Causes of Reduced North Atlantic Storm Activity in a CAM3 Simulation of the Last Glacial Maximum

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(Manuscript submitted 25 July 2008, Revised April 06 2009 )

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ABSTRACT

The aim of this paper is to determine how an atmosphere with enhanced mean state baroclinity can support weaker baroclinic wave activity than an atmosphere with weak mean state baroclinity. As a case study, we consider a Last Glacial Maximum (LGM) model simulation that has been previously documented to have reduced baroclinic storm activity, relative to the modern day climate (simulated by the same model), despite having an enhanced mid-latitude temperature gradient. We evaluate several candidate mechanisms to explain this apparent paradox.

We first perform a linear stability analysis on the jet in the modern day and the LGM simulation; the later has relatively strong barotropic velocity shear. We find the LGM mean state is more unstable to baroclinic disturbances than the modern day mean state although the three-dimensional jet structure does stabilize the LGM jet relative to the Eady growth rate. Next, we use feature tracking to assess the storm track seeding and temporal growth of disturbances. We find that the reduction in LGM eddy activity, relative to the modern day eddy activity, is due to the smaller magnitude of the upper level storms entering the North Atlantic domain in the LGM; while the LGM storms do grow more rapidly in the North Atlantic than their modern day counterparts, the storminess in the LGM is reduced because storms seeding the region of enhanced baroclinity are weaker.
1. Introduction

It is often assumed that the amplitude of the mid-latitude storm activity is proportional to the meridional temperature gradient. This concept is often invoked to explain seasonality of storm activity (Lau 1988 and Chang et al. 2002) which, with some notable exceptions, peaks in the winter, when the meridional temperature gradient is greatest. It is also invoked to explain the localized nature of storm activity (i.e., the storm tracks; Hoskins and Valdes 1990 and Frederiksen 1983) and it is often assumed in simple climate models e.g. (North 1978).

The idea that storm activity is proportional to the background state temperature gradient stems from Eady’s and Charney’s linear theories of baroclinic instability (Eady 1949 and Charney 1947) which state that the growth rate of baroclinic disturbances is proportional to the baroclinicity. Non-linear theories of baroclinic adjustment also suggest that as the meridional gradient in insolation increases, storms will transport more heat in order to bring the atmosphere to a state of neutral baroclinic stability (Stone 1978).

There are only a few studies, observational and numerical, that directly test the hypothesis that systems with larger baroclinicity support more storm activity. Observations of the wintertime North Atlantic show a positive correlation between seasonal averaged anomalies in mean state temperature gradient and synoptic activity (Raible 2007), whereas the North Pacific seems to have the opposite correlation in the middle of the winter (Nakamura 1992). The glacial climate system, which is characterized by large land based ice sheets in the Northern Hemisphere high latitudes and an associated enhanced meridional gradient in absorbed solar radiation relative to the modern day climate system,
provides an excellent setting to test ideas about the relationship between baroclinity and storm activity. It is often stated that, in the glacial climate, “high meridional temperature gradients, would have been associated with more frequent development of mid latitude depressions,” e.g., (Dong and Valdes 1998), though paleoclimate data can not directly verify this hypothesis. Furthermore, studies that have examined the storm activity in simulations of the glacial atmosphere have found either subtle changes in storm activity (Justino et al. 2005 and Kageyama et al. 1999 ) relative to the modern day climate state or, in some cases, reduced storm activity (Hall et al. 1996) in the glacial state.

A recent simulation of the Last Glacial Maximum (LGM) climate system was performed with the National Center for Atmospheric Research’s (NCAR) Community Climate System Model (CCSM3): a state of the art, fully coupled climate model (Otto-Bliesner 2006a). In this simulation, the coupled model was forced with insolation from 21 kyears before present, pre-industrial aerosol concentrations, greenhouse gas concentrations derived from ice core data (Otto-Bliesner 2006b) and ICE-5G ice sheet topography (Peltier 2004). Li (2007) performed further analysis of this simulation and found that storm activity was reduced in the LGM North Atlantic relative to the model’s modern day climatology. She also duplicated this result using the atmospheric component of the coupled model (CAM3) with identical forcings and the interactive ocean replaced by the climatological average SST seasonal cycle taken from the coupled model integration; this result indicated that the reduced storm activity during the LGM simulation was a consequence of the atmospheric boundary conditions and not the coupling process although the dynamical mechanism responsible for the reduced eddy activity was left unexplained.
Li and Battisti (2008) highlighted to co-existence of strong North Atlantic baroclinity and weak storm activity in the coupled CCSM3 LGM simulation and argued that this climate simulation is more consistent with paleoclimate indicators than previous LGM climate simulations; the taller and more expansive Laurentide ice sheet in the updated (ICE-5G) land-ice boundary conditions were found to force the atmospheric circulation into a state not realized in previous simulations. In particular, they found that the tropical sea surface temperatures produced in the coupled LGM simulations exhibited a more meridionally and zonally confined Pacific warm pool, consistent with a growing body of paleoclimatic data. Furthermore, the coupled model produces seasonal ice free regions of the Eastern North Atlantic and Nordic Sea where previous model simulations indicated year round sea ice. These ice free conditions are consistent with multiproxy reconstructions of the glacial ocean.

