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A coupled model study of the last glacial maximum: Was part of the North Atlantic relatively warm?

Chris D. Hewitt[1], Anthony J. Broccoli[2], John F. B. Mitchell[1], Ronald J. Stouffer[2]

Abstract. A coupled ocean-atmosphere general circulation model is used to simulate the climates of today and the last glacial maximum (LGM). The model, which does not require artificial flux adjustments, produces a pattern of cooling at the LGM that is broadly consistent with the findings from simpler models and palaeoclimatic data. However, changes to the ocean circulation produce anomalously warm LGM surface conditions over parts of the North Atlantic, seemingly at odds with palaeoceanographic data. The thermohaline circulation is intensified for several centuries, as is the northward heat transport in the Atlantic equatorward of 55°N, but this may be a transient result. Mechanisms that lead to this response are discussed.

Introduction

General circulation models (GCMs) of the climate system are powerful tools for understanding and predicting climate and climate change on regional and global scales. Reconstructions of the last glacial maximum (LGM) climate of 21 ka from palaeoclimatic data provide an opportunity for an extreme test of a GCM's ability to simulate a change of climate, and allow us to increase our understanding of mechanisms of climate change.

Most earlier attempts to simulate the climate at the LGM have stopped short of including three-dimensional (3-D) representations of both the atmosphere and ocean. The earliest attempts used atmospheric GCMs with reconstructed glacial sea surface temperatures (SSTs) prescribed at their lower boundary [*Gates,* 1976]. Later, the ocean was represented by thermodynamic mixed layer ocean and sea ice models [*Manabe and Broccoli,* 1985]. More recent studies have considered the dynamic response of the ocean, but with a highly simplified energy/moisture balance atmosphere [*Weaver et al.,* 1998], or a more highly parameterized representation of atmospheric dynamics and a 2.5-dimensional ocean [*Ganopolski et al.,* 1998]. The first simulation of the LGM climate by a 3-D ocean-atmosphere GCM [*Bush and Philander,* 1998] was only 15 years in duration (including spin-up), which is very short compared to the adjustment time of the full ocean.

Here we use a fully coupled high resolution ocean-atmosphere GCM (HadCM3), to carry out the first multi-century simulation for the LGM. The model does not require artificial fluxes of heat and water to maintain a stable ocean circulation in present day climate which makes it particularly suitable for studying changes in climate.

The Model and Experimental Design

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.

[1]Met Office, Hadley Centre, Bracknell, UK
[2]GFDL, Princeton, New Jersey

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The sea ice model includes a simple representation of sea ice dynamics. 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].

In the LGM experiment changes are made to the land surface characteristics to include the extensive continental ice sheets, modified coastlines and topographic heights to account for a sea level lowering [*Peltier,* 1994]. Atmospheric CO_2 is lowered from 280 ppmv to 200 ppmv, and a different pattern of insolation is specified due to a change to the earth's orbit.

The model is initialised from present day conditions since there is no three-dimensional global data set of glacial conditions. The time-scale of parts of the deep ocean is so long that it is not possible with current resources to bring the deep ocean to equilibrium with the glacial boundary conditions. To increase the rate of cooling of the ocean, the SSTs are linearly damped toward glacial values for 70 years, a technique known as Haney forcing [*Haney,* 1971]. The dataset of glacial SSTs is obtained from a slab model simulation of the LGM [*Hewitt,* 2000]. The coupled model is run for a further 700 years without the Haney forcing, and the integration is continuing. The 700 year simulation brings the upper ocean close to equilibrium, and produces a relatively stable solution. Below the thermocline the model is still slowly cooling by the end of the simulation — the global volume weighted temperature of the ocean cools by 0.07°C over the final century. The changes averaged over the final 100 years of the LGM simulation are compared to the control simulation in the following sections.

LGM Surface Response

The global mean surface temperature is 3.8°C colder in the LGM experiment compared to the control simulation. The cooling exceeds 10°C at high latitudes and over land, and 20°C over much of the Laurentide and Fennoscandian ice sheets (Figure 1), broadly consistent with previous modelling studies and palaeoclimatic data [*COHMAP members,* 1988]. However, the coupled model does not produce as large a cooling as slab model simulations [*Broccoli and Manabe,* 1987; *Hewitt and Mitchell,* 1997] or the CLIMAP reconstruction [*CLIMAP Project Members,* 1981] across much of the North Atlantic Ocean. In fact the coupled model produces slightly *warmer* SSTs than today over some regions of the North Atlantic, with the warmest anomaly of 1.8°C occurring west of the British Isles at about 25°W.

