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*Christopher D. Hewitt , Ronald J. Stouffer ,
Anthony J. Broccoli , John F. B. Mitchell
and Paul J. Valdes .*

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Christopher D. Hewitt¹, Ronald J. Stouffer², Anthony J. Broccoli²,
John F. B. Mitchell¹ and Paul J. Valdes³.

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¹Met Office, Hadley Centre for Climate Prediction and Research, Bracknell, UK

²NOAA/Geophysical Fluid Dynamics Laboratory, Princeton, USA

³The University of Reading, Department of Meteorology, UK

Note: this technical note concentrates on the experimental design of the HadCM3 simulation of the climate at the Last Glacial Maximum, and the adjustment of the model towards equilibrium. More detailed results from the 1000-year long near equilibrium simulation will be submitted for publication in due course.

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 deep ocean can take 1000 years or longer to respond, which is currently impractical to simulate using a GCM. In the absence of a global reconstruction of the state of the glacial ocean, the model is initialised from present day conditions, and a strategy has been devised to increase the rate of cooling of the ocean. The model displays large trends once the glacial boundary conditions are applied, particularly in the structure of the thermohaline circulation. Mechanisms for these trends are discussed. A lower resolution coupled model with atmospheric CO₂ concentrations halved has been used to carry out millennial timescale simulations to investigate the experimental strategy used for the LGM HadCM3 simulation.

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, 1990; Houghton et al, 1996). 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) is well-suited for palaeoclimate studies. The LGM represents the largest global and regional climate change of recent geologic times, and the main factors responsible for the different climate are believed to be well known. There is an abundance of well-dated palaeoclimatic data available, and it is a period that has been the focus of research using 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).

The large differences between the glacial and present day climates represent perhaps an extreme challenge for a GCM's capability to realistically simulate climate change. Several climate model experiments have already been carried out to simulate the atmospheric response at the LGM (e.g. Gates, 1976; Manabe and Broccoli, 1985; Kutzbach and Guetter, 1986; Lautenschlager and Herterich, 1990; Bush and Philander, 1998; Broccoli, 2000). Since the thermal structure of the thermocline takes decades to respond, and the deep ocean responds

on timescales of centuries and longer, no experiments to date have attempted to simulate the long-term response of both the atmosphere and ocean three-dimensional circulations, and their interaction with sea ice.

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. The advantages of the AGCM and slab model studies was that the simulations could be run long enough to produce a stable (equilibrium) LGM climate simulation, and, by ignoring feedbacks from the deep ocean and ocean dynamics, it was easier to gain an insight into the mechanisms of the atmospheric changes. Such advantages are particularly useful for intercomparing model simulations, as has been done in PMIP for example (Joussaume and Taylor, 1995; Taylor et al, 2000). 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 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) either with prescribed atmospheric boundary forcing (e.g. Seidov and Haupt, 1997; Bigg et al, 1998) or with a simple energy balance model representing the atmosphere (Weaver et al, 1998). However, important feedbacks may also have been neglected in such studies. 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). The EMIC consists of a very coarse resolution atmospheric model (10° latitudinal and 51° longitudinal horizontal resolution) and a zonally averaged ocean model with three separate ocean basins (Atlantic, Indian and Pacific oceans). The advantage of this formulation is that the experiments can be run for many thousands of years to reach an equilibrium state. However, the disadvantages are that the EMIC cannot simulate regional details, the zonal average ocean model cannot simulate the gyre circulations, and the heat and salt transports of the horizontal circulation are parameterized.

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 it has not been a viable proposition to attempt to simulate a quasi-equilibrium LGM climate using a coupled OAGCM. The only previously published coupled OAGCM simulation of the response to glacial boundary conditions consisted of a 15-year simulation initialised from present day conditions (Bush and Philander, 1998). However, the simulation was not long enough to have allowed the thermocline and deep ocean to have adjusted to the glacial conditions.

Increases in computer processing power have now made it possible to run simulations using relatively sophisticated models for many hundreds of years in length, thus allowing much of the ocean to respond to the glacial surface conditions. The predecessors of the current state-of-the-art models needed artificial fluxes (so-called 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.

In this study 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 artificial flux adjustments. This study is the first to attempt to simulate the quasi-equilibrium climate at the LGM using a fully coupled high resolution ocean-atmosphere general circulation model that does not require artificial flux adjustments.

The high-resolution HadCM3 model is computationally expensive and the slow timescales associated with the response of the deep ocean mean that the LGM simulation has not yet reached full equilibrium throughout the deep ocean. To increase the rate of cooling of the ocean the SSTs are linearly damped toward glacial values for 70 years, a technique 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 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.

The HadCM3 model and the LGM experimental design are described in Section 2, as are the GFDL model and the findings from the lower-resolution coupled model experiments. Results from the HadCM3 Haney forced stage are briefly described in Section 3. Once the Haney forcing is removed, the HadCM3 coupled model shows signs of large adjustments as it approaches equilibrium, as described in Section 4. The North Atlantic thermohaline circulation (NATHC) in particular undergoes substantial changes, and is described in some detail in Section 5.

The response of the HadCM3 coupled model to the LGM boundary conditions after 700 years of simulation has already been described in Hewitt et al (2001). The findings presented by Hewitt et al (2001) are consistent with the findings after 1000 years of simulation, and will not be repeated here. The purpose of this technical note is to describe the experimental design employed to bring the model close to equilibrium, and to describe some of the mechanisms responsible for the transient response of the coupled model. A detailed evaluation of the LGM response is beyond the scope of this manuscript.

2 Experimental design

2.1 Met Office coupled model 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. The sea ice model includes a simple thermodynamic budget and the ice moves with the ocean currents. 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 ocean surface fluxes are 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 Boundary conditions

To simulate the glacial climate the model’s boundary conditions need to be modified. Much work has been done to determine the appropriate glacial boundary conditions for GCMs as part of PMIP. The glacial boundary conditions differ from those of the present day by modifying the land surface characteristics to include the extensive continental ice sheets, modified coastlines to account for a sea level lowering, lowering the atmospheric composition of CO_2 , and a different pattern of insolation arising from a change to the earth’s orbit.

The land surface properties in ice-free regions are not changed from the present day properties, although new values of albedo are specified for emerged land points resulting from the sea level lowering. The vegetation and soil properties are only modified at new land points and new ice points. For the ice-covered points appropriate values, as determined from permanent ice points in the control run, are prescribed. For new land points, the zonal average value for each parameter at that particular latitude is applied, or if there are no values in a particular latitude band then the global average value is used instead.

Of the trace greenhouse gases, only CO_2 concentrations are altered in the PMIP experiments and in this study. The CO_2 concentration of the atmosphere is 280 ppmv in the control simulation, and 200 ppmv in the LGM simulations. Recent calculations suggest that the LGM decrease in methane and nitrous oxide would add approximately 16% to the overall radiative forcing from greenhouse gases (Broccoli, 2000).

We have stuck as close as possible to the other PMIP boundary conditions, but the inclusion of the three-dimensional ocean model necessitates some minor changes to the boundary conditions, noted below.

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). 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. These ice points are classified as land points. This produces a glacial land-sea distribution and the location of the LGM ice sheets. This glacial land-sea distribution is similar, but not identical, to the PMIP one.

The ocean's glacial bathymetry is the same as in the control run except for the top 7 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.

River runoff is included in the model using predefined river catchments, and the runoff enters the ocean at coastal outflow points. This presents two problems in the LGM experimental setup since the coastlines have moved, and the change in orography produces different river catchments in some regions. The LGM coastal outflow points are based on a combination of the river catchments from the control simulation for present day with output from a river routing program with the glacial orography.

2.3 Initial Conditions

It is often assumed that the equilibrium state of the model is largely independent of the initial state of the model, within reasonable physical bounds. This assumption is supported by the modelling study of Oglesby et al (1997), for the atmosphere and upper ocean at least, but some studies suggest there may be multiple equilibria for the thermohaline circulation of the ocean (for example, Manabe and Stouffer, 1988). The ideal initial conditions for the LGM experiment would be the “observed” global three-dimensional distribution of the model's prognostic variables, such as pressure, temperature and salinity for the glacial climate, but such data is of course not available. Therefore, in the absence of a global data set of glacial conditions from which to initialise the model, it seems prudent to initialise the model with the only climate state for which there is a global data set of conditions, namely the present day. Oglesby et al (1997) found that it takes longer to spin up a model from a cold state than a warm state. The implication here for cooling down from the present day “warm” state is that the spin-up stage will benefit from the negative forcing that the LGM boundary conditions produce. The forcing will tend to cool the ocean surface which will destabilise the water column and oceanic convection will be enhanced. This will assist in mixing the cold water throughout the ocean. In the case where the model is warming relative to the initial state, convection is suppressed and oceanic mixing will be slowed down which will mean the model will take longer to reach equilibrium.

