Basal temperature evolution of North American ice sheets and implications for the 100-kyr cycle

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[1] We simulate three-dimensional ice temperature fields to examine spatial-temporal history of the subglacial thermal environment during the last glacial cycle. Model results suggest that 60–80% of the Laurentide Ice Sheet was cold-based (frozen to the bed) at the LGM, and therefore unable to undergo large-scale basal flow. The fraction of warm-based ice increases significantly through the ensuing deglaciation, with only 10–20% of the Laurentide Ice Sheet frozen to the bed by 8 kyr BP. This basal thermal evolution, a function of both climatic and ice sheet history, could enable a dynamical switch to widespread basal flow through the deglacial period. Because basal flow has the capacity to evacuate large amounts of ice from the interior of continental ice sheets, creating thin and climatically-vulnerable ice masses, this switch in flow regime may have played a significant role in glacial terminations and the 100-kyr glacial cycle. *INDEX TERMS:* 1620 Global Change: Climate dynamics (3309); 1827 Hydrology: Glaciology (1863); 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology. **Citation:** Marshall, S. J., and P. U. Clark, Basal temperature evolution of North American ice sheets and implications for the 100-kyr cycle, *Geophys. Res. Lett.*, *29*(24), 2214, doi:10.1029/2002GL015192, 2002.

1. Introduction

[2] The origin of the dominant ~100-kyr ice-volume cycle in the absence of substantive radiation forcing remains one of the most vexing questions in climate dynamics [*Imbrie et al.*, 1993]. An important characteristic of the 100-kyr cycle is its asymmetric structure, with long (~90 kyr) fluctuating ice-growth phases followed by rapid (~10 kyr) terminations. To explain this structure, some models suggest that precession and obliquity forcing cause ice sheets to grow to some critical size beyond which they no longer respond linearly to orbital forcing. Rapid terminations then occur through nonlinear feedbacks in the climate system, possibly triggered by the next Northern Hemisphere summer insolation maximum [e.g., *Imbrie et al.*, 1993; *Tarasov and Peltier*, 1997; *Gildor and Tziperman*, 2001].

[3] Previous modeling studies that address West Antarctic ice sheet stability [*MacAyeal*, 1992] and the origin of Heinrich events [*MacAyeal*, 1993; *Marshall and Clarke*, 1997b] have found that basal temperature evolution in ice sheets strongly influences glacier flow dynamics through thermally-enabled instabilities. Here, we use a three-dimensional model of ice sheet thermomechanics [*Marshall and Clarke*, 1997a; *Marshall et al.*, 2000, 2002] to examine the evolution of basal temperature of the North American ice sheets during the last glacial cycle. Although such glaciological models do not yet properly represent essential fast flow processes, we assess whether the timescale of thermal evolution is comparable to that of the 100-kyr cycle, and thus whether thermal enabling of basal flow may be an important feedback in explaining terminations [*Clark et al.*, 1999].

2. Ice Sheet Model

[4] The climate forcing for the glacial cycle simulation [*Marshall et al.*, 2000] is based on LGM and present-day climate simulations from the Canadian Centre for Climate Modelling and Analysis (CCCma) general circulation model [*Vettoretti et al.*, 2000]. We employ a weighted linear combination of these end-member climate fields, with a time-variable weighting based on the GRIP ice core $\delta^{18}O$ record from Summit, Greenland (Figure 1a) [*Dansgaard et al.*, 1993]. The prescribed air temperature history also includes elevation lapse rate effects caused by the ice sheets. Climate fields are updated every 100 years and the glacial cycle simulations run from 120 kyr BP to present, starting with no initial ice and modern-day topography.

[5] Three-dimensional ice dynamics and thermodynamics are simulated in a finite difference model in the style of *Huybrechts* [1990], as detailed in *Marshall and Clarke* [1997a]. The North American grid in these tests has a meridional resolution of 0.5°, a zonal resolution of 1°, 20-level vertical grid. Isostatic effects are modeled based on a spherically symmetric visco-elastic Earth model [*James and Ivins*, 1998], with the bed response updated every 500 years. Viscosity-depth structure and lithosphere thickness are representative of the North American continental interior [*Marshall et al.*, 2002].

