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A tropical mechanism for Northern Hemisphere deglaciation

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[1] We investigate the role of the tropics in the melting and reforming of the Laurentide ice sheet on glacial timescales using an atmospheric general circulation model. It is found that warming of tropical sea surface temperatures (SSTs) from glacial boundary conditions, as observed at the end of glacial periods, causes a large increase in summer temperatures centered over the ice sheet-forming regions of Canada. This high-latitude response to tropical change is due to relatively small changes in the circulation of the extratropical atmosphere, which lead to changes in the vertical profiles of temperature and moisture in the extratropical atmosphere. The maximum perturbation in the summer radiative balance over the Laurentide ice sheet (>25 W/m^2) due to the changes in the local atmospheric water vapor inventory is much larger than that induced by glacial to interglacial changes in atmospheric CO_2. These changes via an atmospheric bridge between the tropics and extratropics represent a mechanism for deglaciations which is consistent with timing constraints. In contrast, a cold perturbation to tropical SST for interglacial boundary conditions results in only very small changes in the delivery of water vapor to the Laurentide region, and therefore almost no cooling over the Canadian region. This implies that tropical SSTs could play a more important role in melting ice sheets in the Northern Hemisphere than in reforming them, possibly providing a mechanism which could help to explain the rapidity of deglaciation relative to glacial inception.

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1. Introduction

[2] Until quite recently, discussion of mechanisms for glacial-timescale climate variability focused on processes in the Northern Hemisphere. Within the last decade, however, analyses of ice core and ocean-sediment records have shown that for the last two deglaciations, changes in temperatures over Antarctica precede changes in North American ice volume by more than 1000 years [*Sowers and Bender*, 1995; *Broecker and Henderson*, 1998; *Henderson and Slowey*, 2000; *Shackleton*, 2000]. Other paleo-records suggest that tropical SST also leads the high-latitude Northern Hemispheric ice volume during deglaciation [*Bard et al.*, 1997 (Somali Basin records); *Schneider et al.*, 1999; *Lea et al.*, 2000; *Nürnberg et al.*, 2000; *Herbert et al.*, 2001; *Seltzer et al.*, 2002; *Visser et al.*, 2003]. This chronology calls into question many of the mechanisms which have been proposed for deglaciation, especially those involving changes in North Atlantic overturning or internal ice sheet instabilities as triggers of deglaciation.

[3] Two classes of theories have been articulated which propose mechanisms nonlocal to the high-latitude northern hemisphere for driving deglaciation. One class of theories [*Cane*, 1998; *Bush and Philander*, 1998; *Clement et al.*, 1999] argues that processes internal to the tropics can generate changes within the tropics which then propagate to the region corresponding to the Laurentide ice sheet through the atmosphere. Another class of theories argues that it is changes in high-latitude Southern Hemisphere insolation that trigger a Southern Ocean change [*Toggweiler et al.*, 1999; *Sigman and Boyle*, 2000], which then propagates to the high-latitude Northern Hemisphere, possibly through control of CO_2.

2. Model Experiments

[4] This study focuses on a mechanism whereby changes in tropical SST can drive changes in air temperature over modern-day Canada. We turn to an atmospheric general circulation (AGCM) study of the LGM [*Lohmann and Lorenz*, 2000] together with two new model runs. The model experiments were performed with the ECHAM3 AGCM [*Roeckner et al.*, 1992], a primitive equation model which includes radiation and a hydrological cycle. In the spectral domain the model equations are solved with a triangular truncation of wave number of 42 (T42), which corresponds to a 2.8 degree grid spacing on which the physical parameterizations are applied. The model uses 19 hybrid levels in the vertical. Four model runs are considered, and they comprise two separate sensitivity studies. The first sensitivity study (deglaciation) consists of a warm tropical perturbation to glacial boundary conditions, and the second (glaciation) consists of a cold perturbation applied to interglacial boundary conditions. A fifth run was analyzed as a complement to the deglaciation study for which a warm perturbation was applied to the North Pacific rather than the tropics. For each experiment, the model is run for 15 years, and an average year is constructed from the last 10 years of the simulation.

