Reduced Atlantic Storminess during Last Glacial Maximum: Evidence from a Coupled Climate Model

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ABSTRACT

The Last Glacial Maximum (LGM), 21 000 yr before present, was the time of maximum land ice extent during the last ice age. A recent simulation of the LGM climate by a state-of-the-art fully coupled global climate model is shown to exhibit strong, steady atmospheric jets and weak transient eddy activity in the Atlantic sector compared to today’s climate. In contrast, previous work based on uncoupled atmospheric model simulations has shown that the LGM jets and eddy activity in the Atlantic sector are similar to those observed today, with the main difference being a northeastward extension of their maxima. The coupled model simulation is shown to agree more with paleoclimate proxy records and thus is taken as the more reliable representation of LGM climate. The existence of this altered atmospheric circulation state during LGM in the model has implications for understanding the stability of glacial climates, for the possibility of multiple atmospheric circulation regimes, and for the interpretation of paleoclimate proxy records.

1. Introduction

The Last Glacial Maximum (LGM) was a cold period approximately 21 000 yr before present (21 ka) when massive ice sheets covered much of the Northern Hemisphere (NH) continents. Paleoclimate proxy records provide valuable information on the climate forcings and climate state at the Last Glacial Maximum, but they are limited in spatial and temporal resolution and can be difficult to interpret, especially if one wishes to deduce circulation features such as flow fields or energy fluxes. Climate models have proved to be increasingly useful as a tool for tackling some of the questions that cannot be answered using proxy data alone. Of interest in this study is the large-scale circulation of the atmosphere during glacial times and how it compares to the circulation observed in the present-day climate. In particular, we direct our attention toward the North Atlantic sector, a region of deep water formation that experienced large, abrupt climate events during the last glacial period.

In the 1990s, the Paleoclimate Model Intercomparison Project (PMIP) was undertaken to evaluate past climates using a collection of climate models. The first phase of the project (PMIP1) comprised, for the most part, uncoupled atmospheric general circulation models forced with LGM boundary conditions: a sea surface temperature (SST) and sea ice cover reconstruction from the Climate: Long-range Investigation, Mapping, and Prediction (CLIMAP) project (CLIMAP 1981) and the ICE-4G land ice reconstruction (Peltier 1994). The resulting simulations offered a glimpse into how atmospheric circulation may have been during LGM. Subsequent studies focused on describing and understanding specific features such as atmospheric heat transport, transient activity, and storm tracks (Hall et al. 1996; Kageyama et al. 1999; Kageyama and Valdes 2000). Their findings have been used to endorse the idea that the glacial world was a stormier world because of stronger equator-to-pole temperature gradients and enhanced production of baroclinic eddies. Revisiting the original studies, however, discloses some often-overlooked details.

Synthesizing results from all the European models in
PMIP1, Kageyama et al. (1999) reported a northeastward extension of both Northern Hemisphere storm tracks across the board and an elongation of the Atlantic storm track in six of the seven models. They note “no systematic increase or decrease in the storminess from the present climate to the last glacial maximum,” although the figures in the paper do seem to indicate an increase in peak low-level transient eddy activity for most of the models (see Kageyama et al. 1999, their Figs. 1, 2, 6, and 7). In the specific case of the U.K. Universities’ Global Atmospheric Modelling Programme (UGAMP) prescribed SST simulation, the increase in Atlantic eddy activity is limited to low levels of the atmosphere and is associated with enhanced low-level baroclinicity, thus suggesting the presence of stronger but shallower synoptic waves (Hall et al. 1996). Furthermore, it was found that changes in the normal modes largely account for changes in the position and dominant wavenumber of the storms but not the amplitude of the actual storm tracks (Kageyama and Valdes 2000).

Included in the analysis of Kageyama et al. (1999) are simulations from select models that were run with a slab ocean, a configuration that allowed SSTs and sea ice to be computed based on their thermal response to the climate forcings. Results from these simulations are somewhat equivocal because the oceanic heat flux that must be specified for a slab ocean was set to present-day values, thereby determining, in large part, the SST and sea ice distributions as well as the location of the storm tracks. However, among slab ocean simulations, there is some support for increased storminess at low to middle levels of the atmosphere during LGM (Dong and Valdes 1998). Going one step further, a study involving an intermediate complexity model with a simple dynamical ocean model reports similar findings (Justino et al. 2005).

Mounting paleoclimate evidence has since made it clear that both the CLIMAP SST reconstruction and the ICE-4G land ice reconstruction used in PMIP1 are flawed. In view of such developments, the question of how jets and storminess changed during LGM is by no means settled. There are now simulations of the LGM by more complex, fully coupled climate models with updated climate forcings based on the paleoclimate record, some of which fall within the framework of the second phase of PMIP (PMIP2). The majority of the literature documenting these simulations is concerned with issues such as atmosphere–ocean processes in the tropics and subtropics, in which the CLIMAP reconstruction is known to have problems (Bush and Philander 1998; Broccoli 2000; Kitoh and Murakami 2002; Timmermann et al. 2004); the general features of LGM climate (Kitoh and Murakami 2001; Kim et al. 2003; Shin et al. 2003; Otto-Bliesner et al. 2006); and the role of ocean dynamics in the maintenance of this climate state (Dong and Valdes 1998; Hewitt et al. 2003). Initial PMIP2 intercomparison studies have found an improvement in many aspects of the new LGM simulations (Kageyama et al. 2006; Braconnot et al. 2007), however, considerable problems remain in reproducing the peak glacial–interglacial temperature change from the Greenland ice cores (Mas- son-Delmotte et al. 2006) and in simulating changes in the ocean thermohaline circulation (Otto-Bliesner et al. 2007).