The glacial cooling produces thicker and more extensive Arctic sea ice than in the control consistent with palaeoclimatic reconstructions [*de Vernal et al.,* 2000]. Winter-time ice more than 1 m thick covers much of the Norwegian Sea, the North Atlantic south of Iceland to 60°N and the Labrador Sea south to the coast of Newfoundland. Much of the glacial sea ice melts in the summer along the coast of the Fennoscandian Ice Sheet and in the interior of the Labrador Sea.

The large cooling in the high-latitudes and over the Northern Hemisphere continents leads to subsidence which generally increases the pressure in these regions [*Rind,* 1998]. The large cool-

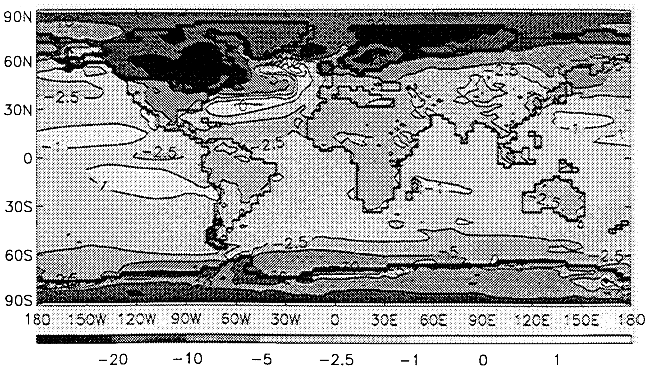

Figure 1. Change in annual mean surface temperature, in °C, LGM–control.

ing over the Laurentide Ice Sheet produces an intense glacial anticyclone over North America (Figure 2a) [*COHMAP members,* 1988] which strengthens the surface winds that blow from the cold ice sheets down the Labrador Sea into the North Atlantic. Further south, changes to the meridional Hadley circulation intensify the Azores High and the Icelandic Low which creates stronger westerly/southwesterly flow over the North Atlantic. The mechanical (wind) forcing produces a stronger and more extensive subpolar gyre at the LGM (Figure 2b).

The colder glacial atmosphere produces a decrease in precipitation (P) over most of the mid- and high-latitudes and over most of the continents, but there is an increase in P over large parts of the tropical and sub-tropical oceans and the mid-latitude North Atlantic. The evaporation (E) decreases almost everywhere, with a large decrease in E in the North Atlantic at about 40°N where there is a large decrease in SSTs. However, the strong surface winds that blow down the Labrador Sea and across the mid-latitude North Atlantic, combined with warm surface temperatures, produce a large increase in E in the Labrador Sea and the mid-latitude North Atlantic. The large increase in E over the Labrador Sea and the mid-latitude North Atlantic leads to a reduction in $P - E$ (Figure 2c) which makes the surface waters more saline in this region leading to an increase in surface density.

LGM Thermohaline Circulation

The structure of the NATHC in the control simulation is shown in Figure 3a and is discussed in more detail elsewhere [*Wood et al.,* 1999; *Gordon et al.,* 2000]. The maximum strength of the overturning remains fairly constant throughout the run at about 20 Sv (Figure 3c).

In the LGM experiment the sinking branch of the NATHC shifts southwards (Figure 3b). The glacial expansion of sea ice in the Nordic Seas provides an insulating blanket on the ocean, reducing the large loss of heat from the ocean to the cold overlying atmosphere in the Nordic Seas. The convection sites shift southward beyond the sea ice margin, south of Iceland, and deep convection occurs across much of the North Atlantic and into the Labrador Sea [*Hewitt,* 2000] where the sea ice changes are not as great as in the Nordic Seas. The southward rearrangement of the sites of convection means that the NATHC does not extend as far north as in the control, and the location of the maximum rate of overturning shifts southwards to 32.5°N (Figure 3b). Consequently, the flow of relatively warm water into the Nordic Seas is reduced. However, there is enough warm water, transported by a glacial anticyclonic gyre circulation in the Nordic Seas, to melt the sea ice in the summer along the coast of the Fennoscandian Ice Sheet. These ice-free waters allow some weak convection to take place in the Norwegian Sea, forced by both heat loss to the cold atmosphere and brine release from sea ice formation. The southward return flow of dense water is from the Norwegian Sea across the Iceland-Scotland Ridge back into the eastern North Atlantic, which is a very different situation from the control where most of the return flow is further west through the Denmark Straits.

