**Climate
of the Past
Discussions**

The key role of topography in altering North Atlantic atmospheric circulation during the last glacial period

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Abstract

The Last Glacial Maximum (LGM; 21 000 years before present) was a period of low atmospheric greenhouse gas concentrations, when vast ice sheets covered large parts of North America and Eurasia. Paleoclimate reconstructions and modeling studies suggest that the atmospheric circulation was substantially altered compared to today, both in terms of its mean state and its variability. Here we present a suite of coupled model simulations designed to investigate both the separate and combined influences of the main LGM boundary condition changes (greenhouse gases, ice sheet topography and ice sheet albedo) on the mean state and variability of the atmospheric circulation as represented by sea level pressure (SLP), 200-hPa zonal wind in the North Atlantic sector. We find that ice sheet topography accounts for most of the simulated change during the LGM. Greenhouse gases and ice sheet albedo affect the SLP gradient in the North Atlantic, but the overall placement of high and low pressure centers is controlled by topography. We also show that North Atlantic sea surface temperatures do not substantially influence the pattern of the climatological-mean SLP field, SLP variability or the position of the North Atlantic jet in the LGM.

1 Introduction

The Last Glacial Maximum (LGM), approximately 21 000 yr before present (21 ka), was the time of maximum land ice volume during the last glacial period (Yokoyama et al., 2000), when global mean surface temperature was 4 to 5 K lower than today (Braconnot et al., 2007; Roche et al., 2007). The main boundary condition changes for the LGM compared to the present climate are in incoming solar radiation, atmospheric greenhouse gas (GHG) concentrations, and the extent and height of continental ice sheets. At the LGM, the seasonal cycle and amplitude of solar radiation received at the top of the atmosphere were similar to today's as the geometry of the Earth's orbit around the Sun was close to its present configuration (Table 1). Therefore, the differences in

insolation alone are small and cannot explain the differences between the LGM and the modern climate, in particular the magnitude of the global mean cooling (Hewitt and Mitchell, 1997; Chiang et al., 2003). On the other hand, the presence of massive ice sheets over North America and Eurasia (Peltier, 2004) and the reduced atmospheric

5　GHG concentrations (Flückiger et al., 1999; Dallenbach et al., 2000; Monnin et al., 2001; Table 1) represent large boundary condition changes for the atmospheric circulation. For example, the atmospheric concentration of carbon dioxide (CO_2) was almost 100 ppm lower than pre-industrial (PI, 1750 AD) values as shown by trace gas analysis from ice cores (Monnin et al., 2001). The volume of water stored in the ice sheets dur-

10　ing the LGM is equivalent to 130 m of relative sea level drop (Yokoyama et al., 2000), which is enough to generate ice sheets over North America and Eurasia up to 5 km thick (Peltier, 2004).

Previous studies have investigated the impact of these boundary conditions on the LGM climate. For example, using atmospheric general circulation models (AGCMs)

15　forced with proxy-based sea surface temperature (SST) reconstructions, Hansen et al. (1984) showed that most of the global mean LGM cooling is due to the presence of ice sheets and increased sea ice cover. The role of ice sheets in LGM simulations has been further investigated using AGCMs coupled to static slab oceans, with a focus on regional versus global effects as well as on the particular importance of ice sheet

20　extent (albedo) for surface cooling (Manabe and Broccoli, 1985; Felzer et al., 1996; Hewitt and Mitchell, 1997; Felzer et al., 1998).

Other studies investigated the atmospheric circulation response to the full range of LGM boundary conditions. An early study using a coarse resolution (8° × 10°) AGCM (GISS model II) found that the various boundary conditions can have opposite

25　influences on the intensity and the location of transient and stationary eddies (Rind, 1987); however, the boundary conditions were applied sequentially rather than tested individually. More recently, Justino et al. (2005) performed a suite of experiments using an Earth system model of intermediate complexity (EMIC) to isolate the individual boundary condition effects. They found that topography is most important for controlling

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large scale atmospheric flow, at least in the simplified, coarse resolution atmosphere of the EMIC (5.6° × 5.6°). Pausata et al. (2009) suggested that ice sheet topography is also the most important factor for reducing extratropical Northern Hemisphere sea level pressure (SLP) variability in a state-of-the-art AGCM (T42, approximately 2.8° × 2.8°)
5   simulation of the LGM.

