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Causes of sudden, short-term changes in ice-stream surface elevation

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[1] Recent satellite-borne observations of Antarctica's ice streams show sudden, spatially confined surface-elevation changes that are interpreted as caused by subglacial water movement. Using a numerical model of idealized ice-stream flow coupled to various simple treatments of subglacial bed conditions, we demonstrate that ice-stream flow dynamics significantly modulates the surface-elevation expression of processes taking place at the ice-stream bed. This modulation means that observed surface-elevation changes do not directly translate to basal-elevation changes, *e.g.* inflation or deflation of subglacial water pockets, of equal magnitude and shape. Thus, subglacial water volume change is not directly proportional to the area integral of surface-elevation changes. Model results show that ambiguities in interpretation of surface elevation changes can be overcome with additional measurements, such as of surface velocity change, and through development of methodology designed to understand transfer of basal change to surface change. **Citation:** Sergienko, O. V., D. R. MacAyeal, and R. A. Bindschadler (2007), Causes of sudden, short-term changes in ice-stream surface elevation, *Geophys. Res. Lett.*, *34*, L22503, doi:10.1029/2007GL031775.

1. Introduction

[2] Recent discoveries of sudden, meter-scale changes in surface elevation over spatially compact areas of Antarctica's ice streams made possible by various satellite-borne instruments suggest the presence of previously unknown sub-ice-stream lakes capable of rapid volume changes [*Gray et al.*, 2005; *Fricker et al.*, 2007]. This suggestion motivates the present study which examines how changes in basal conditions associated with sub-ice-stream lake development and discharge may influence surface elevation and velocity of the ice stream. As demonstrated in previous work [e.g., *Gudmundsson*, 2003; *Raymond and Gudmundsson*, 2005], the transmission of basal variability to the surface is non-linear and complex. The patterns of surface change seen in SAR interferometry or ICESat surface altimetry [*Gray et al.*, 2005; *Fricker et al.*, 2007] are thus not necessarily translatable to simple changes in sub-ice-stream lake extent and volume without consideration of how this translation is also affected by ice-stream dynamics.

[3] The well-known stress balances of ice-stream flow [*Van der Veen*, 1987; *Whillans and Van der Veen*, 1997] prescribe how basal resistance, $\vec{\tau}$, and surface elevation, S, are related via the gravitational driving stress. For example, where basal resistance is reduced, faster ice flow and mass transport cause the flow to reduce ice thickness, thereby reducing driving stress toward a new balance. Accumulation of subglacial water is a well known means to alter basal resistance. Accumulation and discharge of subglacial lakes also adds another complexity: the vertical movement of the lake "roof". When considering the causes of surface-elevation changes revealed by recent observations, it is thus reasonable to expect that changing basal resistance and lake roof elevation will combine to produce superimposed effects on the ice-stream surface elevation.

[4] To aid in the interpretation of recent ice-stream surface elevation changes, we study the effects of three phenomena that may influence ice streams as a result of subglacial water movement: (1) lowering of the ice-stream base in association with lake roof deflation, and (2) decrease and (3) increase of basal resistance independently of lake-volume changes. We use a time-dependant model of ice-stream flow and mass balance to examine these three phenomena in a simple, idealized ice-stream-flow geometry.

2. Model Description

[5] Our analysis is based on a finite-element model (finite-element mesh used in this study is shown in Figure S1 of the auxiliary materials)[1] of two-dimensional, vertically integrated ice-stream flow set in an idealized, rectangular domain Γ in the horizontal x, y plane. The domain dimensions are 250 km along flow and 100 km across flow, and the bed of the ice stream is inclined along the long axis of the rectangular domain, with a slope of 10^{-3} (Figure 1). To represent a compact region over which changes in basal conditions will be modeled, a 10-km diameter circular subdomain, Γ_c, is introduced at a centered location 100 km from the inflow boundary (x = 0 km) and 50 km from the side boundaries (y = 0, 100 km).

[6] The variables which the model determines include the two horizontal velocity components, $u(x, y, t)$ and $v(x, y, t)$ in the x and y directions respectively, and the ice thickness and surface elevation $H(x, y, t)$ and $S(x, y, t) = H(x, y, t) + B(x, y, t)$, respectively. Following common practice [e.g., *MacAyeal*, 1989], the horizontal velocities are assumed to be independent of the vertical coordinate, and the stress-balance is assumed to be quasistatic, and thus independent of time, t. The ice is also assumed to be incompressible and to obey Glen's flow law, described in the present study by a strain-rate dependent effective ice viscosity. The governing

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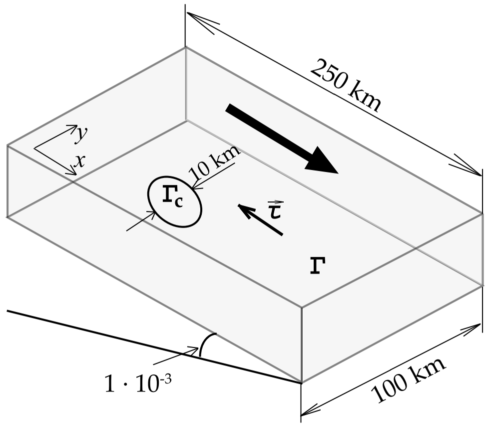