While this simulation is relevant to paleoclimate studies, in the context of the present work, we will examine this simulation as a dynamically self-consistent climate that confounds our first order conceptual understanding of storm track dynamics. Here, we highlight the impact of the LGM boundary conditions on the heat transport and meridional wind variance due to storms: the non-stationary disturbances on time scales between 2 and 8 days (Figure 1). Further definitions of storm activity are provided in Section 2. Diminished storm activity is particularly marked in the North Atlantic Basin where the vertically integrated eddy heat transport was reduced by approximately a factor of two in the basin average. This result is somewhat surprising because the Atlantic jet in the LGM simulation is much stronger and narrower (and more zonally oriented) – implying less stability in the mean flow and more storminess. Indeed, we show below
that in this same region where eddy activity is reduced in the LGM compared to the modern climate the low level baroclinity is enhanced by approximately a factor of two.

Despite the apparent logical simplicity that links a stronger meridional temperature gradient to enhanced storm activity, there are two potential, fundamental, logical flaws in this association that could explain the apparent paradox in this simulation of the glacial climate system. First, factors beyond baroclinity dictate linear stability in a three dimensional system. Second, changes in the seeding of the baroclinically unstable zones can modify the strength of storm activity in the storm track; both initial magnitude and initial structure are relevant (this mechanism was demonstrated, in an idealized context, by Zurita-Gotor and Chang (2005)). A system that starts out small and grows rapidly in a highly unstable system may still end up smaller than a system that starts out large and grows slowly in a more stable system.

Each of the above mentioned processes could potentially break the idealized association between baroclinity and storm activity. In the present work, we will examine the role of each of these processes in the simulation of the LGM climate system and quantify which, if any, of the proposed mechanisms explain the modeled reduction of storm activity relative to the modern day climate system. The model simulation of the modern climate and the LGM climate are discussed in Section 2. Linear stability analysis is applied to the general circulation model’s LGM and modern day mean states to assess the stability of each climate in Section 3. Feature tracking is used to diagnose storm magnitude, frequency, temporal growth and the seeding of the storm track in Section 4. In Section 5, we present an Eulerian analysis of the storm structure and spatial patterns of
variance and link these results to the linear stability analysis and feature tracking results. A discussion section follows.

2. Experiments and Methods

We examine the cause of the reduction in storminess in the North Atlantic LGM simulation discussed in the Introduction. Unfortunately, not all the fields from the coupled integrations were saved at the resolution required for the analysis presented here. Li (2007) showed that the mean state and storm track statistics from the LGM and modern day coupled runs are reproduced in uncoupled model integrations with identical forcings (e.g. ice sheet topography, greenhouse gas concentrations, orbital parameters, etc) to those in the coupled simulation, and the sea surface temperature (SST) is prescribed to be the monthly averaged seasonal cycle taken from the appropriate coupled integration. The LGM eddy statistics and mean state were found to be relatively insensitive to the SSTs used. We therefore choose to present results from the uncoupled runs here.

We compare the LGM run to a modern day run where the model is forced by the observed sea surface temperatures and sea ice concentrations from 1950-1975 and modern day insolation, greenhouse gas concentrations, topography and aerosol concentrations (Hurrell et al. 2006)\(^1\). Sensitivity studies indicate that the statistics of

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\(^1\) While this is not an ideal modern day run for comparison to the uncoupled LGM run because the former will have enhanced low frequency variability due to the inter-annual variability of SSTs (the later is forced by climatological SSTs) there is no impact on the variance at synoptic time scales.
synoptic disturbances in the modern day integration are very similar to those in the modern day climate from the instrumental records (Hurrell et al. 2006).

The LGM and modern day simulations are made using the National Center for Atmospheric Research (NCAR) Community Atmospheric Model version 3 (CAM3), a primitive equation model of spectral resolution T42 on 26 hybrid vertical levels see Collins et al. (2006) for details. At this resolution, the model is found to adequately simulate both the climatology and variability of mid-latitude eddy statistics (Alexander et al. 2006) as assessed by comparison to re-analysis.

We adopt the methodology of Wallace and Gutzler (1981) for isolating the transient eddies from the mean state: the eddy field is defined as the temporally high passed filtered daily data and the mean state as the low passed filtered daily field. Daily averaged fields were high pass filtered using a double-pass Butterworth filter with a cutoff period of 8 days. The wintertime eddy statistics are taken from January 1st to March 31st averaged over 25 seasons. This definition of winter was adopted so the winter season was centered about the month with the maximum in mid-latitude temperature gradient and eddy variance in both runs. It was found that the LGM eddy variance peaked slightly later in the calendar year than their modern day counterpart.

The storm activity is weaker in the LGM compared to that in the modern day run, especially in the Eastern North Atlantic (Figure 1). This conclusion is robust to the choice of cutoff frequency in the temporal filter, the use of instantaneous fields instead of daily averages, the use of a spatial filter instead of a temporal filter, and the choice of field used to quantify the eddy amplitude statistics. Some of the high pass filtered fields show more eddy reduction in the LGM than do others. The eddy sensible heat transport shows
the most reduction followed by meridional wind variance; geopotential variance (Figure 9) shows the least reduction.

In the remainder of this work, we will analyze the cause of the reduced LGM eddy activity, relative to the modern day eddy activity, despite the enhanced mean state baroclinity. The possible explanations we will explore are: enhanced stability due to flow field changes (Section 3), and less frequent and/or smaller magnitude disturbances seeding the Atlantic storm track in the LGM (Section 4).

3. Linear Stability Analysis

A natural starting point for understanding the dynamics of the eddies in the North Atlantic is an examination of the stability of the mean state: will random noise in the atmosphere grow more rapidly on the LGM mean state than on the modern day mean state? To answer this question we calculate the growth rate of the fastest growing normal mode on the climatological mean state. We start by calculating the Eady growth rate (Section 3a) and build towards the three dimensional stability by first including the zonally invariant flow (Section 3b). Finally, we calculate the stability when the three dimensional flow over the Atlantic domain is taken into account in Section 3c.

