Dynamic Effect of Last Glacial Maximum Ice Sheet Topography on the East Asian Summer Monsoon

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ABSTRACT

The effect of ice sheet topography on the East Asian summer monsoon (EASM) during the Last Glacial Maximum is studied using CCSM3 in a hierarchy of model configurations. It is found that receding ice sheets result in a weakened EASM, with the reduced ice sheet thickness playing a major role. The lower ice sheet topography weakens the EASM through shifting the position of the midlatitude jet, and through altering Northern Hemisphere stationary waves. In the jet shifting mechanism, the lowering of ice sheets shifts the westerly jet northward and decreases the westerly jet over the subtropics in summer, which reduces the advection of dry enthalpy and in turn precipitation over the EASM region. In the stationary wave mechanism, the lowering of ice sheets induces an anomalous stationary wave train along the westerly waveguide that propagates into the EASM region, generating an equivalent-barotropic low response; this leads to reduced lower-tropospheric southerlies, which in turn reduces the dry enthalpy advection into East Asia, and hence the EASM precipitation.

1. Introduction

Continental ice sheets have experienced large glacial–interglacial cycles in the last 2 million years during the Pleistocene. At their maximum, such as the Last Glacial Maximum (LGM) (21,000 years ago), the Laurentide Ice Sheet (LIS) covered much of North America and extended southward to approximately 40°N, while the Eurasian Ice Sheet (EIS) covered much of northern Europe. These ice sheets can reach a height of 3–4 km with a total volume equivalent to ~130 m in global sea level. The waxing and waning of the ice sheets have long been hypothesized as a major forcing factor on global climate (e.g., Clark et al. 1999). This hypothesis has been corroborated by climate model simulations, which showed a significant impact of ice sheets on the climate over North America (e.g., Manabe and Broccoli 1985; Anderson et al. 1988), the North Atlantic (e.g., Li and Battisti 2008; Pausata et al. 2011; Hofer et al. 2012; Zhu et al. 2014), and the tropical ocean–atmosphere system (e.g., Chiang et al. 2003; Chiang and Bitz 2005; Lee et al. 2015; Lu et al. 2016).

Observational and modeling studies also suggest a potential role of ice sheets on global monsoon systems, in spite of the recognition that the primary driving forcing factor for monsoon evolution is the orbital forcing and the associated seasonal insolation contrast during the Pleistocene (Kutzbach 1981; Kutzbach and Street-Perrott 1985; Ruddiman 2006; Cheng et al. 2009). In observational studies, traditionally, the impact of ice
volume on monsoons has often been thought of conceptually from a simple thermodynamic view: an ice sheet retreat in the Northern Hemisphere (NH) reduces surface albedo and therefore leads to a warming over the NH, which will enhance the land–sea thermal contrast and, in turn, the summer monsoon. Based on this simple thermodynamic view, ice sheets have been speculated to contribute to the phase difference between some Asian monsoon records and the precessional forcing in the Pleistocene (e.g., Clemens and Prell 1990, 2003), including the Holocene (e.g., Overpeck et al. 1996). This simple thermodynamic view, however, is challenged by recent modeling studies, especially of the East Asian summer monsoon (EASM). In an Earth system model of intermediate complexity (LOVECLIM; Yin et al. 2008, 2009), the retreat of ice sheet weakens the EASM, just opposite to what expected from the simple thermodynamic view. In a study of EASM in MIS-13, Yin et al. (2008) showed that the presence of Eurasian and North American ice sheets together enhances the summer precipitation over East China through a stationary wave train propagating eastward from the ice sheets. This result is further confirmed in an AGCM (ARPEGE) experiment by Sundaram et al. (2012). This result appears to be also consistent with the evolution of the EASM forced by the ice sheet retreat in the last 19,000 years in a coupled GCM CCSM3. Wen et al. (2016) showed that the retreating ice sheet in CCSM3 weakens the EASM monsoon wind over North China. These results pose the question: why is the ice sheet impact on the EASM different from what is expected from the simple thermodynamic view?