The ocean is initialised from rest, with temperature and salinity taken from the Levitus datasets (Levitus and Boyer, 1994; Levitus et al, 1994). However, to take account of there being fewer ocean points in the LGM simulation and hence less volume of water, care has been taken

to conserve the total salt content of the ocean in the initial conditions. The amount of salt that would be lost (at ocean points that become land in the LGM simulation) is calculated and 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. Note that there is no attempt being made to adjust ocean salinity to compensate for the storage of fresh water in ice sheets. 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, the overall salinity of the ocean increases, with the volume-averaged global average salinity increasing from 34.73 practical salinity units (PSU) to 34.90 PSU.

2.4 Timescales of climate response

The glacial boundary conditions are representative of a different climate state than the initial conditions used in this study. The boundary conditions act as a forcing on the model and cause the model to adjust. The model will eventually produce a climate state that is in equilibrium with the boundary conditions. The timescale 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, as will be shown below, a 1000 year simulation will bring much of the system close to equilibrium. 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.

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. 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.

We have made use of long simulations performed using a low-resolution coupled OAGCM developed at GFDL to estimate the timescales of response of the coupled ocean-atmosphere climate system.

2.4.1 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 AOGCM has a global domain with realistic geography. The atmospheric component has nine unevenly spaced vertical levels. The horizontal distributions of the predicted variables are represented by spherical harmonics and associated grid point values with 4.5° latitude by 7.5° longitude spacing (Gordon and Stern, 1982). Sunshine varies seasonally, but not diurnally. Cloud cover is predicted based on relative humidity. A simple land surface model computes the fluxes of heat and water (Manabe, 1969). The oceanic component

of the coupled model employs a finite-difference technique using a grid spacing of 4.5° latitude by 3.75° longitude. There are 12 unevenly spaced levels in the ocean. Subgrid scale tracer mixing occurs on density surfaces (Tziperman and Bryan, 1993). The sea ice model computes ice thickness based on thermodynamic heat balance and movement of the ice by ocean currents. The atmospheric, ocean and sea ice components exchange fluxes of heat, water and momentum 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 timescales. 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.

In the integration described here, the initial conditions are obtained from near the beginning of a 15,000 year control integration. The CO_2 concentration is reduced at a rate of 1% per year (compounded) until the concentration is one half its original value (referred to as “1/2X no acceleration” in Figure 1). For the remainder of this integration, the CO_2 concentration is held fixed at half its normal value. The integration then continues for more than 4000 model years, until the model climate reaches a statistical equilibrium with the changed radiative forcing.

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 localised 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 below.

2.4.2 GFDL low-resolution coupled model results

Figure 1 shows how the global average ocean temperature evolves in time at mid-depths in the ocean (it will be shown below that it is the mid-depths that have the longest timescale of response). At mid-depths the ocean gradually cools by about 1 K over the initial 2000 years of the GFDL control simulation (Figure 1), with a negligible cooling thereafter, indicative that the simulation has reached equilibrium.

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 2000m depth in the Northern Hemisphere ocean, and not the deep ocean (Figure 2). The long timescales occur in both the North Pacific and the North Atlantic oceans. As in the control, it takes about 2000 years for the model to reach its equilibrium state (Figure 1).

2.5 Ocean acceleration techniques

Since it is not computationally practical to integrate HadCM3 for 2000 years a method of accelerating the cooling of the ocean needs to be considered. The method developed for the LGM

experiment in this study involves a Newtonian restoring of the ocean surface heat fluxes to those representing the glacial state, referred to as Haney forcing (Haney, 1971). The heat fluxes are calculated by restoring the model’s SSTs towards a climatology of glacial SSTs obtained from the slab model (see Section 2.6). The net heat flux applied to the ocean is the sum of the heat flux from the atmosphere, $Q_{net}(T)$, and a multiple of the difference between the climatological SST (T_C) and the model predicted SST (T), i.e.,

$$Q_{net}(T) + \lambda(T_C - T) \quad (1)$$

where Q_{net} is the sum of the surface short-wave heating, surface sensible heating, surface latent heating, and surface net long-wave flux.

The size of λ , the Haney coefficient, determines how far the model can deviate from the climatological SSTs — the larger the value the tighter the constraint to the climatology. A large Haney forcing coefficient ($163.76 \text{ Wm}^{-2}\text{K}^{-1}$) is used in the coupled model experiment which corresponds to a relaxation timescale of about two weeks for a 50 m mixed layer.

The glacial SST climatology is that of a slab model simulation for the LGM (Hewitt, 2000) using the same boundary conditions as the coupled model. The method does not provide any information for, or affect, the other forcing fluxes, namely the penetrative solar heat flux, the wind mixing energy, the wind stress, and the fresh water flux. It is also possible to apply a similar methodology for prescribing the freshwater flux, but there is no global LGM salinity reconstruction available to act as a climatology, and the slab model does not provide a proxy for salinity, particularly the salinity flux associated with melting and formation of sea ice.

Two other methods for accelerating the cooling of the ocean were considered. One technique involves formulating a set of equations that have the same steady state (equilibrium) solutions as the primitive equations of the model, but these new equations have a finite difference solution that allows for a longer timestep to be used. This technique, using so-called distorted physics (Bryan, 1984), would allow for multi-century simulations of the ocean to be performed. Fast processes, such as internal gravity waves and Rossby waves, place an upper limit on the timestep permissible. In order to allow a longer timestep, an assumption is made that these fast processes are not thought to have a direct bearing on long timescale adjustments (i.e., seasonal and longer) for climate simulations and so these processes are slowed down. These techniques have not been developed for HadCM3.

The other technique uses periodically synchronous coupling (Sausen and Voss, 1996). In this method the coupled OAGCM is run for several decades, then the coupling is switched off and the (cheaper) OGCM is run for several centuries using the surface forcing (heat fluxes, wind mixing energy, wind stress, and fresh water flux) diagnosed during the previous coupled phase. The atmosphere and ocean are coupled together again and the process is repeated until a stable climate is reached. This method was not used for HadCM3 — the ocean model takes up about half of the computing time because of the relatively high resolution of the ocean compared to the atmosphere. If several iterations of the periodic synchronous coupling were needed to reach a stable climate the computing time needed would be prohibitively large, and there is

no guarantee that once the periodic synchronous coupling stage was finished the coupled model would stay close to that stable climate state, which would then necessitate a further long coupled adjustment period. However, a coupled model that includes a cheap ocean model (relative to the atmosphere model) might be able to successfully use this method to allow the model to approach equilibrium (see Jones and Palmer, 1998 for an example of this using HadAM3 coupled to a low-resolution version of HadOM3).

2.6 Coupled model spin-up

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 a Haney restoring term for SSTs, with the climatological reference SSTs taken from the slab model run in the previous stage. The other forcing fluxes, namely the penetrative solar heat flux, the wind mixing energy, the wind stress, and the fresh water flux, are calculated by the atmosphere model. The ocean model is initialised from the Levitus climatology of temperature and salinity for the present day (Levitus and Boyer, 1994; Levitus et al, 1994). Salinity is increased uniformly to preserve the total volume integrated salinity. The strong relaxation towards the slab model's SSTs means the sea-ice area is closely constrained to the slab model's glacial sea ice simulation. This Haney forced configuration of the coupled 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 initialised from the end of the Haney forced HadCM3H stage and has been run for 1000 years without any Haney restoring, i.e., the model computes all of the surface fluxes as in the HadCM3 control simulation.

A similar experimental methodology is applied to the GFDL low-resolution coupled model with atmospheric CO₂ concentrations halved. The model is run for 70 years with a Haney restoring term for SSTs based on a slab model simulation with CO₂ concentrations halved. The model is then run for a further 1000 years without any Haney restoring. This run is referred to as "1/2X acceleration" in Figure 1.

3 Haney acceleration run

The results from the Haney forced stage of the spinup do not necessarily have much physical basis and so we only include a brief description of the results here. However, the response of the NATHC displays some interesting transient behaviour and is described in some detail in

The HadCM3 control simulation produces a stable and realistic simulation of SST (Gordon et al, 2000). The global volume weighted temperature of the ocean in the control simulation for present day is reasonably stable. On very long timescales there is a small cooling trend of about 0.02°C per century (Figure 3) consistent with the small net heat loss at the top of the atmosphere (TOA) (Figure 4b). The average sea ice depths and total sea ice area are also reasonably stable after an initial adjustment over the first century of the control simulation (Figure 4c–f).

Once the Haney forcing is applied, the model’s SSTs are restored to the colder glacial values. The model then produces a positive net radiative flux at the TOA, with a global average of 2.3 Wm^{-2} over the 70 year run, which would seem to imply that the climate system is warming. However, the Haney forcing introduces an anomalous heat flux at the ocean surface which depends on the difference between the model’s SST and the climatological SST (see Equation 1). These climatological SSTs come from the slab model which has a prescribed corrective heat flux to account for the lack of dynamics in the slab model. In the slab model the global average annual mean TOA radiative flux in the control is 2.6 Wm^{-2} . This is fairly well balanced by the corrective heat flux which has a globally average annual mean value of -2.7 Wm^{-2} . The Haney forced run is therefore being largely forced by the slab model LGM SSTs which are associated with a TOA radiative imbalance of 2.6 Wm^{-2} . The relaxation to the slab model SSTs in the Haney forced run allows the imbalance to be reduced.