The mean states for which the linear stability will be assessed are taken to be the climatological average of the fields from January through March, summarized in Figure 2. The upper level jet and lower level baroclinity are stronger throughout the Atlantic region in the LGM run compared to the control run. In the region of maximum baroclinity, the meridional temperature gradient is twice as large in the LGM mean state.
The jet is also narrower in the LGM with an increased barotropic shear across the core of the jet. While the modern jet has a southwest to northeast tilt, the LGM jet is more zonally oriented.

\[ \sigma_{\text{Eady}} = 0.31 \frac{f}{N} \left( \frac{dU_m}{dz} \right). \]  

\[ (1) \]

Eady’s model starts from the quasi-geostrophic equations on an f plane with a rigid lid, and assumes that the eddies develop on a mean state baroclinity and static stability that is constant in space (Pedlosky 1998). While both the baroclinity and static stability have spatial dependence in both CAM3 runs, we can use the Eady model growth rate to quantitatively assess the growth rate of baroclinic disturbances by assuming the local baroclinity and static stability hold over an infinite horizontal domain. Figure 3 shows the Eady growth rate calculated at each grid point using the static stability and baroclinicty of the CAM3 LGM and modern day mean states evaluated between 700 and 550 hPa.

The Eady growth rate in LGM is enhanced by the stronger baroclinity but diminished by the slightly stronger static stability (not shown): changes in the baroclinity play the dominant role in determining the Eady growth rate. Thus, the Eady model of
linear stability suggests that the CAM3 LGM North Atlantic jet is approximately 80% more unstable than its modern day counterpart over the jet core.

**b. Linear Stability of Zonally Invariant Mean States**

The Eady growth rate calculation does not take into account the meridional structure of the jet. Linear theory suggests that the narrowness and enhanced barotropic velocity curvature of the jet in the LGM compared to the modern day could potentially modify the growth rates anticipated from the mean state baroclinity alone (Ioannou and Lindzen 1986 and James 1987) because the growing disturbances experience a mean state with a larger meridional vorticity gradient. Eddies in an enlarged meridional vorticity gradient have less westward tilt with height, which reduces the efficiency of eddy heat transport and the baroclinic energy conversion; hence, it diminishes the linear growth rate relative to the Eady growth rate.

To evaluate the impact of the meridional jet structure on the (baroclinic) stability of the flow, we use a linear, two layer, quasi-geostrophic, beta-channel model (Holton 1992) applied to the mean state Atlantic jet core cross sections at the 450hPa and 900hPa levels for both the LGM and control runs shown in Figure 2. We choose the linear, two-layer quasi-geostrophic model because it is the simplest model of baroclinic instability which is capable of including the effects of the meridional and zonal jet structure, and baroclinic wave development in this model has been found to be very similar to wave development in more complicated primitive equations models (Esler and Haynes, 1999). Channel boundaries were placed at 20° N and 70° N where the prescribed boundary
conditions are zero perturbation flow through the boundary and zero perturbation vorticity at the channel wall. In the zonal direction, the domain is periodic.

The meridional velocity of the mean state was taken to be zero. The mean state zonal velocity and temperature \((U(\lambda, p)\) and \(T(\lambda, p)\)) are taken to be the meridional cross sections at the longitude of maximum jet strength. Additional model runs were performed using zonally invariant mean states corresponding to the jet cross sections with the largest barotropic shear (which are located downstream of the longitude with the maximum jet speed in both the modern day and LGM simulations) and similar results were obtained. The model is initialized with random noise and allowed to run for 30 days with no frictional dissipation. The linear growth of the perturbation is calculated by projecting the perturbation tendency onto the perturbation structure and by calculating the exponential growth of the spatial standard deviation of the perturbation over the second half of the run. Both estimates give comparable results.

An ensemble of five runs for each mean state predicts the most unstable mode has zonal wavelength of order 5,000 km, corresponding to a zonal wavenumber between five and six. The most unstable mode in the LGM climate is slightly more meridionally confined to the jet core (not shown). The linear growth of the LGM run is 0.51 per day as compared to 0.33 per day for the modern climatology. In both cases, inclusion of the beta plane and the meridional structure of the jet, substantially reduces the growth rate from that predicted by the Eady model. Furthermore, using a realistic two dimensional jet structure reduces the growth rate in the LGM more than it does in the modern day mean state. Thus, the enhanced barotropic shear of the LGM jet reduces the growth rate of baroclinic instabilities relative to the modern day run. Nevertheless, the net effect of
baroclinity and the zonally symmetric flow changes renders the two-dimensional mean state substantially more unstable in the LGM than the modern day mean state.

c. Linear Stability of Three-Dimensional Mean States

In this section, we explore the impact of zonal asymmetries (i.e. the localized nature of the jet) on the linear stability of the LGM and modern day mean states. The model used is identical to that discussed in the previous subsection except that now the mean state has zonal structure. We wish to focus on the stability of the Atlantic jet alone, without imposing zonal boundaries that would be physically unrealistic. Therefore, we compose the mean state as follows: (i) the zonally variant jet taken from the GCM over the Atlantic basin (100W to 10W) is placed in the (zonal) center of the linear model domain and composes half the domain, from point B to C in Figure 4; (ii) the Atlantic jet exit (point C) is linearly interpolated into the GCM zonal mean jet (point A on the right side) over the eastern quarter of the domain, between points C and A on the right side of Figure 4; (iii) the GCM zonal mean jet (point A on the left side) is linearly interpolated into the Atlantic jet entrance (point B) over the western quarter of the domain, between point A and B in Figure 4; (iv) the domain is taken to be zonally periodic, such that point A on the left and right side of Figure 4 are the same point. Linear damping is included outside of the Atlantic domain because we want the storm growth to be determined solely by the mean state over the Atlantic.