2.1. Imposed Boundary Conditions

[5] SST reconstructions for the last glacial maximum are controversial [*Bard*, 1999; *Crowley*, 2000; *Rind and Peteet*, 1985; *Broccoli and Marciniak*, 1996] and one must be wary of the limitations of any such reconstruction used as a lower boundary condition for AGCM experiments. There is growing evidence that the CLIMAP reconstruction [*CLIMAP Project Members*, 1981] gives tropical SSTs that have a warm bias when compared to SSTs inferred from other paleo-proxies [*Lea et al.*, 2000; *Nürnberg et al.*, 2000; *Crowley*, 2000; *Guilderson et al.*, 1994; *Mix et al.*, 1999]. In order to adjust our SST boundary conditions used in the modeling study of deglaciation to these new results for glacial SSTs, we took the CLIMAP SST reconstruction and uniformly subtracted 3°C in the tropics (between 30°N and 30°S). It has been shown that the resulting last glacial maximum climate using these SSTs provides a climate which is more consistent with terrestrial and marine proxy data [*Lohmann and Lorenz*, 2000].

[6] The model experiment we shall denote as LGM uses boundary conditions corresponding to the last glacial maximum, and LGM_WTP (last glacial maximum with warm tropical perturbation) differs

from LGM only in a uniform axisymmetric 3°C warm perturbation between 30°N and 30°S, with a transition zone over several grid points between 30° and 40° latitude in either hemisphere. Likewise the run IGL (interglacial) uses modern boundary conditions and IGL_CTP (interglacial with cold tropical perturbation) has a uniform 3°C cold tropical perturbation, also applied through the SST boundary condition. The fifth run, called LGM_WPAC (last glacial maximum with a warm Pacific), adds a warm perturbation over the North Pacific in addition to the LGM_WTP SST. This warming amounts to 3.5°C and is centered at 52°N with a smooth transition zone from 40°N to the Bering Strait.

2.2. Model Response

[7] The imposed warming of the glacial tropics by 3°C causes a significant model response in summer surface (2 meter) air temperature over North America (Figure 1a). This response is in excess of 3°C over most of the Laurentide ice sheet, with maximum values in excess of 6°C. The fact that the thermal response over the Laurentide ice sheet is larger than the tropical perturbation attests to the sensitivity of this area to tropical changes. Temperature differences between the runs are smaller during winter than during summer, and thus not only the mean surface temperature over the Laurentide ice sheet, but also the seasonal temperature cycle increases in response to an axisymmetrical tropical SST perturbation. The warming observed in tropical proxy records at the end of glacial periods (*Bard et al.* [1997] (Somali Basin Records), *Lea et al.* [2000], and *Nürnberg et al.* [2000]) is therefore expected to cause significant warming over glacial ice sheets, and this provides a viable mechanism for deglaciation.

[8] In the opposite case, when a cool tropical perturbation is applied to modern (interglacial) SST boundary conditions, the sensitivity of summer temperatures over the Hudson Bay region is much smaller (Figure 1b). The cooling over Canada is typically less than 1°C, and is thus weaker than the tropical SST perturbation. The seasonal cycle in surface air temperature is increased by the imposed tropical cooling, especially for the region surrounding Alaska (i.e., the region north of 60°N, and west of 100°W), where the winter temperatures are up to 10°C cooler in winter. However, in eastern Canada the difference in the seasonal cycle is not as pronounced, and the annual mean surface temperature differences over the region surrounding Hudson Bay are still only of order 1°C.

[9] The effect of adding an additional warm perturbation to the extratropical Pacific can be seen in Figure 1c (LGM_WPAC minus LGM_WTP). The temperature response over the Laurentide ice sheet is of order 1°C, and is thus much smaller than the response detected in Figure 1a. Here we see that even an imposed warming of the ocean immediately adjacent to the ice sheet does not induce a temperature change comparable with that due to changes of the Pacific warm pool.