The large relaxation coefficient used in the Haney forced run constrains the SSTs to be close to those of the slab model LGM SSTs. Most of the surface of the ocean cools to within 1 K of the forcing SSTs within the first year (Figure 5). The SSTs in the North Atlantic differ most compared to the restoring SSTs, partly because these are the regions where the slab model SSTs differ most to Levitus, and partly because of the large horizontal and vertical heat transports and strong vertical mixing in this region. The upper ocean also responds quickly, with much of the cooling occurring in the first few years (Figure 6a). Progressively deeper down in the ocean the response becomes less rapid (Figures 6b–d).

4 Approach to equilibrium

Once the Haney forcing is “switched off”, i.e., the SSTs of the model are not restored to climatological SSTs, the model may still exhibit noticeable trends (for example in temperature, salinity, and sea ice amounts), and so a further stage is necessary to allow the coupled model to continue to adjust to the LGM boundary conditions. Figures 3, 4 and 6 clearly show that the ocean is not at equilibrium by the end of the Haney forced stage, and that the model does continue to adjust to the glacial boundary conditions after the Haney forced stage.

The Haney forced run 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 ocean dynamics and the response of the ocean below the mixed layer. It was not possible at the outset to predict whether the coupled model would be generally

warmer or colder than the slab model at equilibrium at the surface before embarking on the coupled model experiment, and so the point at which to stop the Haney forcing was not well-defined.

Figure 1 suggests that without the initial Haney forced 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 Haney forced stage. Since the LGM experiment uses a different climate forcing and a different model we cannot quantify the acceleration the Haney forced 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 Haney forced LGM experiment has a similar, if not larger, acceleration towards equilibrium.

4.1 Global average trends

Once the Haney forcing is switched off the global average surface air temperature rapidly increases by a few-tenths of a degree and then gradually cools throughout the rest of the HadCM3 coupled model LGM run (Figure 4). The average sea ice depth and areal coverage change markedly over the first few centuries of the adjustment phase. In the Southern Hemisphere the sea ice depth and coverage rapidly increase, and in the Northern Hemisphere the sea ice depth and coverage tend to decrease. By the end of the run the sea ice shows signs of having stabilised.

The global average radiative imbalance at the TOA rapidly reduces to about -0.5 Wm^{-2} in the first few hundred years and then gradually decreases throughout the rest of the run. The negative radiative imbalance is indicative of a cooling, and this is consistent with the gradual cooling seen in the volume average temperature of the ocean (Figure 3). The global average ocean temperature in the mixed layer gradually cools throughout the first part of the adjustment phase and shows signs of stabilising by the end of the run (Figure 6a). Deeper down, for example at 447 m (Figure 6b), the global average ocean temperature initially warms for about a century relative to the Haney forced stage, and then gradually cools and shows signs of stabilising by the end of the run. In the deep ocean (Figures 6c and d) the model is still slowly cooling by the end of the run. At abyssal depths the small rate of cooling is similar to that in the control.

4.2 Temperature trends in the ocean basins

Although the ocean generally cools during the adjustment phase when considered globally, the individual ocean basins do not follow this simple cooling trend (Figure 7).

The Atlantic Ocean and the Southern Ocean become very cold throughout much of their depths during the Haney forced run due to the cold surface forcing which is convectively mixed vertically down (Figure 7a and b). Once the Haney forcing is switched off, most of the Atlantic and Southern Oceans do not maintain this cold state and initially warm (Figure 7). Over subsequent centuries the warming decreases and eventually much of the Atlantic Ocean cools, particularly at depth as Antarctic Bottom Water penetrates into the basin.

The Arctic Ocean initially has a cold, i.e. sub-zero, fresh surface layer, about 200 m thick, due to sea-ice processes. Below the surface layer, relatively warm and salty water flows into the Arctic Basin from the North Atlantic, occupying a layer about 600 m thick. At low temperatures the low salinity surface water has a low density and so floats on top of the North Atlantic water, forming a halocline. Below the level where the North Atlantic supplies warm water, the deep Arctic water is cold. Once the Haney forcing is applied the surface waters become more dense, due to a combination of colder SSTs and an increase in surface salinity. The surface salinity increases are largely due to a reduction in the amount of water entering the Arctic Ocean from rivers and a reduction in summer sea-ice melt. The denser surface waters break down the halocline and the upper ocean cools (Figure 7a). However, as we shall see in Section 5, a further consequence of the Haney forcing is that the North Atlantic thermohaline circulation becomes much stronger than at present day and more heat is transported to high latitudes. This produces a warming of the cold deep Arctic waters, and this warming persists, but becomes less pronounced, throughout the adjustment phase which also has a strong thermohaline circulation. The upper Arctic ocean waters also warm throughout the adjustment phase once the Haney forcing is switched off.

The Pacific Ocean (Figure 7) and Indian Ocean (not shown) do not cool as much as the Atlantic and Southern Oceans in the Haney forced run, partly because the SST forcing is not as large over these ocean basins. Once the Haney forcing is switched off, the Pacific and Indian ocean basins continue to cool during the adjustment phase, which dominates the global average response described above. By the end of the simulation the cooling trends are almost imperceptible, particularly in the deep ocean (Figure 7l). As with the Atlantic, the upper ocean in the Northern Pacific warms during the adjustment phase because the coupled model does not produce SSTs as cold as the slab model.

The atmosphere responds quickly to changes at the ocean surface. However, the gradual changes deeper in the ocean influence the changes at the ocean surface. For example, the SSTs in the tropical East Pacific gradually cool, by a degree or two, over the first four centuries of the HadCM3 LGM simulation (Figure 8), after which time the cooling becomes negligible. Therefore, the response of the atmosphere, for example the surface winds, may also adjust over the first few centuries of the simulation, and one must be cautious in performing and interpreting short experiments that have large drifts in the deep ocean.

5 North Atlantic Thermohaline Circulation

The ocean is generally stably stratified with less dense water overlying denser water, which inhibits convection, but convective instabilities do occur in the ocean. The ocean's thermohaline circulation is driven by horizontal density gradients, initiated by vertical density gradients resulting from deep convection at a few sites in the North Atlantic Ocean and Southern Ocean. Convection can be forced by an increase in a water parcel's density due to heat loss or increased salinity, making the parcel more negatively buoyant. There are only a few regions in the ocean

where the surface conditions are extreme enough for convection to be strong enough to penetrate down to the deep ocean, and these conditions are intermittent. The thermohaline circulation is strongly linked to these sites of deep convection, which occur in the Labrador Sea, the Nordic Seas (the Greenland, Iceland and Norwegian Seas), the Ross Sea, and the Weddell Sea.

The buoyancy flux B at the ocean surface, as used in the ocean model, is given by Gill (1982) as

$$B = \alpha C_W^{-1} g Q + \beta s g W \quad (2)$$

where Q is the heat flux into the ocean surface, W is the freshwater flux into the ocean surface, s is the surface salinity, α and β and the thermal and salinity expansion coefficients of sea water at the surface respectively given by $\alpha = \left(-\frac{1}{\rho} \frac{\partial \rho}{\partial T}\right)$, $\beta = \left(\frac{1}{\rho} \frac{\partial \rho}{\partial s}\right)$, T is the surface temperature and ρ is the surface density.

The freshwater flux W is the balance between precipitation P , evaporation E , coastal runoff from rivers R , and the salinity change from ice melting or forming or subliming. Note that the derivative of density with respect to temperature $\frac{\partial \rho}{\partial T}$ is generally negative which accounts for the minus sign in the thermal expansion coefficient.

In the model the heat flux Q is calculated from a balance of several terms. The dominant terms are, in order of decreasing magnitude, the Haney forcing term in the Haney forced segment (F , which is the anomalous heat flux applied at the ocean surface, $\lambda(T_C - T)$ in Equation 1), the sum of all the heat fluxes from the atmosphere (HTN), and the solar radiation that penetrates through the ocean surface (PEN). The remaining terms account for heat fluxes from sea ice processes and snow falling into open water, and are relatively small on the global scale.

In the mid- and high-latitudes the buoyancy flux is generally dominated by the heat flux component (Figure 9 shows the North Atlantic) but the freshwater flux can dominate in some coastal regions where rivers produce a significant input of fresh water and in regions where sea ice melts or forms. In the Tropics and extra-Tropics the freshwater flux is important, especially in the low-latitude North Atlantic Ocean where the high evaporation causes the surface waters to become more salty.

In the regions where there is a large negative buoyancy flux and deep water is formed, the relatively high density of the deep water creates a horizontal pressure (density) gradient which produces equatorward flow, as a deep western boundary current, whose position is influenced by the bottom topography. This deep ocean pressure gradient is an important factor in driving the whole thermohaline-induced overturning circulation (Stommel, 1961) — the equatorward flow produces a downward flow at high latitudes by virtue of mass conservation, which then similarly produces a poleward surface flow of warm salty subtropical water. To complete the circulation there is slow upwelling in the interior of the ocean. The surface water's high salinity helps to keep the deep water's density relatively high. The North Atlantic deep water is made up from a combination of deeply convected water in the Labrador Sea and from dense water from convection in the Nordic Seas that overflows across the Greenland-Iceland-Scotland (GIS) ridge.

