Stationary Wave Reflection as a Mechanism for Zonalizing the Atlantic Winter Jet at the LGM

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ABSTRACT

Current estimates of the height of the Laurentide Ice Sheet (LIS) at the Last Glacial Maximum (LGM) range from around 3000 to 4500 m. Modeling studies of the LGM, using low-end estimates of the LIS height, show a relatively weak and northeastward-tilted winter jet in the North Atlantic, similar to the modern jet, while simulations with high-end LIS elevations show a much more intense and zonally oriented jet. Here, an explanation for this response of the Atlantic circulation is sought using a sequence of LGM simulations spanning a broad range of LIS elevations. It is found that increasing LIS height favors planetary wave breaking and nonlinear reflection in the subtropical North Atlantic. For high LIS elevations, planetary wave reflection becomes sufficiently prevalent that a poleward-directed flux of wave activity appears in the climatology over the midlatitude North Atlantic. This entails a zonalization of the stationary wave phase lines and thus of the midlatitude jet.

1. Introduction

The Last Glacial Maximum (LGM)—about 25 000–18 000 yr before present (25–18 kyr BP)—marks the apex of the last glacial cycle (about 115–12 kyr BP), when massive ice sheets expanded over the Northern Hemisphere continents. The Laurentide Ice Sheet (LIS) was by far the largest and covered most of the North American continent poleward of ~40°N. While the horizontal boundaries of the LIS are well constrained by field observations (e.g., Kleman et al. 2013), only much weaker constraints can be placed on its thickness. As a result, reconstructions of the LIS’s maximum elevation range from around 3000 to 4500 m [see Ullman et al. (2014) and references therein]. Simulations of the LGM using a range of climate models robustly show that topographic variations of this magnitude have a major impact on the atmospheric circulation over the North Atlantic, with the jet becoming stronger and more zonally oriented as the LIS elevation increases (Li and Battisti 2008; Kageyama et al. 2013; Ullman et al. 2014; Merz et al. 2015; Pausata and Löfverström 2015). The mean tilt of the jet axis is of fundamental importance to the climate of the North Atlantic and Europe, as it determines the orientation of the storm track and has a major impact on the distribution of surface temperature, precipitation, and wind. In the modern climate, the jet and the surface wind stress maximum extend northeastward across the Atlantic basin, yielding an ocean gyre circulation that efficiently carries warm and saline surface water poleward to the subpolar North Atlantic, contributing to the maintenance of high salinities and a vigorous meridional overturning circulation (Warren 1983; Czaja 2009; Nilsson et al. 2013). Changes in the tilt of the Atlantic jet could thus also have major impacts on the ocean circulation during the LGM.

Here, we conduct a series of LGM simulations spanning a wide range of LIS elevations (section 2) with the aim of exploring the mechanisms leading to the strengthening and zonalization of the Atlantic jet. We find that the stationary planetary waves excited by the LIS topography play a central role in governing this response. The stationary waves excited by midlatitude topography and diabatic heating propagate...
predominantly equatorward and are absorbed in the subtropics (Held 1983; Held et al. 2002). These equatorward-propagating planetary waves have southwest–northeast-tilted phase lines, which contribute to the observed poleward tilt of the modern Atlantic jet (Brayshaw et al. 2009). When the undulations in the potential vorticity (PV) field induced by a Rossby wave become large—as typically happens near a critical line, where the wave’s phase speed matches the background wind speed—the wave may break, irreversibly mixing and locally homogenizing the PV distribution (McIntyre and Palmer 1983). This generally results in the wave being absorbed and giving up its wave activity to the mean flow, but sometimes it is instead reflected back into the midlatitudes. Such nonlinear wave reflection has been found in idealized (Brunet and Haynes 1996; Wang and Kushner 2010; Magnudottir and Haynes 1999) and full-complexity models (Walker and Magnudottir 2003) forced by sufficiently high topography. It has also been detected in observations (Abatzoglou and Magnudottir 2004, 2006b); it turns out that planetary wave reflection is fairly common, occurring as often as every third wave breaking event in the Northern Hemisphere. The poleward-propagating reflected waves have phase lines that are tilted in the opposite direction to the incident waves. The orientation of the climatological stationary wave phase lines thus depends on the relative strength of the equatorward- and poleward-propagating wave bundles; if their amplitudes are comparable, the mean phase lines will be roughly zonal in the region of overlap.

As will be shown in sections 3 and 4, this is what happens in the North Atlantic in the LGM simulations, leading to the zonalization of the jet. Section 5 provides a time-varying perspective on the relation between wave breaking and nonlinear reflection. Section 6 discusses the relative roles of topography and sea ice boundary conditions and the impact of horizontal resolution for our results, while section 7 deals with the response of the Atlantic storm track. Some further implications of the results are discussed in section 8, and section 9 summarizes our conclusions.