Conceptually, there are potentially several problems with the simple thermodynamic view. First, many modeling studies suggest that the dynamic effects of the ice sheet topography, rather than the simple thermodynamic effects of the ice sheet albedo, are the dominant impacting factor of the ice sheet on the atmospheric and oceanic circulation and, in turn, regional climate in the NH (e.g., Li and Battisti 2008; Pausata et al. 2011; Lee et al. 2015; Lu et al. 2016). The large and tall LIS splits the westerly jet over North America, leading to a southward displacement of the jet stream and storm track locally over North America and the nearby downstream region over the North Atlantic (Manabe and Broccoli 1985; Anderson et al. 1988; Hofer et al. 2012; Ullman et al. 2014). The ice sheet topography can also exert a significant impact remotely on the climate and climate variability over the tropical Pacific (Lee et al. 2015; Lu et al. 2016) and the EASM (Yin et al. 2008, 2009; Sundaram et al. 2012). Second, the direct thermodynamic effect (e.g., warming associated with the reduced albedo and ice sheet thickness) may be reduced or even overwhelmed by cooling caused by the removal of the ice sheet. Indeed, in recent fully coupled ocean–atmosphere model simulations, the lowering of NH ice sheets leads to a broad cooling over the North Atlantic, the North Pacific, and northeastern Asia, in spite of a local warming over the North America continent and the midlatitude North Atlantic and North Pacific; this broad cooling appears to be caused by the changed planetary wave and sea ice expansion (Zhu et al. 2014; Ullman et al. 2014; Lu et al. 2016). Third, recent fully coupled ocean–atmosphere model simulations showed that a retreat of ice sheet can also induce a secondary dynamic impact associated with the change of the Atlantic meridional overturning circulation (AMOC). Indeed, in some CGCMs, a lowering or removal of LIS tends to weaken the AMOC (Zhu et al. 2014; Ullman et al. 2014; Klockmann et al. 2016; Sherriff-Tadano et al. 2018; Galbraith and de Lavergne 2019), which will then induce a bipolar seesaw response in SST and associated climate changes. Finally, it should be pointed out that, in the real world reconstructions, it is difficult to separate the ice sheet effect from the CO2 effect. Indeed, during the major glacial cycles of the last 800,000 years, ice volume roughly covaries with the atmospheric CO2 concentration level, with an interglacial (glacial) period corresponding to higher CO2/less ice sheet (lower CO2/more ice sheet) (e.g., Petit et al. 1999; Shackleton 2000). This covariation makes it difficult to isolate the climate impact of the ice sheet from that of CO2 in the reconstructions. The isolation is made even more difficult by the potentially opposite effects of the ice sheet and CO2 on some regional climate systems. For example, a retreat of ice sheet weakens the AMOC, opposite to the enhancement effect of AMOC by the slow deglacial increase of CO2 (Zhu et al. 2014; Ullman et al. 2014; Klockmann et al. 2016; Sherriff-Tadano et al. 2018; Galbraith and de Lavergne 2019); retreating glacial ice sheets tend to intensify El Niño–Southern Oscillation (ENSO), whereas increasing deglacial atmospheric CO2 concentrations weaken ENSO (Liu et al. 2014; Lu et al. 2019); and a retreat of ice sheet weakens the EASM, opposite to the strengthened EASM by the deglacial rise of CO2 level (Wen et al. 2016).

Here, we will focus on the impact of ice sheet topography on the EASM during the LGM. This is a follow-up study of Wen et al. (2016), which has shown a decreasing EASM in response to the ice sheet retreat in the transient TRACE-ICE simulation in CCSM3. There is, however, no analysis of the mechanism of the response of the EASM. The objective of this paper is a thorough understanding of the mechanism of the ice sheet impact on the EASM in CCSM3. This is done with further analyses and sensitivity experiments in a hierarchy of models. Consistent with Yin et al. (2008, 2009) and Sundaram et al. (2012), our experiments show that the ice sheet retreat
decreases the EASM; furthermore, this impact is caused primarily dynamically by the orographic effect of the NH ice sheets via jet migration and stationary Rossby waves. The paper is arranged as follows. After describing our model and experiments in section 2, we present the general climate response over the NH in section 3. In section 4, we show that the removal of NH ice sheet topography in summer decreases the zonal-mean zonal wind over the subtropics and generates an equivalent-barotropic low geopotential height anomaly accompanied by reduced lower-tropospheric southerlies over East Asia. In section 5, we show how the atmospheric circulation changes discussed in section 4 contribute to the reduced EASM precipitation from the perspective of moist static energy. Separate influences of the Laurentide and Eurasian ice sheets on the EASM and role of the ocean are presented in sections 6 and 7, respectively. A summary is given in section 8.

2. Model and experiments

We study the impact of ice sheet on the EASM using CCSM3 (Collins et al. 2006). CCSM3 is composed of the Community Atmospheric Model version 3 (CAM3) at T31 resolution (96 × 48 horizontal grids cells, approximately 3.75° resolution) with 26 hybrid coordinate vertical levels, the Community Land Model version 3 with dynamic vegetation also at T31 resolution, the Parallel Ocean Program (POP) at a nominal 3° horizontal resolution with 25 vertical levels, and the Community Sea Ice Model version 5, a dynamic–thermodynamic sea ice model on the same grid as the ocean model. The model simulates the current climatology well in both the summer and winter. CCSM3 at T31 resolution simulates the EASM reasonably well in terms of the atmospheric wind field, with the EASM characterized by a southeasterly penetrating deep into northern China [see supplementary Figs. 1a,c in Wen et al. (2016)]. Figure 1 shows the present-day summer stationary waves in the slab ocean version of CCSM3, which will be used for sensitivity experiments, and the NCEP–DOE Reanalysis II data (Kanamitsu et al. 2002). It is seen that both the spatial patterns and amplitudes of the stationary waves are reproduced by the model. Notably, the summer stationary waves feature a baroclinic structure, with surface ridges and upper-level troughs over the Pacific and Atlantic and a surface trough and an upper-level ridge over Asia, which is likely induced mainly by diabatic heating [as in Ting (1994)].