The seasonally ice-free waters, the weak convection and the anticyclonic gyre in the Nordic Seas in the LGM experiment are all supported by palaeoceanographic data [*Veum et al.,* 1992; *Lassen et al.,* 1999]. The southward shift of the convection sites are consistent with other modelling studies and palaeoceanographic data [*Sarnthein et al.,* 1994; *Ganopolski et al.,* 1998].

The strength of the NATHC also increases. The cold glacial winds increase the vertical air–sea temperature gradient across the Labrador Sea, the mid-latitude North Atlantic, and along the coast of Western Europe. Consequently, the loss of heat from the ocean to the atmosphere is increased which leads to strong convective mixing and dense deep water formation in these regions, which drives a strong overturning cell (Figure 3b). The reduction in $P - E$ described above also contributes to this strong convective mixing. The maximum strength of the overturning circulation increases to 27 Sv over about a century and remains high thereafter (Figure 3c), for reasons described below.

The existence of a thermohaline circulation at the LGM is supported by palaeoceanographic data [*Boyle and Keigwin,* 1982; *Veum et al.,* 1992; *McCave et al.,* 1995] and other modelling studies [*Ganopolski et al.,* 1998; *Weaver et al.,* 1998], but such studies sug-

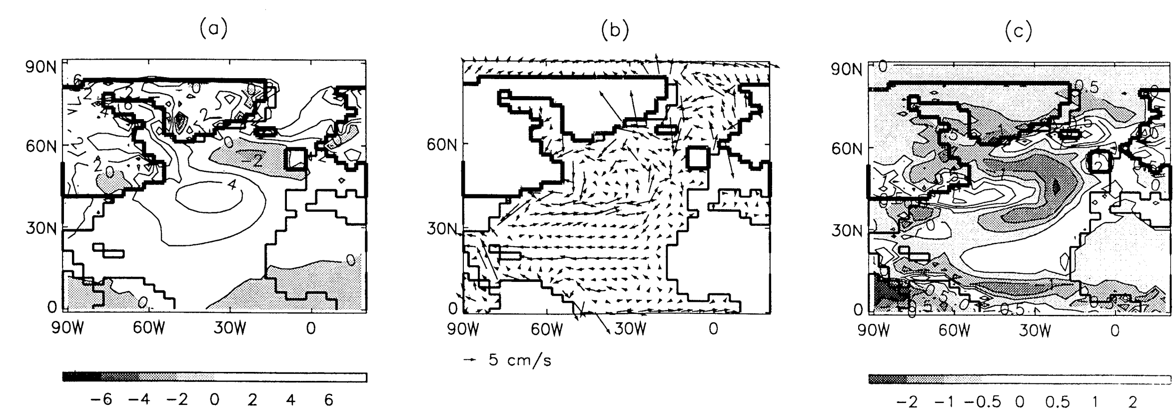

Figure 2. Annual mean changes, LGM–control. (a) Pressure at mean sea level, in mb. (b) Ocean surface currents, in cm s^{-1}. Note: the velocity vectors (arrows) are only plotted at every 3rd gridbox for clarity. (c) P–E, in mm day^{-1}.