2. Experiments and methods

2a. Model and setup

We employ the National Center for Atmospheric Research (NCAR) Community Atmosphere Model, version 3.0 (CAM3.0) (Collins et al. 2004, 2006), using a spectral dynamical core with T85 (~1.4°) horizontal resolution and 26 hybrid sigma–pressure levels in the vertical. The atmospheric model is coupled to a mixed-layer ocean (implying interactive SSTs and sea ice extent) with a specified, monthly varying ocean heat transport convergence field (i.e., a $q$ flux) derived from an equilibrated LGM simulation (Brandefelt and Otto-Bliesner 2009) with the fully coupled NCAR Community Climate System Model, version 3 (CCSM3). We use the LGM ice sheet reconstruction presented by Kleman et al. (2013) in which the LIS has a maximum elevation of around 3400 m. Orbital parameters and greenhouse gas concentrations are set to their estimated LGM values, while the vegetation cover in nonglaciated areas is prescribed as the modern distribution. The reader is referred to Löffverström et al. (2014) for further details about the simulation setup.

We conduct a sequence of simulations all using identical LGM conditions but with increasing height of the North American and Eurasian ice sheets (Greenland is kept at a constant height). The height is varied by scaling the reference ice sheet topography by a uniform constant so that the horizontal extent and morphology do not change. The height of the LIS in our simulations ranges from approximately 1700 m in the lowest case up to 5100 m in the highest case (which corresponds to a scaling factor of 0.5 and 1.5, respectively). A simulation with zero-elevation ice sheets (scaling factor 0.0) was also carried out in order to qualitatively isolate the albedo effect of the ice sheet. To provide a present-day baseline climate for comparison, we also carried out a simulation using modern topography and prescribed monthly varying climatological sea surface temperature and sea ice taken from observations. The analysis is based on 30-yr winter climatologies [December–February (DJF)], calculated after the model climate has reached statistical equilibrium.

2b. Wave activity flux

We use the wave activity flux diagnostic introduced by Plumb (1985), which is a generalized Eliassen–Palm flux (Eliassen and Palm 1961; Andrews and McIntyre 1976) indicating the propagation of stationary waves in three dimensions. It is applicable in the Wentzel–Kramers–Brillouin–Jeffreys (WKBJ) limit, a method for finding approximate solutions to linear differential equations of almost-plane quasigeostrophic waves on a slowly varying background flow. The flux vectors are approximately parallel to the group velocity and orthogonal to the phase lines of the stationary waves. Note that the wave activity flux derived by Plumb (1985) is a three-dimensional vector quantity, though we only consider the horizontal components in this study.
\[
F_{x(y)} \sim \left\{ \frac{1}{2a^2 \cos^2 \phi} \left[ \left( \frac{\partial \psi^*}{\partial \lambda} \right)^2 - \psi^* \frac{\partial^2 \psi^*}{\partial \lambda^2} \right] - \frac{1}{2a^2 \cos \phi} \left[ \frac{\partial \psi^*}{\partial \lambda} \frac{\partial \psi^*}{\partial \phi} - \psi^* \frac{\partial^2 \psi^*}{\partial \lambda \partial \phi} \right] \right\}, \tag{1}
\]

where \(\psi^*\) is the zonally asymmetric component of the streamfunction, \(\phi\) and \(\lambda\) are the latitude and longitude, respectively, and \(a\) is the radius of Earth.

c. Rossby wave breaking analysis

Rossby wave breaking is detected using the algorithm presented in Barnes and Hartmann (2012), which identifies overturning absolute vorticity contours on pressure surfaces; the algorithm yields similar results to those of methods based on isentropic potential vorticity. A wave breaking event is identified when a single meridian intersects an overturning absolute vorticity contour three times. All overturning vorticity contours within 500 km of one another are considered to be part of the same wave breaking event, and if several overturnings have geographical centers within 2000 km of one another, they are grouped under a single wave-breaking event. The location of the wave breaking event is then defined as the centroid of all points on all absolute vorticity contours that are part of the episode. The algorithm further enforces a minimum 5° longitudinal width of the overturning to qualify as a wave breaking event. Cyclonic and anticyclonic wave breaking (respectively, CWB and AWB) events are identified separately by determining the direction in which the vorticity contours overturn. The absolute vorticity field is truncated at T15 to reduce noise in the unfiltered fields (e.g., Postel and Hitchman 1999; Masato et al. 2012) and to ensure that only overturnings of planetary scale are detected. Reasonable changes in the truncation threshold were found not to affect the wave breaking distributions produced by the algorithm (Barnes and Hartmann 2012). We base this part of the analysis on 10 yr of daily wave breaking data computed from high-frequency (6 hourly) instantaneous wind on the 300-hPa surface.

3. Climatological jet and PV

The climatological circulation response to increased ice sheet elevation in the Atlantic sector is depicted in Fig. 1, which compares the observed winter climatology [as represented by the ERA-Interim product (Dee et al. 2011)] with the present-day and LGM simulations. As is well known, the present-day Atlantic midlatitude jet (Fig. 1a) is relatively weak, distinctly tilted in the southwest–northeast direction, and well separated from the subtropical jet over North Africa, yielding a double-jet structure in the eastern Atlantic basin. The associated PV field (Fig. 1b) presents a generally weak meridional gradient between the subtropics and mid-latitudes, though somewhat steeper in the western Atlantic, where the jet is stronger. The present-day simulation captures all these qualitative features (Figs. 1c,d), though the midlatitude Atlantic jet is generally about 10 m s\(^{-1}\) too strong and slightly more zonal than in observations.