[6] Ice temperature T is solved from the 3D energy balance for the advection-diffusion system:

$$\frac{\partial T}{\partial t} = -v_k \frac{\partial T}{\partial x_k} + \kappa(T)\frac{\partial^2 T}{\partial z^2} + \frac{1}{\rho c}\frac{\partial k}{\partial T}\left(\frac{\partial T}{\partial z}\right)^2 + \frac{\Phi}{\rho c}, \quad (1)$$

where v_k is the ice velocity field, Φ describes strain heating due to the internal deformation of ice, and ρ, c, k, and κ are the density, heat capacity, thermal conductivity, and thermal diffusivity of ice. Horizontal temperature diffusion is neglected in ice sheet models because vertical temperature gradients are typically three orders of magnitude higher.

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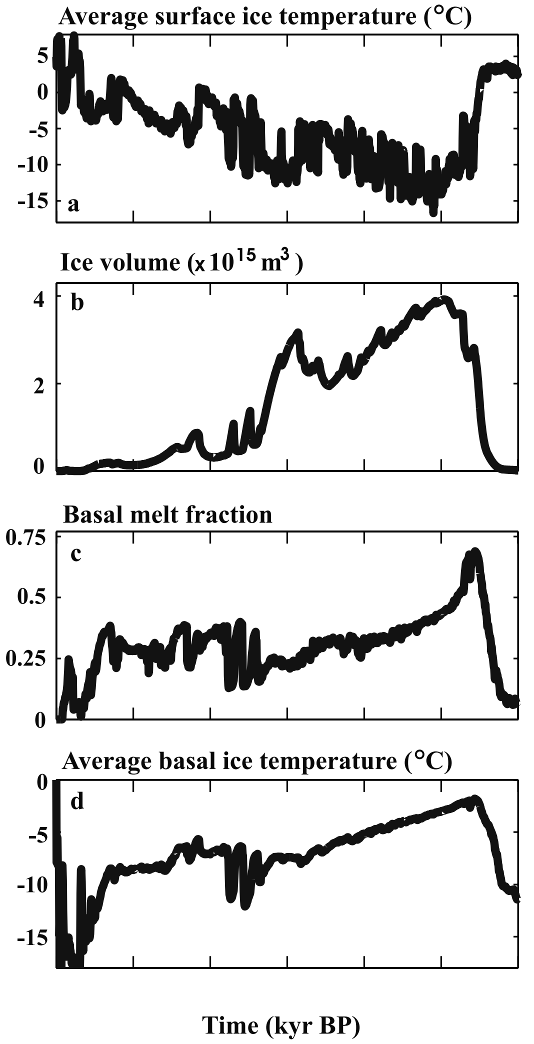

Figure 1. Time series of modeled North American ice sheet evolution through the last glacial cycle, 120 ka to present. (a) Mean air temperature over the region of the North American ice sheets (°C). (b) Ice sheet volume (10^{15} m^3). (c) Mean basal ice temperature (°C). (d) Area of the ice sheet sole at the pressure melting point (%).

[7] Upper surface temperatures in the model are specified based on mean monthly air temperature. We add a source term to account for the latent energy of refreezing where this occurs in ice sheet accumulation areas [e.g., *Reeh*, 1991]. Temperature at the basal boundary, T_b, is freely solved from (1), with geothermal and frictional heat flux from sliding (if applicable) implemented via the Neumann boundary condition $k\partial_z T_b = -(Q_G + v_{bj}\tau_{bj})$, where Q_G is the geothermal heat flux, v_{bj} is basal ice velocity and τ_{bj} is basal shear stress. If basal ice reaches the pressure-melting point, the energy balance on the right-hand-side of (1) is calculated but temperatures are pinned at the pressure melting point and surplus (or deficit) energy is directed to basal meltwater production (or refreezing).