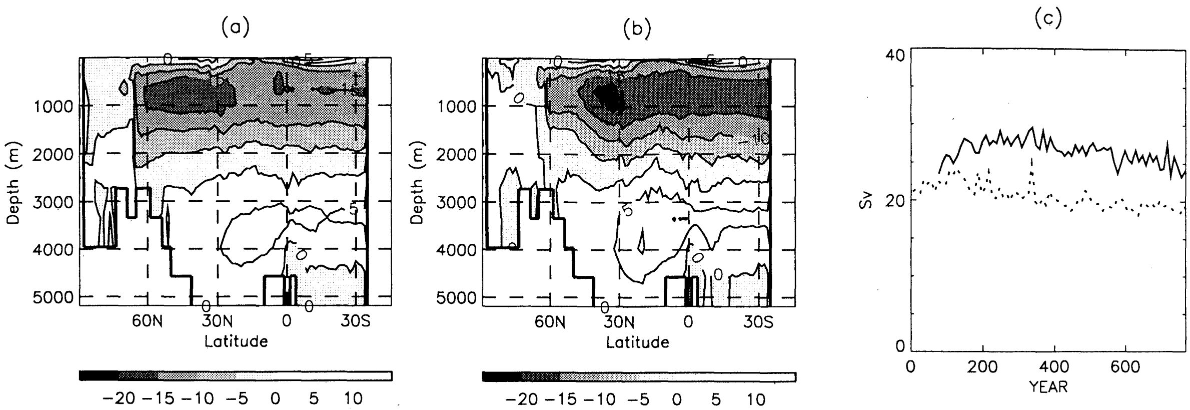

Figure 3. (a) and (b) show depth-latitude cross-sections of Atlantic meridional overturning streamfunction, in Sv. Positive values indicate clockwise flow in the plane shown on the panels. (a) Control simulation. (b) LGM simulation. (c) Timeseries of maximum strength of the streamfunction. Dashed line is the control, solid line is the LGM simulation after the 70 year Haney forced stage.

gest that the present day NADW was replaced with glacial North Atlantic intermediate water (GNAIW). The coupled model results presented here are seemingly at odds with these studies, since the deep water formation does not shoal appreciably. However, there are two points to consider. Firstly, the model's present day deep water in the North Atlantic (Figure 3a) does not penetrate as deeply as observations suggest, and instead occurs at depths similar to those proposed for GNAIW. Secondly, the Atlantic deep ocean is still cooling after 700 years, and AABW is penetrating into the North Atlantic. The deep ocean may become sufficiently dense to prevent the North Atlantic surface waters from sinking as deeply. Palaeoceanographic data supports the possibility of a stronger thermohaline circulation at the LGM [*Veum et al.,* 1992; *McCave et al.,* 1995] as simulated by the model, but the magnitude of the increase is difficult to verify.

Ocean Heat Transports

The LGM Atlantic Ocean transports more heat northwards than the control throughout the tropics and mid-latitudes up to 55°N, beyond which the heat transports are reduced (Figure 4). The increased heat transport throughout the tropics is a consequence of the stronger meridional overturning circulation. Further north, in the mid-latitudes, more heat is transported by the strong subpolar gyre circulation, partly at the expense of the heat transport by the meridional circulation. The stronger subpolar gyre, combined with the stronger circulation in the tropics increases the heat transport up to 55°N.

The strong overturning cell is self-maintaining due to a positive feedback from the increased northward heat transport in the North

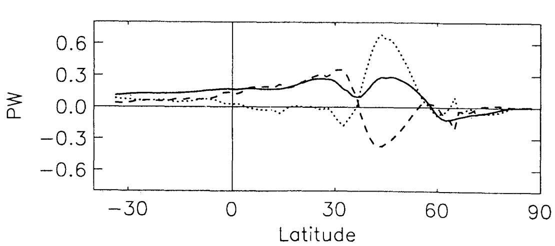

Figure 4. LGM–control zonal mean northward ocean heat transport (solid line) in the Atlantic Ocean, in PW, and contributions from the meridional circulation (dashed line) and gyre circulation (defined as the deviation from the zonal mean, dotted lines).

Atlantic. The warm mid-latitude surface waters, combined with the strong surface winds, increase the amount of evaporation to the atmosphere which increases the salinity and hence density of the surface waters, strengthening the NATHC. The warm surface waters also increase the air–sea temperature gradient described above.

Northward advection of warm water from the tropical Atlantic reduces the surface density in the North Atlantic, and this could weaken the NATHC and therefore the northward advection of warm water. However, this negative feedback is damped by the presence of the atmosphere which absorbs some of the extra heat, and then advects the heat away and radiates some to space. In the LGM run the increased overturning *is* accompanied by an increase to the amount of heat transported northwards in the Atlantic (Figure 4) in the Gulf Stream, but the increased heat loss to the extremely cold atmosphere at mid-latitudes weakens this feedback, and the salinity increases prevent the water from becoming less dense.