The LGM simulations (Figs. 1e–j) show a dramatic response to increasing ice sheet elevation: the Atlantic jet becomes progressively narrower, stronger, and more zonal as the topography is raised, while meridional PV gradients become much tighter; especially in the western basin, where a steplike gradient separates two zones of almost homogeneous PV. Note the presence of anticyclonically overturning subtropical PV contours in the two higher LIS simulations (Figs. 1h,j), indicative of the key role played by Rossby wave breaking in the reorganization of the circulation, as further discussed in section 5.

Thick red contours in the left column of Fig. 1 show the winter-mean sea ice margin, taken as the 50% sea ice fraction contour. The extensive North Atlantic sea ice cover in the LGM simulations resembles that in the CLIMAP (1976) reconstruction, which shows a largely zonal ice margin at around 40°–45°N. A similar sea ice distribution was also found in an equilibrated LGM simulation using the fully coupled version of the model, CCSM3 (Brandefelt and Otto-Bliesner 2009, their Fig. 2), on which we base our ocean heat flux prescription. The extensive sea ice cover appears unrealistic in comparison with more recent proxy data, showing evidence of perennially open water to 50°–55°N in the central and eastern Atlantic (De Vernal et al. 2005, 2006; Kucera et al. 2005a; MARGO 2009). A discussion of how sea ice affects our results will follow in section 6.

Figure 2 shows the response of the jet’s meridional tilt to increasing ice sheet elevation. The meridional tilt is measured by the latitude difference in the jet axis (location of maximum wind speed) across the longitude interval 55°–10°W. The jet tilt is insensitive to ice sheet height at elevations below about 2500 m, but at higher elevations the eastern tip migrates equatorward, making the jet more zonal; the western end of the jet is less sensitive and remains at roughly the same latitude throughout. In the reference LGM simulation (with a 3400-m LIS) the jet axis is perfectly zonal, with no meridional tilt across the basin, while in the highest LIS cases the jet axis has a slight negative tilt (i.e., the jet axis is at a lower latitude in the eastern side of the basin than in the west). Interestingly, the jet tilt is sensitive to LIS
4. Stationary waves

Figure 3 explores the response of the climatological stationary waves and their associated wave activity flux $F_s$ [Eq. (1)] to increasing ice sheet elevation. In the reanalysis (Fig. 3a), $F_s$ points southeastward everywhere in the Atlantic basin, implying predominantly equatorward propagation of stationary waves. The modern simulation (Fig. 3b) shows a similar pattern of stationary waves and wave activity flux vectors; the low-elevation LGM simulation (Fig. 3c) is qualitatively similar to the modern, but with some subtle differences. In particular, the modern wave activity flux in the North American–Atlantic sector is mostly concentrated in an arc over Canada and into the central Atlantic, while the low-elevation LGM simulation shows an additional wave flux bundle directed southeastward over the continent and into the subtropical western Atlantic.

As ice sheet height increases further, stationary wave amplitudes grow (Figs. 3e,g), and the southeastward flux bundle strengthens more rapidly than that over Canada; this preference for equatorward propagation of waves excited by high topography is also observed in idealized modeling studies (e.g., Ringler and Cook 1997, 1999). In the reference LGM run (Fig. 3e), $F_s$ drops to zero in the central and eastern midlatitude Atlantic, and the streamfunction phase lines are zonal. In the highest-elevation run (Fig. 3g), the streamfunction phase lines have a slightly negative tilt (similar to the negative jet axis tilt seen in Fig. 2), and there is strong poleward-oriented wave activity flux in the northeastern Atlantic,
suggesting a transition into a state where low-latitude reflection in the central Atlantic becomes prevalent.

To further explore the structural response of the stationary wave field to the increasing mechanical wave forcing, we show in Figs. 3d, 3f, and 3h the eddy streamfunction and $F_s$ anomalies after subtracting the climatology of a simulation with zero-elevation ice sheets (but including the ice sheet’s albedo effect). In the linear limit, this anomaly gives the topographically forced response; note, however, that the flow dynamics is nonlinear and that SST and sea ice cover also change between the simulations, so the response should be interpreted with care. In any event, the anomaly in response to a low-elevation ice sheet (Fig. 3d) shows a strengthening of the equatorward stationary wave flux, as would be expected from more vigorous wave excitation by increased midlatitude topography. The response resembles that found in idealized studies when increasing the topographic forcing of an isolated midlatitude mountain in the near-linear regime [cf. Fig. 7 in Cook and Held (1992)]; there is a general strengthening of the streamfunction anomalies and an eastward expansion of the leeside cyclone to the east of the mountain.