In this study, we examine the atmospheric circulation of the LGM in a PMIP2 simulation from a state-of-the-art, fully coupled climate model, the Community Climate System Model, version 3 (CCSM3), that was developed at the National Center for Atmospheric Research (NCAR). The simulation was set up and performed by Otto-Bliesner et al. (2006) to contrast preindustrial, mid-Holocene, and LGM climates. One major objective of the present study is to characterize the mean state and variability of atmospheric circulation in the Atlantic sector during the LGM by examining the jet and transient eddies, with an emphasis on identifying differences between the LGM and present-day climates.

Interestingly, when the CCSM3 is forced with PMIP2 boundary conditions, the simulated LGM climate shows a strong, steady Atlantic jet and enhanced low-level baroclinicity, but diminished wintertime eddy activity at all levels of the atmosphere compared to the present day. These results appear to contradict the atmosphere-only simulations from PMIP1, and furthermore, they suggest the existence of an altered atmospheric circulation regime during the LGM compared to present day (PD). As an initial illustration of the differences, Fig. 1 shows contours of total horizontal wind speed at 200 mb for PD and LGM simulations from a selection of the higher-resolution PMIP1 models (details in Table 1) and from the CCSM3 coupled model. For each uncoupled model (Figs. 1c–h), the LGM simulation (forced with PMIP1 boundary conditions) produces an Atlantic jet with strength and orientation (southwest–northeast tilt) comparable to the present-day jet (both simulated and observed) but with a northeastward extension. This is in stark contrast to the CCSM3 LGM simulation (Fig. 1j), which features a strong Atlantic jet with a more zonal orientation. Two of the PMIP1 slab ocean simulations are also shown (Figs. 1a,b). Although there are difficulties in interpret-
ing these simulations, they do serve to emphasize the qualitative difference between the CCSM3 LGM simulation and all the LGM simulations that use PMIP1 boundary conditions.

The paper is organized as follows. A brief description of the model and methods are contained in section 2. Section 3 presents the model results indicating that, during LGM, the atmosphere exhibited a strong, steady mean circulation with decreased eddy activity. Section 4 discusses some possible mechanisms for the suppression of eddy activity. Section 5 summarizes evidence, based on paleoclimate observations, that the CCSM3 simulation produces a more realistic representation of LGM climate than the PMIP1 simulations; the focus is on the roles of land ice forcing and the SST/sea ice distributions. Finally, the main results of this work are summarized in section 6.

2. Model description and methods

We investigate changes between PD and LGM climates as simulated by the CCSM3 (Collins et al. 2006a), a global coupled atmosphere–ocean–sea ice–land surface climate model developed at NCAR. The setup of and results from these model simulations are documented in detail in Collins et al. (2006a) and Otto-Bieser et al. (2006). Briefly, the CCSM3 comprises the primitive equation Community Atmosphere Model, version 3 (CAM3) at T42 horizontal resolution with 26 hybrid coordinate vertical levels (Collins et al. 2006b); a
land model with land cover and plant functional types, 
prognostic soil and snow temperature, and a river rout-
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LMD5.3 Y 50 64 11 Grid point
CAM3 N 64 128 26 Spectral

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criterion. As an additional check, we repeated the eddy 
analysis on the full 40-yr dataset and found no change 
to the main results of the study. Eddy fields were 
filtered with a sixth order high-pass Butterworth filter to 
emphasize the variability at periods less than 8 days. 
Such filters are often used in the study of storm tracks, 
with the high-pass cutoff varying between 6 and 10 days 
(e.g., Nakamura 1992; Trebonth 1991; Yin 2002). The 
exact choice of filter is relatively unimportant because 
baroclinic waves dominate the eddy statistics at these 
synoptic time scales.

3. Last Glacial Maximum climate in the coupled 
model

Of interest here is the large-scale atmospheric 
circulation in the model’s simulations of the LGM and PD 
climates. We will focus on the Northern Hemisphere 
Atlantic sector, in which differences in forcings 
between the LGM and PD, and consequently circulation 
features, are most dramatic. All differences between 
the LGM and PD climates discussed here are signific-
ant at the 95% confidence level unless otherwise noted.

a. Circulation and heat transport

Upper-level zonal wind and geopotential height pro-
vide a useful broad-brush picture of large-scale flow 
characteristics of the atmosphere. Figure 2 shows win-
tertime maps of these two fields for the LGM and PD 
simulations. Under the LGM forcings described in the 
previous section, we observe an enhanced stationary 
wave associated with the Laurentide ice sheet covering 
most of North America and a stronger, more zonal jet 
in the Atlantic sector downstream of the ice sheet. The 
stronger winds during LGM are consistent with the 
stronger equator-to-pole surface temperature gradient 
seen in Fig. 3. Changes in the Pacific are more subtle, 
showing a slight equatorward shift of the jet (Fig. 3) and 
the development of a split flow over Siberia, which is 
evident from the region of weak easterlies outlined by 
the thick zero contour in Fig. 2b.