First, we analyze a transient simulation of the global climate of the last 19 000 years forced only by the deglacial evolution of the ice sheets (TRACE-ICE; He et al. 2013) in CCSM3. In TRACE-ICE, the ice sheet configuration is altered every 1000 years from 19 to 17 ka BP and every 500 years after 17 ka BP based on the ICE-5G reconstruction (Peltier 2004), with all other forcing factors and boundary conditions fixed at their values at

![Figure 1](image-url)
Table 1. Effects in sensitive experiments in the SLAB model.

<table>
<thead>
<tr>
<th>Effect</th>
<th>Experiment differences</th>
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<tr>
<td>noLIS&amp;noEIS</td>
<td>(0L + 0E) – (100L + 100E)</td>
</tr>
<tr>
<td>noLIS_withEIS</td>
<td>(0L + 100E) – (100L + 100E)</td>
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<tr>
<td>noEIS_withLIS</td>
<td>(100L + 0E) – (100L + 100E)</td>
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<tr>
<td>noLIS_withoutEIS</td>
<td>(0L + 0E) – (100L + 0E)</td>
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<tr>
<td>noEIS_withoutLIS</td>
<td>(0L + 0E) – (0L + 100E)</td>
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19 ka. Second, we assess the climatic effect of ice sheet topography in CCSM3 with six sensitivity experiments, in which ice sheet thickness in the Northern Hemisphere is set at 100%, 80%, 60%, 40%, 20%, and 0% of the LGM ice sheet thickness value (i.e., elevation differences between LGM and preindustrial levels). In all simulations, orbital parameters, greenhouse gas concentrations, coastlines, and sea level remain the same as those at 22 ka. All simulations are initialized with the LGM vegetation. Each simulation is integrated for 200 years, and the last 50 years are used as model climatology for study.

To further isolate the role of oceanic dynamic response, we conducted two sensitivity experiments in the atmosphere model CAM3 coupled with a slab ocean model (SLAB). The slab ocean contains an ocean mixed layer coupled with a thermodynamic sea ice, with the mixed layer depth and global climatology heat flux (Q flux) prescribed as the seasonal cycle of their values in the 100% fully coupled simulation. Therefore, SLAB considers the effect of local oceanic thermodynamic response by fixing the oceanic heat transport as the climatology. The difference between the fully coupled CCSM3 and SLAB is the effect of ocean dynamics. Two sets of sensitivity experiments are performed in SLAB with different combinations of 100% and 0% heights of the two major ice sheets in the form 100L + 100E, 0L + 0E (Table 1), where letters L and E indicate the LIS and EIS, respectively. For example, 100L + 100E represents the experiment with 100% LIS and 100% EIS, while 0L + 0E represents the experiment with 0% LIS and 0% EIS. We can obtain the combined effects of LIS and EIS on the EASM as the difference between 0L + 0E and 100L + 100E experiments (“noLIS&noEIS”) as shown in Table 1. For the SLAB model, each simulation is integrated for 100 years, and the last 50 years are analyzed as model climatology. Note that the extent of ice sheet in all of the sensitivity experiments in CCSM3 and SLAB is the same as in the LGM climate by keeping land surface properties the same as in 100%, as such the albedo effect of the ice sheet remains the same in all experiments and thus is removed between different experiments.

3. Climate response to ice sheet forcing

Our experiments show that a deglacial ice sheet retreat weakens the EASM, with the reduced ice sheet thickness playing a major role. Figure 2a shows the evolution of ice sheet volume in TRACE-ICE experiment. The volume of ice sheets in the Northern Hemisphere decreases by an equivalent sea level of about 117 m from 19 to 7 ka BP. In contrast, the changes in the ice sheet volume in the Southern Hemisphere are much smaller, only of 20-m sea level rise. We choose the EASM index as the summer (June–August) southerly anomalies at 850 hPa averaged in East China (110°–120°E, 20°–40°N; 18 grid points) (Wen et al. 2016) and calculate this index in the TRACE-ICE simulation (red line in Fig. 2b) as well as the sensitivity experiments in the fully coupled model (black circles in Fig. 2b) and SLAB model (100L + 100E and 0L + 0E; blue squares in Fig. 2b). The red line indicates that the EASM weakens in response to retreating ice sheets (reduced albedo and reduced ice sheet thickness). As the ice sheet thickness decreases from 100% to 0% of the LGM level in the sensitivity experiments (black circles, associated with the effect of reduced ice sheet thickness only), the strength of EASM shows a weakening trend similar to that in the TRACE-ICE simulation (the red line), suggesting the dominant role of ice sheet topography on the EASM.