An ensemble of 12 40 day runs, initialized with random noise, were performed for each the modern day and LGM mean state. The leading mode grows exponentially in time, never converging on a steady solution. The magnitude of the modal (exponential)
growth is estimated as described in Section 3b. The fastest growing mode in the LGM mean state grows 30% faster than its modern day counterpart.

The spatial structure of the most unstable mode on the three dimensional mean states show some differences between the LGM and modern day. In both cases, the zonal wavelength of the most unstable mode is on order 5,000 km, which corresponds to a planetary wavenumber 5 disturbance at a 45° latitude. The upper level disturbance is, on average, displaced 550 km to the west of the lower level disturbance in the LGM run as compared to 700 km in the modern day run. From an energetics perspective, the modal perturbations in each run could grow more rapidly if their structures had more westward tilt with height. However, the difference in the mean state jet renders a reduction in the westward tilt with height of the LGM modes relative to the modern day modes. We can estimate the impact of the changes in vertical eddy structure on growth rate of the most unstable mode by noting that, in 2-layer quasi-geostrophic theory, the magnitude of the heat transport relative to the geopotential perturbation magnitude scales as the sine of the phase difference between the upper and lower level perturbations. Using the vertical tilts from above, we calculate that, per unit of perturbation magnitude, the LGM baroclinic mode will transport 30% less heat than its modern day counterpart. We can think of this result as a modification to the anticipated Eady growth rate, since the baroclinic generation that is needed for the eddy potential energy reservoir to grow is linearly proportional to the eddy heat transport. Therefore, the Eady mode expectation that the LGM mean state is 80% more unstable to baroclinic disturbances than the modern day mean state, should be reduced by approximately 30% to take into account the changes in eddy structure in the three dimensional mean states. This modified Eady growth rate
suggests that the LGM mean state is approximately 25% more unstable, in general agreement with the numerical results of the three-dimensional model.

An alternative approach to estimating the stability of the mean state is to follow the center of each perturbation (in the ensemble of 2-layer, linear, quasigeostrophic model runs) as it moves through the domain, keeping track of the magnitude as a function of time and location. The instantaneous linear growth rate of each perturbation is calculated as

\[
\frac{1}{Z'} \frac{DZ'}{Dt}
\]

where the tendency is calculated in the reference frame moving with disturbance, using finite differencing in time. The ensemble average growth rate as a function of location (Figure 5) suggests that linear perturbations grow more rapidly throughout the Atlantic when the 2-layer model is linearized about the LGM mean states, as compared to the modern day mean state. In both runs, the maximum linear growth rate is achieved at approximately 70°W, at the East coast of the United States and approximately 10 degrees of longitude upstream of the zone of maximum baroclinity. Calculating the eddy growth rate in this fashion is directly comparable to the estimates of eddy growth rates in the full model simulations that will be presented in Section 4.

4. Feature Tracking Eddy statistics

The results of the previous section suggest that the growth rate of the fastest growing mode is enhanced in the Atlantic in the LGM run (relative to the modern day
run). We now ask: does this expectation come true in the GCM simulations of the LGM and modern day? That is, in the ensemble mean, do disturbances grow more rapidly (in time) in the LGM simulation (as compared to the modern day simulation).

To answer this question, we apply feature tracking analysis: identifying and following synoptic features in the model simulations as they move through the domain and keeping track of their magnitude and location. The feature tracking approach addresses storm growth more directly, as the temporal development of each individual perturbation can be calculated and ensemble mean behavior can then be assessed.

Though, in the present climate, areas of enhanced Eulerian variances are associated with the areas of high feature tracking densities (Hoskins and Hodges 2002), the two perspectives address different conceptual definitions of a storm track. From the Eulerian perspective, a storm track is a region of enhanced variance in a band pass filtered field. This definition assumes that large variance is associated with organized synoptic systems passing through the domain. In actuality, the Eulerian methodology does not distinguish between (i) random, unorganized noise, (ii) frequent but small, organized synoptic systems, and (iii) infrequent, large, organized storms. The feature tracking framework is better for determining whether (changes in) eddy variance reflect (changes in) the number of storms, the magnitude of the storms, and the growth and decay of storms within a given region. The feature tracking perspective is viewed here as a complimentary perspective to the Eulerian analysis (Section 5).

In this section, we first discuss the algorithm we use to identify the regions of growth and decay of storms (i.e. band pass filtered systems) and apply it to evaluate
differences in the seeding of the baroclinically unstable regions in the CAM LGM and modern day runs.