The change in atmospheric circulation over the At-
lantic sector during LGM is particularly striking. There 
is an inverse relationship between maximum jet 
strength and jet width in January for the PD simulation 
(red circles in Fig. 4), in which jet strength $u_{\text{ATL}}$ is 
the maximum zonal wind in the sector and jet width is 
the latitude range over which $u_{\text{ATL}}$ decreases to half its 
maximum value. The relationship is not evident in the 
LGM climate (blue triangles), which furthermore in-
habits a sector of width–strength space separate from
the PD climate. The LGM jet is 20% stronger on average, with half the variability in maximum speeds of the PD jet; it is also 30% narrower, with almost 6 times less variability in width. These results are robust to the choice of month(s) used to define winter and the exact longitude range used to define the Atlantic jet.

Figure 5 shows implied annual meridional energy transports calculated from the model output. The total (atmosphere plus ocean) heat transport (RT) by the climate system is determined by integrating the top-of-atmosphere (TOA) radiation imbalance $R_{\text{TOA}}$ over all longitudes $\theta$ from the North Pole to each latitude $\phi$:

$$RT(\phi) = R_E^2 \int_0^{2\pi} \int_{\phi}^{\pi/2} R_{\text{TOA}}(\phi', \theta') \cos \phi' \, d\phi' \, d\theta'.$$

Finally, the ocean heat transport (OT) is calculated as the residual:

$$OT = RT - AT.$$

From Fig. 5b, the discrepancy between the model’s PD total heat transport (black line) and the satellite-derived radiatively required total heat transport from Trenberth and Caron (2001) (filled gray area) is less than 0.5 PW, or 10% of the maximum heat transport.

We find that the amount of heat transported toward the poles is remarkably similar in the PD and LGM simulations. Between the two model simulations, the
LGM does, indeed, have slightly more transport, with the atmosphere helping to increase the peak NH value to 6.33 ± 0.01 PW, about 0.3 PW greater than in PD. The robustness of the total heat transport curve found across the CCSM3’s simulations of PD and LGM climates is corroborated by other coupled climate models (Hewitt et al. 2003; Shin et al. 2003) and from uncoupled general circulation models using a prescribed sea surface temperature forcing (Hall et al. 1996), with the magnitude of the LGM increase varying from 0.2 to 1.5 PW for peak NH values.

All else being equal, a back-of-the-envelope calculation tells us that this relatively modest 0.3 PW boost in heat transport at 35°N translates to a 3 W m$^{-2}$ boost in heating rate north of this latitude circle. Assuming a midrange climate sensitivity of 0.5°C (W m$^{-2}$)$^{-1}$, this is equivalent to a 1.5°C warming of the mid- to high-latitude regions. Clearly, all else is not equal and the actual surface temperature difference between the two climates poleward of 35°N is closer to 10°C. Note that technically, the 0.5°C (W m$^{-2}$)$^{-1}$ value is a global climate sensitivity and should not be used to estimate the response of the polar cap to increased heat flux from the lower latitudes. However, the polar cap, as defined in our calculation (35°N to the North Pole), is a large region over which there is a net TOA radiation loss to space. Furthermore, the use of this global climate sensitivity to estimate the contribution of heat transport changes to the simulated temperature change is supported by an experiment in Seager et al. (2002). In this
experiment, ocean heat transport was turned off, and this lead to a 1.3 PW reduction in energy that moved across 35°N, corresponding to a 13 W m⁻² decrease in heating rate north of this latitude and a 6°C cooling of the mid- to high-latitude regions.

Upon closer inspection, there are interesting differences in the partitioning of these heat fluxes in the PD and LGM simulations. Overall, the net (atmosphere plus ocean) transport in the NH during LGM is very similar to that in the PD. The total AT in the LGM is also comparable to that in the PD, but there is a large difference in the balance of processes responsible for this AT. The pronounced ridging forced by the Laurentide ice sheet enhances the dry stationary wave heat transport contribution $\bar{w}^* T^*$ to AT in the LGM compared to PD (Fig. 5c). By inference, the transient heat transport by eddies must be diminished in this region. We note that there is a slight decrease in latent energy transport in the LGM simulation (not shown), an effect that is to be expected in a drier climate (Held and Soden 2006); this change is small compared to the dry stationary wave change. In a global view, these results are in fact in agreement with the energy budget analysis of the UGAMP LGM simulation by Hall et al. (1996). Although they observe an increase in transient heat transport at low levels, there is a compensating decrease aloft such that the column-integrated transient eddy transport is smaller during LGM.

b. Transient eddy activity

We can diagnose the eddy activity associated with this reduction in energy transport by the transient eddies in the LGM by calculating high-pass filtered quantities such as low-level temperature flux $(u'T')$, upper-level momentum flux $(u'v')$, and upper-level eddy kinetic energy $[(u'^2 + v'^2)/2]$. In addition to presenting maps of these eddy statistics, we use several metrics to quantify the steadiness of the atmospheric circulation at low levels and aloft during the winter season. The first metric is the maximum zonal wind at 200 mb in the Atlantic sector 15°–65°N, 90°W–0°. The second metric is the kinetic energy associated with the departure of the 200-mb flow from climatological monthly means, averaged over the Atlantic sector:

$$KE_{\text{ATL}} = \frac{1}{2 A_{\text{ATL}}} \int_{A_{\text{ATL}}} (u_{200} - \bar{u}_{200})^2 + (v_{200} - \bar{v}_{200})^2 \, dA,$$

(4)
where $u$ and $v$ are monthly-mean fields, $A$ represents area, and overbars indicate climatological means. KEI_{ATL} thus gives an indication of the interannual variability of the upper level flow. The final metric is the northward eddy heat flux at 850 mb averaged over the Atlantic sector:

$$\nu T_{ATL} = \frac{1}{A_{ATL}} \int_{A_{ATL}} \nu_{850} T_{850} \, dA,$$

where primes indicate daily fields that have been high-pass filtered to retain variability at periods less than 8 days.