Advection of salinity can also provide a positive feedback. The NATHC transports salt northwards in the upper ocean and southwards in the deep ocean. Even though the strength of the NATHC increases in the LGM experiment, the northward salt transport actually decreases. The increase in P–E over the tropical Atlantic Ocean reduces the salinity of the upper ocean, opposing the effect of the increase in surface flow.

Discussion

The increased northward heat transport in the Atlantic Ocean in the coupled model produces a smaller cooling than slab models and palaeodata over the mid-latitude North Atlantic. The regions of warming in the coupled model are caused by stronger and warmer currents in the North Atlantic, in particular the Gulf Stream and the eastern section of the subpolar gyre which bring extra heat northwards along the coast of Western Europe, westwards across the mid-latitude North Atlantic and up into the Labrador Sea and Baffin Bay. Slab models cannot simulate changes to the amount of heat transported by the ocean because they cannot simulate changes to ocean currents, so they tend to cool everywhere in response to the large atmospheric cooling.

The stronger subpolar gyre and strong dense Labrador Current cause the Gulf Stream to shift south in the central and eastern Atlantic basin, producing relatively cold conditions where the LGM Labrador Current replaces the present day Gulf Stream. This shift is not associated with a shift of the line of zero Ekman pumping, as has been suggested [*Keffer et al.,* 1988]. The stronger LGM overturning cell in the tropics and the stronger subpolar gyre produce a stronger glacial Gulf Stream which transports more heat and produces the relatively warm conditions where the glacial Gulf Stream flows. Many of the differences between the HadCM3 simulation and a similar experiment using a slab model are associated with these changes in the Gulf Stream and Labrador Current. Such changes cannot be simulated in simpler ocean models.

The regions of warming in the coupled model are at odds with most palaeoclimatic datasets, although there are a few studies that suppport the existence of warm SSTs in the North Atlantic at the LGM [*Crowley,* 1981; *Rosell-Melé and Comes,* 1999; *de Vernal et al.,* 2000]. Since the model is still cooling these anomalously warm regions may be transient features. However, the warm regions were formed during the first century of the simulation and have been maintained over the following six centuries. Therefore, they represent a quasi-stable state in which relatively warm SSTs can be maintained in a cold glacial environment, brought about by the mechanisms discussed.

This study has focused on the unexpected response over the North Atlantic which has led us to investigate mechanisms of climate change involving the thermohaline circulation, an issue of interest for the prediction of future climate change. The response in other regions generally agrees more favourably with palaeodata [*Hewitt,* 2000]. Coupled ocean-atmosphere processes in the equatorial east Pacific for example produce good agreement with palaeodata, where enhanced easterly atmospheric flow in the Trade Wind region leads to greater oceanic upwelling, and cooler conditions than models that do not represent ocean dynamics, such as slab models.

Since the response we have described may be transient, evaluating the results in more detail with palaeodata is problematical. We have therefore concentrated on mechanisms of climate change. A more detailed comparison with palaeodata will be performed once the simulation is closer to equilibrium.

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References

Boyle, E. A. and L. D. Keigwin, Deep circulation of the North Atlantic over the last 200,000 years: Geochemical evidence, *Science*, *218*, 784–787, 1982.

Broccoli, A. J. and S. Manabe, The influence of continental ice, atmospheric CO_2, and land albedo on the climate of the last glacial maximum, *Climate Dyn.*, *1*, 87–99, 1987.

Bush, A. B. G. and S. G. H. Philander, The role of ocean-atmosphere interactions in tropical cooling during the last glacial maximum, *Science*, *279*, 1341–1344, 1998.

CLIMAP Project Members, Seasonal reconstruction of the earth's surface at the last glacial maximum, Geological Society of America Map and Chart Series, **MC-36**, 1981.

COHMAP members, Climatic changes of the last 18,000 years: Observations and model simulations, *Science*, *241*, 1043–1052, 1988.