[10] Despite the fact that the magnitude of the imposed perturbation in the meridional SST gradient is the same, but opposite in sign, for both the deglaciation (Figure 1a) and glaciation (Figure 1b) sensitivity studies, the structure and amplitude of the summer response over the region surrounding Hudson Bay is quite different. In order to understand why the extratropical response is stronger for the deglaciation scenario, we consider the changes in atmospheric circulation and radiative balance associated with the surface temperature perturbations shown in Figure 1 (case LGM_WTP minus LGM).

[11] A warm tropical SST perturbation under glacial conditions causes a reduction in the flow of relatively cold and dry air from the Arctic Ocean onto the North American continent, and a corresponding increase in the supply of relatively warm and moist air from the North Pacific (Figure 2). Although these changes in circulation are not large relative to the mean circulation, the resulting differences in water vapor transport are sufficient to impact the summer radiative balance over the Laurentide region. The warm tropical perturbation under glacial conditions also leads to changes in incoming (backward) longwave radiation at the surface (Figure 3). The differences in this field represent differences in the atmospheric water vapor inventory, and the differences are maximum over the Laurentide ice sheet. The position of the

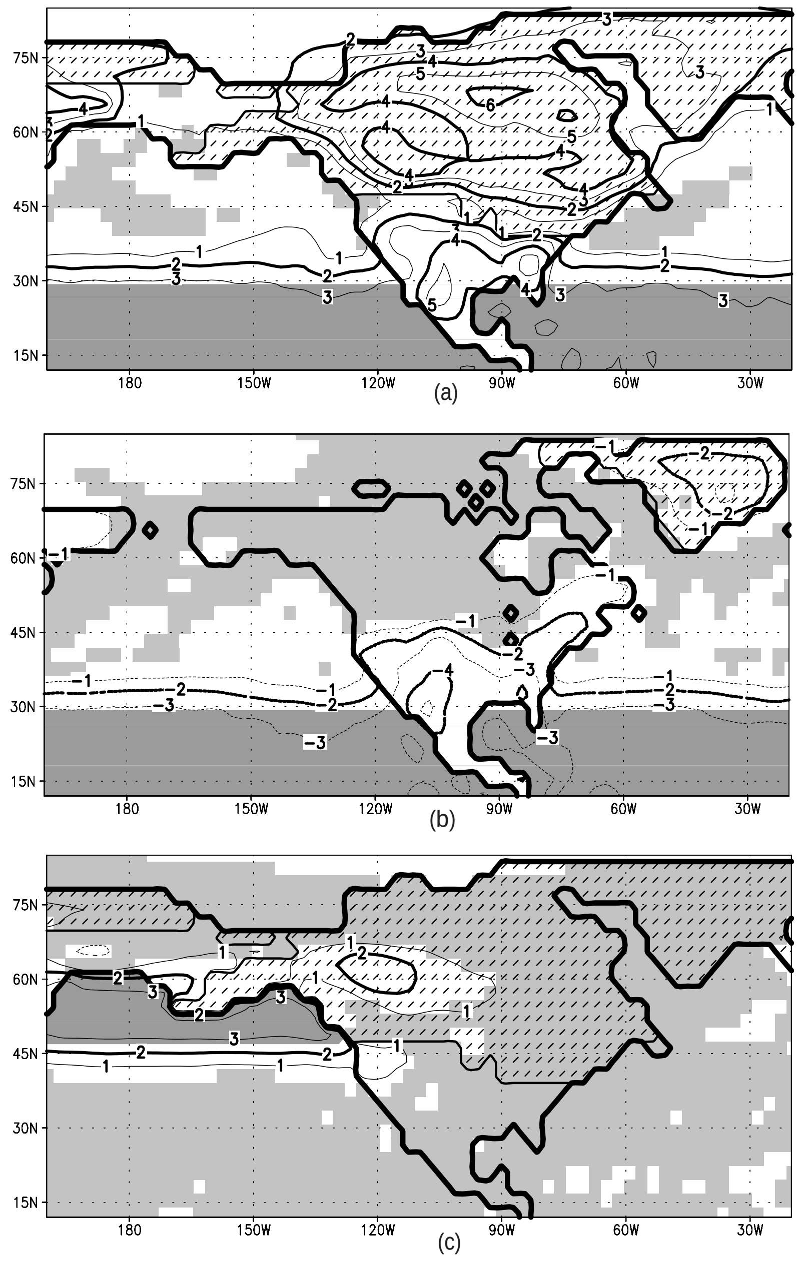

Figure 1. ECHAM3 model results for the North American region: surface air temperature differences caused by (a) a warming of tropical SST from glacial conditions (LGM_WTP minus LGM); (b) a cooling of tropical SST from interglacial conditions (IGL_CTP minus IGL), (c) an additional warm perturbation to the North Pacific (LGM_WPAC minus LGM_WTP). All panels show model results for summer, i.e., mean of June, July, and August. The units are °K. Stipling indicates the location of continental ice in the AGCM. Model experiments have been run for 15 years where the first 5 years