\textbf{a. Tracking Methodology}

The tracking algorithm we use was developed by Kevin Hodges and is well documented in Hodges (1994, 1995, and 1999). In this analysis, we will track features in the 2-8 day band-pass filtered geopotential height field at 300 hPa and 700 hPa. We elect to use a temporal filter, as opposed to the spatial filter suggested by other studies (Anderson et al. 2003) prior to application of the tracking algorithm because the commonly used spatial filters do not remove large magnitude, stationary geopotential features in the mid-latitude North Atlantic domain and therefore artificially bias cyclone magnitudes to high values on the poleward side of the jet (Donohoe and Battisti, 2009). Furthermore, we chose to use the temporal band pass filter for feature tracking analysis to be consistent with the definition of storm activity presented in the other sections of this study. Daily band pass filtered fields at 300 hPa and 700 hPa are put into the tracking algorithm. We identify a feature if it achieves a threshold of 30 (60) meters geopotential height at 700 (300) hPa. We define a track only if the feature travels more than 200 km and exists for more than three days. Only tracks over the winter season (November-March) are discussed below.

\textbf{b. Tracking Results}

Storm track statistics for the LGM and modern day integrations are shown in Figure 6. All averages are calculated from 25 winters of data. (All results in this Section
are also found using random sub-samples of 12 winters of data). The growth rate of each individual perturbation is calculated using equation (2) and downstream, first order finite differencing. Individual storms show large variability in growth rate and amplitude at each longitude and some individual systems are not displayed on the plots in Figure 6.

At 700 hPa fewer storms enter the western Atlantic (longitude 100°W to 70°W, the region where storms grow rapidly), in the LGM compared to the modern day. In contrast, the modern day run has more lower level disturbances entering the Atlantic that subsequently grow relatively weakly compared to the LGM case, if at all. The average growth rate is actually negative over the jet entrance in the modern day run, reflecting a somewhat bimodal storm growth distribution in this region, consisting of weakly growing storms and more rapidly decaying storms (not shown). It is noteworthy that, from a feature tracking perspective, there is no reduction in storm geopotential magnitude at 700 hPa as LGM and modern day systems have comparable magnitudes in the Western Atlantic and LGM systems tend to be larger over the Eastern Atlantic.

The results at 300 hPa show a large deficit of storms in the LGM run in the Atlantic jet entrance region as well as a reduction in the average amplitude of the storms. On average, systems grow much more rapidly over the baroclinic region in the LGM than the modern day simulation (both temporally and spatially) and achieve comparable magnitudes on the eastern side of the Atlantic (~40°W). This result agrees well with the band pass filtered heat transport (Figure 1) which shows a deficit of eddy heat transport over the western Atlantic in the LGM run and an enhancement in eddy heat transport in the eastern Atlantic during the LGM compared to the modern day run.
In addition to the average storm having a smaller magnitude, the middle panel of Figure 6 also indicates that there are fewer storms seeding the jet entrance region (longitude 90°W to 70°W), during the LGM. The number, size and spatial structures of storms in this region are critical to the storm statistics in the Atlantic because this region is just upstream of the greatest baroclinic region, where the strongest storm growth occurs (Section 3c). We can explore this idea further by looking at a histogram of the size of disturbances that are advected into the Atlantic domain at 300 hPa, shown in the top panel of Figure 7. The number of storms that enter the Atlantic domain is 17% higher in the modern day run relative to the LGM run and the distribution of storms entering the domain shifts towards larger magnitudes. The mean of the distribution is 43% larger for the modern day run compared to the LGM run. Thus, even though the LGM storms grow more rapidly as they traverse across the Atlantic, they start with very weak upper level seeds relative to their counterparts in the modern day run.

The distribution of storm growth is shown in the middle panel of Figure 7 as a probability distribution function. Only storms that grew for at least one day after they entered the Atlantic domain were considered. This eliminated approximately 30% of identified features in the Atlantic domain that were either decaying upon entrance or spent less than two consecutive days in the Atlantic. There is a clear shift of the modal growth rate of disturbances during the LGM at both 300 hPa and 700 hPa (not shown) towards higher growth rates. This result is in qualitative agreement with the conclusions of Section 3.

The feature tracking framework is arguably the best way to determine the growth rate of storms. The growth histogram indicates that there is no single growth rate of
synoptic disturbances and the modal value of growth rate is consistent with the linear growth rate of instabilities on a three-dimensional mean state: 0.35 and 0.25 per day for the LGM and modern, respectively. The spread in growth rates (as well as the growth rates that exceed that of the most unstable normal mode) likely results from a combination of spatial variations in the initial disturbances that project differently on the most unstable linear mode, non-linear process not captured in our linear model, and diabatic heating and frictional effects.

The probability distribution function of the duration of storm growth (shown in the bottom panel of Figure 7) indicates a slight shift towards greater growth duration within the Atlantic for the LGM run relative to the modern day run. This result agrees with the idea that the storms in the modern day run are more mature and closer to beginning their decay phase upon entrance to the Atlantic domain. This tendency, along with the enhanced LGM growth rate would act to increase the amplitude of the disturbance within the Atlantic domain. Nonetheless, the upper panel of Figure 7 indicates that, over most of the Atlantic domain, the weaker amplitude and smaller number of the seeds in the LGM has a greater effect on the maximum amplitude obtained by the storms than does the increased growth rate and growth period of the disturbances.

To more formally test this conclusion, we can multiply the 300 hPa distribution of seeds in each run by the probability distribution function of storm growth and growth duration to determine an expected distribution of maximum storm size. Here, we are assuming that upper level geopotential disturbances entering the storm track are the major source of perturbations to the baroclinic zone. The ensemble of the expected behaviors at 300 hPa for the ensemble of all 300 hPa seeds gives the expected distribution function of
maximum obtained storm size within the Atlantic. Those results (not shown) qualitatively reproduce the 300 hPa storm size distribution observed in the GCM with larger average storms in the modern day run relative to the LGM run. This suggests that the reduced upper level perturbation size in the Atlantic storm track in the LGM relative to the modern day results from weaker and less frequent upper levels seeding despite the enhanced growth rate.