Concentrating on boreal winter in the NH, the eddy fields reveal an LGM climate that is more quiescent than the PD (Fig. 6). In the Atlantic sector, the reduction in eddy activity from PD to LGM is observed at both low levels (15% decrease in sector-averaged $v' T'$) and upper levels [30% decrease in sector-averaged $(u'^2 + v'^2)/2]$. Compared to PD, the LGM jets are strong and narrow (Fig. 3), and the eddy fluxes occupy a narrower latitudinal band hugging the axis of the jet core rather than a broad band perched on the poleward flank of the jet (e.g., compare Figs. 6b and 6c). The differences between the two climates can also be seen in Fig. 4, in which January poleward heat fluxes in the LGM simulation (blue triangles) span a narrow range of smaller values than in the PD simulation (red circles). Although we will not discuss the Pacific sector, it, too, exhibits changes in jet structure and eddy fluxes.

The measures of atmospheric flow in Table 2 provide another way to compare the Atlantic sector in today’s climate and glacial climates. The weak eddy activity in the LGM simulation coexists with a stronger, narrower Atlantic jet, as shown in Fig. 4. From this plot, we inferred a less variable upper-level flow field during LGM compared to PD. The winter season flow metrics provide additional and more direct ways to evaluate the steadiness of the LGM jet. From Table 2, we see a 35% decrease in monthly departures of kinetic energy from its climatological mean state (KEI_{ATL} in column three). Together, these results are consistent with the picture of global heat transport in Fig. 5, in which a slight overall increase in meridional energy flux during LGM is achieved by a greatly enhanced stationary wave, with the implication that the contribution from transient eddies must be weaker.

As a final remark on this topic, we note that the gross structure of the Atlantic jet and eddies described here for the PD simulation is consistent with observational data from the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis (Kalnay et al. 1996) and the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005). Both NCEP (not shown) and ERA-40 (Figs. 6a,d,g) show a broad wintertime jet with a southwest–northeast tilt and eddy activity peaking poleward of the jet, just as we have seen for the PD simulation (Figs. 6 b,e,h). One feature that does not compare well is the absolute jet strength, which the model overestimates in each winter month (as seen for January in Fig. 4) such that the winter season average of $u_{ATL}$ is 20% too strong (Table 2). However, the variability in the large-scale flow field in the PD simulation is comparable to that in the two reanalyses products; furthermore, the PD variability is substantially differ-
ent from that in the LGM simulation. For example, \( u_{ATL} \) spans a range of approximately 25 m s\(^{-1}\) in both the observations and in 50 yr of the PD simulation (Fig. 4), whereas the range in jet strengths in the LGM simulation is approximately half this value. Also, the variability as measured by the poleward heat flux \( vT_{ATL} \) and kinetic energy \( KEI_{ATL} \) indices (Table 2) appear to show characteristic values for the PD climate (3.42–3.56 K m s\(^{-1}\), 31–34 m\(^2\) s\(^{-2}\)) that are distinct from those of the LGM climate (2.93 K m s\(^{-1}\), 22 m\(^2\) s\(^{-2}\)).

c. Baroclinicity of the atmosphere

The result of diminished meridional heat transport in a climate with sharper temperature gradients and stronger jets (Figs. 2–4) is somewhat surprising. To express this apparent paradox in more quantitative terms, we use the Eady growth rate parameter (Eady 1949) formulated by Lindzen and Farrell (1980) as a means of predicting transient behavior from the time mean flow. The parameter is the growth rate of the fastest-growing Eady mode and is given by

\[
\sigma = 0.31 \frac{f}{N} \frac{\partial u_h}{\partial z},
\]

where \( f \) is the Coriolis parameter, \( N \) is the Brunt–Väisälä buoyancy frequency, and \( \partial u_h/\partial z \) is the vertical
shear of the horizontal wind component. We calculate this growth rate for the 850–700-mb layer (Fig. 7).

For each climate, regions of high Eady growth rate correspond in general to regions of enhanced eddy activity; across different climate states, however, Eady growth rates increase in regions where eddy amplitudes decrease. For example, there is a general increase in midlatitude growth rates during LGM relative to PD, especially over the Atlantic, which suggests that the glacial climate should be stormier than today’s climate. That the eddy diagnostics indicate decreased eddy activity during LGM despite more baroclinic conditions (larger $\sigma$) points to the complicated nature of the relationship between transient activity and the mean flow.