To understand the response of the EASM to ice sheet topography, we first briefly discuss the hemispheric climate response in the NH to ice sheet orographic forcing, as shown in Figs. 3a,b and 4a (experiment 0%–100% in the fully coupled model). As expected, as the ice sheet...
thickness decreases from 100% to 0%, northern North America features a notable warming associated with the lapse rate (Fig. 3b). Interestingly, however, the rest of the NH, such as Alaska, Beringia, and northeastern Asia, is dominated by a cooling response, consistent with annual mean surface temperature anomalies in Lee et al. (2015) and Ullman et al. (2014). We also find a cooling anomaly southeast of the Eurasian Ice Sheet, also

FIG. 3. Anomalies induced by reduced NH ice sheet thickness in summer in (a),(b) fully coupled experiments (0% minus 100%), (c),(d) SLAB experiments [(0L + 0E) minus (100L + 100E)], and (e),(f) AGCM experiments [(0L + 0E_AGCM) minus (100L + 100E_AGCM)] for the following atmospheric variables in JJA: (left) eddy geopotential height at 200 hPa (Z200; shading; gpm) and (right) precipitation (shading; mm day\(^{-1}\)) and surface air temperature (SAT; contours; K). Contours in (a), (c), and (e) represent Z200 climatology at LGM in corresponding models; contour intervals are 30 gpm; dashed contours indicate negative values. In (b), (d), and (f), red solid contours are for SAT anomalies of 1, 2, 3, 6, 10, and 15 K, indicating warming anomalies; blue dashed contours are for SAT anomalies −1, −2, −3, −6, −10, and −15 K, indicating cooling anomalies. The model domains for calculating the EASM meridional wind index at 850 hPa (110°E–120°E, 20°–40°N) are represented in the blue boxes.

FIG. 4. Wind anomalies at 850 hPa (vectors; m s\(^{-1}\)) induced by reduced NH ice sheet topography in summer in (a),(b) fully coupled experiments (0% minus 100%), (c),(d) SLAB experiments [(0L + 0E) minus (100L + 100E)], and (e),(f) AGCM experiments [(0L + 0E_AGCM) minus (100L + 100E_AGCM)].
consistent with previous modeling studies, such as Manabe and Broccoli (1985), Abe-Ouchi et al. (2007), and Löfverström et al. (2014). In the absence of NH ice sheet topography, the Northern Hemisphere upper-tropospheric circulation is less meandering (Löfverström et al. 2014), as shown by the opposite sign of the climatology and anomaly of eddy geopotential height at 200 hPa (Fig. 3a). The reduction of the ice sheet topography generates a negative geopotential height anomaly over western Canada and positive anomalies over Greenland, the United States, and southeastern Canada at 200 hPa, suggesting a weakened climatological stationary wave [Fig. 3a, consistent with Fig. 7c in Hoskins and Karoly (1981)]. The geopotential height anomalies show an anomalous wave train with negative anomalies over the North Atlantic and the eastern European plain and positive anomalies over Europe, Lake Baikal, and the North Pacific. The East Asian region (around 110°–130°E, 30°–60°N) exhibits a relatively lower height in the 0% simulation relative to the 100% simulation, associated with the reduced precipitation (Fig. 3b) and the reduced southerlies at 850 hPa (Fig. 4a). Therefore, the EASM weakens in response to the reduced NH ice-sheet thickness in both the precipitation and wind anomalies.

Ocean dynamics plays a relatively minor role in the ice sheet impact in the coupled model. Indeed, when the ice sheet thickness is reduced from 100% to 0%, EASM weakens in SLAB by a similar magnitude (blue squares in Fig. 2b) as in the fully coupled CCSM3 (black circles in Fig. 2b). Furthermore, the overall climate response over the NH land to ice sheet orographic forcing is similar in SLAB (Figs. 3c,d and 4b) and CCSM3 (Figs. 3a,b and 4a). Therefore, we will focus on the direct orographic influence of the ice sheets by analyzing SLAB simulations in sections 4, 5, and 6, leaving the discussion on the additional indirect influence associated with ocean dynamics to section 7.