5.1 Haney acceleration run

The HadCM3 control simulation produces a meridional overturning cell in the upper North Atlantic (the NATHC) with deep water formation at high northern latitudes (Figure 10a). The maximum strength of the overturning in the NATHC is about 19 Sv, 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 Haney forcing produces a strong overturning cell characterised by a greater formation of very dense deep water which fills much of the Atlantic basin, preventing AABW from penetrating into the Atlantic basin. By the end of the Haney forced stage (Figure 10b) the maximum strength of the overturning cell is about 50 Sv, at a depth of about 1500 m around 30°N, and still extends down to the deep ocean, reducing the intrusion of AABW that is seen in the control. This result is strikingly similar to that found by Bigg et al (1998) who forced an ocean GCM with atmospheric conditions taken from the LGM palaeoclimate simulation of Hall et al (1996).

While the maximum strength of the NATHC is fairly constant in the control, there is considerable variability in the LGM simulations. Once the Haney forcing is applied to the model the maximum overturning rapidly increases to about 70 Sv in the first decade and then slowly decreases to about 50 Sv over the next six decades (Figure 10c). Once the Haney forcing is switched off the maximum overturning rapidly decreases again, as will be discussed below.

An important process for driving the overturning circulation is the north-south density gradient in the deep ocean (as discussed above). As a diagnostic tool we shall consider the density gradient ΔD between a region in the South Atlantic D_S and a region in the North Atlantic D_N , with the deep flow from north to south being caused by the lower density in the South Atlantic deep ocean compared to the North Atlantic deep ocean. It has been shown that the strength of the NATHC in HadCM3 is proportional to such a density gradient (Thorpe et al, 2001).

The diagnostic ΔD in the following discussion is calculated at the model level below the level of maximum overturning. D_N is taken as the average density of four model rows in the North Atlantic Ocean (from 60°W – 0°, and 58.125°N – 61.875°N) at a depth of 1501m in the control simulation and 2116m in the HadCM3H experiment. D_S is the average density of four model rows in the South Atlantic Ocean (from 60°W – 5°E, and 31.875°S – 28.125°S) at the same depths as for D_N .

Rapid, i.e. on timescales of less than a decade, changes in this density gradient ΔD in the Haney forced run correlate very well with changes in the strength of the NATHC (Figures 11a and b). There are two distinct stages to the transient response of the NATHC in the Haney forced run. Firstly the deep ocean density gradient between the North and South Atlantic more than doubles over the initial few years of the Haney forced run, and the overturning circulation strengthens in response to this. The density gradient then gradually weakens and consequently, after a few years lag, the overturning circulation weakens. Note that we are concerned with *changes* to the density gradient in this discussion — the actual density gradient in the control is not directly comparable to that of the Haney forced run because the densities are calculated at

different depths.

The initial driving force for the changes to the density gradient in the Haney forced run are changes to the density of the North Atlantic (D_N , Figure 11d) which is forced by changes in the surface buoyancy flux (Figure 11c). However, it is important to realise that D_S is not independent of D_N . Density anomalies in the North Atlantic will be advected to the South Atlantic (over decadal timescales by the deep western boundary currents driven by the NATHC) since the density of the North Atlantic is higher than the South Atlantic D_S . This will generally increase the salinity of the South Atlantic in time (Figure 11d) which acts as a negative feedback on the system, subsequently weakening the density gradient and gradually reducing the strength of the NATHC.

Figure 11e shows that the freshwater flux W does not vary greatly in the Haney forced run (nor for that matter in the control run). It is the heat flux Q which produces the large negative buoyancy flux in the Haney forced run, and the rapid transient changes in the heat flux are responsible for the transient changes in the buoyancy flux, and hence ultimately changes to the strength of the NATHC.

Figure 11f shows that the initial large transient changes in the Haney forcing term are responsible for the rapid transient changes in the heat flux. The Haney forcing is very large at the start of the run because the Levitus SSTs in the North Atlantic are very different to the slab model LGM SSTs, which leads to a large negative heat flux in this region ($T_C - T < 0$ in Equation 1) producing a cooling. As the model cools (Figure 11g), the difference between the model SSTs and the slab model LGM SSTs decreases and so therefore does the Haney forcing term, and in turn so does the negative heat flux, the negative buoyancy flux and the strength of the NATHC.

After about three years, once the SSTs have decreased to be close to the forcing SSTs, the above factors become relatively weak and the North Atlantic density stops increasing. However, as already stated, advection of this high density water to the South Atlantic increases the density of the deep ocean in the South Atlantic which slowly weakens the meridional density gradient and this weakens the NATHC.

5.2 Approach to equilibrium

We can test the above conclusions concerning the effect of Haney forcing on the changes to the NATHC by analysing the response of the model once the Haney forcing is switched off.

Once the Haney forcing is switched off the initial transient nature of the strength of the NATHC also has two distinct stages (Figure 12a). In the first decade the overturning cell rapidly weakens from about 50 Sv to 22 Sv in response to a reduction in the density gradient between the North and South Atlantic. The density gradient subsequently gradually increases over the following two decades and the strength of the overturning cell also slowly increases, to about 27 Sv. Again, it is *changes* to the density gradient that are producing changes to the strength of the NATHC — the actual density gradient in the control is not directly comparable to that of the LGM run because the density of the Atlantic in the two simulations differs due to different

temperature and salinity profiles.

The rapid weakening of the density gradient in the first few years of the LGM run is due to a large decrease in the density of the North Atlantic (Figure 12d). Once the Haney forcing is switched off the model does not maintain the very cold North Atlantic SSTs and the sea surface starts to warm (Figure 12g). The Haney forcing produced a large negative buoyancy flux in the Haney forced run, and without this additional negative buoyancy flux the surface waters become less negatively buoyant (Figure 12e), and the relatively high density of the deep ocean in the North Atlantic at the end of the Haney forced run is not maintained.

The gradual strengthening of the NATHC over the following decades is due to the gradual increase in the density gradient (Figure 12b), driven by a combination of a gradual decrease in the density of the South Atlantic deep ocean and a slight increase in the density of the North Atlantic deep ocean (Figure 12d). As stated earlier, the density of the deep ocean in the South Atlantic is to some extent dependent on the density of the deep ocean in the North Atlantic, due to advection, and so as the density of the North Atlantic decreases from the relatively high values at the end of the Haney forced run, so does the density of the South Atlantic. The North Atlantic density subsequently increases slightly in response to an increase in the negative surface buoyancy flux (Figure 12e), driven by increased sensible heat loss to the cold atmosphere as the SSTs warm (Figure 12f).

6 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 (2001). The purpose of the current paper is to describe the experimental design employed to bring the model close to equilibrium, and to describe some of the mechanisms responsible for the transient response of the coupled model. A detailed evaluation of the LGM response is beyond the scope of this paper.

The coupled model is initialised from present day conditions since there is no three-dimensional global data set of glacial conditions available to use. 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 timescales 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 back to glacial SSTs, as determined from a slab model LGM experiment, for 70 years. This Haney forced stage is not run to equilibrium because the slab model LGM SSTs, are not expected to be the same as the equilibrium coupled model LGM SSTs.

In the second stage of the experiment the Haney forcing is “switched off”. The atmosphere responds quickly, i.e. within a decade, to the glacial boundary conditions. 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 also 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. However, we feel that one should be cautious of performing and interpreting short experiments that have large drifts in the deep ocean. We have shown that the large initial drifts in the deep ocean affect the response at the ocean surface. 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 timescales 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 Haney forcing used in the LGM experimental design produced an enhanced cooling that would take several hundred years longer without the initial Haney forced stage, representing a considerable saving in computer time.

The model displays large trends once the glacial boundary conditions are applied, particularly in the structure of the thermohaline circulation. Changes to the density gradient between the North Atlantic and South Atlantic produce changes to the strength of the North Atlantic thermohaline circulation. In the Haney forced run the relaxation towards very cold SSTs in the high latitudes of the North Atlantic produces a large negative buoyancy flux and very dense waters in the North Atlantic. This leads to a continual strengthening of the NATHC throughout the first decade of the run, with a maximum overturning of about 70 Sv compared to about 20 Sv in the control. However, two factors weaken the NATHC over the following decades. Firstly, the difference between the model’s SSTs and the restoring SSTs decreases and so less heat is extracted from the ocean which reduces the negative buoyancy flux. Secondly, advection of the dense water from the North Atlantic to the South Atlantic weakens the density gradient. The combination of these factors gradually weakens the maximum strength of the NATHC to about 50 Sv by the end of the Haney forced stage.

Once the Haney forcing is switched off the coupled model does not maintain the cold North Atlantic SSTs and the sea surface starts to warm. This reduces the density of the North Atlantic and the NATHC rapidly weakens as the density gradient weakens, with a maximum overturning of about 20 Sv. Over the subsequent decades the NATHC gradually strengthens again to about 27 Sv. The strengthening is predominantly due to an increase in the density of the North Atlantic waters due to heat loss from the ocean to the cold atmosphere.