4. Suppression of eddy activity

The LGM climate as simulated by the CCSM3 coupled model exhibits stronger jets but weaker storms. This quiescent glacial climate is counter to the intuitive idea that a more vigorously circulating atmosphere should produce a more tempestuous world. It is also counter to more theoretically based indicators such as the stronger meridional temperature gradient and the increased low-level baroclinicity. However, a range of factors not captured by the Eady growth rate parameter also affect the life cycle of eddies. These factors range from diabatic heating (Hoskins and Valdes 1990) and seeding effects (Zurita-Gotor and Chang 2005) to “governors” that limit the ability of eddies to tap into the available baroclinicity (James 1987; Lee and Kim 2003; Harnik and Chang 2004). In this section, we discuss whether such factors may play a role in reducing the eddy activity in the coupled model LGM simulation. A prerequisite to this discussion is an assessment of the structure of the eddies.

a. Eddy structure

To study the structure of eddies, we use one-point regression analysis (Lim and Wallace 1991; Chang 1993), in which a regression coefficient $b(i)$ is calculated for a specific pressure, latitude, and longitude location $i$ by linearly regressing the input variable of interest $y(i)$ against a reference time series $x_i$:

$$ b(i) = \frac{\sum_{i=1}^{N} y_i x_{i}'}{\left(\sum_{i=1}^{N} x_{i}'^2\right)^{1/2}}. \quad (7) $$

Here, $N$ is the number of time samples and primes denote daily fields that have been high-pass filtered to retain variability at periods less than 8 days.

We examine regression structures at 200 and 850 mb in the CCSM3 simulations of PD and LGM climates (Fig. 8). The reference time series against which variables are linearly regressed is the 200-mb meridional wind field $u_{200}$ at 43°N, 28°W for the PD climate and at 38°N, 28°W for the LGM climate. These points were selected based on the location of maximum variance in the DJFM $u_{200}$ field. We normalize the $b(i)$ coefficients such that the regression maps show amplitude, expressed in the units of the variable of interest, per standard deviation of the reference time series. Structures calculated using reference time series at different locations, reference time series based on different fields, and alternate definitions of the winter season are similar to these and are not shown here.

The eddy structure in both the PD and LGM climates resemble localized wave trains. The LGM eddy amplitudes peak just south of 40°N, the latitude marked by the solid gray line in Fig. 8, whereas the PD eddy amplitudes peak just north of this latitude. Although meridional wind variations $v'$ are much stronger aloft (Figs. 8a,d) than at low levels (Figs. 8b,e) in both simulations (note the different contour intervals), there is a greater discrepancy in the LGM. Thus, eddies in glacial times appear to be trapped closer to the tropopause, with weaker low-level disturbance amplitudes. Furthermore, though the LGM has slightly weaker eddy amplitudes, the eddy structures are correlated over greater distances and are more persistent, as seen in lag regres-
sion analyses, in which the time series \( y_i(i) \) is shifted by a certain number of days relative to \( x_i \) (not shown). The larger spatial correlations are, in fact, consistent with the eastward extension of storm tracks seen in the PMIP3 simulations of Kageyama et al. (1999), however, the location and strength of the storm tracks themselves are not. These results suggest that differences in eddy structure may be an important part of why the LGM simulation exhibits such weak eddy activity compared to the PD simulation. There is a large body of theoretical work on why eddy activity might be suppressed in conditions that appear conducive to eddy growth. Next, we examine the relevance of these hypotheses for the reduction of eddy activity in the LGM.

b. Discussion

One established idea in the literature is the barotropic governor mechanism, in which horizontal shear is thought to limit baroclinic instability and hence storm growth (James 1987). The sharpening of the Atlantic jet (Fig. 3) makes this an attractive candidate, but some telltale signs of the barotropic governor are absent. If the barotropic governor were in operation, we would expect the eddies to bend and elongate into a boomerang or banana shape because they are deformed by the strongly sheared mean flow; this characteristic shape is not evident in the LGM simulation (Fig. 8).

Another explanation for the suppression of eddy activity is inspired by the existence of a modern analog to the strong jet–weak storms dichotomy, namely the Pacific midwinter suppression of storms (Nakamura 1992) that occurs every January when the jet is at its strongest (Fig. 9). According to Yin (2002), this phenomenon has to do with the southward migration of tropical convection in the western Pacific and the related strengthening of the subtropical jet during boreal winter. Yin (2002) diagnosed that these changes increase upper-level static stability and lower the tropopause height in a manner that discourages the growth of storms. In the modern climate, the climatological latitude of the Atlantic jet is 2°–3° north of the climatological latitude of the Pacific jet in winter, perhaps positioning it too far poleward to...
be affected by changes in tropical heating. In the LGM simulation, however, the more zonal orientation of the Atlantic jet pulls it toward the equator, in closer proximity to the tropical convection sites. Preliminary inspection reveals little difference in tropical heating or precipitation between the LGM and PD simulations in the Atlantic sector. Nonetheless, subtle changes in the tropics can have a large impact, so a more thorough investigation should be carried out before this mechanism for jet stability can be rejected.