4. Atmospheric circulation changes in response to the direct orographic influence of the ice sheets

To understand the direct orographic influence of the ice sheets on the EASM, we show the three-dimensional structure of the global atmospheric anomalies after the removal of both LIS and EIS, that is, the “noLIS_noEIS” effect calculated as the difference of experiments (0L + 0E) − (100L + 100E) (Table 1). The diminished topography of NH ice sheets leads to a weaker and more northward Asian-Pacific jet and North Atlantic jet [Figs. 5a,c,e; consistent with Löfverström et al. (2014)]. The difference of zonal winds in the troposphere shows a dipole mode anomaly: zonal winds are strengthened in high latitudes and weakened in the subtropics (the latitude of EASM), with the wind anomaly centers located over the North Pacific and North America (Figs. 5e,f).
experiment compared to a southwest–northeast tilt in the 100L + 100E experiment at 200 and 850 hPa (Figs. 5a–d; consistent with Löfverström et al. 2014).

When the ice sheet thickness is removed, the NH upper-tropospheric circulation is less meandering and an anomalous stationary wave train is generated in the midlatitudes at 200 hPa (Fig. 5e) similar to that in the fully coupled simulations (Fig. 3a), confirming the dominant role of the direct orographic influence of ice sheet. The geopotential height anomaly shows clearly a midlatitude stationary wave train with four anticyclonic anomalies lying over the United States and southeastern Canada, Europe, Lake Baikal, and the midlatitude North Pacific and four cyclonic anomalies over western Canada, the North Atlantic, the east European plain, and East Asia. These anomalies exhibit an equivalent barotropic structure, as seen in the zonal–vertical section of eddy geopotential height anomalies (Fig. 6a). Figure 6a shows a clear wave train of zonal wavenumber 4. Forced stationary Rossby waves excited by ice sheet topography can radiate wave energy to the east of the ice sheet and into the far field. Following the theory of Hoskins and Ambrizzi (1993), Rossby waves can be trapped in westerly jets as a waveguide. The perturbations over/near the LIS and EIS are transmitted to East Asia along the waveguide. The energy of the stationary wave propagates eastward into East Asia with an equivalent-barotropic low response (Figs. 5e,f; Figs. 6a,b), accompanied by reduced lower-tropospheric southerlies and a weakened EASM (Fig. 6b).

At 850 hPa, a low eddy geopotential height anomaly is generated over the East China Sea (Fig. 7a), accompanied by a weakened EASM characterized by reduced lower-tropospheric southerlies and, in turn, reduced moisture import (Fig. 7c), descending air (Fig. 7b), and anomalous dryness (Fig. 7d) in East Asia.

5. Atmospheric dynamics of the reduced EASM precipitation

We now focus on the mechanism of the weakening EASM from the moist static energy (MSE) perspective and show that the reduced horizontal advection of dry enthalpy contributes predominantly to the decreased precipitation in the EASM region. Sampe and Xie (2010) used thermodynamic energy equation at 500 hPa to illustrate the dominant role of midtropospheric temperature advection in inducing ascending motion along the mei-yu–baiu rainband. Previous work has emphasized the advantages of the MSE budget over the dry thermodynamic equation in the understanding of tropical–subtropical circulations, because the MSE perspective takes into account the effects of temperature, humidity, and diabatic processes as well as the interaction between moist convection and the circulation comprehensively (e.g., Chou and Neelin 2003; Neelin 2007). Chen and Bordoni (2014) indicated that positive horizontal moist enthalpy advection, primarily dry enthalpy advection, plays an essential role in sustaining the mei-yu rainfall band in a region of otherwise negative net energy input into the atmosphere. This role of dry enthalpy advection is also found in all our simulations (not shown). By comparing vertical integral of horizontal dry enthalpy advection anomaly with the separate contributions from the dry enthalpy (Fig. 8b) and latent energy (Fig. 8c), it is clear that dry enthalpy advection anomaly dominates the MSE advection anomaly and in turn the ascending motion and precipitation changes in the EASM region (Figs. 7b,d and 8b). This effect of column-integrated horizontal dry enthalpy advection anomaly (Fig. 8a) with the separate contributions from the dry enthalpy (Fig. 8b) and latent energy (Fig. 8c), it is clear that dry enthalpy advection anomaly dominates the MSE advection anomaly and in turn the ascending motion and precipitation changes in the EASM region. The advection of dry enthalpy can be further shown to be caused by the stationary eddy–mean flow interaction. We decompose the horizontal dry enthalpy advection anomaly into mean, transient, and stationary eddy fluxes following Chen and Bordoni (2014). Therefore, the
response of the dry enthalpy advection can be decomposed as

\[ \langle \delta_{C_p} \nabla \cdot \nabla T \rangle = \delta(\langle C_p \nabla \cdot \nabla T \rangle) + \delta(\langle C_p \nabla \cdot \nabla T^* \rangle) \\
+ \delta(\langle C_p \nabla^* \cdot \nabla \nabla^* \rangle) + \delta(\langle C_p \nabla^* \cdot \nabla \nabla \nabla^* \rangle) \\
+ \delta(\langle C_p \nabla^* \cdot \nabla T \rangle). \]  