5. Eulerian Analysis of Storms in the CAM3 Runs

In this section, we analyze the structure of the eddies and the Eulerian variance of band pass filtered fields and look for further evidence of the processes discussed in sections 3 and 4. In particular, we ask if the changes in eddy structure between the LGM and modern day simulations conform to our expectations of reduced vertical tilt and therefore less efficient baroclinic generation of energy identified in section 3. We also ask if we can reconcile the Eulerian variance maps for each simulation with the conclusion from the feature tracking analysis (Section 4) that eddies grow more rapidly in the LGM simulation.

a. Eddy Structures from One Point Correlation Maps

Band pass filtered Eulerian fields can be used to assess the structure of eddies by use of a one point correlation map (Wallace and Gutzler 1981 and Blackmon et al. 1984). The analysis is done on the JFM band pass filtered meridional wind. The reference point for regression is on the 550 hPa pressure surface in the jet entrance region at the latitude of maximum variance and is shown with a black star in each panel of Figure 8. The time
series at the reference point is regressed onto the time series of the same field at all locations on both the 700 hPa and 400 hPa pressure surfaces. This method allows both the horizontal and vertical structure of the eddies to be visualized.

The eddies are more meridionally confined in the LGM than in the modern day simulation, consistent with a narrower zone of baroclinity in the mean state and the more meridionally confined Eulerian band pass variances. The zonal wavelength of the eddies is approximately 4,000 km and is nearly unchanged between runs. On the eastern side of the Atlantic, the LGM eddies remain more coherent with the perturbation at the reference point, suggesting that the storm track remains somewhat organized into Western Europe in the LGM run while the modern run storm track looses its organization by this point. This is consistent with the spatial differences in the Eulerian eddy variance (Figure 1 and Figure 9).

The vertical structure of the eddies shows the anticipated westward tilt with height that is characteristic of growing baroclinic eddies. A quantitative analysis of the vertical tilt was preformed by taking the spatial map of the one point correlation field at 700 hPa and shifting it west until a maximum correlation with the one point correlation field at 400 hPa was achieved. This analysis found the vertical tilt of synoptic eddies between 700 hPa and 400 hPa was 3.3 degrees longitude for the LGM and 4.7 degrees longitude for the modern day run. For reference, these vertical tilts are approximately 50% less than the vertical tilts of the most unstable modes on the three dimensional LGM and modern day mean states respectively (Section 3c). Qualitatively, the decrease in the vertical tilt of the most unstable linear mode in the LGM compared to the modern day mean state is consistent with the differences in the vertical tilt shown in the regression maps. This
suggests that the three dimensional jet structure in the LGM dynamically limits the ability of the eddies to extract energy from the enhanced baroclinity of the mean state and provides a partial explanation for why the storm activity in the LGM relative to the modern day does not follow our expectations based on the Eady model, but it does not explain the reduced storm activity in the LGM.

b. Linear Growth Rate from Eulerian Variance

From the Eulerian perspective, the tendency of any eddy field must be zero in the time mean in order for the field to remain finite. Therefore, any processes that support eddy growth locally must be compensated by advection of the eddy variance plus dissipation. In this framework, regions that support eddy growth lead to strong spatial gradients in eddy variance leading to large variance downstream of the regions of growth; the ensemble mean growth rate fields should be in quadrature phase with the Eulerian variance field.

Here, we attempt to quantify the relationship between the spatial gradient of Eulerian variance and the magnitude of ensemble mean eddy growth in the reference frame moving with the passing disturbances. We assume that the geopotential disturbances are modal (shape preserving and exponentially growing) and propagate with the same zonal phase speed at all levels with the functional form

\[ Z'(x, y, t) = F(y)e^{i(ks - \sigma t + \sigma x)} \]  

where \( \omega/k \) is the zonal phase speed of the disturbance and \( \sigma \) is the linear growth rate.

The local (Eulerian) and feature tracking reference frames are related by

\[ \frac{DZ'}{Dt} = \frac{dZ'}{dt} + \frac{\omega}{k} \frac{dZ'}{dx}. \]
The left hand side denotes the tendency of the eddy geopotential ($Z'$) in the reference frame following the disturbance. We can convert this to an ensemble mean linear growth rate (equation (2)) by dividing by $Z'$ and averaging in time (over an ensemble of passing features). The right hand side of the equation can be manipulated by multiplying both terms by $Z'/Z'$ and by using the identities

$$ Z' \frac{dZ'}{dx} = \frac{1}{2} \frac{dZ'^2}{dx} \quad \text{and} \quad Z' \frac{dZ'}{dt} = \frac{1}{2} \frac{dZ'^2}{dt} \quad (5) $$

prior to time averaging (averaging over an ensemble of perturbations). The resulting expression for the ensemble mean feature growth rate is

$$ \overline{\frac{1}{Z'} \frac{dZ'}{dt}} = \left( \frac{\omega}{2k} \right) \overline{\frac{1}{Z'^2}} \frac{d(Z'^2)}{dx} + \left( \frac{1}{2} \right) \overline{\frac{1}{Z'^2}} \frac{d(Z'^2)}{dt} \quad (6) $$

where overbars denote time averages. The second term on the right is zero and thus the ensemble mean growth of disturbances can be calculated from the spatial map of the Eulerian variance if the phase speed of the disturbances is known. Here, we assume the zonal phase speed is vertically and temporally invariant and equal to the zonal velocity at the steering level, which will be taken to be 700hPa (James 1995). Though the above derivation assumed that the perturbations consisted of a single waveform, the result holds for a superposition of traveling waves, provided they are non-dispersive.

