Sub-catchment melt and long-term stability of ice stream D, West Antarctica

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[1] The apparent long-term persistence and short-term variability of the Siple Coast ice streams of West Antarctica are tied to regional thermal conditions and local basal lubrication. Numerical simulations indicate that the flux of latent heat in a throughgoing hydrologic system fed by melt beneath thick inland ice maintains the lubrication of fast-moving modern and ice-age ice streams despite their tendency to freeze to the bed, and would allow additional thinning and grounding-line retreat. *INDEX TERMS:* 1863 Hydrology: Snow and ice (1827); 3210 Mathematical Geophysics: Modeling; 9310 Information Related to Geographic Region: Antarctica; 1827 Hydrology: Glaciology (1863)

1. Introduction

[2] Changes in the West Antarctic ice sheet could affect the local environment and global sea level. Attention is focused on the fast-moving ice streams that discharge most of the ice. Those ice streams crossing the Siple Coast have exhibited startling unsteadiness at least since the LGM (last glacial maximum) [*Alley and Bindschadler*, 2001]. Yet, despite major ice-age advance of grounded (non-floating) ice to near the continental-shelf edge that approximately doubled the extent of the grounded ice and was followed by similar retreat, these ice streams maintained roughly their same locations and their anomalously low-elevation, low-slope surface profiles [*Conway et al.*, 1999; *Anderson and Shipp*, 2001].

[3] Inland ice flows meters to tens of meters per year by internal deformation or slow basal motion. Flow accelerates to hundreds of meters per year in the modern ice streams, which slide over or deform subglacial till softened by high-pressure water [*Blankenship et al.*, 1986; *Alley*, 1989; *Engelhardt and Kamb*, 1998; *Tulaczyk et al.*, 2000a; *Kamb*, 2001]. The thinning associated with this acceleration moves cold surface ice downward, steepening the basal temperature gradient and tending to freeze the bed and slow the rapid motion [*Hulbe and MacAyeal*, 1999]. Ice-stream models assuming only local basal heat and water balance, with freezing initially supplied by water extraction that strengthens subglacial sediments, reproduce the marked unsteadiness of the ice streams [*Tulaczyk et al.*, 2000b]; this suggests that the ice streams live balanced on the edge between freezing and thawing, perhaps in a self-organized-critical state [*Clarke and Marshall*, 1998].

[4] However, data indicate that melt beneath thicker inland ice feeds a throughgoing sub-ice-stream water system [*Engelhardt and Kamb*, 1997, 1998; *Kamb*, 2001] comprising widespread, thin, interconnected, ice-contact braided channels in which water pressure increases with water flux [*Walder and Fowler*, 1994; *Catania and Paola*, 2001]. A > 10 m layer of debris-rich ice [*Engelhardt*, 2001] frozen on the base of ice stream C is too thick to have grown since the shutdown of that ice stream ~150 years ago [*Retzlaff and Bentley*, 1993; *Vogel and Tulaczyk*, 2001], or to have been extracted from the still-soft hence not-dewatered sediments under the ice stream [*Kamb*, 2001]; therefore, freeze-on of the throughgoing water occurred for some time without stopping the ice stream. Drilling records [*Engelhardt*, 2001] provide evidence of similar frozen-on layers beneath fast-moving ice streams D and B (the latter now called Whillans Ice Stream). Ice from diverse settings exhibits similar freeze-on without stopping (e.g. [*Lawson et al.*, 1998]).

[5] Because of the orders-of-magnitude mismatch between the subannual to few-annual time scale for adjustment of the basal water system and the centennial or longer time scale for ice-stream thinning and basal conductive cooling, a throughgoing water system may allow strong unsteadiness, especially as the water system is subject to piracy or other diversion events [*Alley et al.*, 1994]. We thus hypothesize that the long-term persistence ($\mathcal{O}$(10,000) yr) and short-term variability ($\mathcal{O}$(100) yr) of the Siple Coast ice streams are both linked to the throughgoing hydrological system, which provides sufficient latent heat to keep the ice streams from the critical edge of freezing to their beds. We test this hypothesis through a suite of model experiments for ice stream D.