(1)

where \( \delta(\cdot) \) is defined as the difference of \((0L + 0E) - (100L + 100E) \) and \( \langle \cdot \rangle \) is defined as a vertical mass integral [i.e., \( \int \langle \cdot \rangle dp/g \)]. The prime denotes the deviation from the monthly time mean for each individual year (e.g., \( u' = u - \bar{u} \)), and the star denotes the deviation from the global zonal mean (e.g., \( u^* = u - [u] \), where the square brackets indicate the global zonal mean). The five terms on the right side represent, in order, the zonal-mean dry enthalpy advection by the zonal-mean flow, the stationary eddy dry enthalpy by the zonal-mean flow, the zonal-mean dry enthalpy by the stationary eddy velocity, the advection of the stationary eddy dry enthalpy by the stationary eddy velocity, and the advection of the transient eddy dry enthalpy by the transient eddies.

In the region associated with EASM precipitation (110°–120°E), we calculate the zonal-mean anomaly of each term.
in Eq. (1) (Fig. 9). When the ice sheet is lowered, total dry enthalpy advection $-\delta \langle c_p \mathbf{v} \cdot \nabla T \rangle$ reduction between 27° and 37°N (black line) results mainly from the reduction in the advection of the stationary eddy dry enthalpy by the zonal-mean flow $-\delta \langle c_p \mathbf{v} \cdot \nabla T^c \rangle$ (red line) and the advection of the zonal-mean dry enthalpy by the stationary eddy velocity $-\delta \langle c_p \mathbf{v}^c \cdot \nabla T \rangle$ (blue line). These two terms represent stationary eddy–mean zonal flow interactions.

The response of the stationary eddy–mean zonal flow interaction can be further decomposed as follows:

$$\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle = \langle c_p \delta ([\mathbf{v} \cdot \nabla T^c]) \rangle + \langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle,$$

$$\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle = \langle c_p \delta ([\mathbf{v} \cdot \nabla T^c]) \rangle + \langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle,$$

where the subscript $c$ stands for the climatology in the 100L + 100E simulation. The $[\mathbf{v} \cdot \nabla T^c]$ term approximately equals the $[\mathbf{v} \cdot \nabla T]$ term, while the $\mathbf{v} \cdot \nabla T^c$ term equals the $(\mathbf{v} \cdot \nabla T)$ term. The total difference in $-\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle$ (Fig. 10a, black line) is mainly due to decrease in zonal-mean zonal wind in the troposphere, $\delta \langle \mathbf{u} \rangle$ in $-\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle$ (Fig. 10a, red line). Figure 11a shows that the reduction in $-\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle$ occurs at all levels below 200 hPa, suggesting the importance of the middle and upper troposphere in addition to midtropospheric temperature advection anomaly mentioned by Sampe and Xie (2010). The term $\delta \langle \mathbf{v} \cdot \nabla T \rangle^c$ is negative in the troposphere because of warmer land than sea in summer (Fig. 11b). Therefore, the decrease in zonal-mean zonal wind in the troposphere ($\delta \langle \mathbf{u} \rangle$; Fig. 11c) contributes to the negative anomaly of $-\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle$. The total difference in $-\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle$ (Fig. 10b, black line) is mainly due to decrease in meridional stationary eddy wind, $\delta \mathbf{v}^c$ in $-\langle c_p \delta ([\mathbf{v} \cdot \nabla T]) \rangle$ (Fig. 10b, red line), in the lower troposphere (Figs. 11d,f).

In short, the lowering of NH ice sheets decreases EASM precipitation through two mechanisms: the jet...
shifting mechanism and the stationary wave mechanism. In the jet shifting mechanism, the removal of ice sheet topography shifts the jet northward and reduces the mean westerly jet \( \bar{u} \) in the subtropical region, which leads to a reduction of the dry enthalpy advection in the whole troposphere over East Asia, and in turn a reduction of the EASM precipitation. In the stationary wave mechanism, the lowering of ice sheet topography induces an anomalous stationary wave response, such that the meridional stationary eddy velocity \( v^* \) is reduced over East Asia, also leading to a reduction of dry enthalpy advection and in turn rainfall of EASM. This stationary eddy mechanism has been discussed by Chen and Bordoni (2014) in their study of the role of Tibetan Plateau on the formation of the MB front. The role of the westerly jet in the EASM has also been discussed in general recently (Molnar et al. 2010; Chiang et al. 2015), although how the ice sheet affects the jet and in turn downstream EASM has not been discussed specifically.