are taken as time for model spin-up. The last 10 years have been used for evaluating the mean and the variance of temperature. A student t-test has been performed to assess the significance of the differences. Areas where temperature differences are not significant to a 95% level are lightly shaded. Dark shading highlights areas where the SST perturbations are supplied. For the 5 model runs discussed in this paper, 3 correspond to runs discussed in *Lohmann and Lorenz* [2000]. LGM is their LGM.N, LGM_WTP is their LGM.O, and IGL corresponds to their CTL run.

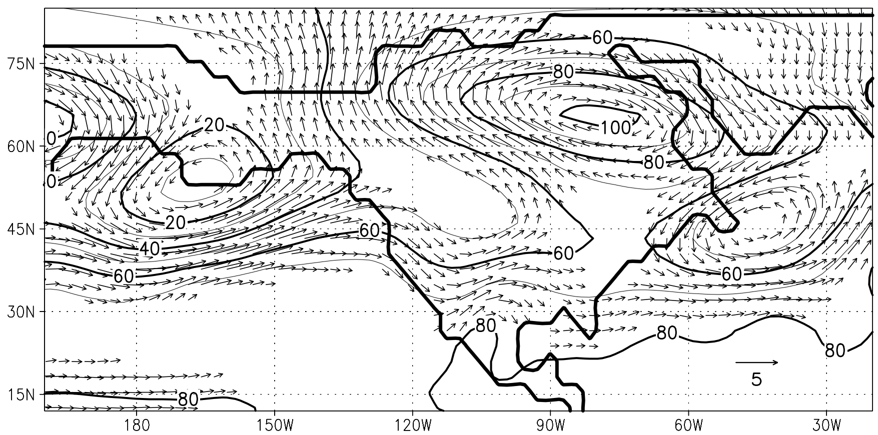

Figure 2. The difference in circulation and geopotential height for the 500 hPa surface (LGM_WTP minus LGM) averaged over the summer months (June, July, August) is shown. The scale for the wind vectors is shown in the lower right hand corner (units of meters/second). The units for geopotential height are meters.

region of maximum incoming (backward) longwave radiation corresponds roughly to the position of the 5°C contour shown in Figure 1a. The maximum values (in excess of 25 W/m^2) are much larger than those induced by glacial to interglacial changes in atmospheric CO_2.

[12] Surface air temperature differences for the 3 sensitivity studies along a track (position of track shown in Figure 3), running from the North Pacific over the Laurentide ice sheet, to the region corresponding to maximum temperature differences in Figure 1a, reveal important differences between the sensitivity studies. For a warm tropical perturbation under glacial conditions (LGM_WTP minus LGM), the temperature differences are only 0.2°C at sea level, due to the SST boundary condition used over the open ocean. Moving to the northeast along the track, the temperature difference increases with altitude. In fact, the temperature continues to increase even after the maximum altitude (780 hPa) has been reached. This reveals that although the temperature differences tend to be larger with altitude, this is not the only control on the temperature difference.