2. Model Description

[6] The PSU/UofC flowline model [*Parizek*, 2000; *Parizek et al.*, in review] is sufficiently fast to allow many integrations using various parameterizations over glacial-interglacial cycles. The 2-D (vertical and along flow) finite-element model is based on *MacAyeal* [1996]. Ice dynamics uses a diffusion formulation of the thin-ice approximation [*Hutter*, 1983] in which only the horizontal shear stress drives ice deformation, set proportional to the third power of stress [*Budd and Jacks*, 1989]. Sliding is parameterized as linear in basal shear stress with an adjustable prefactor set higher for ice streams and lower for inland ice to simulate the effects of changing geology. The thermal subroutine includes horizontal and vertical advection, vertical diffusion, and viscous dissipation in ice, and vertical diffusion in bedrock with a geothermal-flux boundary condition 1120 m below the ice. Isostatic depression is calculated with an elastic lithosphere and relaxed asthenosphere [*Le Meur and Huybrechts*, 1996].

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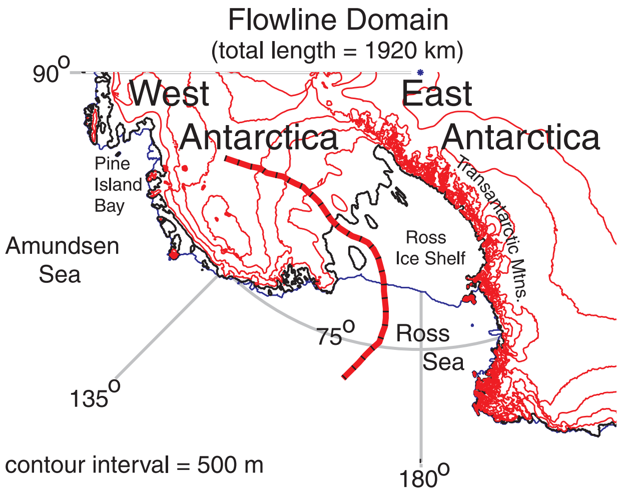

Figure 1. Flowline-model domain, from divide down ice stream D and across Ross embayment to the continental-shelf edge, from BEDMAP data set (sponsored by the Scientific Committee on Antarctic Research and coordinated by the British Antarctic Survey, Cambridge, England).

[7] Air temperatures are initialized with modern values, and adjusted for elevation change (using the annual-average adiabatic lapse rate for West Antarctica of 5.1°K/km [*Fortuin and Oerlemans*, 1990]) and climate change (using the 10–90 ka Byrd Station ice-isotopic record, dated by *Blunier and Brook* [2001], converted to temperature change at 0.6 per mil/°C, extended across the glacial cycle using Byrd's modern −28°C from 0–10 ka and a linear trend to this value from 90–130 ka as displayed in Figure 3a). Sea-level changes affecting elevation and lithospheric loading are taken from SPECMAP [*Imbrie et al.*, 1984]. The accumulation-rate parameterization assumes a 55% decrease for a 10°C cooling, initialized from modern values [*Huybrechts*, 1991].

3. Ice-Stream-D Simulations

[8] Eight simulations considered here follow a flow path down ice stream D (Figure 1), and begin with a thermal spinup from 150–130 ka under modern conditions and initially linear temperature gradients. Because the model does not treat ice-shelf dynamics, currently floating ice is artificially grounded on a very slippery bed (see Figure 2a). Sensitivity experiments tested reasonable ranges of the geothermal flux, basal sliding conditions, and accumulation rate on a domain divided into three distinct basal regions: "sheet" (0–360 km), "stream" (360–760 km), and "shelf" (extended region (760–1580 km) of enhanced sliding). Six runs permit sliding in the "stream" and "shelf" zones regardless of the thermal conditions at the bed. This ensures a thin, flat profile over these regions in agreement with geologic evidence [*Conway et al.*, 1999; *Anderson and Shipp*, 2001], thereby encouraging modeled basal freeze-on. In these runs, sliding

