and sea ice extents during glacial times using an OAGCM, which has not been shown before.

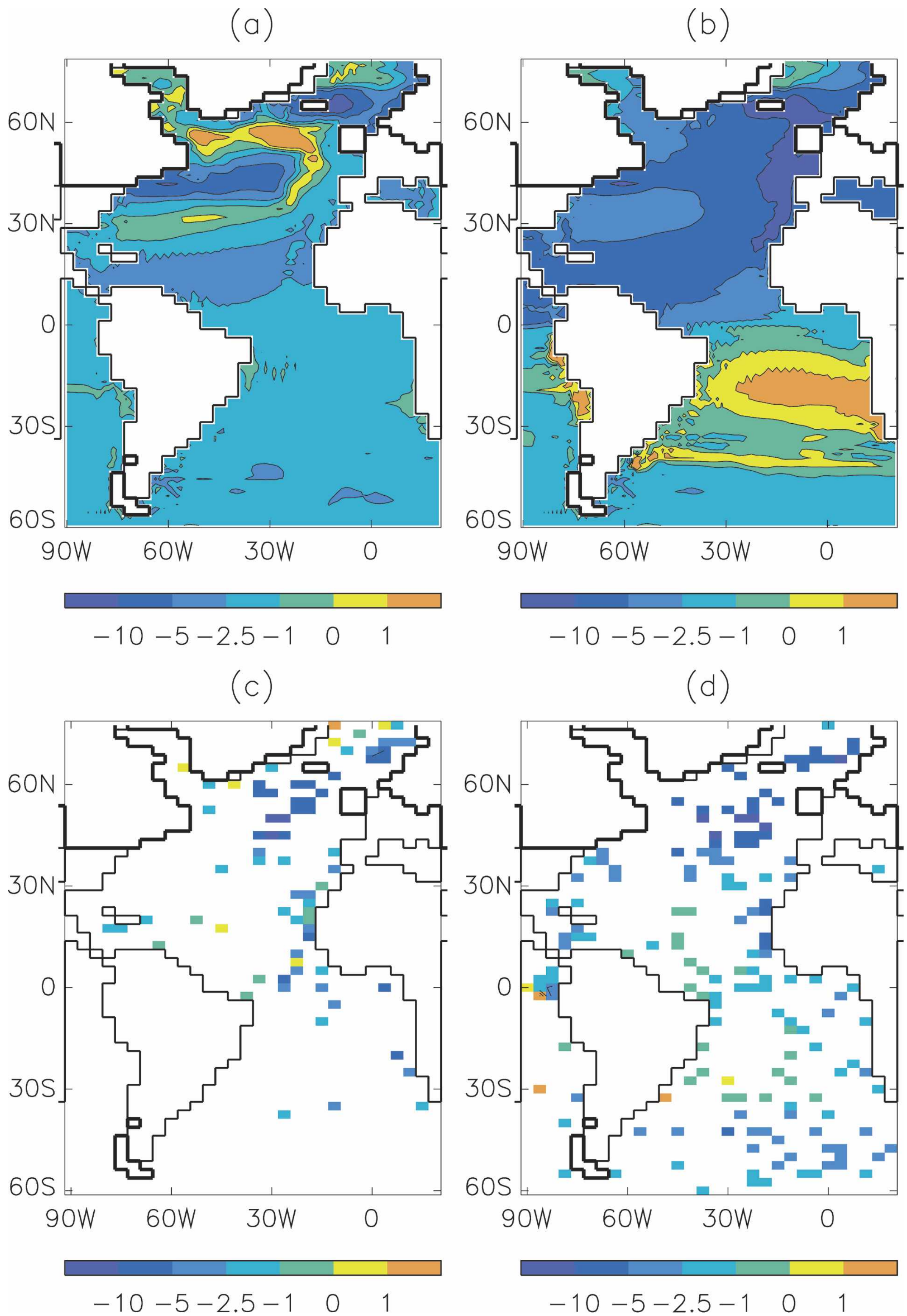

FIG. 7. The change in annual average ocean temperature (in K) for the (a) *LGM* minus modern ocean temperature at 5-m depth, (b) *LGM_FW* minus modern ocean temperature at 5-m depth, (c) GLAMAP LGM minus modern SSTs plotted at the model grid boxes where GLAMAP cores are located, and (d) CLIMAP LGM minus modern SSTs plotted at the model grid boxes where CLIMAP cores are located. Note that the GLAMAP and CLIMAP "annual averages" have been constructed by averaging the northern winter and summer reconstructions.