6. Separate influences of the Laurentide and Eurasian Ice Sheets on EASM

To identify the separate effects of LIS and EIS on EASM, we also conducted two experiments in SLAB denoted as 0L + 100E and 100L + 0E in which either topography of LIS or EIS is removed, respectively. At LGM, we can consider the effect of the removal of LIS only as the experiment difference of (0L + 100E) − (100L + 100E) or noLIS_withEIS, and the removal of EIS only as (100L + 0E) − (100L + 100E), or noEIS_withLIS. (Table 1).

Our experiments show that the lowering of LIS and EIS both play an important role in the weakening of EASM. The lowering of LIS and EIS both contribute to the equivalent-barotropic low response and reduced lower-tropospheric southerlies over East Asia (Figs. 12a–d and 13a,e). In the meantime, only the lowering of LIS induces the northward shift of the westerly jet and decreased zonal-mean zonal wind over the subtropics (Figs. 12a,b), because the LIS is in the latitude range of the jet and therefore has a strong impact on the jet. Therefore, the lowering of LIS weakens the EASM through shifting the jet and altering stationary waves, while the lowering of EIS weakens the EASM only through altering stationary waves.

Yin et al. (2008) and Sundaram et al. (2012) showed that EASM precipitation decreases mainly due to the removal of Eurasian ice sheet (EA case), contrary to our model results. However, there is a difference in the experimental setup. To make a more qualitative comparison with their studies, we also consider the separate effects of the removal of LIS and EIS without the presence of the other ice sheet as the experiment difference of (0L + 0E) − (100L + 0E) and (0L + 0E) − (0L + 100E), or noLIS_withoutEIS and noEIS_withoutLIS (Table 1). The experiment noEIS_withoutLIS is a purely idealized experiment with no real world relevance, which is
similar to the EA case in Yin et al. (2008) and Sundaram et al. (2012). It is seen that the anomalies in the effects of noLIS_withoutEIS (comparing Figs. 12e,f with Figs. 12a,b, and Figs. 13e,f with Figs. 13a,b) are featured with decreased zonal-mean zonal wind and less precipitation (similar to noLIS_with EIS). However, stationary wave patterns east of EIS are different in noLIS_withoutEIS (absence of equivalent-barotropic low response and reduced lower-tropospheric southerlies). The anomalies in the effects of noEIS_withoutLIS are not significant (Figs. 13g,h), which is different from the dominant role of Eurasian ice sheet shown in Yin et al. (2008) and Sundaram et al. (2012).

Fig. 11. Vertical profile of variables over 110°–120°E, 27°–37°N in JJA: (a) $-c_p \delta[\bar{u}] \cdot \partial T \bar{c} / \partial x$, (b) $\partial T \bar{c} / \partial x$, (c) $\delta[\bar{u}]$, (d) $-c_p \delta \bar{v} \cdot [\partial T \bar{c} / \partial y]$, (e) $[\partial T \bar{c} / \partial y]$, and (f) $\delta \bar{v}$. The subscript $c$ stands for the climatology in the 100L + 100E simulation; $\delta(\cdot)$ is defined as the difference of $(0L + 0E) - (100L + 100E)$. The star denotes the deviation from the global zonal mean; e.g., $u^* = u - [u]$. 

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Several reasons may account for this difference, as follows. First is the model dynamics. The model they employed—LOVECLIM—is of intermediate complexity with quasigeostrophic atmospheric dynamics, and with coarse vertical resolution (three levels) and imposed stratification. CCSM3, on the other hand is a full general circulation model with finer resolution in the vertical. Second, the ice sheet configuration is different. Their North American ice sheet is located more north and its height is only about half of our LIS. The position of their Eurasian ice sheet is about 20° east of our EIS.

7. Role of the ocean

The indirect dynamic effect associated with the AMOC change caused by the lowering of ice sheets on EASM is not important. This can be seen from the similar responses in fully coupled CCSM3 and SLAB in section 3. Previous studies show that the lowering of ice sheets tends to weaken the AMOC (Eisenman et al. 2009; Zhang et al. 2014; Zhu et al. 2014; Sherriff-Tadano et al. 2018) and a decrease in the AMOC strength can lead to a decrease in the EASM (Fallah et al. 2016; Wen et al. 2016). Thus, the lowering of ice sheets may weaken EASM through the effect of AMOC change. We also simulated weakened AMOC in response to the lowering of the NH ice sheets: maximum AMOC decreases from 11.56 Sv (1 Sv = 10⁶ m³ s⁻¹) (100%) to 7.64 Sv (0%) in the fully coupled sensitivity experiments, which is consistent with Zhu et al. (2014), who showed that AMOC weakened significantly (by 41%) in response to the de-glacial ice sheet retreat. A weakened AMOC can also be validated from a typical dipole SST anomaly: cooling in the North Atlantic and weaker warming at the equator and the South Atlantic (Fig. 14c). We use differences of the SST anomalies in the sensitivity experiments (Fig. 14c) between the fully coupled experiments (0%–100%; Fig. 14a) and the SLAB experiments [(0L + 0E) – (100L + 100E); Fig. 14b] to represent the SST changes induced by ocean dynamics. However, the weakened AMOC seems to have limited impacts on the EASM compared to the direct orographic effect of the ice sheets.