Most studies that have examined the influence of LGM boundary conditions on the atmospheric circulation are limited by the absence of a fully interactive ocean. The usual approaches are to use a slab ocean model, in which the ocean is highly simpli-fied, or to do an AGCM study, in which the ocean is treated as a boundary condition
10 rather than as an interactive component of the climate system that itself adjusts to inso-lation, ice sheet and GHG changes. As an example of the second approach, Pausata et al. (2009) performed uncoupled experiments using an AGCM forced by combinations of LGM and PI boundary conditions, including prescribed SSTs from existing coupled LGM and PI simulations. They used these experiments to suggest that topography
15 controls extratropical variability. A more definitive conclusion based on the uncoupled experiments was not possible because the prescribed SSTs include the influence of forcing by insolation, GHG concentrations and ice sheets (i.e., the SSTs are the result of these forcings on ocean circulation, as calculated by the coupled model). It has even been argued that the dynamical ocean response to reduced GHG concentrations
20 alone can produce much of the high latitude cooling and sea ice expansion produced in a coupled model (e.g., Kim, 2004).

A fully interactive ocean allows for a dynamical ocean response that can potentially feed back onto the atmospheric circulation and its variability. The current study aims to investigate further the dominance of LGM topography in determining the atmospheric
25 response suggested by Pausata et al. (2009) by clearly separating the large-scale atmospheric response to changes in GHG concentration, topography and albedo in a fully coupled general circulation model.

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2   Model and experiments

The model used in the current study is the Institut Pierre-Simon Laplace coupled AOGCM (IPSL-CM4-V1). This model took part in the Paleoclimate Modelling Intercomparison Project phase II (PMIP2), which coordinated simulations for the LGM, mid-
⁵ Holocene and PI climate states (http://pmip2.lsce.ipsl.fr, Braconnot et al., 2007). The atmospheric component has a horizontal resolution of 2.5° × 3.75° on a regular latitude by longitude grid and 19 vertical levels. The ocean component has a horizontal resolution of about 2° × 2° with a refined grid near the equator and 31 vertical levels. In this study, the control simulation is the PI climate in which insolation corresponds to
¹⁰ a 1950 AD orbital configuration and the GHG concentrations correspond to those that existed around 1750 AD (Table 1). For the LGM simulation (hereafter "full LGM"), the orbital configuration is set to 21 ka and GHG concentrations are lower, corresponding to the LGM values inferred from the Greenland and Antarctic ice cores (Flückiger et al., 1999; Dallenbach et al., 2000; Monnin et al., 2001). In the full LGM, the topography
¹⁵ is set following ICE-5G reconstruction (Peltier, 2004). For both simulations, a modern distribution of vegetation is prescribed (except in the glaciated LGM grid boxes).

In addition to the control (PI) and full LGM simulations, four sensitivity experiments have been performed (Kageyama et al., In prep.) in which each LGM boundary condition has been added separately to the PI control simulation: (1) LGM GHG con-
²⁰ centrations (LGMghg); (2) LGM land albedo with PI topography (LGMalb); (3) LGM topography with PI albedo (LGMtopo); and (4) LGM topography and albedo (LGMice). Table 2 summarizes the boundary conditions used in the sensitivity experiments.

Each equilibrium experiment is 500 years long, and the last 100 years of monthly SLP, 200-hPa zonal wind and surface air temperature data from 20°–90° N are ana-
²⁵ lyzed. Anomalies are calculated based on monthly departures from the climatological-mean seasonal cycle, focusing the analysis on the interannual variability. Standard Empirical Orthogonal Function (EOF)/Principal Component (PC) analysis has been used to calculate the leading mode of SLP variability in the North Atlantic.

3   Results

The results presented here describe the effects of the LGM boundary conditions on the mean state and variability of the Northern Hemisphere atmospheric circulation, and are divided into three parts. The first section investigates the suggestion of Pausata et al. (2009) that most key features of the large-scale LGM SLP response are determined by topography. To allow for a direct comparison with the previous study, changes in the spatial distribution of annual mean SLP and in the leading patterns of North Atlantic SLP variability are analyzed in the various sensitivity experiments. The second section extends the analysis from the surface to the upper troposphere by examining the 200-hPa wind response. The third section shows the surface temperature response to various LGM boundary conditions and discusses the influence of changed SST on the climatological-mean SLP and its interannual variability.

3.1   Effects of LGM boundary conditions on SLP

A robust feature of LGM model simulations is a southward shift of the climatological-mean SLP pattern in the extratropical Northern Hemisphere (e.g., Pausata et al., 2009; Laîné et al., 2009). A southward shift of the climatological high (hereafter "subtropical high") and low (hereafter "subpolar low") SLP centers in the North Atlantic (i.e., the Icelandic low and Azores high) and North Pacific (i.e., the Aleutian low and North Pacific high) occurs in the LGM relative to the PI simulation (see Fig. 1 as well as Fig. 1 in Pausata et al., 2009). The climatological-mean SLP gradient between the subtropical high and subpolar low weakens in the North Pacific, whereas it strengthens substantially in the North Atlantic in the LGM simulation compared to the PI simulation (Table 3).