Once the NATHC has strengthened, feedbacks maintain this stronger overturning cell over the following centuries. The stronger overturning cell transports warm, salty water from the Tropics to the high latitude North Atlantic. The overlying air masses are very cold, originating from the surrounding continental ice sheets, and so the ocean loses large amounts of heat to the atmosphere, producing a negative buoyancy flux. This negative buoyancy flux combined with the relatively saline water is what maintains the strong overturning cell in the LGM experiment, which in turn maintains the supply of warm, salty water (Hewitt et al, 2001).

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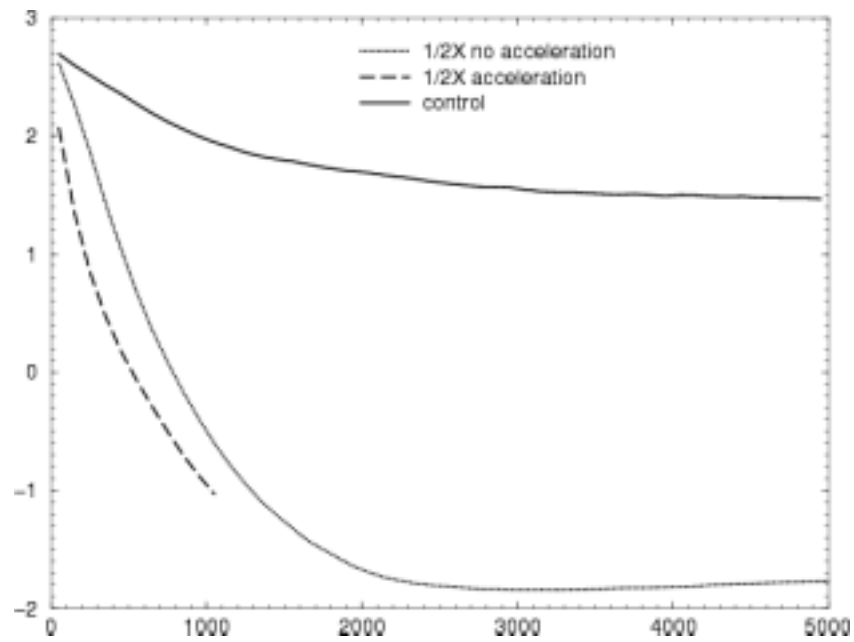


Figure 1: Timeseries of global average ocean temperature at mid-depth (2228 m) from the GFDL runs.

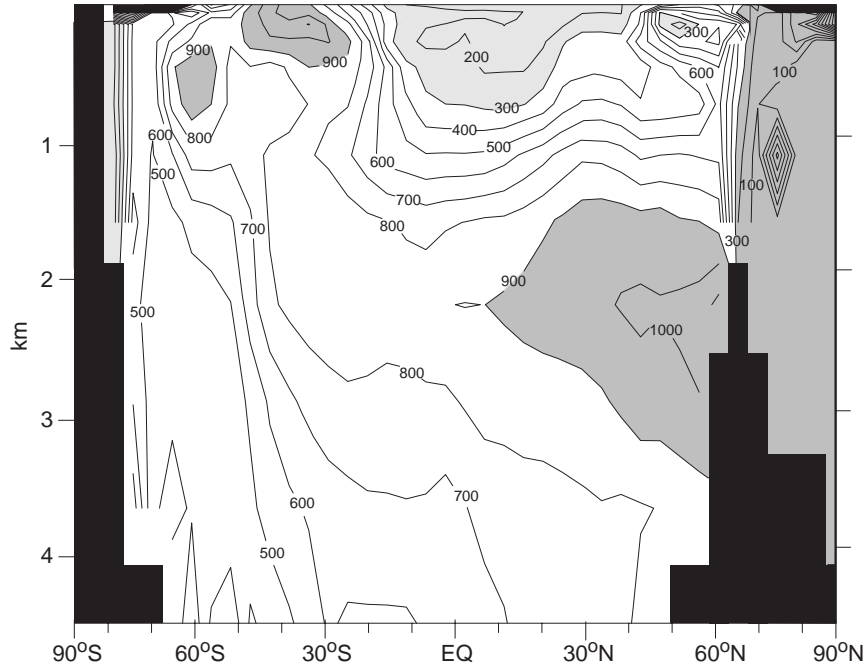


Figure 2: The time (years) to reach 70% of the total, equilibrium response from the beginning of the integration (the first occurrence). The fraction of the total response is computed by dividing the response at any given time by the total response at each grid location. The zonal mean, latitude versus depth response time scales are shown.

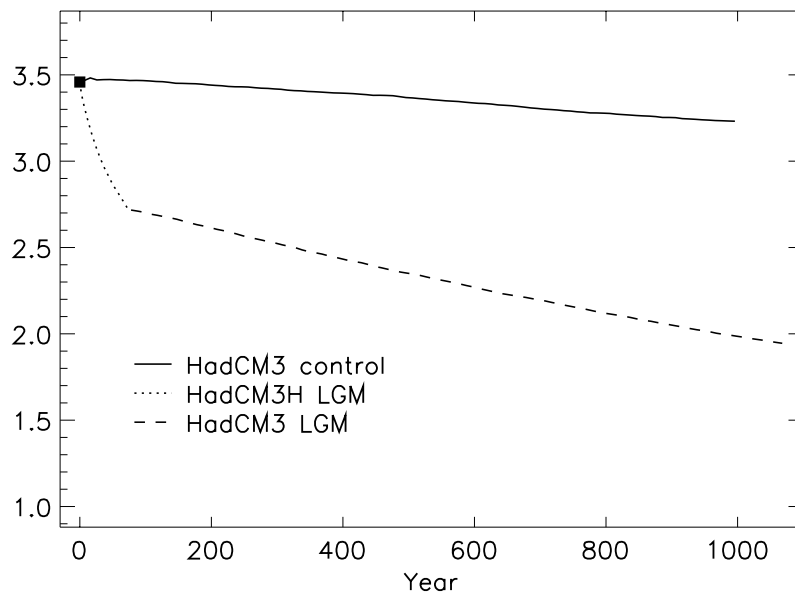


Figure 3: Timeseries of volume averaged ocean temperature, in °C. The black square denotes the initial condition.

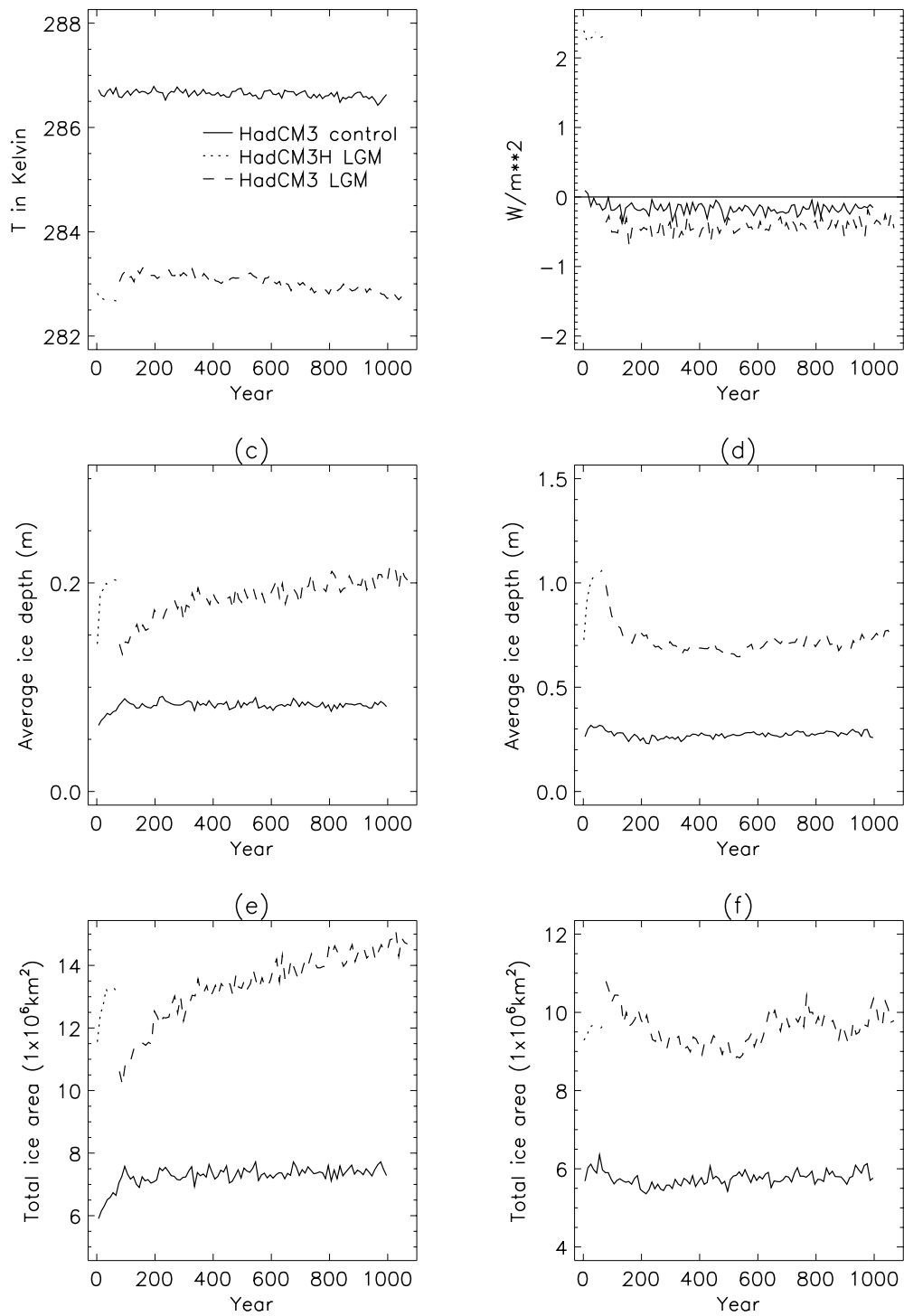


Figure 4: Timeseries of decadal average quantities simulated by the coupled model. (a) Global mean surface air temperature. (b) Global mean net downward radiation at the TOA. (c) Southern Hemisphere sea ice depth. (d) N. Hem. sea ice depth. (e) S. Hem. total ice area. (f) N. Hem. total ice area.