Indeed, the total ocean effect, including both thermal and dynamic, is not important. This can be seen in the AGCM experiment (CAM3 with fixed SST and sea ice) without the ocean effect at all (Figs. 3e,f and 4c). We repeat the sensitivity experiments in the atmosphere alone model, with the atmosphere general circulation model CAM3 forced by the seasonal cycle of SST and sea ice cover prescribed from the 100% fully coupled
FIG. 13. Anomalies of (a),(c),(e),(g) low-level wind (vectors; m s$^{-1}$) and (b),(d),(f),(h) precipitation (shading; mm day$^{-1}$) in JJA. The effect represented by each row is the same as in Fig. 11. Pink shadings and black dots represent, respectively, meridional wind and precipitation anomalies that are statistically significant at a 95% confidence level.
simulation as 100L_100E_AGCM, 0L_0E_AGCM. Therefore, the difference between SLAB and AGCM simulations shows the effect of local ocean thermodynamics, and the difference between CCSM3 and AGCM simulations gives the total effect of the ocean. In all the three model configurations, the lowering of ice sheets induces weakened EASM monsoon wind (Fig. 4) and reduced precipitation (Figs. 3b,d,f). In contrast, Sundaram et al. (2012) considered the effect of ocean as a significant link in how ice sheets affect EASM. They argued that the presence of ice sheets over North America and Eurasia during MIS-13 induces a positive phase of the winter North Atlantic Oscillation (NAO)-like feature. The ocean allows the NAO like feature to persist until summer and excite a wave train. In our model, these oceanic effects do not seem to be important.

In summary, our results suggest that the direct ice sheet orographic effect on the EASM is dominant while the indirect effects of ocean dynamics and thermodynamics are not important.

8. Summary

Our CCSM3 model study suggests that receding ice sheets weaken the EASM with the ice sheet topography playing a major role. Two mechanisms are implicated in the connection between the ice sheet and EASM, as shown schematically in Fig. 15. In the jet shifting mechanism, the removal of ice sheet topography in the Northern Hemisphere induces a northward shift of westerly jet and decreased zonal-mean zonal wind over the subtropics in summer; the reduced jet decreases the advection of dry enthalpy into East Asia, and therefore reduces EASM rainfall there. Meanwhile, the lowering of ice sheet induces an anomalous stationary Rossby wave train from North America all the way to East Asia, along the westerly jet waveguide; the wave propagation causes an equivalent-barotropic low geopotential height anomaly over East Asia accompanied by reduced lower-tropospheric southerlies, which in turn reduces the advection of dry enthalpy and hence EASM rainfall.

There are still outstanding questions about the role of the ice sheet topography on climate and the EASM. We note that in a recent study (Roberts et al. 2019), the global effect of the Laurentide Ice Sheet on summer stationary waves is contributed significantly not only by the topography effect, but also by the surface albedo effect. Our experiments show a similar impact of the combined topography and albedo effect compared to their results (figures not shown). However, in our experiments, the topographic effect is dominant, while the

1. the jet shifting mechanism

- removal of NH ice sheet topography
- teleconnection
- jet stream northward shift
- decreased zonal mean zonal wind over subtropics
- an equivalent-barotropic low response over East Asia
- reduced low level southerly winds over East Asia
- reduced horizontal dry enthalpy advection
- less EASM precipitation

2. the stationary wave mechanism

Fig. 15. Schematic of the influence of the removal of NH continental ice sheet topography on EASM. The orange color represents a weakened EASM.
albedo effect plays a minor role. Further studies are needed to understand the difference among different models. In addition to stationary wave change, there may remain alternative mechanisms which could explain the reduced lower-tropospheric southerlies over East Asia in response to the removal of NH continental ice sheet topography. For example, it is possible some disturbance caused by LIS can propagate into the tropical Atlantic first and then to the Asian sector in the equatorial Kelvin wave, which can also affect EASM (Ueda et al. 2009). We plan to further study these mechanisms in the future.

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Data availability. The NCEP–DOE Reanalysis II can be obtained from https://psl.noaa.gov/data/gridded/data.ncep.reanalysis2.html. The ICE-5G reconstruction data-set is available at https://pml.ncep.noaa.gov/products/ice/ice5g.html. The NCEP–DOE Reanalysis (R-2). The ICE-5G reconstruction data-set is available at https://pml.ncep.noaa.gov/products/ice/ice5g.html. The data underlying the figures from this manuscript are available at https://doi.org/10.7910/DVN/1ZHRSD.

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