An examination of the sensitivity experiments reveals that topography is the boundary condition that alone creates a SLP distribution most similar to the full LGM simulation (compare LGM and LGMtopo in Fig. 1; also shown in black contours for the North Atlantic sector only in Fig. 2). However, the SLP gradient in both the Atlantic and Pacific sectors is weaker in the LGMtopo compared to the full LGM experiment (Table 3).

Conversely, the LGMghg and LGMalb simulations exhibit stronger SLP gradients, but much less change (compared to the control simulation) in the location of the subtropical highs and subpolar lows. Thus, the main effect of both GHG and albedo changes at the LGM is to increase the midlatitude SLP gradient, whereas ice sheet topography

5 shifts the location of the climatological SLP features (Fig. 1 and Table 3).

As also shown in Pausata et al. (2009), another distinctive characteristic of LGM model simulations is a reduction in the North Atlantic interannual SLP variability and in the prominence of the leading pattern of SLP variability relative to the PI climate. The sensitivity experiments presented here reveal that these changes in SLP variability

10 are driven almost entirely by changes in topography (Figs. 1 and 2). Topography is responsible for the change in the pattern of SLP variability, producing a broad decrease in both interannual variability at high latitudes (Fig. 1) and total interannual variance explained by the leading mode (EOF1, Fig. 2), as well as a less defined North Atlantic dipole in EOF1 (Fig. 2). In contrast, both reduced GHG concentrations and increased

15 albedo at the LGM produce a small increase in the SLP standard deviation at mid-to-high latitudes (Fig. 1) and a stronger North Atlantic dipole pattern in EOF1 (Fig. 2). In fact, EOF1 of the PI, LGMghg and LGMalb experiments (i.e., those with control climate topography, left panels of Fig. 2) resembles the latitudinal shifting of atmospheric mass described by the well-known North Atlantic Oscillation (NAO; Walker and Bliss, 1932).

20 In LGM, LGMice and LGMtopo (i.e., the experiments with LGM topography), EOF1 represents mostly a change in the intensity and longitudinal extent of the Icelandic low instead. Similar changes in the NAO/Arctic Oscillation were found in a study by Rivière et al. (2010) by examining the nature and the location of Rossby wave breaking events in both PI and LGM simulations.

25 In summary, these coupled model experiments support the suggestion of Pausata et al. (2009), based on uncoupled AGCM experiments that topography is key in altering the main features of the SLP field. In the full LGM simulation, the pattern of the climatological-mean SLP field, of the SLP variability, and of the leading pattern of SLP variability (Table 3 and Fig. 2) are set almost entirely by the ice sheet topography.

The GHG and ice sheet albedo boundary conditions are associated with changes to the SLP gradient and relatively smaller changes to the SLP variability and the leading pattern of SLP variability.

3.2 Effects of LGM boundary conditions in 200-hPa zonal winds

In this section, we analyze the 200-hPa zonal wind in order to better characterize the atmospheric circulation response to LGM boundary conditions through the depth of the troposphere.

In the full LGM simulation, the 200-hPa zonal wind has a stronger, sharper maximum than in the control PI simulation (Fig. 3). The broader PI maximum results from a more tilted SW–NE orientation of the Atlantic jet and greater variability in the jet position around its climatological mean. This result is valid for both the annual and winter mean analyses, and has been verified by comparing maps of the climatological-mean response and its interannual variability.

During winter, LGM-PI changes in 200-hPa zonal wind are almost entirely explained by topography changes (LGMtopo, Fig. 3). In fact, the shape of the zonally averaged zonal wind profiles and the positions of the maxima are nearly identical in the LGMtopo experiment and the full LGM simulation. The maximum in the annual mean profile is, however, weaker in the LGMtopo experiment than in the full LGM. This is due to the fact that over North America in summer, LGMtopo has a much lower albedo than the full LGM simulation with its perennial land ice cover. As a consequence, LGMtopo exhibits large surface heating in summer, leading to a weaker meridional temperature gradient and (via the thermal wind relationship) weaker upper level winds. In winter, there is less discrepancy because the presence of snow cover makes the albedo of LGMtopo similar to the albedo of the full LGM simulation. Even though the LGM topography leads to most of the sharpening and strengthening of the Atlantic zonal mean wind profile, reduced GHG concentrations and increased albedo during the LGM do also contribute to strengthened upper tropospheric winds during summer (Fig. 3).

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As seen in Sect. 3.1, SLP mean state and variability changes are most strongly associated with changes in topography. This section has shown that topography is also the leading factor in determining the upper tropospheric zonal wind field (orientation, shape and intensity) but that LGM GHG and albedo changes also play a role during the summer months.