In a related vein, Lee and Kim (2003) used idealized numerical experiments to demonstrate that flow regimes with strong (weak) subtropical jets exhibit weaker (stronger) eddy activity located just (far) poleward of the jet core. In a series of experiments, they varied the strength of the tropical convective heating and the high latitude cooling to strengthen or weaken the subtropical jet (Lee and Kim 2003; Son and Lee 2005). They showed that the strong subtropical jet regime supported more isotropic eddies and weaker barotropic conversion than the weak subtropical jet regime. The LGM simulation exhibits several features in common with their strong subtropical jet case, including relatively weak eddy fluxes located at roughly the same latitude as the tropospheric jet maximum (Fig. 6) and a slightly more isotropic eddy structure (Fig. 8). Rather than requiring changes in tropical heating, this idea allows for the possibility of a strong jet regime to be set up by other ways, for example, by topographic effects from the Laurentide or additional high-latitude cooling because of extensive sea ice cover in the North Atlantic.

The Lorenz energy cycle (Lorenz 1955) provides further insight into why the eddy activity in the LGM simulation is suppressed compared to the PD simulation. The terms in the energy budget equation describe how energy is exchanged between the mean flow and the transient eddies as the eddy disturbances grow and decay. Different suppression mechanisms affect different parts of the eddy life cycle, and thus should have different signatures on these budget terms. Preliminary results (A. Donohoe 2007, personal communication) indicate that the LGM Atlantic mean state does indeed support faster eddy growth because of its sharper temperature gradients, despite competing influences from the increased static stability and the strong barotropic shear of the jet. However, the baroclinic conversion of potential energy into eddy kinetic energy is greatly reduced in the LGM compared to PD. The explanation for the reduced eddy activity appears to be related to the fact that there is substantially less seeding of the Atlantic baroclinic zone by upstream disturbances in the LGM, and that the disturbances that do enter the baroclinic zone are weak. The cause of the reduced seeding is under continuing investigation.

5. Evaluation of the CCSM3 simulation

The PMIP1 simulations of LGM climate share certain atmospheric flow characteristics that are very similar to those in the present-day climate, whereas the coupled model simulation of LGM exhibits fundamental differences when compared to the present-day climate. Specifically, in the PMIP1 models, the amount of storminess in the Atlantic sector during LGM is comparable to that in the PD climate, with slight increases in peak low-level transient eddy activity during the LGM (Kageyama et al. 1999). Furthermore, in the
PMIP1 LGM simulations, the Atlantic jet shows the same northeastward extension as the Atlantic storm track with little change in strength (Fig. 1). Conversely, the CCSM3 LGM simulation exhibits a strong, steady jet with zonal orientation and weaker transient eddies at all levels of the atmosphere (Fig. 6).

We have reproduced the PMIP1 results with CAM3, the atmospheric component of the CCSM3, to illustrate more clearly the differences between the circulation simulated by the coupled model and the generic result obtained by the suite of uncoupled PMIP1 models (Fig. 10). The atmosphere model has now been forced with SST, sea ice and land ice prescribed exactly as specified in the PMIP1 experimental setup. The resulting simulation is the same that appears in Fig. 1h and hereafter will be referred to as PMIPa, where “a” means “atmosphere only.” Compared to the coupled model LGM simulation (Fig. 10c,d), PMIPa produces a stormier Atlantic sector and a weaker jet that exhibits a marked southwest-to-northeast tilt (Fig. 10e,f). Indeed, the eddy heat transport and jet orientation in PMIPa bear a closer resemblance to the coupled PD simulation (Fig. 10a,b) than to the coupled LGM simulation.

The source of the differences between the LGM climate simulated by the uncoupled models and the LGM climate simulated by the coupled model lies at the earth’s surface. Recall that the uncoupled PMIP1 models experience or “see” one set of prescribed land ice and SST/sea ice boundary conditions. The coupled model experiences a somewhat different set of land ice boundary conditions, those of the improved ICE-5G (Peltier 2004) reconstruction. In addition, the coupled model has interactive ocean and sea ice components that calculate their own SST/sea ice fields. The fact that CCSM3’s atmosphere model achieves a PMIP1-like LGM climate when run in the PMIP1 configuration is further support that these land and sea surface forcings, rather than a disparity in models or atmosphere model physics, are, indeed, key to creating the different LGM climates (Fig. 1h).

The question becomes one of evaluating which simulated LGM climate, the uncoupled PMIP1-like LGM or the coupled CCSM3-like LGM, is a more realistic representation of the actual climate of the Last Glacial Maximum. Detailed comparisons have been performed between PMIP1 model simulations and a selection of the fully coupled model simulations, including CCSM3, that participated in PMIP2. These studies conclude that the fully coupled PMIP2 simulations using ICE-5G land ice are in better agreement with the paleoclimate observations currently available, particularly in terms of surface air temperature, SST, and tropical climate (Braconnot et al. 2007), as well as regional temperatures over northern Eurasia and the North Atlantic.

Fig. 10. Wintertime jet position and eddy diagnostics in CCSM3 simulations compared to the simulation PMIPa reproducing the PMIP1 results. Colors show (left) eddy temperature flux $v'T$ (K m s$^{-1}$) at 850 mb and (right) zonal momentum flux $u'v'$ (m$^2$ s$^{-2}$) at 200 mb, all for DJFM. Black contours show zonal wind at 250 mb (10 m s$^{-1}$ contours starting at 30 m s$^{-1}$). All eddy fields are calculated from daily data that have been high-pass filtered to retain variability at periods less than 8 days.
Kageyama et al. (2006). These intercomparisons include a number of partially coupled simulations from PMIP1 in which SSTs and sea ice were calculated by a slab ocean model. These will not be included in our discussion because their use of present-day oceanic heat transport in the slab ocean heavily influences the resulting SST patterns.