[13] For a warm perturbation to the North Pacific under glacial maximum conditions (LGM_WPAC and LGM), it is clear from Figure 4 that the differences decrease as one moves inland along

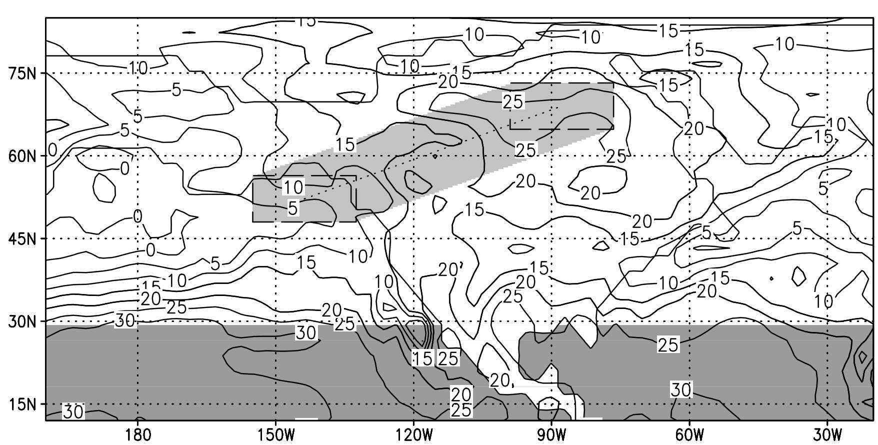

Figure 3. The perturbation in longwave backward radiation (LGM_WTP minus LGM) averaged over the summer months (June, July, August) is shown in units of W/m^2. This quantity reflects changes in the column inventory of atmospheric water vapor, and the maximum values (>25 W/m^2) over Northern Canada correspond to the regions of maximum surface temperature perturbation shown in Figure 1a.

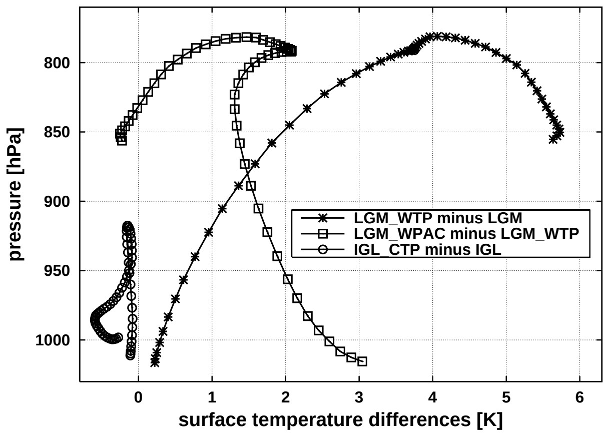

Figure 4. Profiles of surface temperature differences (average over June, July, August) along the track indicated by the lightly shaded region in Figure 3 are shown. The track extends from the North Pacific to Northern Canada, and connects the regions of minimum and maximum extratropical temperature perturbations seen in Figure 1a. The surface temperatures (units of K) are shown as they vary with surface pressure (units of hPa) along the track. (Abbreviations in legend are as follows: LGM_WTP, last glacial maximum with warm tropical perturbation; LGM, last glacial maximum; LGM_WPAC, last glacial maximum with a warm Pacific; IGL_CTP, interglacial with cold tropical perturabation; IGL, interglacial.)

the track. The surface air temperature difference of 3°C reduces strongly when moving inland and the temperature differences are very close to zero at the northeastern extreme of the track. For a cold tropical perturbation to modern boundary conditions (IGL_CTP and IGL) the temperature differences are everywhere very small, and do not change with altitude as one traverses the transect of Figure 3. Note that the effective range of pressures (vertical axis) is much smaller for this sensitivity study, due to the absence of the Laurentide ice sheet.