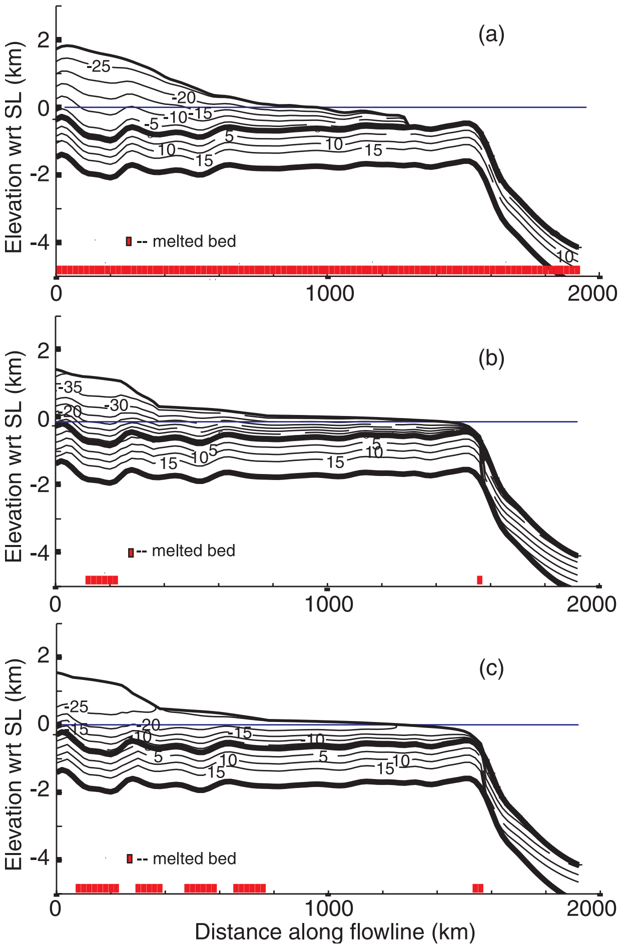

Figure 2. (a) Initial conditions including isotherms along flowline through ice stream D. (b) LGM (22.1 ka); and (c) modern ice-sheet reconstructions with basal melting, from sixth simulation, Table 1.

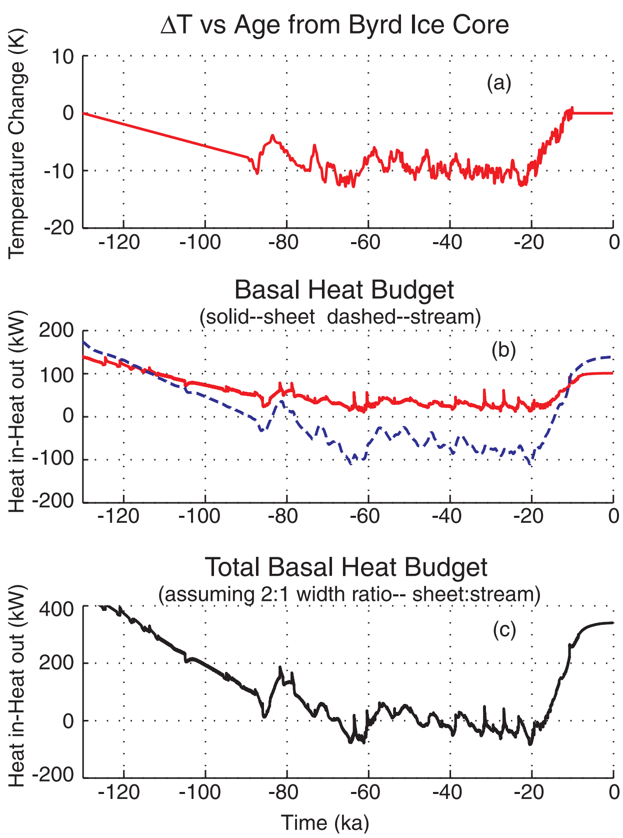

Figure 3. (a) 130-kyr temperature forcing. (b)–(c) History of basal heat budget, for second simulation, Table 1: (b) sums over "sheet" and "stream" regions; (c) area-weighted sum of the curves in (b), assuming catchment sheet is twice as wide as stream.