In the reference LIS case (Fig. 3f), the anomalies indicate stronger stationary waves that propagate along a more southeastward trajectory over North America and into the subtropical central Atlantic. In addition, there is now a poleward-propagating wave train that appears to emanate in the eastern Atlantic and propagates in an arc over Europe and into the tropics over the Middle East. We interpret this wave train as resulting from nonlinear Rossby wave reflection in the subtropical Atlantic (Brunet and Haynes 1996; Magnusdottir and Haynes 1999; Walker and Magnusdottir 2003; Abatzoglou and Magnusdottir 2004, 2006b). A very similar pattern but with much larger amplitude is seen in the high-LIS simulation (Fig. 3h).

The anomaly fields thus suggest that wave reflection plays an important role in determining the stationary wave response in the reference LIS simulation, even though climatological wave reflection in the eastern Atlantic is not directly visible in the full fields shown in Fig. 3e. This occurs because the poleward-propagating wave flux has an intensity comparable to that of the equatorward-propagating wave flux that would otherwise exclusively occupy the eastern Atlantic; these two wave bundles effectively cancel in the time mean in the region where they overlap, resulting in zero net meridional wave flux and thus zonally oriented phase lines in the midlatitude central and eastern Atlantic. Note however that the equatorward-propagating part of the wave train over Eurasia is visible in the full fields, as this part of the reflected wave field occupies a region with little background wave activity and is therefore not eliminated in the time mean.

5. Statistics of wave breaking and reflection events

Rossby wave reflection, wave breaking, and local homogenization of the PV field are interrelated phenomena generally associated with processes occurring on subseasonal to synoptic time scales. In this section, we examine how the statistics of these time-varying phenomena respond to increasing ice sheet elevation and how such changes relate to the climatological response discussed above. Rossby wave breaking typically occurs on the flanks of the jet where the wind speed is lower and the waves meet their critical lines. AWB is more frequent on the equatorward flank of the jet, both because the background flow is anticyclonic there and because the equatorward-propagating waves, which predominate there, have southwest–northeast-tilted phase lines; vice versa, CWB occurs more frequently on the poleward flank of the jet (Thorncroft et al. 1993; Riviére 2009; Barnes and Hartmann 2011). Furthermore, nonlinear Rossby wave reflection is associated with anticyclonic wave breaking in the low latitudes, with the wave breaking region acting as a source of poleward-propagating wave activity (e.g., Brunet and Haynes 1996; Wang and Kushner 2010).

Figure 4 shows the spatial distribution of Rossby wave breaking frequency in our simulations, identified as
described in section 2c. In all our simulations [and in reanalysis data (Barnes and Hartmann 2012)], Rossby wave breaking in the North Atlantic occurs predominantly in a band on the equatorward flank of the jet where it is mostly anticyclonic. As LIS height increases and the jet becomes more zonal, this band of AWB becomes correspondingly more zonal and also much more concentrated in the central Atlantic, precisely in the region where the climatological flow shows permanently overturning vorticity contours (thick black lines in Figs. 4c,d).

To study the relation between Rossby wave breaking and reflection, we first count the total number of AWB events occurring within the red box in Fig. 4; this box roughly captures the area where stationary waves excited over the North American continent are most likely to break in the North Atlantic (as is evident by the overturning vorticity contours in Figs. 4c and 4d). For each AWB event identified within the box, we then examine the meridional component of the wave activity flux $F_{sy}$ averaged over the area of the central Atlantic where climatological poleward wave activity flux is most prevalent in the high-LIS runs, identified by the green box in Fig. 3g. We define a “reflective” AWB event as one for which $F_{sy}$ averaged over the green box becomes positive at any instant (using 6-hourly data) in the 3 days following initial detection of the AWB event. Note that this definition makes no distinction as to the relative strength of the reflection events. A 3-day time window was selected to ensure that all wave reflection events are captured. A too-narrow time interval implies that wave reflection events can be missed (time scale of wave breaking to reflection is of order 2 days), whereas a too-long time interval implies that several wave breaking and reflection events can happen within the search period and therefore contaminate the statistics.

Results are shown in Table 1. The total number of AWB events identified in the southwest Atlantic box increases with the LIS elevation, and the AWB distribution shifts and becomes more localized in the central part of the box (Fig. 4). At the same time, the fraction of reflective events increases from about 60% in the modern simulation to 90% with the highest LIS elevation. The subtropical Atlantic thus appears to transition into a state where almost all AWB events are followed by wave reflection.
On the basis of previous work examining stationary wave reflection in time-varying three-dimensional flows (e.g., Walker and Magnusdottir 2003), we hypothesize that a large fraction of the reflective AWB events are initiated by pulses of quasi-stationary wave activity that move off the North American continent, causing wave breaking and reflection when they enter the subtropics in the southwestern North Atlantic. To test this hypothesis, we detect pulses of wave activity over North America by defining a 6-hourly wave amplitude index as the spatial root-mean-square value of the 300-hPa eddy meridional wind field over a box covering the southern part of North America (20°–50°N, 130°–70°W). We then search for times when this index exceeds a fixed threshold (15 m s⁻¹; approximately one standard deviation below the mean wind in all LGM simulations; right column in Table 1) and then determine how many of the identified wave pulses are followed within 3 days by initiation of a reflective AWB event, defined as above.