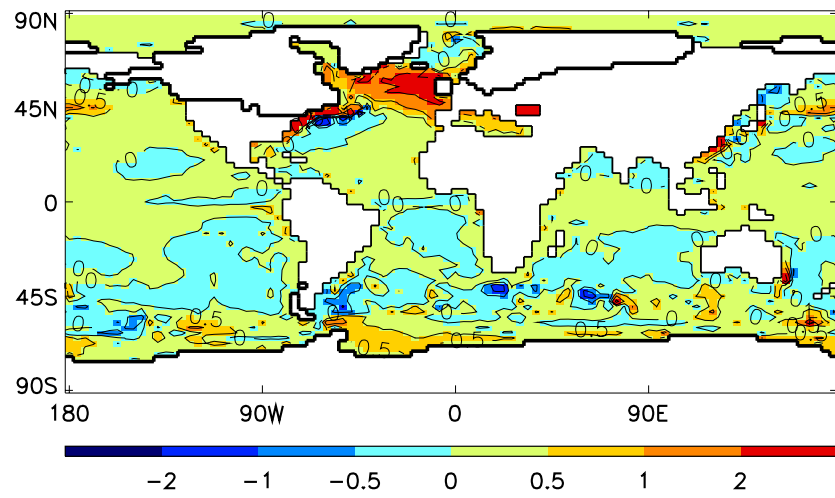


Figure 5: Difference between the ocean surface temperature from the first year of the Haney forced run and the annual average restoring SSTs taken from the slab model LGM run, in K. The LGM ice sheets are marked by the thick black lines.

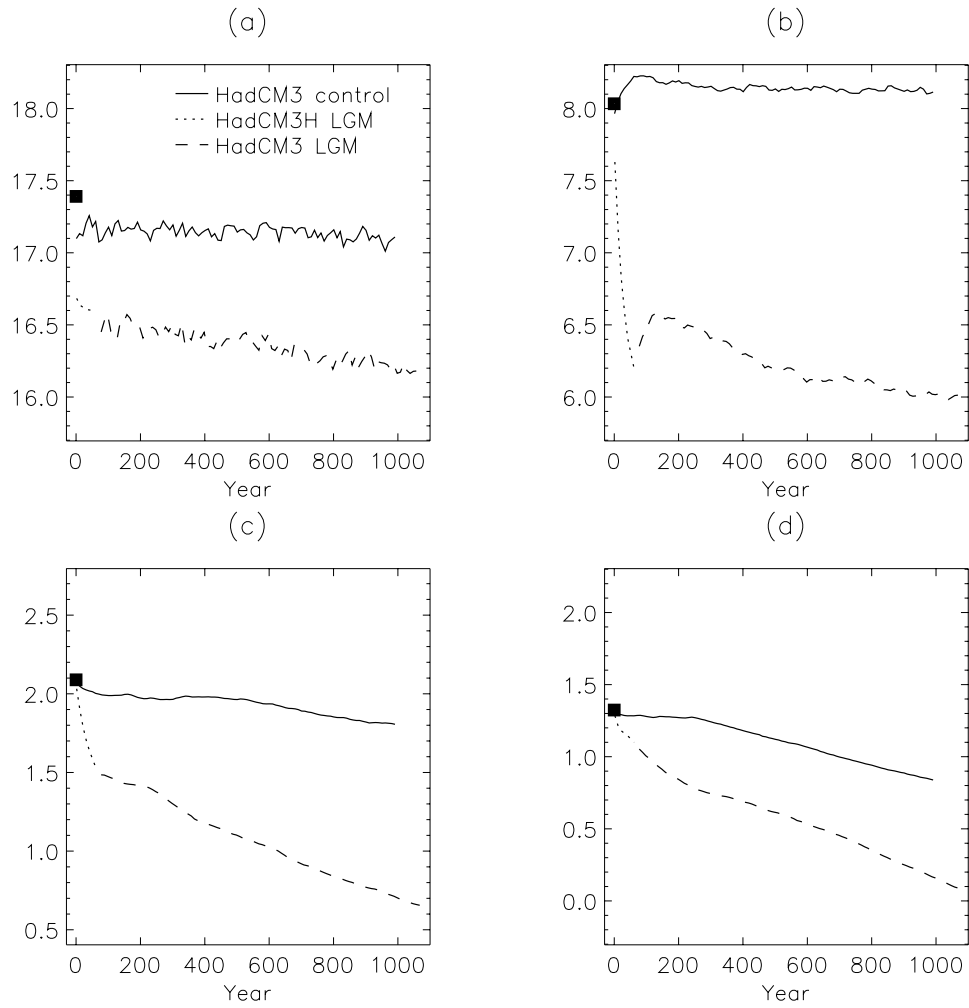


Figure 6: Timeseries of the global average ocean temperature, in $^{\circ}\text{C}$, at certain depths. The black square denotes the initial value used for the LGM runs. (a) 48 m. (b) 447 m. (c) 2116 m. (d) 3347 m.