3.3 Surface temperature

In the coupled experiments performed in this study, the SST field is calculated as part of the model's response to the imposed GHG, topography and/or albedo boundary conditions. SST is known to be an important influence on the atmosphere, able to alter features of its mean state, its variability, and the persistence of atmospheric anomalies (e.g., Czaja and Frankignoul, 1999; Watanabe and Kimoto, 2000; Frankignoul and Kestenare, 2005; Rhines et al., 2008). Thus, an examination of the surface temperature response in each of the sensitivity experiments is of interest to our discussion about atmospheric variability.

The full LGM simulation shows generally cooler surface air temperatures (SATs) relative to PI (Fig. 4) and a corresponding SST cooling (not shown) over the Atlantic basin. The cooling is concentrated over the ice sheets due to the effects of albedo and altitude, and at high latitudes.The high-latitude cooling is evident in all the sensitivity experiments, including LGMghg, and is mainly due to sea ice expansion. Both the high elevation (LGMtopo) and high albedo (LGMalb) of the ice sheets contribute to the surface cooling over the continents.

Surprisingly, there is widespread warming over the North Atlantic (as well as of the underlying SSTs, not shown) in the LGMtopo experiment relative to the PI simulation (Fig. 4). We note that LGMtopo shows increased northward heat transport compared to the PI simulation in the Atlantic sector by the atmosphere and by the ocean (Fig. 5). The full LGM simulation also shows increased Atlantic heat transports, but does not exhibit surface warming compared to the PI simulation (Fig. 4), probably because of

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the compensating cooling effect of lower GHG concentrations and increased albedo at high latitudes.

Even though the LGMtopo experiment has a substantially warmer North Atlantic, the large-scale atmospheric circulation features of LGMtopo resemble those of the full
5 LGM simulation, as shown in the previous sections. That the surface temperature distributions are substantially different in the LGMtopo and full LGM simulations, yet the atmospheric circulation and its variability are quite similar (e.g., Fig. 2), suggests that altered SST and associated feedbacks do not greatly affect the mean atmospheric circulation in the presence of large topographic features like the Laurentide ice sheets.
10 However, as also suggested by Pausata et al. (2009), SST and sea ice changes do affect SLP variability and the variance associated with the PC1 timeseries, particularly in winter (not shown).

4 Discussion

We have presented results from sensitivity experiments carried out with a coupled
15 AOGCM showing the impact of various LGM boundary conditions on the mean state and variability of the atmospheric circulation. Each LGM boundary condition has been applied separately to investigate its individual impact relative to a control PI simulation.

Our study shows that ice sheet topography plays a dominant role in altering the mean state and variability of the atmospheric circulation, particularly in the North Atlantic. The
20 other boundary conditions, such as GHG concentrations and land ice albedo, have a much smaller influence. However, reduced GHG concentrations do cause anomalies across the Atlantic basin that result in a more zonally oriented 200-hPa wind structure in the LGMghg experiment than in the PI simulation (Fig. 6). A stronger and more zonally oriented jet is consistent with stronger meridional temperature gradients and
25 with stronger North Atlantic SLP gradients (Table 3) in the LGMghg experiment. In a world with increased GHG concentrations and presumably reduced near surface

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meridional temperature gradients, one might thus anticipate a less zonally oriented (more tilted) North Atlantic jet.

The fact that ice sheet topography appears to be such a strong control on the large scale atmospheric circulation is consistent with previous model studies, even when dif-
5 ferent ice sheet reconstructions (e.g., ICE-4G, Peltier, 1994 instead of ICE-5G Peltier, 2004) are used (Kageyama et al., 1999; Justino et al., 2005; Li, 2007; Li and Battisti, 2008; Pausata et al., 2009). Given the remarkable discrepancy (up to 2 km of thickness) in the Laurentide ice sheet between ICE-4G and ICE-5G, it would be interesting to further investigate the possibility of thresholds in ice sheet elevation necessary to
10 produce a significant response in atmospheric circulation.

The impact of topography on SSTs is somewhat unexpected. LGMtopo shows a widespread surface warming over the North Atlantic relative to the PI simulation, as discussed in Sect. 3.3, but the 200-hPa winds, the pattern of SLP climatology and the pattern of SLP variability are very similar to the full LGM simulation. Other studies us-
15 ing either an AGCM only (e.g., Manabe and Broccoli, 1985), or an AGCM coupled to a slab ocean (e.g., Rind, 1987; Felzer et al., 1996, 1998) have already shown warming in the presence of North American and Fennoscandian ice sheets, particularly in the regions south of the ice sheets and in the northernmost North Atlantic, east of Greenland. They attributed this localized warming to thermal anticyclones (over the ice
20 sheets) that cause subsidence and reduce summer cloud cover. These studies also noted a generally deeper Icelandic low, which increases southwesterly winds across the Atlantic Ocean during winter and accounts for the warming in the northernmost North Atlantic.