[8] Basal flow v_{bj} occurs in regions where the ice sheet is warm-based and overrides deformable sediments [*Marshall et al.*, 2000]. The locations of basal flow therefore vary spatially and temporally, largely governed by the basal thermal evolution. Where subglacial conditions permit, a simple basal flow parameterization calculates v_{bj} in proportion to gravitational driving stress, scaled to give maximum basal velocities of several 100 m/yr.

3. Results

[9] We evaluate a suite of 33 simulations of the last glacial cycle that result in good agreement between modelled geographic extent of LGM North American ice cover and the ice extent reconstructed from glacial geology [*Dyke and Prest*, 1987]. Results of one such simulation (Figure 1) show that, within 5,000 years of ice-sheet inception, the warm-based area of the ice sheets reaches ~35%, where it then remains (with some variability) for ~70 kyr. Subsequently, the percentage area of warm-based ice begins to increase exponentially, reaching 70% by 12 kyr BP. This temperature evolution is largely governed by the timescale of thermal diffusion in ice. New ice that moves into an area (via ice advance or *in situ* accumulation) generally begins as a thin, cold-based cover. As the ice thickens with time, insulation from cold surface temperatures increases and the base is gradually warmed through a combination of geothermal and deformational (strain) heating.

[10] Horizontal and vertical advection are also significant in the 3D energy balance (1), of the same order of magnitude as the vertical diffusion term. Thermal advection therefore offers the main complication to generalized predictions of basal temperature zonation. Cold surface ice is advected downward, promoting cold-based conditions in the ice sheet interior where vertical "downwelling" is highest. In the model, this advective cooling creates perpetual cold-based conditions in ice divide regions in northern Quebec, Keewatin, the Foxe Basin, the high Arctic, and Hudson Bay (Figure 2).

[11] In addition to this vertical advection structure, cold advection "plumes" are typically found downstream of the ice divide in ice sheets and in polythermal glaciers, particularly in active systems. The strain heating term, Φ, is also of importance in regions of steep and active ice, acting as an internal heat source that is concentrated at the bed. For standard n = 3 power-law rheologies used to describe ice deformation [*Paterson*, 1994], strain heating increases in proportion to shear stress to the 4th power. This heat source plays an important role in generating warm-based conditions in the ice sheet periphery and in topographic channels (e.g., Hudson Strait).

[12] Uncertainties in surface climate forcing offer a potential source of error to basal temperature reconstructions. On timescales of 10^3 to 10^4 yr, the strength of vertical advection in interior regions of the ice sheet is proportional to the temporally averaged snow accumulation rate. Modeled basal temperatures are therefore sensitive to paleoprecipitation rates, which are poorly constrained during the glacial period in North America. Air temperatures also influence basal temperature, although the ice sheet acts as a low-pass filter in propagating surface temperatures downward and downstream. Advective and diffusive timescales are of order 10^4 yr, sufficient time for high-frequency surface temperature variations to diffuse (Figures 1a and 1c), suggesting that only the average surface temperature on these timescales influences basal temperature evolution.