Results are again shown in Table 1. The amplitude threshold of 15 m s⁻¹ gives 18 ± 1 wave pulse events per season across all simulations. This stable number belies the fact that the underlying amplitude index distribution actually changes markedly as LIS elevation increases, with high amplitudes becoming much more frequent at the expense of lower amplitudes (right column in Table 1). It therefore appears that, while LIS elevation does not affect the number of high-amplitude pulses (which is presumably set upstream by conditions in the Pacific), it does skew their amplitude toward higher values. In addition, the number of wave pulses leading to reflective AWB increases with increasing LIS elevation, though it remains roughly stable as a fraction of all reflective AWB events (around 55%). In summary, rising LIS elevation leads to increased AWB and reflection, but we find no clear connection to the amplitude of incoming wave pulses. Instead, we surmise that the changing mean state of the subtropical Atlantic—in particular, the increasingly homogenized PV field there—facilitates AWB and reflection. There is some circularity in this argument, since it is likely that the increased wave breaking itself helps homogenize the PV field. Disentangling this wave–mean flow interaction problem would require more detailed study that we leave to future work.

Finally, we study the spatiotemporal evolution of a wave pulse–reflective AWB event by computing composites over a large number of such events. For clarity, we focus on large-amplitude events by selecting wave pulses that exceed a 30 m s⁻¹ threshold. Figure 5 shows composites of w and 10-day low-pass-filtered meridional wind from the high-LIS simulation; lag 0 is the time at which the wave amplitude index peaks. The climatology has been subtracted to more clearly visualize the characteristics of the event.

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In the early stages of the life cycle, we see a wave packet over the eastern Pacific and North America with southeastward group velocity (as indicated by Fₓ). By lag 0, the packet is centered over North America, but its leading edge has already reached the western Atlantic, causing AWB there, as indicated by the overturning of absolute vorticity contours (thick black line in Fig. 5b). We also see poleward-propagating wave activity emitted from the wave breaking region, the classic signature of nonlinear reflection. Note that the waves within the packet show almost no phase propagation, but the packet envelope propagates rapidly downstream at around 40°–45° longitude day⁻¹. The wave packet is refracted toward low latitudes over Europe and the Middle East and is eventually absorbed in the subtropics.


Table 1. Statistics of wave breaking, reflection, and wave pulses. AWB denotes the total number of AWB events detected in the region indicated by the red box in Fig. 4; AWB + R denotes the number of reflective AWB events (in parentheses as a percentage of AWB); N denotes the number of wave pulses exceeding a threshold of 15 m s\(^{-1}\) [computed as the spatial root-mean-square value of the 300-hPa instantaneous (6 hourly) eddy meridional wind field over 20\(^\circ\)—50\(^\circ\)N, 130\(^\circ\)—70\(^\circ\)W]; N + AWB denotes the number of wave pulses that are followed by AWB events (in parentheses as a percentage of the total number of AWB events); N + AWB + R denotes the number of such pulses associated with reflective AWB events (in parentheses as a percentage of AWB + R); A ± σ denotes the mean amplitude and standard deviation (m s\(^{-1}\)) of wave pulses exceeding the 15 m s\(^{-1}\) threshold.

<table>
<thead>
<tr>
<th></th>
<th>AWB</th>
<th>AWB + R</th>
<th>N</th>
<th>N + AWB</th>
<th>N + AWB + R</th>
<th>A ± σ</th>
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<tr>
<td>Modern</td>
<td>17.4</td>
<td>10.4 (60%)</td>
<td>17.4</td>
<td>9.0 (52%)</td>
<td>5.9 (57%)</td>
<td>17.0 ± 4.1</td>
</tr>
<tr>
<td>LIS(_{2400})</td>
<td>18.0</td>
<td>9.9 (55%)</td>
<td>18.8</td>
<td>8.1 (45%)</td>
<td>5.3 (54%)</td>
<td>19.4 ± 5.0</td>
</tr>
<tr>
<td>LIS(_{4400})</td>
<td>22.3</td>
<td>16.8 (75%)</td>
<td>18.7</td>
<td>10.7 (48%)</td>
<td>9.3 (55%)</td>
<td>20.4 ± 5.5</td>
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<tr>
<td>LIS(_{5100})</td>
<td>23.8</td>
<td>21.2 (89%)</td>
<td>18.8</td>
<td>12.0 (50%)</td>
<td>11.2 (53%)</td>
<td>21.2 ± 6.0</td>
</tr>
<tr>
<td>LGM SST, present-day topography</td>
<td>18.2</td>
<td>11.0 (60%)</td>
<td>17.2</td>
<td>8.2 (45%)</td>
<td>6.3 (57%)</td>
<td>20.0 ± 5.0</td>
</tr>
<tr>
<td>Present-day SST, LGM topography</td>
<td>26.9</td>
<td>17.7 (66%)</td>
<td>17.8</td>
<td>10.9 (41%)</td>
<td>8.3 (47%)</td>
<td>17.1 ± 4.7</td>
</tr>
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</table>

over South Asia. Overall, this life cycle is similar to reflective Rossby wave breaking events documented in both idealized modeling studies and observations (Walker and Magnusdottir 2003; Abatzoglou and Magnusdottir 2004, 2006b,a).