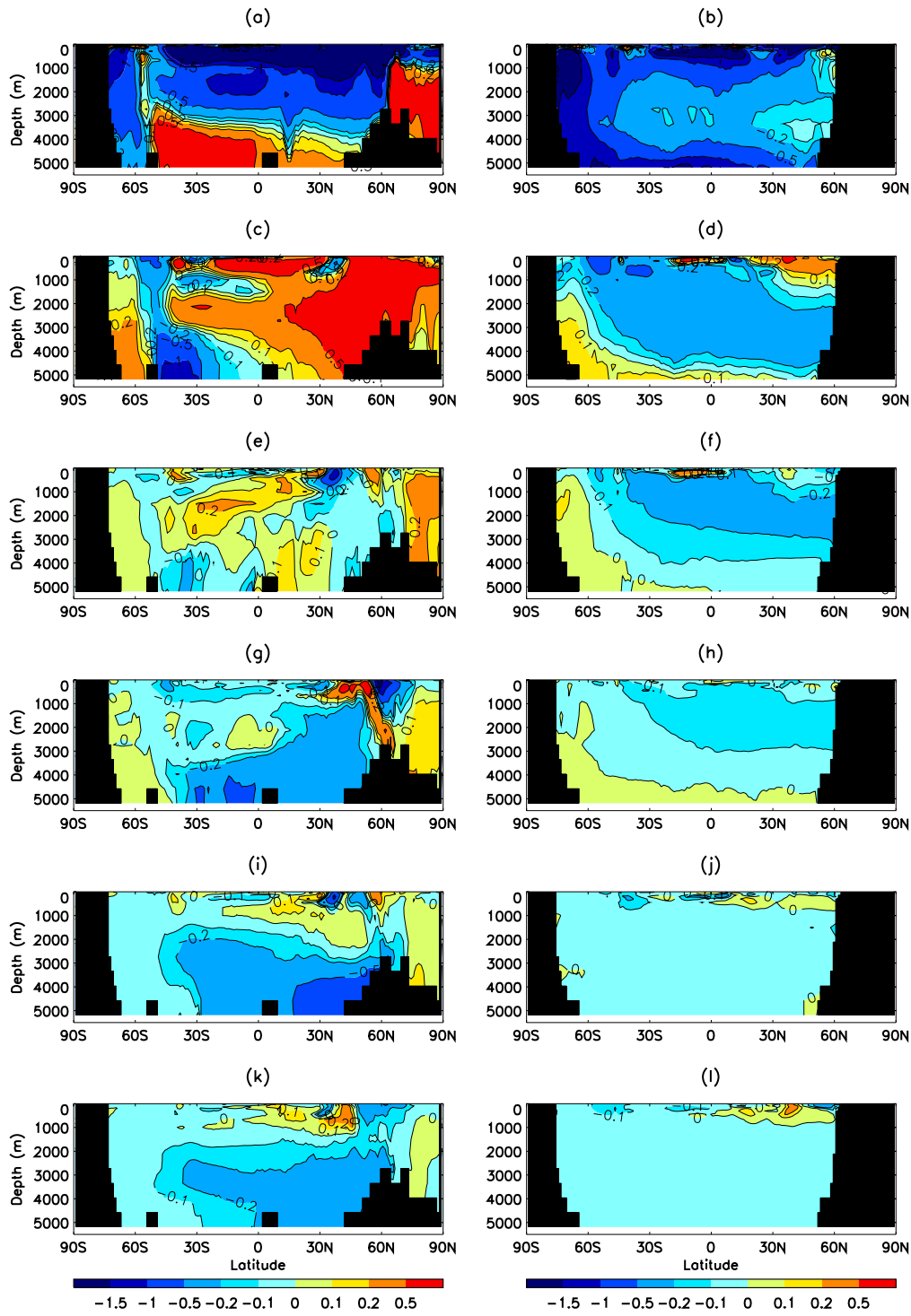


Figure 7: Change in ocean temperature, in K, zonally averaged over the Atlantic and Arctic Basins (left column) and the Pacific Basin (right column). The top row shows the last decade of the Haney forced run – climatology (Levitus 1994 annual mean). Subsequent rows show the drift over consecutive centuries of the LGM run starting with (c) and (d) showing 2nd Century (C) – the 1st C of the LGM run, (e) and (f) show 4th C – 3rd C. (g) and (h) show 6th C – 5th C. (i) and (j) show 8th C – 7th C. (k) and (l) show 10th C – 9th C.

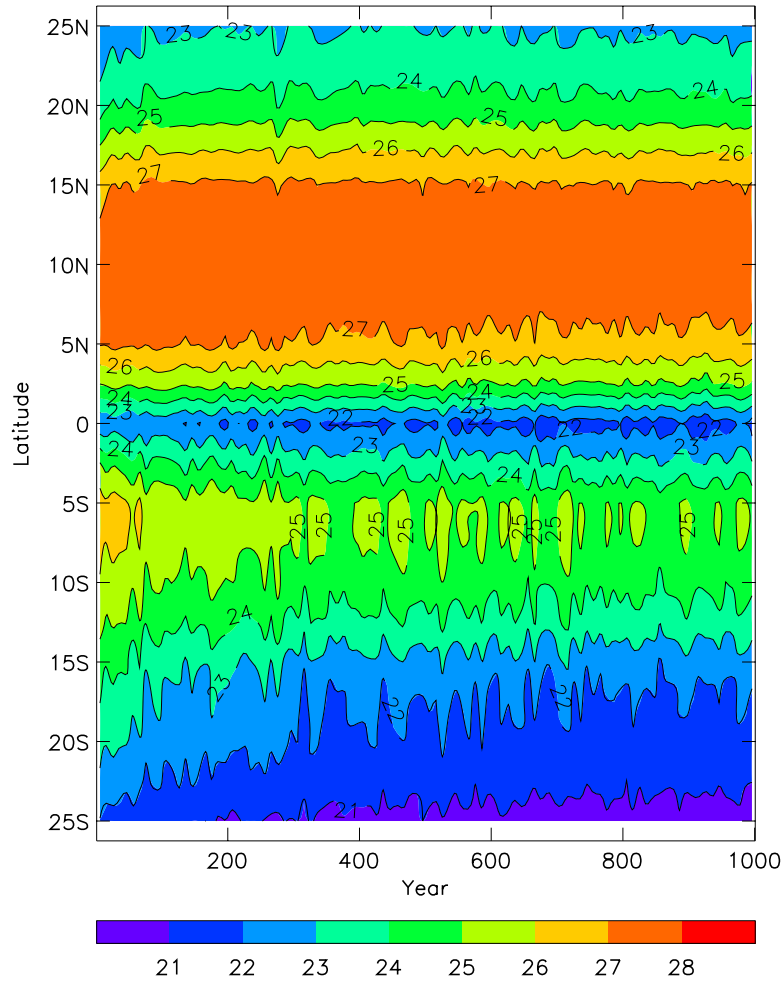


Figure 8: Time versus latitude plot of SST from the HadCM3 LGM run averaged between 90°W and 120°W for the latitude belt 25°S to 25°N. Contours every °C.

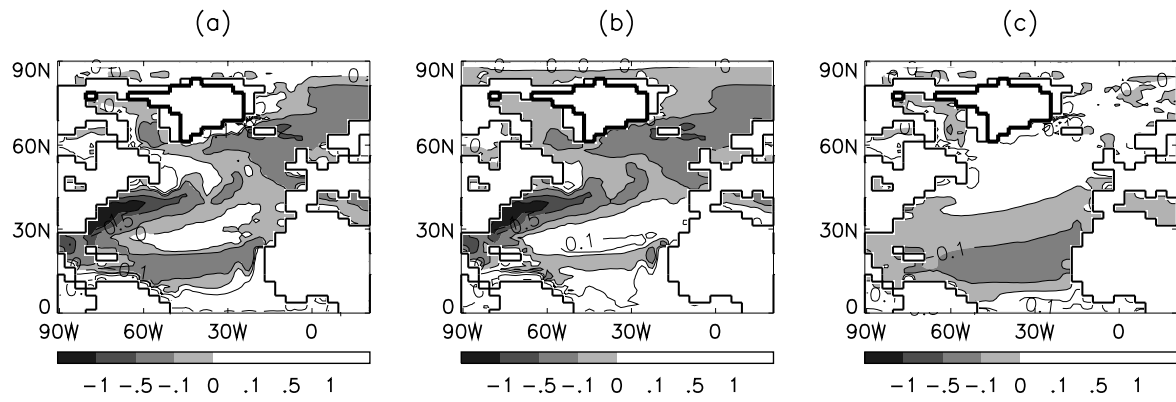


Figure 9: Surface buoyancy flux in the North Atlantic from the HadCM3 control simulation, in $10^{-4} \text{ kg m}^{-1} \text{ s}^{-3}$. (a) Total buoyancy flux. (b) Surface heat flux component. (c) Freshwater flux component.

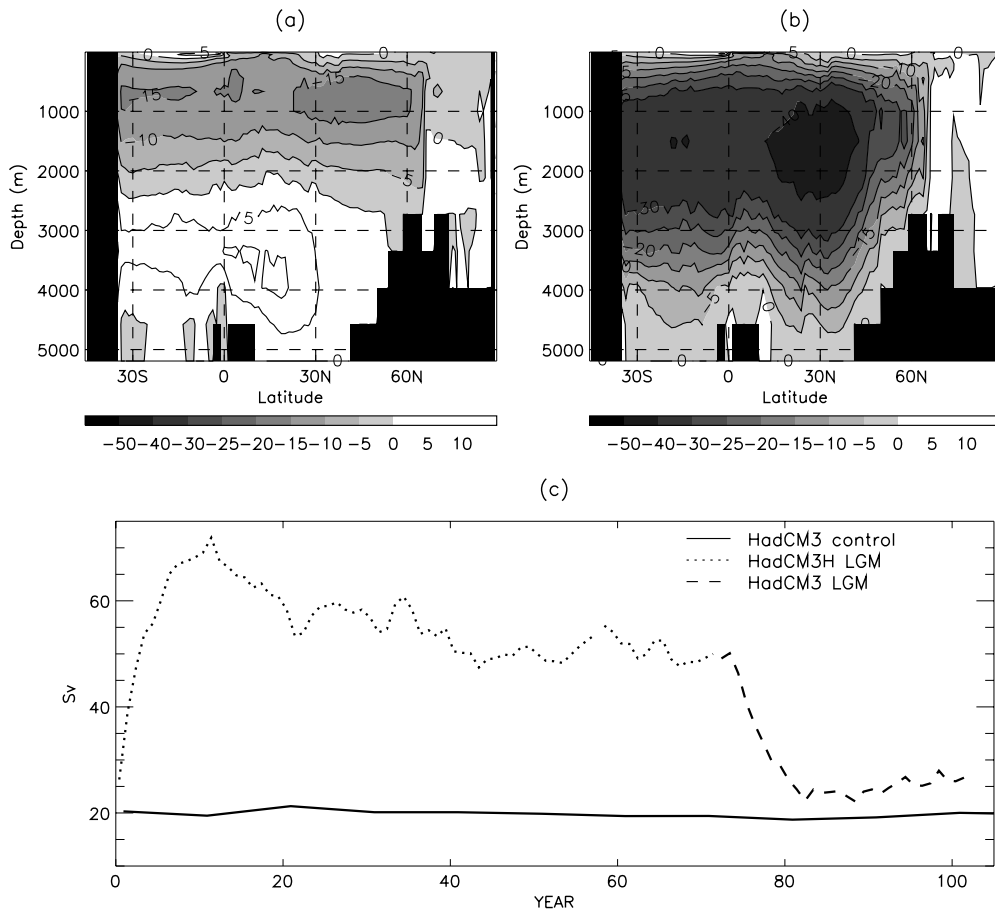


Figure 10: North Atlantic meridional overturning streamfunction, in Sv. Negative values denote clockwise circulation in the plane shown on the figure. (a) Depth-latitude cross-section for HadCM3 control. (b) Depth-latitude cross-section for final decade of HadCM3H experiment. (c) Timeseries of the magnitude of the maximum overturning. Only the first 30 years of the HadCM3 LGM run are shown.