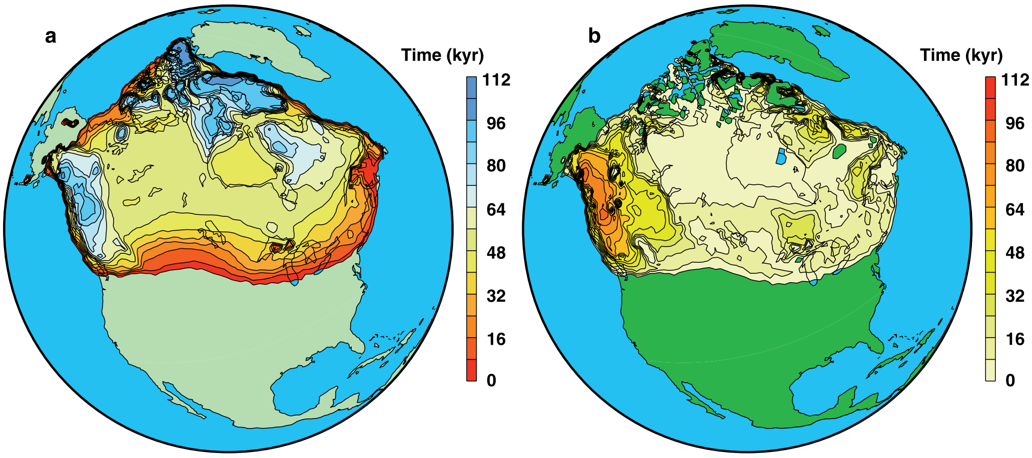

Figure 2. Diagnostics from the full (120-kyr) glacial cycle. (a) Occupation time of ice (kyr). (b) Duration of warm-based ice-cover (kyr). Note that the background colour – green for land and blue for water – appears in (b) if the ice did not reach the melting point in the model at any time during the last glacial cycle (0 kyr warm-based duration).

[13] Basal temperature trends also lag surface temperature influences by this timescale. Increased warm-based conditions during deglaciation (Figure 1d) are thus not a result of climatic warming. Rather, these are attributable to: (i) increased diffusive heating at the base, due to thickening ice in the buildup to LGM, (ii) increased strain heating due to this thickening, and (iii) decreased cooling from vertical advection, due to extremely dry conditions in the ice sheet interior over the mature LGM ice sheets. The buildup to thick, high-elevation ice sheets from 35–20 kyr plays an important role in all three effects.

3.1. Model Sensitivity

[14] Thus far, we have discussed results from a single model realization, with parameter settings that produce geologically reasonable LGM reconstructions of ice area. There are many uncertainties in the subcomponents of the model, however, particularly with respect to the paleoclimatic history. Figure 3 presents a range of basal temperature reconstructions drawn from a suite of climatic sensitivity experiments that produce reasonable LGM ice sheets in North America [*Marshall et al.*, 2002]. Within these results, the area of the ice sheet sole at the melting point ranges from 8–42% at LGM. Despite this variability, trends are similar in all simulations that produced acceptable LGM ice sheets; much of the ice sheet is cold-based at LGM, followed by a dramatic increase in percent melt fraction during late glacial stages.

[15] Latent heat transfer associated with routing and storage of basal meltwater is a potentially important term in the thermal balance that is not considered here [*Parizek et al.*, in press]. Thermal inertia provided by meltwater supply to a region can help to maintain warm-based ice conditions and basal flow, suppressing the tendency of fast-flowing sectors of an ice sheet to shut down as they thin and advectively cool.

3.2. Discussion

[16] Ice sheet deglaciation involves an amount of energy larger than that provided directly from high-latitude radiation forcing associated with orbital variations. Internal glaciologic, isostatic, and climatic feedbacks are thus essential to explain the deglaciation [*Imbrie et al.*, 1993].

[17] Ice dynamics can be an effective feedback for deglaciation if flow mechanisms support significant and sustained increases in ice flux from central regions of the ice sheet. Increased transport of ice to the ice sheet ablation area would deflate the ice sheet interior and accelerate total ablation by availing of the excess solar energy available for snow and ice melt at the low elevations and latitudes of the southern ice sheet limits. Calving on the marine and lacustrine margins also provides an essentially unlimited sink.