The following sections will point out certain aspects of the boundary conditions and model results where the CCSM3 simulation shows marked improvements, based on the paleoclimate record, over the PMIP1 simulations. (For a comprehensive treatment of this topic, the reader is referred to the intercomparison studies mentioned in this paper.) Also discussed are the potential consequences of these features for atmospheric dynamics in general and for the Atlantic jet and storm track in particular.

a. Land ice

The ICE-4G (Peltier 1994) and ICE-5G (Peltier 2004) reconstructions of deglaciation history are based in part on the theory of glacial isostatic adjustment, a process by which the external surface load of the continental ice sheets is compensated by changes in the surface of the earth. ICE-5G is widely considered to be an improvement over ICE-4G: its land ice reconstruction accounts for new refinements in the reconstruction of global sea level, and it also corrects many of the regional shortcomings of ICE-4G (see Peltier 2004, their references). As a result, at the time of the LGM, ICE-5G has more land-based ice than did ICE-4G (approximately 12 m sea level equivalent). In ICE-5G, there is much less land ice on Greenland and the Eurasian continent, whereas the Laurentide ice sheet complex over North America is significantly larger in volume compared to ICE-4G (15 m sea level equivalent, or 25% of the Laurentide ice volume), with the bulk of the extra ice forming the Keewatin Dome west of Hudson Bay (Fig. 11). The addition of this dome reconciles the original ICE-4G results with more recent observations, namely the large crustal uplift rates near Yellowknife, Canada, (Argus et al. 1999) and gravity measurements south and west of Hudson Bay (Lambert et al. 2001). A higher Laurentide ice sheet forces a stronger stationary wave pattern in the Northern Hemisphere (Otto-Bliesner et al. 2006), and we expect changes in surface temperatures and in the interannual variability and seasonal cycle of surface temperature. In general, these features are better simulated in the PMIP2 runs using ICE-5G than in the PMIP1 runs using ICE-4G, with the exception of western European winter temperatures, which are still underestimated compared to pollen data (Kageyama et al. 2006). The altered stationary wave and the physical barrier of this larger ice sheet may also contribute to reduced seeding of the Atlantic jet (section 4b), and hence explain both the strengthening and stabilization observed in the LGM simulation.

b. SST and sea ice

The CCSM3 coupled model provides a self-consistent simulation of the Last Glacial Maximum climate. It produces a climatological SST distribution and sea ice coverage that differ from those determined in CLIMAP and used as boundary conditions in PMIP1 (CLIMAP 1981). At first glance, one might expect the observation-based CLIMAP dataset to provide a more trustworthy picture of the glacial ocean than a model.
CLIMAP produced the first SST maps of the glacial ocean by using transfer functions to translate population distributions of fossil plankton species found in ice age marine sediments (CLIMAP 1981) into sea surface temperatures. The transfer functions (Imbrie and Kipp 1971) were derived from knowledge of how these same plankton species are distributed in today’s ocean. In the intervening decades, however, the proliferation of sediment core data and the advent of new statistical and geochemical methodologies for reconstructing past SSTs have challenged some of the assumptions and results of the CLIMAP method. Moreover, modeling studies have shown that the consequences of such SST errors on the atmospheric circulation are potentially dramatic, whether the errors themselves are in the tropics or in high latitudes (Yin and Battisti 2001; Toracinta et al. 2003; Hostetler et al. 2006).

Though the data still fall short of providing definitive SST and sea ice distributions in the glacial ocean, in regions where there is growing consensus, the CCSM3 simulation shows a clear improvement over the CLIMAP reconstruction. Among the more problematic areas of CLIMAP are the lack of cooling in tropical and subtropical regions, and the extensive sea ice in the North Atlantic (Fig. 12a). These features are very different in the CCSM3 simulation (Fig. 12b), and in fact in the coupled PMIP2 simulations in general (Kageyama et al. 2006; Braconnot et al. 2007, and references therein).

An intercomparison of eight coupled model simulations of the LGM shows a 1°–2°C annual mean cooling over the global tropics, which is in agreement with proxy data (Braconnot et al. 2007, their references). But more relevant than changes in the absolute temperature field in the tropics and subtropics are changes in the distribution of SSTs and SST gradients in the tropics. Evaluating whether CLIMAP or the CCSM3 simulation does a better job in this respect is difficult given the uncertainties in the SST reconstructions and the differences between them. However, several interesting features are robust enough in the simulation and the records to merit mention. At the Last Glacial Maximum, the Pacific warm pool is contracted meridionally in the CCSM3 simulation (Fig. 12), a feature that is consistent with the foraminifera census data (Trend-Staïd and Prell 2002; Kucera et al. 2005); in contrast, CLIMAP has a warm pool that is expanded meridionally compared to the present-day climate. Several studies point to stronger zonal gradients across the tropical Pacific (Lea et al. 2000; Trend-Staïd and Prell 2002; Kucera et al. 2005), but the reconstructed SST patterns themselves are quite dissimilar. Nevertheless, the CCSM3 simulation does exhibit stronger zonal and meridional tropical SST gradients compared to CLIMAP.