Some aspects of the HadCM3 simulation of the climate at the LGM (compared to a simulation of the modern climate) were unexpected, for example, slightly warmer sea surface temperatures in parts of the northern North Atlantic (Hewitt et al. 2001) and a slight strengthening of the THC. Confusingly, the HadCM3 LGM results were in agreement with the interpretations of some proxy records and some models but in disagreement with others. When the freshwater was applied to the LGM simulation, the model produced results more in line with those models and paleodata that show a relatively cold surface climate and a weak THC. The freshwater forcing dramatically weakened the thermohaline circulation in the North Atlantic, had a major effect on the simulated glacial temperatures particularly around the North Atlantic, and produced a large expansion of Northern Hemisphere sea ice, with the marginal sea ice zone expanding out to the ice-rafted detritus belt associated with Heinrich events (Ruddiman 1977; Grousset et al. 2001).

One possible conclusion is that different modes existed during glacial times, potentially forced by variations in freshwater discharge from the ice sheets. One mode might have been characterized by relatively moderate temperatures in the North Atlantic, a modest expanse of sea ice, and an active thermohaline circulation, as suggested by the reconstructions of Rosell-Melé and Comes (1999), GLAMAP 2000 (Sarnthein et al. 2003b), and Yu et al. (1996), respectively. The other mode, perhaps forced by the input of freshwater from the ice sheets, might have been characterized by a very cold North Atlantic, large areal cover of sea ice, and a sluggish thermohaline circulation, as suggested by the reconstructions of CLIMAP Project Members (1981) and Lynch-Stieglitz et al. (1999) and also seen in simulations using an intermediate complexity climate model (Rahmstorf 2002).

This idea of multiple modes of the THC has been put forward by several authors (e.g., Alley et al. 1999; Rahmstorf 2002), with a common classification into the following three modes: a "modern mode" with deep water formation in both the Nordic Seas and farther south in the North Atlantic, a "glacial mode" with much reduced deep water formation in the Nordic Seas but continued deep–intermediate water formation in the North Atlantic, and a "Heinrich mode" with greatly reduced deep–intermediate water formation [such as found in measurements of a kinematic proxy for the THC by McManus et al. (2004)]. The HadCM3 *LGM* results discussed here would correspond to the "glacial mode" and the *LGM_FW* results would correspond to the "Heinrich mode." The HadCM3 simulation of the present-day climate (not the subject of this paper) would correspond to the "modern mode."

The LGM should not be seen as a climatic extreme everywhere—a view pointed out by McManus et al. (2004)—and we should consider the potential shortcomings of the experimental design, the models used to simulate the climate, and the proxy records used to reconstruct the climate. However, as an idealized modeling study, these results show that very different conditions can be simulated over and around the North Atlantic by applying plausible freshwater fluxes. Previous coupled OAGCM glacial simulations have used nonvarying climate forcing, but in order to simulate different climate states during the glacial period, different climate forcing (perhaps time varying) may be needed to reproduce some of the reconstructed LGM climatic features.

Experiments such as this support the interpretation of paleoevidence in suggesting that large climate changes, concentrated in the Atlantic region but with worldwide influence, could have occurred in the past in response to an influx of freshwater from the melting of ice sheets. This suggests the possibility that future greenhouse warming may disrupt the THC by the melting of the Greenland ice sheet (Church et al. 2001; Gregory et al. 2004). However, even in extreme scenarios, that process takes many centuries, producing a freshwater influx of a few 0.01 Sv (more than an order of magnitude smaller than the forcing applied here). In future climate change, ice sheet meltwater is likely to be a less important source of high-latitude freshwater input than increased precipitation or Arctic river inflow, which might amount to fluxes on the order of 0.1 Sv. In recent model intercomparison studies, Stouffer et al. (2006) show that an influx of this size weakens the THC, with a climatic impact that is significant but smaller than in our experiment, while Gregory et al. (2005) find that this cooling due to freshwater input is outweighed by the greenhouse warming in scenarios of CO_2 increase.

In terms of future work, this study only uses one climate model and one experimental design. What effect would different boundary conditions have on the results [e.g., a different ice sheet, different freshwater forcing (including different river pathways), interactive vegetation, and atmospheric dust]? Work is now underway in the second phase of the Paleoclimate Modelling Intercomparison Project (PMIP2; Harrison et al. 2002) and the European Community (EC)-funded Models and Observations to Test Climate Feedbacks (MOTIF) project to investigate the effect of LGM boundary conditions on coupled climate models by using a common experimental design for all LGM simulations as well as sensitivity experiments to freshwater input. The Multiproxy Approach for the Reconstruction of the Glacial Ocean surface (MARGO) project is aiming to improve the reconstructions from proxy records (Kucera et al. 2005), and model intercomparison projects such as the Coupled Model Intercomparison Project (CMIP; Stouffer et al. 2006) and PMIP2 are aiming to improve our understanding of the models with the intent of improving the models themselves.

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