[18] Thermally enabled basal flow represents one potential feedback that could cause an increase in ice flux relative to frozen-bed conditions [*MacAyeal*, 1992, 1993; *Marshall and Clarke*, 1997b]. We find that the timescale of bed temperature evolution modeled here is directly comparable to that of the 100-kyr glacial cycle, indicating that thermal enabling of basal flow may have been an important feedback in causing terminations. In particular, we note that the long growth phase of ice volume to the LGM was associated with widespread frozen-bed conditions; a substantial

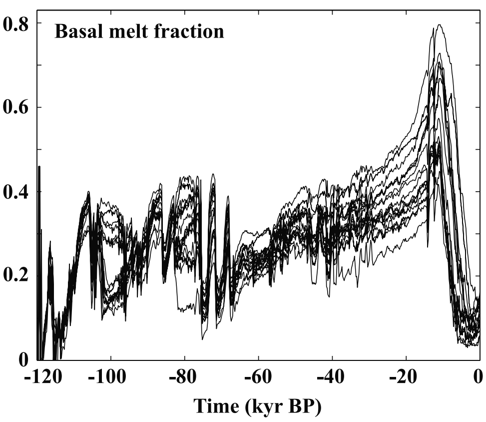

Figure 3. Range of modeled basal temperature evolution in the last glacial cycle. This suite of simulations is representative of the span of climate and basal-sliding parameterizations that produced reasonable LGM ice sheet area; tests that over- or under-predicted LGM southern ice extent are excluded.

increase in basal melt fraction only occurred late in the glacial cycle, with the most rapid rate of increase occurring after the LGM.

[19] While we cannot make firm conclusions without a rigorous model of basal flow processes, these results clearly suggest the potential for increased dynamical activity of the North American ice sheets in response to this thermal evolution. Available supporting evidence of enhanced ice dynamics is suggested by the thin, low-sloping, and mobile peripheral areas of the Laurentide Ice Sheet (LIS) during the deglaciation [*Clark*, 1994], and geomorphic records indicating widespread basal thawing of the LIS interior following the LGM [*Kleman and Hättestrand*, 1999].

[20] Our results indicate that the advective and diffusive timescales (of order 10^4 yr) are comparable to climate forcing arising from precession (23 kyr) and obliquity (41 kyr); the signal of higher frequency forcing will diffuse. Because surface boundary conditions affect the basal temperature through diffusive and advective heat transfer, a change in the amplitude of surface forcing will cause a change in the timing of significant changes in the thermal regime. The way in which surface climate forcing is communicated to the bed also depends on the ice sheet dynamical history. We thus expect that the exact timing of thermal evolution will exhibit some variability from one glacial cycle to the next in response to variability in the amplitude of orbital forcing. This may explain the variable time interval between terminations [*Pisias et al.*, 1991; *Raymo*, 1997]. Moreover, our results suggest that thermal enabling of basal flow does not occur in response to surface warming, which may explain why the timing of the Termination II occurred earlier than predicted by orbital forcing [*Gallup et al.*, 2002].

[21] Results suggest that basal temperature evolution plays an important role in setting the stage for glacial termination. To confirm this hypothesis, model studies need improved basal process physics to incorporate the glaciological mechanisms associated with ice sheet instability (surging, streaming flow).

4. Conclusions

[22] Our simulations suggest that a substantial fraction (60% to 80%) of the ice sheet was frozen to the bed for the first 75 kyr of the glacial cycle, thus strongly limiting basal flow. Subsequent doubling of the area of warm-based ice in response to ice sheet thickening and expansion and to the reduction in downward advection of cold ice may have enabled broad increases in geologically- and hydrologically-mediated fast ice flow during the last deglaciation. Increased dynamical activity of the ice sheet would lead to net thinning of the ice sheet interior and the transport of large amounts of ice into regions of intense ablation both south of the ice sheet and at the marine margins (via calving). This has the potential to provide a strong positive feedback on deglaciation.

[23] The timescale of basal temperature evolution is of the same order as the 100-kyr glacial cycle, suggesting that the establishment of warm-based ice over a large enough area of the ice sheet bed may have influenced the timing of deglaciation. Our results thus reinforce the notion that at a mature point in their life cycle, 100-kyr ice sheets become independent of orbital forcing and affect their own demise through internal feedbacks.

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