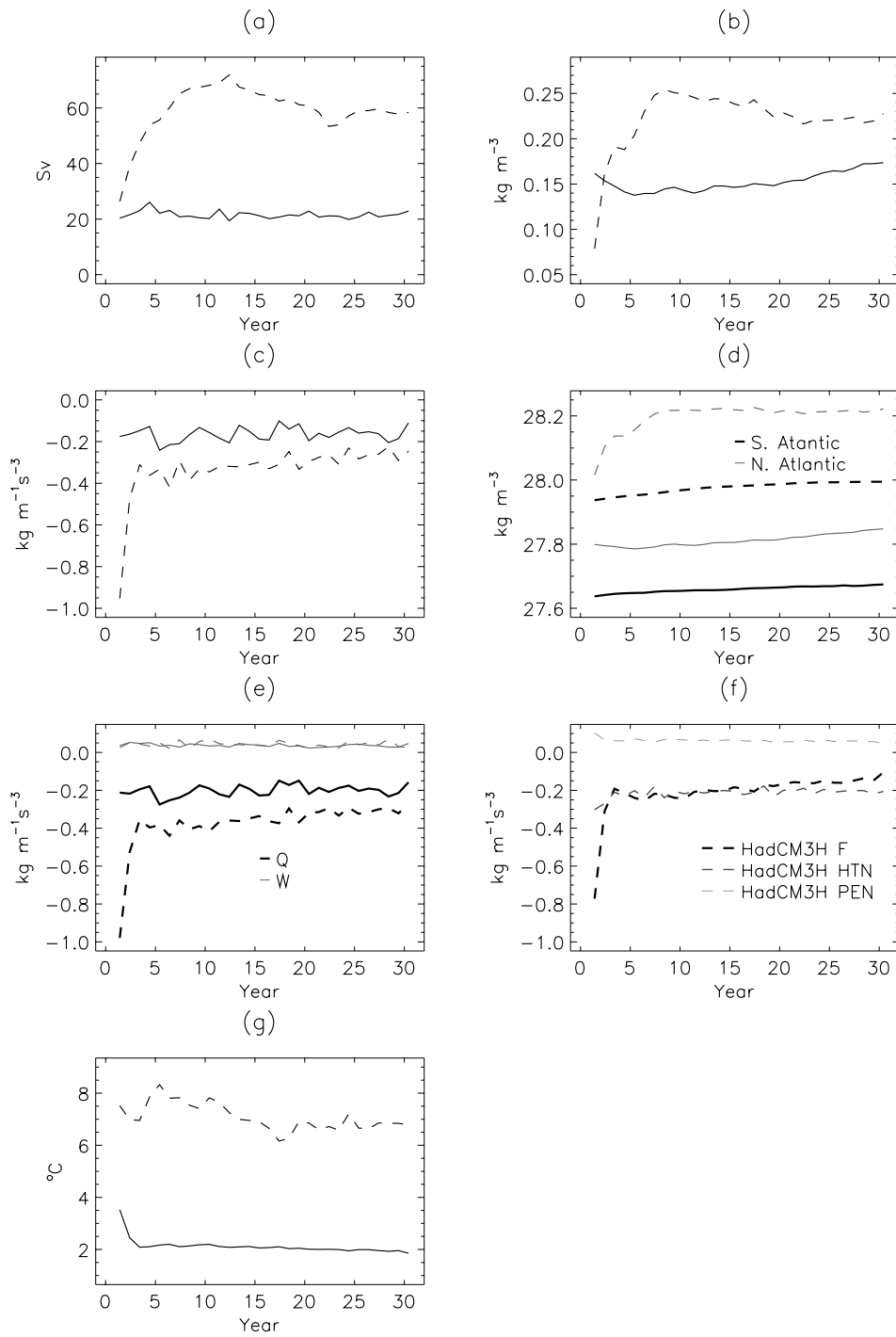


Figure 11: Timeseries of annual means from the first 30 years of the HadCM3 control run (solid lines throughout), and the first 30 years from the Haney forced run (dashed lines throughout). (a) Maximum strength of the North Atlantic overturning streamfunction. (b) North-south density gradient ΔD (See text). (c) N. Atlantic surface buoyancy flux. (d) Deep ocean density. (e) Heat flux Q and freshwater flux W components of the N. Atlantic surface buoyancy flux. (f) Main components of the heat flux term for the HadCM3H experiment. (g) N. Atlantic SST. See text for more details.

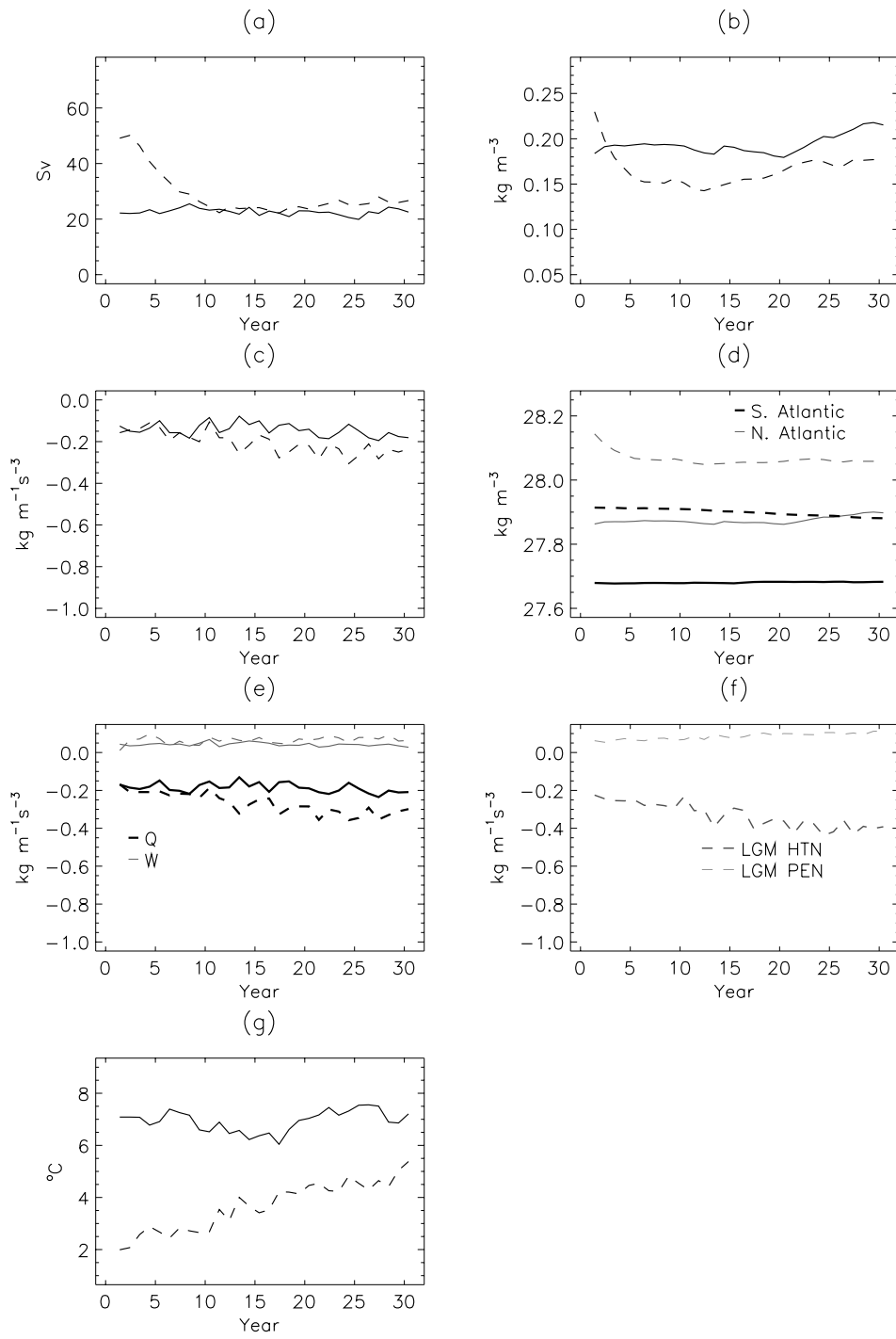


Figure 12: Timeseries of annual means from the first 30 years of the HadCM3 LGM run (dashed lines throughout), and years 71 to 100 from the HadCM3 control run (solid lines throughout). (a) Maximum strength of the North Atlantic overturning streamfunction. (b) North-south density gradient ΔD . (c) N. Atlantic surface buoyancy flux. (d) Deep ocean density. (e) Heat flux Q and freshwater flux W components of the N. Atlantic surface buoyancy flux. (f) Main components of the heat flux term for the HadCM3H experiment. (g) N. Atlantic SST. See text for more details.