The variance in the vertically integrated JFM geopotential height perturbations are assessed over the core of the jet by taking the maximum three gridpoint values at each longitude (this definition avoids topographical domain discontinuities); the zonal gradient in the geopotential height variance is assessed as the downstream finite difference of the field averaged over the jet core. The zonal velocity at the steering level is evaluated as the
mean of the 700 hPa zonal velocity at the same three latitudes (at each longitude) used to evaluate geopotential height variance. Figure 9 shows both the spatial map of the geopotential variance and the linear growth rate calculated using Equation (6).

In the vicinity of the jet core, the analysis of the spatial derivative of the geopotential variance suggests that the growth rate is enhanced during the LGM relative to the modern day. These analysis further supports the conclusions reached in Section 2, that the LGM mean state is more unstable and suggests that the linear growth rate calculated from quasi-geostrophic theory may be applicable to storm growth in CAM3. The conclusions reached here also coincide with the feature tracking results. More importantly, this result further emphasizes that the magnitude of eddies is not dictated solely by their growth rate; it also depends critically on the magnitude of the pre-existing disturbances as they enter the core of the jet, where the mean state is most unstable.

6. Discussion

A recent simulation of the Last Glacial Maximum climate system has challenged our understanding that a climate system with a stronger meridional temperature gradient will support more storm activity. We have discussed several processes that could cause a climate system with a stronger temperature gradient to have less storm activity and we have assessed the role of these processes in the LGM and modern day climates simulated by CAM3.

In the introduction, it was stated that a climate system with a stronger temperature gradient but weaker storms must violate one of two common assumptions:

1.) The instability of the mean state is enhanced
2.) The size and frequency of systems seeding the baroclinic region has not changed systematically.

Several lines of evidence suggest that assumption 1 is valid and that assumption 2, however, seems to break down.

*Assumption 1*: The foundation of the idea that storm activity is proportional to the meridional temperature gradient stems from Eady’s and Charney’s models of baroclinic instability which have the growth rate of synoptic disturbances being proportional to the temperature gradient. These models are linearized about a one dimensional mean state. We applied the same (quasi-geostrophic) physics as in the Eady and Charney models to evaluate the stability of more realistic mean states, representative of the modern and LGM climate system. We found that the enhanced static stability, narrowness of the jet, and finite length of the jet all act to systematically decrease the instability of the LGM climate state relative to the modern day climate state. Nonetheless, our analysis of the spatial structure of the Eulerian variance and the stability of the leading normal mode suggests that the stronger temperature gradient seen in the CAM3 LGM simulation supports increased baroclinic growth, compared to the modern day simulation.

The best way to estimate storm growth is in a feature tracking frame of reference: follow the storms as they move through the region of high baroclinity (the jet region), and measure the change in amplitude as a function of time. We performed these, storm-centric, analyses for the modern day and LGM experiments and confirmed that storms grew more rapidly in the LGM run. Results obtained using this feature tracking approach
agreed well with the linear stability analysis and the analysis of the spatial gradient of Eulerian variance. *Hence, we can conclude, quite confidently, that the LGM mean state is more baroclinically unstable than the modern day mean state.*

**Assumption 2:** The feature tracking analysis shows there are fewer storms seeding the low levels of the Atlantic and both fewer and significantly smaller magnitude storms seeding the upper levels. Hence, although storms grow more rapidly in the LGM (and last longer), the number and magnitude of initial disturbances seeding the baroclinic region are fewer and smaller amplitude than in the modern climate. The later plays a greater role. *Hence, the explanation for the diminished eddy activity in the LGM Atlantic relative to the modern climate is weaker seeding of the baroclinic zone (entrance of the jet), relative to the modern day run.*

Several other mechanisms have been proposed to explain diminished eddy activity in the LGM climate system. A series of idealized general circulation model runs by Caballero and Langen (2005) and O’Gorman and Schneider (2008) have shown that eddy kinetic energy is reduced in climate states with cold mean temperatures. We find that this mechanism can not explain the reduced LGM eddy activity in our simulation; calculations (not shown) indicate that the expected reduction in eddy activity resulting from the colder mean temperature is overwhelmed by the expected enhancement of eddy activity resulting from the enhanced meridional temperature gradient. Furthermore, CCSM3 LGM integrations with a smaller Laurentide ice sheet have simulated cold mean temperatures and enhanced storm activity (Li, 2007) indicating that the reduced storm
activity discussed in this paper is a result of the specified ice sheet topography and not the climate system’s mean temperature.

It has also been proposed that a narrower, more barotropic jet such as the North Atlantic LGM jet, should lead to more efficient barotropic conversion of eddy kinetic energy to mean state potential energy (Feldstein and Held 1989, Simmons and Hoskins, 1980, and Nakamura 1993), the primary mechanism by which synoptic eddies decay (Oort 1964). This enhanced decay mechanism, could lead to a diminished LGM kinetic energy reservoir just as a bucket with a large hole holds less water in equilibrium with a constant inflow rate than its counterpart with a smaller hole. An analysis of the eddy kinetic energy budget (not shown) in the LGM and modern day simulation suggests that differences in the barotropic conversion are small and work in the wrong direction to explain the reduced eddy activity in the LGM (there is more barotropic decay, averaged over the North Atlantic in the modern day simulation as compared to the LGM). This result agrees with Laine et al. (2008) who also preformed an eddy kinetic energy budget on a suite of LGM simulations and found weaker barotropic decay in the LGM simulations (A. Laine, personal communication, 1/30/2009).