6. Sensitivity to surface boundary conditions and model resolution

a. Relative roles of SST, sea ice, and ice sheets

The simulations discussed so far have evolving SST and sea ice, and it is important to understand how much the changes in these boundary conditions affect the circulation compared with the direct effect of the ice sheets. To do this, we conducted two sensitivity simulations using prescribed monthly varying climatological SST and sea ice. In one simulation, we specify present-day topography but prescribe SST and sea ice fields taken from the high-LIS LGM simulation; in the other, the two are switched (high-LIS LGM topography with modern SST and sea ice). Both sensitivity experiments use LGM greenhouse gas concentrations and orbital forcing.

Results are shown in Fig. 6. In the simulation with LGM SST and sea ice (top row), the Atlantic jet is closely aligned with the sea ice line and is stronger, more zonal (Fig. 2), and extends farther east than in the present-day simulation; and there is some strengthening of the meridional PV gradients (cf. Figs. 1c,d and 6a,b). However, there are no climatologically overturning PV contours, and the wave activity flux (Fig. 6c) shows no indication of Rossby wave reflection. On the other hand, the simulation with modern ocean conditions but high-elevation ice sheets (Fig. 6, bottom row) shows a more zonalized jet (Fig. 2), which is much more intense—particularly in the western Atlantic—and shifted substantially to the south of its modern counterpart. The PV field now shows clear climatological overturning contours in the subtropics, and the wave activity flux is close to zero in the midlatitude central/eastern Atlantic, similar to what is seen in the reference LGM simulation.

Overall, it appears that both the ice sheets and the ocean surface conditions contribute to the response, but the ice sheets contribute more of the qualitative characteristics of the full LGM simulation. Specifically, the results suggest that the stationary waves generated by the LIS are essential in driving wave breaking, jet sharpening, and nonlinear reflection in the western part of the basin (also seen in Table 1), while the southward-shifted LGM sea ice margin helps strengthen the eastern tip of the jet and extend it over Europe.

b. Model resolution

Our analysis emphasizes the role of planetary wave breaking and regional homogenization of the subtropical PV field as a precondition for stationary wave reflection in the central Atlantic. Since wave breaking and PV mixing are associated with small-scale filament structures, it is conceivable that the results presented above may be sensitive to the horizontal resolution of the simulations. Therefore, conditions for planetary wave reflection may not occur as frequently—possibly not at all—in simulations on coarse horizontal grids; Dong and Valdes (2000) and Polvani et al. (2004) showed that the horizontal resolution can give rise to large differences both in the climate variability and in the climatological mean state, as Rossby waves and local climate features become poorly represented at sufficiently low horizontal resolutions.

To investigate this issue, we conducted a sequence of simulations with progressively coarser horizontal resolutions. All simulations use high-elevation LGM ice sheets and prescribed SST and sea ice taken from the high-elevation slab-ocean run. The boundary conditions are spectrally interpolated to match the resolution of
each simulation. Climatologies for these simulations are presented in Fig. 7. At T42 and T31 resolution, the results are very similar to those at our reference T85 resolution (cf. Figs. 7a–f with Figs. 1i,j and 3g). When the resolution is further reduced to T21, however, the circulation is considerably distorted: the Atlantic jet is still zonal but considerably weaker than at higher resolution, the climatological overturning PV signature in the southwest Atlantic is wiped out, and the poleward flux of wave activity in the eastern Atlantic is absent. This more muted response may partly be as a result of the lower elevation of the ice sheets at lower resolution but is likely a small effect—maximum LIS elevation drops by only 200 m (from 5100 to 4900 m), going from T85 to T21. It appears, therefore, that there is a minimum resolution between T21 and T31 required to adequately capture Rossby wave breaking and reflection dynamics in this model. A similar resolution limit was also found in previous work using a more idealized model configuration (Magnusdottir and Haynes 1999).

7. Response of the North Atlantic storm track

The climatological circulation response to increasing ice sheet elevation is accompanied by important changes in synoptic-scale variability and storm tracks. Figure 8 compares the upper-tropospheric 2.5–6-day bandpassed eddy kinetic energy (EKE; computed from 10 years of 6-hourly instantaneous data on the 300-hPa surface) and precipitation in the present-day and LGM simulations. The modern Atlantic storm track is meridionally tilted following the mean jet, and both EKE and precipitation maximize over the western and central Atlantic. As ice sheet elevation increases and the jet becomes more zonal, the storm track becomes correspondingly more zonal and also more meridionally confined. In addition, maximum EKE shifts downstream and peaks over southern Europe, while precipitation maximizes over Iberia.