3. Discussion

[14] The enhanced sensitivity of surface air temperatures over the Laurentide region to nonlocal tropical SST perturbations for the deglaciation scenario (Figure 1a) relative to the tropical cooling scenario (Figure 1b) is due to differences in the response of the extratropical atmospheric circulation to the tropical SST perturbations. Associated with the enhanced extratropical circulation response for the deglaciation scenario is an anomalous advection of warm and moist air off the North Pacific and onto the North American continent. This leads to a reduction in the environmental lapse rate of the extratropical atmosphere over North America. The lapse rate affects the surface temperature in the presence of the Laurentide Ice sheet and provides a mechanism which is capable of amplifying the tropical warming.

[15] The remote effect of tropical SSTs on Laurentide ice sheet summer temperatures provides a strong non-linearity in the climate system. When the global climate is in a cold phase of the 100 kyr cycle, the vertical orographic expression of the ice cover over North America is in excess of 2 km, and a warming of tropical ocean temperatures may be able to accomplish a net melting of the continental ice during summer months. This mechanism is significantly diminished when the global climate is in a warm phase, as most of eastern Canada is very close to sea level. Stated differently, our results imply that the orographic perturbation posed by the Laurentide ice sheet results in an amplification of atmospheric teleconnections between the tropics and the extratropics.

[16] Our interpretation of the numerical experiments as a possible deglaciation scenario rests upon two assumptions, (1) that tropical warming occurs before ice sheet melting commences and (2) that it is summer temperatures over modern-day Canada which are the most important factor in determining the growth and decay of ice sheets. The first of these assumptions is supported by observations in the tropical Pacific [*Lea et al.*, 2000], tropical Indian (Somali Basin records [*Bard et al.*, 1997]) and tropical Atlantic [*Nürnberg et al.*, 2000] Oceans, which all reveal tropical SSTs leading ice volume changes at deglaciation (Figure 5). For the cases of the Pacific and the Atlantic, the timing of the SST warming relative to the meltwater pulses of *Fairbanks et al.* [1989] is shown in Figure 5a. By the time of the first meltwater pulse, the Pacific SSTs have completed more than half of their 3°C glacial to interglacial warming. Over the last 4 deglaciations (Figure 5b), the *Lea et al.* [2000] data reveals that the SST (red) leads ice volume (blue). As both the SST and ice volume