It is also possible that the relative stability of the LGM and modern day jets is inadequately assessed by the linear stability analysis used in this study. For example, Swanson and Pierrehumbert (1994) find that the largest amplitude non-linear disturbances on a baroclinic jet are leading edge waves with group speeds comparable to the upper level jet speed. However, the group speed of the leading edge wave has been found to be a discontinuous function of the potential vorticity gradient (Goodman et al. 2002). Because the LGM potential vorticity gradient is strongly enhanced relative to the
modern day, it is possible that LGM storms inhabit a different regime of non-linear wave packet evolution. However, we have shown that the ensemble growth of storms in LGM and modern day simulations is consistent with the predictions of linear stability theory; Ockham’s razor suggests that there is no need to reach for non-linear theories to explain the reduced LGM eddy activity.

We have given some thought to performing some GCM experiments with additional seeding upstream of the jets in the LGM world to further test our conclusion that the diminished eddy variance is due to reduced seeding at the jet entrance. There are numerous obstacles to designing such experiments not least of which is how to determine the forcing required to maintain the three dimensional mean state and how to accurately prescribe the spatial-temporal structure of the additional seeding.

The dynamical cause for reduced seeding in the LGM is currently under investigation. In the modern climate, the seeding of the Atlantic storm track includes upper level disturbances that come from the debris of storms exiting the Pacific storm track (Hoskins and Hodges 2002). In the LGM simulation, the large Laurentide ice sheet sets up a strong stationary wave pattern featuring a ridge over North America. Our leading hypothesis is that this circulation steers upper level disturbances poleward of the Atlantic baroclinic zone, reducing the connection between the Pacific and Atlantic storm tracks. A second hypothesis is the Laurentide Ice sheet inhibits the formation of upper level disturbances being shed off the northern end of the Canadian Rockies, commonly referred to as Alberta Clippers (Mercer 2007) in the LGM run.
Acknowledgments. We thank Camille Li for the use of her CAM3 model runs and her advice and insight. We thank Kevin Hodges for the use of his storm tracking algorithm and Marc Michelsen for his technical support. This manuscript benefited from the suggestions of three anonymous reviewers. This work was supported by the Gary Comer Abrupt Climate Change Fellowship, the National Science Foundation’s Graduate Student Fellowship program and National Science Foundation Grant ATM-0502204.
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Figure Captions

Figure 1. Vertically averaged band pass filtered meridional heat transport (colors, left panels, K m/s) and mean state zonal velocity at 400 hPa (contours, in m/s) for the LGM run (top panel), modern run (bottom panel). Vertically integrated, band pass filtered merdional velocity variance (colors, right panels, m^2/s^2) and mean state zonal velocity at 400 hPa (contours, in m/s). Results are shown for the January, February, March (JFM) season.

Figure 2. Vertical cross section of zonal velocity (contours, in 10 m/s intervals) and temperature (colors, in Kelvin) for the longitude of maximum zonal velocity (left panel) for both the modern day and LGM JFM mean states (Li, 2007). Plan view of the thermal wind between 900 hPa and 450 hPa (m/s) over the Atlantic domain (right panel).
Figure 3. Eady growth rate (colors in fractional growth per day) and 400 hPa zonal velocity (contours, in m/s) for the LGM (upper panel), the modern day simulated mean states (lower panel).

Figure 4. Zonally periodic mean state Atlantic jet zonal velocities used in the 2-layer linear stability model. The upper level (450 hPa) LGM mean state is shown for illustrative purposes. Point A (on both the left and right side) is the zonal mean jet. Point B is the Atlantic jet entrance. Point C is the Atlantic jet exit. See text for a discussion of how the domain is constructed.

Figure 5. Ensemble growth of central geopotential height perturbations (fraction per day) in the 2-layer quasigeostrophic model linearized about the LGM and modern day mean states as a function of zonal location for the ensemble of LGM (dotted line) and modern day (solid line) runs.

Figure 6. Feature tracking magnitude of storms (measured by central geopotential, top panel), number of identified storms (middle panel), and temporal growth rate of storms (lower panel) at 700 hPa (left panel) and 300 hPa (right panel). The lines in the top and bottom panels are the mean at each longitude: the red lines are for the modern day run and the blue lines are for the LGM run. Small dots are values for individual storms; LGM individual storms are displaced 1° longitude to the East for visual purposes. Some individual storms are outside of the plotted domain.
Figure 7. Histogram of storm magnitude entering the Atlantic domain (i.e., crossing 100°W of longitude between 30°N and 60°N) at 300 hPa for LGM (solid line) and modern day (dotted line) runs (top panel). Probability distribution functions of storm growth during the growing phase (middle panel) and time spent in growing phase (lower panel).

Figure 8. JFM one point regression maps of band pass filtered meridional velocity on to the 550 hPa meridional velocity at the reference point (shown with a black dot). The colored field is the meridional velocity at 700 hPa and the contoured field is the same quantity at 400 hPa. The upper panel is the LGM run and the lower panel is the modern day run. For each experiment, the regression values reflect a one standard deviation perturbation in the reference point velocity.

Figure 9. Bandpass filtered JFM vertically averaged geopotential height variance for the LGM run (top) and modern day (middle) runs (m²). Contours in the upper panels denote the climatological JFM zonal wind at 400 hPa. The bottom panel shows the linear growth rates calculated from the upper panels via Equation (6).
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