Previous work (Li and Battisti 2008; Donohoe and Battisti 2009) found a similar reorganization of the storm track in LGM simulations (which used the same atmospheric model used here, but at lower T42 resolution and coupled to a dynamic ocean). Our results differ in one respect, however: while we find a clear increase in maximum EKE as the climate transitions to the LGM, the previous work found a decrease. This discrepancy is most likely related to sea ice cover, which is much less extensive in the previous simulations than in ours [Fig. 12 in Li and Battisti (2008)]. In our simulations, the ice margin is coaxial with the storm track, and the strong baroclinicity implied by the large temperature gradients across the sea ice margin is likely to drive increased storm-track activity. This hypothesis is supported by examination of the EKE in the simulation using LGM ice sheets but modern SST and sea ice (see section 6), which shows a storm track displaced toward southern Europe, but with lower EKE than in the modern simulation (not shown).

8. Discussion and implications

Our key result is that, as the LGM ice sheet elevation increases, nonlinear Rossby wave reflection in the Atlantic becomes increasingly prevalent, and this leads to a zonalization of the midlatitude jet, particularly in the central and eastern parts of the basin. We can understand how wave reflection zonalizes the Atlantic jet as follows: The incident and reflected planetary waves have roughly equal but opposite phase tilts (Held et al. 2002) and therefore largely cancel in the time mean in
the region where they overlap. In a situation where the incident and reflected waves are of comparable intensities, the climatological stationary wave phase lines—and therefore the jet—have zero tilt (i.e., they are zonal); this is precisely the situation in our reference LGM simulation. In the high-elevation simulations, reflection dominates over the incident flux, and the jet acquires a slightly negative (northwest–southeast) tilt. This interpretation is different from that in previous studies, which attribute the zonalization of the jet at the LGM to the influence of transient eddy feedbacks and equatorward-shifted eddy momentum flux convergence as a result of increased frequency of CWB (Laine et al. 2009; Rivièr et al. 2010; Merz et al. 2015). It has been shown that an equatorward-displaced jet stream favors CWB, which helps maintain the jet at a more equatorward location (Rivièr 2009; Rivièr et al. 2010; Drouard et al. 2013). These two different mechanisms for jet zonalization are not mutually exclusive and may operate simultaneously. In fact, our high–ice sheet simulations show some increase in CWB on the poleward flank of the jet (discernible as a region of increased wave breaking frequency over the eastern Atlantic and Europe in Figs. 4c and 4d), which may help shift the eastern tip of the jet southward.

Our results further suggest that both the surface boundary conditions (topography and sea surface) and the horizontal resolution in the model simulation are important for planetary wave reflection. This may explain why planetary wave reflection has been reported in some studies of flow–topography interactions (Waugh et al. 1994; Brunet and Haynes 1996; Magnusdottir and Haynes 1999; Wang and Kushner 2010), whereas it is absent in other studies using a similar experiment setup (e.g., Cook and Held 1992; Ringler and Cook 1997). The lowest resolution where we see a poleward-propagating climatological wave train with high-LIS forcing is T31, which corresponds to a zonal grid spacing of about 300 km in midlatitudes. The climatological wave reflection is absent when using an identical forcing at T21 resolution (a zonal grid spacing of about 450 km in midlatitudes; see Fig. 7). Incidentally, the grid resolution in both Cook and Held (1992) and Ringler and Cook (1997) is comparable to or even coarser than that in our lowest-resolution case and may, therefore, place them in a regime where planetary wave reflection does not occur frequently enough to leave a signature on the climatological fields. Magnusdottir and Haynes (1999) found a similar limit at T31 resolution in their model and argue that the increased dissipation needed for numerical stability at low horizontal resolutions transforms the subtropics into a persistent wave absorber.

The results presented above—in particular, the storm-track responses—have important implications for the LGM climate of Europe and the North Atlantic in general. It is therefore important to assess their robustness and the extent to which they are supported by existing reconstructions using proxy data. While there are no direct proxies for the upper-tropospheric winds that are the main focus in this paper, hydroclimatic proxies can give indications as to changes in storm-track position and intensity. Oster et al. (2015) showed evidence, based on proxy data, of a southward shift of the Pacific storm track and a cyclone path that appears to have been predominantly to the south of the LGM LIS, broadly consistent with our modeling results. Similarly, Beghin et al. (2015) showed evidence from proxy data of a wetter climate in southwestern Europe and northwestern Africa at the LGM compared to the present day, suggesting an equatorward shift of the Atlantic storm track.