Table 1. Thermal-Budget Simulations

Simulation	Heat Budget (kW)		
	Sheet	Stream	Total
C1(T) = 0.5e−3	105	140	350 Modern
C2 = 5e−3	25	−95	−45 LGM
C3 = 50e−3			
Γ = 71.1			
110% Adot	100	140	340 Modern
	35	−80	−10 LGM
90% Adot	105	135	345 Modern
	25	−105	−55 LGM
Γ = 60	50	70	170 Modern
	15	−185	−155 LGM
Γ = 100	305	370	980 Modern
	165	125	455 LGM
C2 = 10e−3	110	30	250 Modern
C3 = 100e−3	15	−270	−240 LGM
C1(T, heat) = 0.5e−3	125	155	405 Modern
C2(T, heat) = 5e−3	30	0	60 LGM
C3(T, heat) = 50e−3			
C1(T, heat) = 0.5e−3	140	55	335 Modern
C2(T, heat) = 10e−3	0	−30	−30 LGM
C3(T, heat) = 100e−3			

Details of eight simulations. C1, C2, and C3 are sheet, stream and shelf sliding factors (m/yr/Pa), applied either: always [C]; when thawed locally [C(T)]; or, when thawed including upstream water [C(T, heat)]. Γ is geothermal flux (mW/m^2), and Adot accumulation rate. Budgets are calculated over the modern sheet and stream domains. Total balances assume 2:1 width ratio for sheet:stream.

in the "sheet" domain occurs only when the local basal temperature is at the pressure melting point.

[9] In the other two runs, local sliding occurs if and only if the local basal heat balance reaches the pressure melting point, or, if melting upglacier stores enough latent heat to balance any thermal deficit at the point of interest. Assuming that a throughgoing hydrologic system links the thermal states of inland and ice-stream beds on timescales much shorter than the timestep sizes for either the dynamic or the longer thermodynamic routines, this excess heat is *instantly* available to downstream locations until consumed by freezing. For calculation of available basal water only, the sheet region is assumed to be twice as wide as the stream and shelf based on modern observations [*Joughin et al.*, 1999].

4. Results

[10] Simulated basal melting is concentrated beneath the ice-stream catchment where thick ice separates the cold surface from the bed, and in bedrock overdeepenings where vertical stretching reduces downward motion of cold ice, as illustrated in Figures 2b and 2c. Reduced accumulation acts to warm the bed during the ice age, but the colder temperatures (Figure 3a) and decreased frictional heating from reduced mass outflux cool the bed more.

[11] As shown in Table 1, none of the sensitivity tests approaches freeze-on for modern ice streams, and all produce enough meltwater to allow grounding of the modern ice shelf without freezing to the bed. However, all simulations are close to freeze-on during the LGM at the modern grounding line and would have had difficulty maintaining streaming flow to the observed LGM grounding line. Because the 2-D model lacks ice-flow convergence from catchment to stream, simulated inland ice is thinner than observed (cf. modern profiles in Figures 2a and 2c), causing the model to underestimate basal melting there and perhaps explaining why LGM ice streams persisted whereas the modeled ones are close to freezing to the bed (e.g. Figures 3b and 3c). Furthermore, five of the eight experiments predict that sufficient lubrication for thawed basal conditions out to the LGM grounding line begins sometime between 10–13 ka, in close agreement with the onset of deglaciation [*Conway et al.*, 1999].

5. Discussion and Conclusions

[12] Numerical results suggest that widespread basal melting beneath the ice-stream catchments and in bedrock overdeepenings maintains the lubrication and existence of both modern and glacial ice streams. The local heat balance tends to freeze ice streams to their beds, but sufficient melt occurs inland and in overdeepenings to buffer this tendency (esp. now vs. LGM). Lower LGM accumulation rates, plus field data suggesting similar LGM ice-stream geometry to today [*Conway et al.*, 1999; *Anderson and Shipp*, 2001], indicate slower LGM ice-stream flow, consistent with modeled LGM reduction in production of lubricating basal meltwater. A coupled stream-shelf model could be used to test whether the deglacial increase in lubricating water contributed to minor thinning near the grounding line, triggering deglaciation. Because modern production of basal meltwater inland exceeds consumption by freezing to ice streams in model and data [*Kamb*, 2001], the basal thermal budget does not prevent further thinning and grounding-line retreat. However, because local thermal deficits promote basal freeze-on (esp. on topographic highs), observed short-term variability is likely to persist, arising from the presence/absence of local lubrication tied to the pattern of the hydrologic system.

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