Figure 1. Idealized ice-stream geometry. Flow is directed along the inclination of the basal plane. Subdomain $\Gamma_c \in \Gamma$, represents the location of basal condition perturbations associated with subglacial lake drainage or changes in basal resistance.

stress-balance equations used to solve for u and v as a function of $H(x, y, t)$ and $S(x, y, t)$ are:

$$\frac{\partial}{\partial x}\left[2\nu H\left(2\frac{\partial u}{\partial x}+\frac{\partial v}{\partial y}\right)\right]+\frac{\partial}{\partial y}\left[\nu H\left(\frac{\partial u}{\partial y}+\frac{\partial v}{\partial x}\right)\right]=\rho g H\frac{\partial S}{\partial x}-\tau_u, \quad (1)$$

$$\frac{\partial}{\partial x}\left[\nu H\left(\frac{\partial u}{\partial y}+\frac{\partial v}{\partial x}\right)\right]+\frac{\partial}{\partial y}\left[2\nu H\left(\frac{\partial u}{\partial x}+2\frac{\partial v}{\partial y}\right)\right]=\rho g H\frac{\partial S}{\partial y}-\tau_v, \quad (2)$$

where $\rho = 910$ kg m^{-3} is ice density, $g = 9.81$ m s^{-2} is the acceleration due to gravity, ν is the effective, strain-rate dependent ice viscosity representing Glen's flow law given by

$$\nu = \frac{D}{2\left[\left(\frac{\partial u}{\partial x}\right)^2+\left(\frac{\partial v}{\partial y}\right)^2+\frac{1}{4}\left(\frac{\partial u}{\partial y}+\frac{\partial v}{\partial x}\right)^2+\frac{\partial u}{\partial x}\frac{\partial v}{\partial y}\right]^{\frac{n-1}{2n}}}, \quad (3)$$

where $D = 1.68 \cdot 10^8$ Pa s$^{1/3}$ is a vertically-averaged ice stiffness parameter, $n = 3$ is the power-law flow exponent, and τ_u and τ_v are x and y components of the basal resistance, defined by

$$\begin{aligned}\tau_u &= -T\frac{u}{\sqrt{u^2+v^2}},\\ \tau_v &= -T\frac{v}{\sqrt{u^2+v^2}},\end{aligned} \quad (4)$$

and where T is a basal-resistance constant. Except within the subdomain Γ_c, T is specified to be 10 kPa uniformly throughout the domain Γ, a value that roughly reproduces characteristic basal shear stress under fast moving ice streams in West Antarctica [*Joughin et al.*, 2004]. Equations (3) and (4) express basal resistance as plastic basal reology. Experiments with viscous basal rheology produce results similar to ones presented here.

[7] The governing mass-balance equation is

$$\frac{\partial H}{\partial t}+\vec{\nabla}\cdot(\vec{\mathbf{v}}H)=\dot{A}+\dot{B}, \quad (5)$$

where $\vec{\nabla}$ is the two-dimensional divergence operator. In the present study we assume no net ablation/accumulation at the surface and melting/refreezing at the base, thus the right hand side of equation (5) is zero in all experiments.

[8] Boundary conditions on horizontal borders of Γ are specified to introduce a channel-like flow that is simple and representative of typical ice-stream conditions. At the two side boundaries, y = 0, 100 km (see Figure 1), u and v are set to 0. At the upstream and downstream boundaries, no-jump conditions are specified for the vertically integrated forces in the x and y directions. The mass-balance boundary conditions are specified as follows. The ice thickness at the upstream boundary is constant $H(x = 0, y) = 1400$ m, mass flux at the two side boundaries at y = 0, 100 km is zero, and at the downstream, outflow boundary mass flux has no jump.

3.4. Conclusions

[22] Surface elevation changes observed in all experiments demonstrate the importance of ice-stream dynamics in defining the complexity of ice stream response to changing basal conditions. Three major conclusions can be drawn from this study. First, surface elevation changes could be caused by variations in basal traction as well as by changes in sub-ice-stream lake volume. Second, ice surface response to any of such changes is complex and does not directly inform an observer about either the nature or magnitude of those changes. Third, simultaneous measurement of surface velocity would help to distinguish between surface elevation changes due to basal traction effects and those due to subglacial lake volume changes.

[23] Cross-sections of surface elevation changes obtained from the model experiments are designed to mimic the way ice-stream-surface elevation has been observed in satellite data. These cross-sections show that observed surface change [e.g., *Gray et al.*, 2005; *Fricker et al.*, 2007] is not a direct measure of the changing elevation of sub-ice-stream lake roof elevation. It is thus possible to misinterpret, for example, an observation of $\Delta S < 0$ as signifying a reduction in lake volume, when in reality the observation may indicate a change (of either sign) of basal resistance.

[24] One possible means of differentiating between lake-drainage events and events associated with changing basal resistance is to simultaneously observe ice velocity changes. There are clear differences in the spatial pattern of velocity change in response to these two kinds of basal forcing. In the case of a lake volume change, there is a dipole structure of velocity change over the lake. In the case of the basal resistance change, the velocity change is of one sign and is distributed over an area that is significantly larger than the area of basal change. Another distinctive feature is magnitude of velocity changes. In the case of lowering ice base it is small (~0.9% of initial velocity). In the case of the variations in basal resistance it is much larger (~15%) and would be easily detected in repeated velocity measurements.