Synthesis data reconstructions of the Atlantic sea ice cover suggest a general southward expansion compared to the present, but perennially open waters as far north as 55°N in the eastern basin (Kucera et al. 2005b; De Vernal et al. 2005, 2006; MARGO 2009). The reconstructions thus have considerably less sea ice in the eastern Atlantic than obtained in these experiments.
Nevertheless, several fully coupled models forced by the high PMIP2 LIS (Peltier 2004) show a strong and largely zonal jet (e.g., Li and Battisti 2008; Kageyama et al. 2013; Ullman et al. 2014; Löfverström 2014; Merz et al. 2015), even with a more realistic sea ice cover in the eastern North Atlantic and the North Sea (Kageyama et al. 2006). At the same time, models forced by lower ice sheets typically have a weaker and more meridionally tilted Atlantic jet (Kageyama and Valdes 2000; Li and Battisti 2008; Ullman et al. 2014; Kageyama et al. 2013; Merz et al. 2015). Our study thus confirms these results and further demonstrates that the ice sheet

![Figure 7](image_url)

**FIG. 7.** As in Fig. 6, but for runs with high-elevation LGM ice sheets and prescribed LGM SST and sea ice using (a)–(c) T42 (~2.8°), (d)–(f) T31 (~3.8°), and (g)–(i) T21 (~5.6°) horizontal resolution.

![Figure 8](image_url)

**FIG. 8.** Winter (DJF) climatologies of (left) transient (2.5–6-day bandpassed) eddy kinetic energy at 300 hPa (m² s⁻²) and (right) precipitation (mm day⁻¹) for (a),(b) the modern simulation and LGM simulations with (c),(d) low–, (e),(f) reference–, and (g),(h) high–ice sheet topography. Gray contours show the DJF climatology of zonal wind at 300 hPa (10 m s⁻¹ contour interval starting at 30 m s⁻¹).
height is the main driver of the orientation of the Atlantic jet, which in turn focuses attention on the large uncertainty of LIS height reconstructions. The LIS reconstruction used here is structurally similar to both the PMIP2 and the PMIP3 reconstructions (Peltier 2004; Abe-Ouchi et al. 2015), with the highest point in a dome centered over Canada. To what extent this topographic outline is important for the results has yet to be determined.

In any event, a zonalization of the Atlantic jet is likely to have strong impacts on the North Atlantic glacial climate. It has been hypothesized that a stable zonal Atlantic LGM jet may suppress climatic feedbacks underlying the millennial-scale Dansgaard–Oeschger (DO) warming events (e.g., Seager and Battisti 2007), which are inferred from the Greenland oxygen isotope records (e.g., Andersen et al. 2004). DO events were frequent in the pre-LGM period, but they appear not to have occurred around the LGM (e.g., Ditlevsen et al. 2005). The mechanisms of the DO events are not fully understood but clearly involve large and rapid sea ice cover reductions in the North Atlantic (Li et al. 2010; Dokken et al. 2013). The height of the LIS could well play a critical role in this context: modeling studies indicate that the interplay between the atmospheric circulation, the sea ice cover, and the Atlantic meridional overturning circulation (AMOC) is sensitive to the height of the LIS (e.g., Eisenman et al. 2009; Zhang et al. 2014). From the present-day differences in jet orientations and meridional overturning circulations in the North Pacific and Atlantic, one may speculate that a more zonal Atlantic jet would diminish the AMOC (Warren 1983). However, the current LGM studies based on climate models give no firm answer to this question, as they disagree on to what degree the Atlantic jet was zonalized and whether the AMOC was weaker or stronger than today (e.g., Weber et al. 2007; Eisenman et al. 2009; Zhang et al. 2014; Brady et al. 2013). The mean tilt of the Atlantic jet axis is also of fundamental importance for the evolution of the Eurasian ice sheet (EIS), as the jet position largely determines where the synoptic systems make landfall over the continent. As shown by Löverström et al. (2014), a meridionally tilted jet yields a significant amount of precipitation on the EIS western and southwestern slopes. However, with a zonal jet, the precipitation maximum is shifted to the southern parts of the continent and thus largely deprives the ice sheet of winter precipitation. The geological data also suggest that the EIS was at a maximum volume about 60 kyr BP (Kleman et al. 2013; Svendsen et al. 2004), which is approximately halfway through the glacial cycle, and that the center of mass progressively shifted southwestward in time. At the LGM, the ice sheet even covered the British Isles (Bradwell et al. 2008; Svendsen et al. 2004; Kleman et al. 2013).

9. Summary and conclusions

We have investigated the response of the large-scale atmospheric circulation in the North Atlantic to the height of the LGM ice sheets. We find that, as the height of the LIS topography increases in our simulations, the Atlantic jet becomes progressively stronger, sharper, and more zonally oriented. Our interpretation of this response emphasizes the role of stationary Rossby waves and planetary wave reflection. As ice sheet height increases, stationary waves of increasing amplitude drive enhanced wave breaking where they enter the tropics in the southwest Atlantic. This intensifies the subtropical “surf zone” and locally homogenizes the PV distribution, expelling gradients to the midlatitudes and forming a steplike PV structure associated with a sharp, intense midlatitude jet. Nonlinear Rossby wave reflection out of this region of homogenized PV generates a poleward flux of wave activity, which superposes on the background equatorward wave activity flux in the eastern Atlantic and promotes a zonalization of the streamfunction phase lines and, therefore, of the jet in that region. The role of nonlinear Rossby wave reflection in explaining key features of past climates has not been previously explored and deserves further study.

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