The warming observed in the LGMtopo experiment is much broader than the warm-
25 ing observed in these previous studies, suggesting that the proposed explanations (subsidence downstream of ice sheets and deepening of the Icelandic low) are insufficient to explain the response described in our study. The changes in ocean heat transport into the North Atlantic (Fig. 5) are a more likely explanation in our coupled model setup.

Pausata et al. (2009) showed in a series of AGCM experiments that SST plays only a minor role in determining interannual SLP variability in both pre-industrial and glacial experiments (see Fig. 6 in Pausata et al., 2009). In other words, a direct link between SST changes and fundamental features of the atmospheric circulation is not clear, in the presence or absence of large Northern Hemisphere ice sheets. Presumed links between SST and atmospheric circulation in a future enhanced GHG world may be similarly obfuscated.

The relatively weak response of the atmospheric circulation mean state to SST changes combined with the fact that topographic boundary condition is treated in a slightly different manner in different models highlights a major challenge in model intercomparison studies. For example, Pausata et al. (2009) found that (1) the mean and variability characteristics of the SLP field were different in different coupled climate model simulations of the LGM, and (2) an atmospheric model forced with SST distributions taken from two different coupled simulations produces remarkably similar SLP fields. Pausata et al. (2009) speculated that SLP, and indeed the large-scale atmospheric flow in general, are model-dependent for drastically altered climate states, but provided no explanation. The results of this study suggest that how different atmospheric models treat topography is crucial for understanding the model-specific response.

5 Conclusions

In this study, we analyze how gross features of the atmospheric general circulation are altered by Last Glacial Maximum (LGM) boundary conditions in a coupled climate model. The main findings are:

- Ice sheet topography sets many features of the SLP field, including the location of the subtropical highs and subpolar lows, the interannual variability, and the leading pattern of variability. Topography further controls the position, strength and shape of the zonal wind field at 200 hPa, especially during winter.

– Applying LGM GHG and ice sheet albedo boundary conditions to a PI simulation increases the gradient between subpolar low and subtropical high SLP centers in the North Atlantic.

– SST does not play a dominant role in setting the climatological-mean SLP pattern or 200-hPa zonal winds, the interannual variability of SLP, or the leading mode of SLP variability.

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Table 1. Boundary conditions for pre-industrial (PI) and Last Glacial Maximum (LGM) simulations. The ranges in square brackets represent the range of possible values for the obliquity and eccentricity of Earth's orbit.

	Ice sheet	Vegetation	CO_2 (ppmv)	CH_4 (ppbv)	N_2O (ppbv)	Obliquity (°) [22.0–24.5]	Angular precession (°)	Eccentricity [0.0034–0.058]
PI	Modern	Modern	280	760	270	23.4	102	0.017
LGM	ICE-5G	Modern	185	350	200	22.9	114	0.019

Table 2. Boundary conditions used in the sensitivity experiments.

	LGMghg	LGMalb	LGMtopo	LGMice
GHGs	LGM	PI	PI	PI
Albedo	PI	LGM	PI	LGM
Topography	PI	PI	LGM	LGM

Table 3. Annual-mean SLP gradient (Δ SLP) between subpolar lows and subtropical highs in the Atlantic and Pacific sectors for the control (PI) simulation, and the differences between each sensitivity experiment and the PI simulation. Positive differences indicate a stronger SLP gradient.

SLP (hPa)	PI	LGM	LGMghg	LGMalb	LGMtopo	LGMice
			Anomalies			
Gradient–Atlantic (Pa km^{-1})	0.50	+0.21	+0.06	+0.18	0.0	+0.19
Gradient–Pacific (Pa km^{-1})	0.59	−0.04	−0.05	−0.02	−0.15	−0.06

Table 4. Annual-mean SLP gradient between subpolar lows and subtropical highs in the Atlantic and Pacific basins for the control (PI) simulation, and the differences between each sensitivity experiment and the PI simulation. Positive differences indicate a stronger SLP gradient.