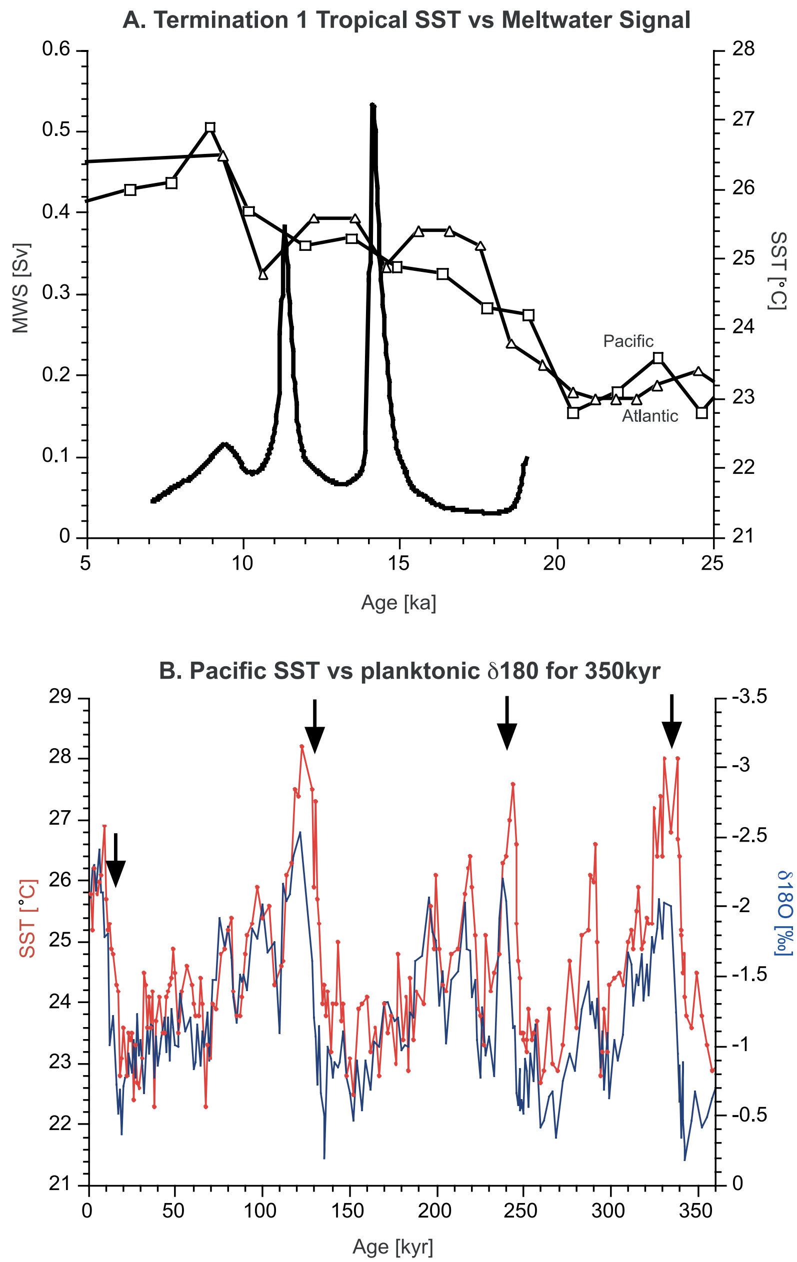

Figure 5. Paleodata demonstrating the lead of tropical warming over ice sheet collapse at deglaciations. (a) Tropical SST records from the Pacific [*Lea et al.*, 2000] (Core TR163-19 90°57′W, 2°16′N, 2348 m) and the Atlantic [*Schneider et al.*, 1999] (Core GeoB 1105-4 1°40s, 12°26′W, 3225 m) compared to the sea level-driven meltwater curve for the last deglaciation [*Fairbanks*, 1989]. (b) 350 kyr record of SST based on foraminiferal Mg/Ca from *Lea et al.* [2000] plotted against $\delta^{18}O$ for the same core. Although this planktic $\delta^{18}O$ curve contains a component of SST, the clear lead of SST over $\delta^{18}O$ at each of the four terminations (highlighted by arrows) demonstrates that the ice-volume component of $\delta^{18}O$ lags behind SST. Note that a similar lag is not generally observed on the return to cool conditions.

proxies are taken from the same cores, this phasing signal is not age model dependent.

[17] The second assumption follows that of *Milankovitch* [1930] who argued that for the net snow balance (melt versus accumulation) of glacial-interglacial changes, summer melt is more important than winter accumulation. The ECHAM3 model does not represent the positive feedback of reduced albedo due to ice melting associated with the

temperature increase. This ice-albedo feedback would serve to enhance the differences seen in Figure 1a.

[18] We have only considered spatially uniform perturbations to tropical SST. *Yin and Battisti* [2001] discuss atmospheric model experiments conducted with the Community Climate Model Version 3 (CCM3), which show that the pattern of extratropical atmospheric circulation in the presence of a Laurentide ice sheet is significantly more sensitive to changes in the pattern of tropical SSTs than to spatially uniform changes in average tropical SST. Despite the sensitivity of the extratropical atmospheric circulation, however, surface temperatures over the Laurentide ice sheet were not significantly affected by changes in the pattern of SST [see *Yin and Battisti*, 2001, Figure 11b]. This suggests that introducing spatial variability to tropical SST change will not alter the basic conclusions of the model results presented here.

[19] Our study has taken tropical warming as the starting point for Northern Hemisphere deglaciation, i.e., the time before 15 ka BP in the last deglaciation (Figure 5a). This is reasonable, as there is considerable evidence that tropical warming leads ice sheet melting, but it begs the question of which mechanism causes this warming. One possible contender is a direct tropical response to insolation. *Clement et al.* [1999] have shown, with a coupled model of intermediate complexity for the equatorial Pacific, that although precession changes the seasonality (but not the annual mean) of insolation in the tropics, the coupled system is capable of rectifying these seasonal changes, leading to a change in the mean state. The timing of the changes in their model are consistent with the observed changes for the last deglaciation (Figure 5a). A sizable tropical SST response to the precessional component of Milankovich variations is also seen in the three-dimensional global coupled ECBilt model [*Tuenter et al.*, 2003], thus indicating that the rectification effect seen by *Clement et al.* [1999] is not model dependent.