In the northern subtropics, the SST simulated by the CCSM3 is more similar to the reconstruction of Trend-Staïd and Prell (2002) than is the CLIMAP product, including more zonal isotherms in the regions of the Kuroshio and Gulf Stream current.

Fig. 12. Sea surface conditions during LGM. (a) CLIMAP reconstruction compared to PD observations. (b) LGM as simulated by the coupled model forced by ICE-5G and 21 ka insolation and greenhouse gases compared to PD observations. (c) LGM compared to CLIMAP. Contours in top panels show annual mean SST (°C), and colors show differences (°C). The white contours mark the 50% sea ice concentration line for August (dashed) and February (solid). Observed SSTs and sea ice are taken from the period 1943–2001 from the extended reconstructed sea surface temperature (ERSST; Smith and Reynolds 2003) and HadISST (Raynor et al. 2003) reconstructions.
Another region in which there are significant differences between the CLIMAP reconstruction and the CCSM3-simulated SSTs is the North Atlantic (Fig. 13). Compared to CLIMAP, the simulated winter SSTs are 3°–5°C warmer along the west coast of Norway and 1°–3°C warmer in a broad swath of the Atlantic Ocean just south of the Greenland–Scotland ridge. These areas remain ice free in winter in CCSM3 and in other PMIP2 simulations (Kageyama et al. 2006), whereas in CLIMAP, perennial sea ice cover extends as far south as 50°N. The presence of open ocean and relatively high temperatures in portions of the eastern North Atlantic and, at least seasonally, in the Nordic Seas is supported by a large body of proxy data, much of which has been synthesized by the Multiproxy Approach for the Reconstruction of the Glacial Ocean (MARGO) project (additional information is available at http://margo.pangaea.de/). These data include estimates of SST and sea ice extent based on coccoliths (Hebbeln et al. 1994), biomarker pigments (Rosell-Melé and Koç 1997), alkenones (Rosell-Melé and Comès 1999), dinoflagellates (de Vernal and Hillaire-Marcel 2000), and foraminifera (Plaumann et al. 2003; Sarnthein et al. 2003; Meland et al. 2005). The simulated North Atlantic conditions in CCSM3 are further corroborated by other coupled atmosphere–ocean models (Hewitt et al. 2003; Shin et al. 2003) and by inferred North Atlantic circulation patterns from foraminiferal assemblages (Lassen et al. 1999).

CLIMAP has very extensive sea ice in the Atlantic, with the winter ice edge reaching at least 45°N over the entire basin (Fig. 12). Consequently, as the storms track northeastward into the Nordic Seas in the PMIP1 simulations, they have access to the strong temperature gradients at the sea ice edge. In the CCSM3 coupled simulation of the LGM, sea ice in the western Atlantic spreads to similarly southern latitudes in winter, but in the eastern Atlantic the ice edge never ventures past 60°N (Fig. 12). With the sea ice so far north, the strongest midlatitude temperature gradient in the eastern Atlantic is now associated with the band of rapidly changing SSTs around 35°–40°N (see Kageyama et al. 2006; Fig. 12). The location and orientation of the jet and storm track in the coupled simulation are consistent with this SST gradient, which clearly points to its importance in determining the state of the Atlantic atmospheric circulation. However, the SSTs are most likely not the whole story. The size of the Laurentide ice sheet upstream seems to exert a large influence on the location of the North Atlantic storm track and jet and on the SST distribution in this region (Li 2007).
6. Concluding remarks

We have presented evidence for an altered atmospheric circulation regime in the Atlantic sector during the Last Glacial Maximum from a global coupled climate model. This LGM Atlantic circulation is characterized by a stronger, more zonally oriented Atlantic jet and yet reduced storminess. In other words, the coupled model produces a glacial climate that is quiescent compared to today’s climate. Analysis of the eddy structures suggests that the barotropic governor is unlikely to play a role in suppressing the storms. Reduced seeding of the Atlantic jet is currently under investigation as the most likely cause of the decrease in storminess during LGM.

The LGM climate simulated by the coupled model is remarkably different from the LGM climate simulated by atmosphere models forced with PMIP1 boundary conditions. Whereas the coupled simulation exhibits a strong, zonal jet with reduced eddies, the uncoupled simulations feature a North Atlantic circulation regime that is similar to the one in today’s climate. We have found that this discrepancy can be traced to differences in the ice sheet topography and sea surface conditions and not to differences in the atmosphere models and model physics. The LGM land ice was updated from the ICE-4G reconstruction in PMIP1 to the improved ICE-5G reconstruction in the coupled simulation. In addition, the SST and sea ice distributions seen by the atmosphere were prescribed to CLIMAP in PMIP1 but calculated by sophisticated ocean and sea ice models in CCSM3. The SST and sea ice distributions simulated by the coupled model are more consistent with the proxy data than are the prescribed SST and sea ice boundary conditions used in PMIP1. Thus, to the extent that changes in atmospheric circulation are linked to changes in orographic forcing and sea surface conditions, the CCSM3 coupled model simulation can be regarded as a better estimate of the LGM climate.

The existence of a different Atlantic atmospheric circulation regime during LGM has implications for our understanding of global heat transport and the stability of glacial climates. In particular, it can offer new outlooks on modes of climate variability that appear to be unique to glacial times, such as the abrupt Dansgaard-Oeschger warming events recorded in Greenland’s ice cores.

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