	∇ SLP (Pa km^{-1})	∇ SLP difference (Pa km^{-1})				
	PI	LGM	LGMghg	LGMalb	LGMtopo	LGMice
Atlantic	0.50	+0.21	+0.06	+0.18	0.0	+0.19
Pacific	0.59	−0.04	−0.05	−0.02	−0.15	−0.06
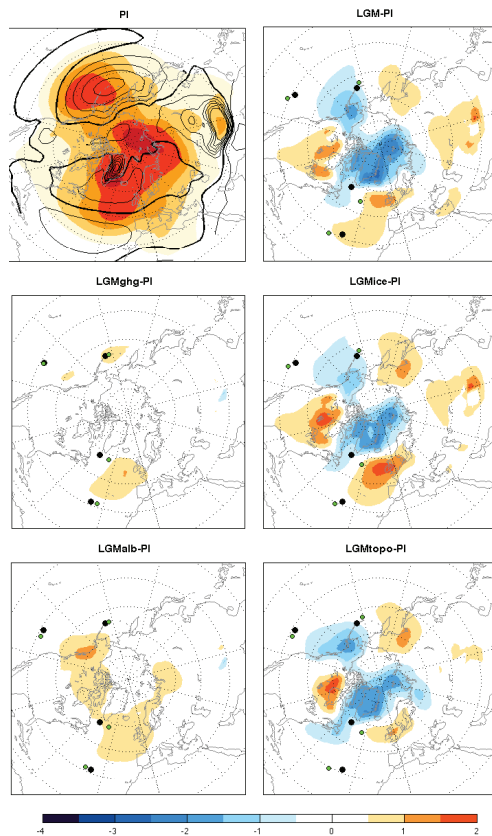
Fig. 1. The top left panel shows the climatological-mean SLP (contours: 4 hPa interval from 1000 to 1040 hPa; bold contour denotes 1016 hPa) and standard deviation of monthly SLP anomalies (colored shading: 1 hPa interval) using data from all months in the control (PI) simulation. The other panels show the difference in the monthly SLP standard deviation between the indicated experiment and the control simulation (PI). Dots show the locations of the subpolar lows and subtropical highs in the Pacific and Atlantic sectors for the control simulation (black) and the indicated experiment (green).
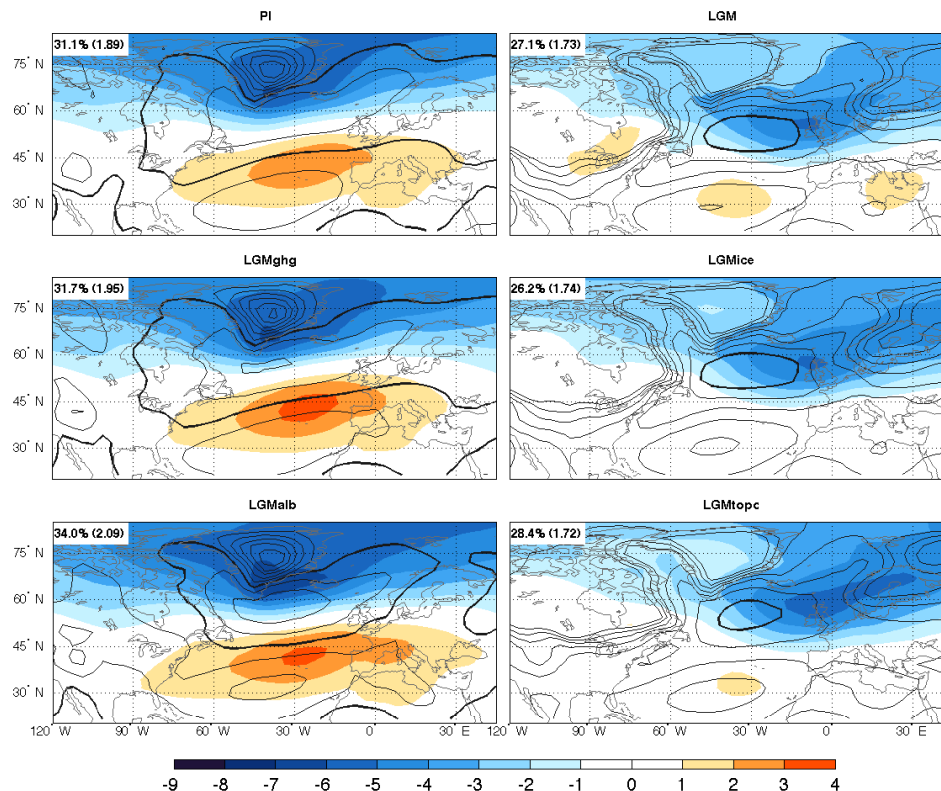
Fig. 2. Leading EOF of monthly SLP anomalies (colored shading: hPa/standard deviation of PC) using data from all months and SLP climatology (contours: 4 hPa interval from 1000 to 1040 hPa; higher values omitted for clarity; bold contour denotes 1016 hPa) in the North Atlantic sector for the control (PI) and LGM simulations, and for the sensitivity experiments. Numbers show the amount of variance explained by the first mode both as a percentage of the total variance (λ_1) and as a standard deviation σ_{NA} in hPa $\left(\sqrt{\lambda_1 \sigma_{NA}^2}\right)$.