[20] Another possible mechanism to cause warming of the tropics is that of CO_2-induced greenhouse warming. It is difficult to reconcile the observation that deglacial changes in CO_2 lead changes in northern-hemispheric ice volume without invoking changes in the Southern Ocean since it is only in high latitudes that the atmosphere can interact with the deep ocean CO_2 reservoir [*Broecker and Henderson*, 1998]. Our proposed tropical mechanism for deglaciation would still function if tropical warming was induced by such a Southern Hemisphere controlled CO_2 change. The lead of Southern Ocean SSTs over Northern Hemispheric ice volume was first documented by *Hays et al.* [1976], and later verified by *Pichon et al.* [1992] and *Mashiotta et al.* [1999]. Indeed, the apparent synchroneity of temperature changes between the tropics and the high-latitude Southern Hemisphere [*Lea et al.*, 2000] suggests that there is interaction between the Southern Ocean and the tropics on Milankovitch timescales.

[21] In arguing for our proposed mechanism, we have used output from an AGCM (ECHAM3) that was configured to conform to the Paleo Model Intercomparison Project (PMIP) [*Joussaume and Taylor*, 2000] protocols. As such, the model representation of the mass balance over the Laurentide ice sheet is not sufficient to resolve the differences in accumulation resulting from the sensitivity study. Air temperature certainly plays an important role in mass balance, but precipitation budgets also need to be considered. *Pollard* [2000] has pointed out that none of the PMIP models (including ECHAM3) are capable of maintaining a positive mass balance over the Laurentide ice sheet when applying CLIMAP SST boundary conditions (LGM_WTP in our notation). Thus our proposed mechanism provides a plausible explanation why the PMIP models using the warm Pacific SSTs as given by the CLIMAP reconstruction show a negative mass balance over the Laurentide ice sheet area.

4. Conclusions

[22] We have presented a mechanism for deglaciation whereby changes in tropical SSTs can impact the stability of the Laurentide ice sheet via an atmospheric bridge. Our tropical mechanism for Northern Hemispheric deglaciation accounts for the main shortcomings of theories which rely on

internal ice sheet dynamics to explain deglaciation [*Tarasov and Peltier*, 1997], namely the problem that the changes in the tropics and high-latitude Southern Hemisphere occur before ice volume changes [*Sowers and Bender*, 1995; *Broecker and Henderson*, 1998; *Shackleton*, 2000]. Our deglaciation scenario is furthermore consistent with sedimentological data from lakes in the tropical Andes indicating that tropical warming occurred several thousand years before the Boelling-Alleroed warming in the Northern Hemisphere [*Seltzer et al.*, 2002].

[23] The suggested mechanism is distinct from the ENSO (El Niño/Southern Oscillation) teleconnection mechanism that has previously been proposed for a tropical role in glacial cycles [*Cane*, 1998]. According to that mechanism, ENSO-like changes in the tropical SST pattern induce large-scale changes in the extra-tropical atmospheric circulation, which themselves induce large-scale changes in surface air temperatures over Hudson Bay, thus modulating continental ice accumulation. In contrast, the mechanism discussed here does not rely on changes in the tropical SST pattern, but rather a spatially uniform perturbation to tropical SSTs. These changes alter the summer radiative budget over the Laurentide ice sheet by impacting atmospheric water vapor budgets which is associated with relatively small perturbations to the atmospheric circulation. Both mechanisms could be important on glacial timescales but, unlike the ENSO teleconnections mechanism, our proposed mechanism exhibits a marked asymmetry in its effectiveness between glaciation and deglaciation scenarios.

Acknowledgments

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