Crowley, T. J., Temperature and circulation changes in the eastern North Atlantic during the last 150,000 years: Evidence from the planktonic foraminiferal record, *Marine Micropaleontology*, *6*, 97–129, 1981.

de Vernal, A., C. Hillaire-Marcel, J.-L. Turon, and J. Matthiessen, Reconstruction of sea-surface temperature, salinity and sea ice cover in the northern North Atlantic during the last glacial maximum based on dinocyst assemblages, *Canadian Journal of Earth Sciences*, *37*, 725–750, 2000.

Ganopolski, A., S. Rahmstorf, V. Petoukhov, and M. Claussen, Simulation of modern and glacial climates with a coupled global model of intermediate complexity, *Nature*, *391*, 351–356, 1998.

Gates, W. L., Modelling the ice-age climate, *Science*, *191*, 1138–1144, 1976.

Gordon, C., C. Cooper, C. A. Senior, H. Banks, J. M. Gregory, T. C. Johns, J. F. B. Mitchell, and R. A. Wood, The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model without flux adjustments, *Climate Dyn.*, *16*, 147–168, 2000.

Haney, R. L., Surface thermal boundary conditions for ocean circulation models., *J. Phys. Oceanogr.*, *1*, 241–248, 1971.

Hewitt, C. D., A study of the mechanisms of climate change at the last glacial maximum, Ph.D. thesis, University of Reading, 178 pp., 2000.

Hewitt, C. D. and J. F. B. Mitchell, Radiative forcing and response of a GCM to ice age boundary conditions: Cloud feedback and climate sensitivity, *Climate Dyn.*, *13*, 821–834, 1997.

Keffer, T., D. G. Martinson, and B. H. Corliss, The position of the Gulf Stream during Quaternary glaciations, *Science*, *241*, 440–442, 1988.

Lassen, S., E. Jansen, K. L. Knudsen, A. Kuijpers, M. Kristensen, and K. Christensen, Northeast Atlantic sea surface circulation during the past 30-10 ^{14}C kyr B.P., *Paleoceanography*, *14*, 616–625, 1999.

Manabe, S. and A. J. Broccoli, A comparison of climate model sensitivity with data from the last glacial maximum, *J. Atmos. Sci.*, *42*, 2643–2651, 1985.

McCave, I. N., B. Manighetti, and N. A. S. Beveridge, Circulation in the glacial North Atlantic inferred from grain-size measurements, *Nature*, *374*, 149–151, 1995.

Peltier, W. R., Ice age palaeotopography, *Science*, *265*, 195–201, 1994.

Rind, D., Latitudinal temperature gradients and climate change, *J. Geophys. Res.*, *103*, 5943–5971, 1998.

Rosell-Melé, A. and P. Comes, Evidence for a warm last glacial maximum in the Nordic seas or an example of shortcomings in UK37′ and UK37 to estimate low sea surface temperature?, *Paleoceanography*, *14*, 770–776, 1999.

Sarnthein, M., K. Winn, S. J. A. Jung, J.-C. Duplessy, L. Labeyrie, H. Erlenkeuser, and G. Ganssen, Changes in east Atlantic deepwater circulation over the last 30,000 years: Eight time slice reconstructions, *Paleoceanography*, *9*, 209–267, 1994.

Veum, T., E. Jansen, M. Arnold, I. Beyer, and J.-C. Duplessy, Water mass exchange between the North Atlantic and the Norwegian Sea during the past 28,000 years, *Nature*, *356*, 783–785, 1992.

Weaver, A. J., M. Eby, A. F. Fanning, and E. C. Wiebe, Simulated influence of cardon dioxide, orbital forcing and ice sheets on the climate of the Last Glacial Maximum, *Nature*, *394*, 847–853, 1998.

Wood, R. A., A. B. Keen, J. F. B. Mitchell, and J. M. Gregory, Changing spatial structure of the thermohaline circulation in response to atmospheric CO_2 forcing in a climate model, *Nature*, *399*, 572–575, 1999.

C. D. Hewitt and J. F. B. Mitchell, Met Office, Hadley Centre, London Road, Bracknell, RG12 2SY, UK. (e-mail: cdhewitt@meto.gov.uk; jfbmitchell@meto.gov.uk)

A. J. Broccoli and R. J. Stouffer, Geophysical Fluid Dynamics Laboratory, P.O. Box 308 Princeton, NJ 08542, USA. (e-mail: ajb@gfdl.noaa.gov; rjs@gfdl.noaa.gov)

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