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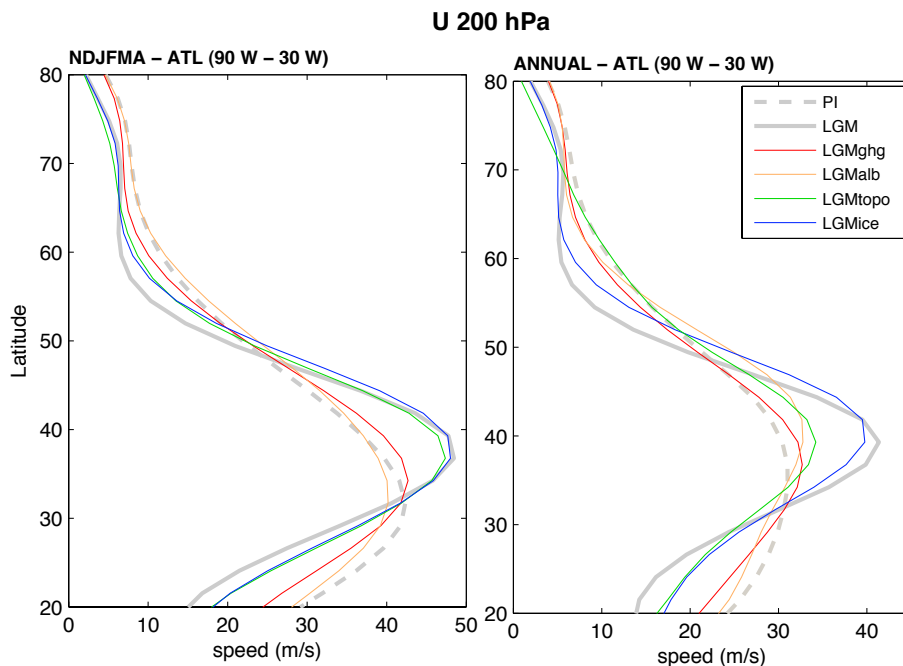
Fig. 3. Climatological-mean zonal wind speed distribution at 200-hPa averaged over the longitudes (90° W–30° W) in the Atlantic sector for the different experiments in winter (November to April, left) and for all months of the year (right).
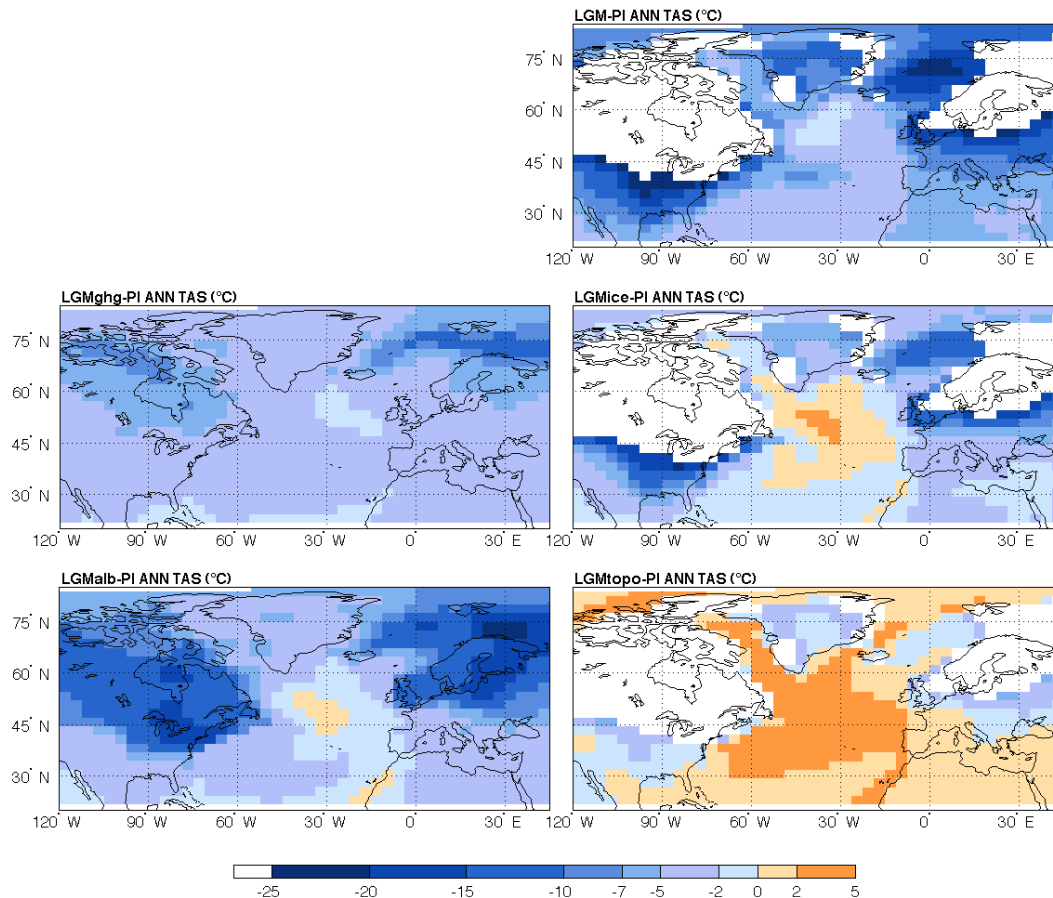
Fig. 4. Anomalies in annual mean surface air temperature (SAT) relative to the PI climate for each sensitivity experiment. Areas where the topography difference between the experiments and the PI is greater than 500 m are masked out in white.
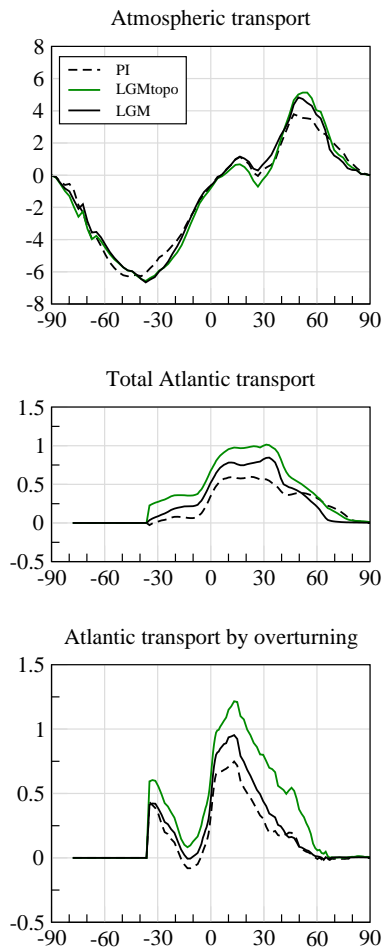
Fig. 5. Northward heat transport in the atmosphere (upper panel), in the North Atlantic ocean (middle panel), and by the North Atlantic overturning circulation (lower panel).

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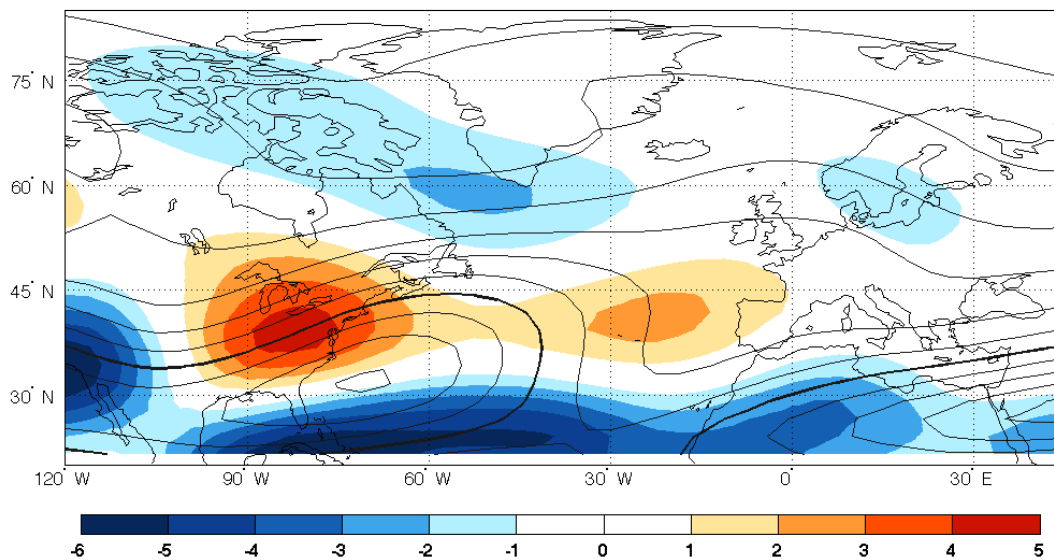
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Fig. 6. Winter (November to April) 200-hPa zonal winds for the control (PI) simulation (contours: $5\,\mathrm{m\,s}^{-1}$ interval from 5 to $50\,\mathrm{m\,s}^{-1}$; bold contour at $35\,\mathrm{m\,s}^{-1}$) and the difference (shading) between the LGMghg experiment and the control run (LGMghg-PI).