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

O. V. Sergienko,[1] D. R. MacAyeal,[2] and R. A. Bindschadler[3]

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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

[1]Oak Ridge Associated Universities at NASA Goddard Space Flight Center, Greenbelt, Maryland, USA.
[2]Department of the Geophysical Sciences, University of Chicago, Chicago, Illinois, USA.
[3]NASA Goddard Space Flight Center, Greenbelt, Maryland, USA.

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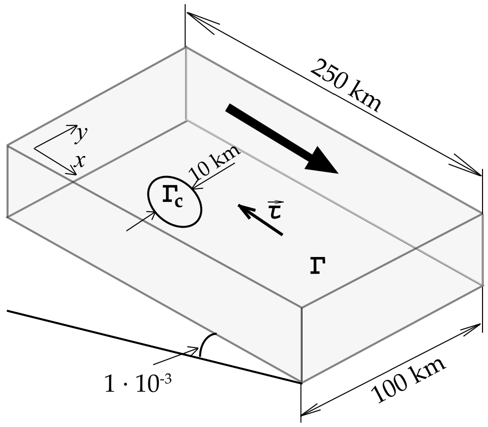

Figure 1. Idealized ice-stream geometry. Flow is directed along the inclination of the basal plane. Subdomain $\Gamma_c \in \Gamma$, represents the location of basal condition perturbations associated with subglacial lake drainage or changes in basal resistance.

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

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

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

where ρ = 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

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

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

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

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

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

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.

[9] All model experiments are transient. Their initial conditions are steady-state configurations obtained by joint iterative solution of the stress-balance and mass-balance equations with the $\frac{\partial H}{\partial t}$ term set to zero in equation (5). The full, time-dependant model equations are run for a 10-year period to produce the results of each model experiment. A 10-year period is chosen because this time scale is consistent with the period over which observations are made by the various satellite missions.

[10] Throughout the 10-year period and at its end, ice velocity and surface elevation are compared with their initial, steady-state values. We compare ice-stream surface elevation, $\Delta S(x, y, t > 0) = S(x, y, t) - S(x, y, 0)$, and velocity magnitude, $\Delta V = \sqrt{u(x,y,t)^2+v(x,y,t)^2} - \sqrt{u(x,y,0)^2+v(x,y,0)^2}$.

3. Model Experiments

[11] The goal of the study is to assess ambiguities in the interpretation of ice-stream surface elevation changes in the simplest, most direct manner possible. Three experiments, denoted A, B and C, are designed for this purpose.

3.1. Experiment A: Draining Lake

[12] Experiment A, the "draining lake experiment", aims to simulate surface elevation changes produced by gradual reduction in sub-ice-stream lake volume, represented by the gradual drop-down of the lake roof. In this experiment, the basal resistance parameter T within the circular subdomain Γ_c is maintained at 0 kPa to determine the steady-state initial condition, and kept at 0 kPa for the time-evolution of the ice stream over the 10-year duration of the experiment. The choice of T = 0, both in development of the initial condition and after the lake discharges, allows separation of the effects of lowering ice-stream basal elevation (lowering lake roof) from the effects of changing basal resistance. To simulate the changing volume of the lake, the basal elevation B within Γ_c held at the large-scale inclined value during the calculation of the steady-state initial condition, is gradually reduced during $t > 0$ with a rate 2 m yr^{-1} during first 5 model years and then is kept constant during next 5 model years. To avoid sharp discontinuities, the reduction of

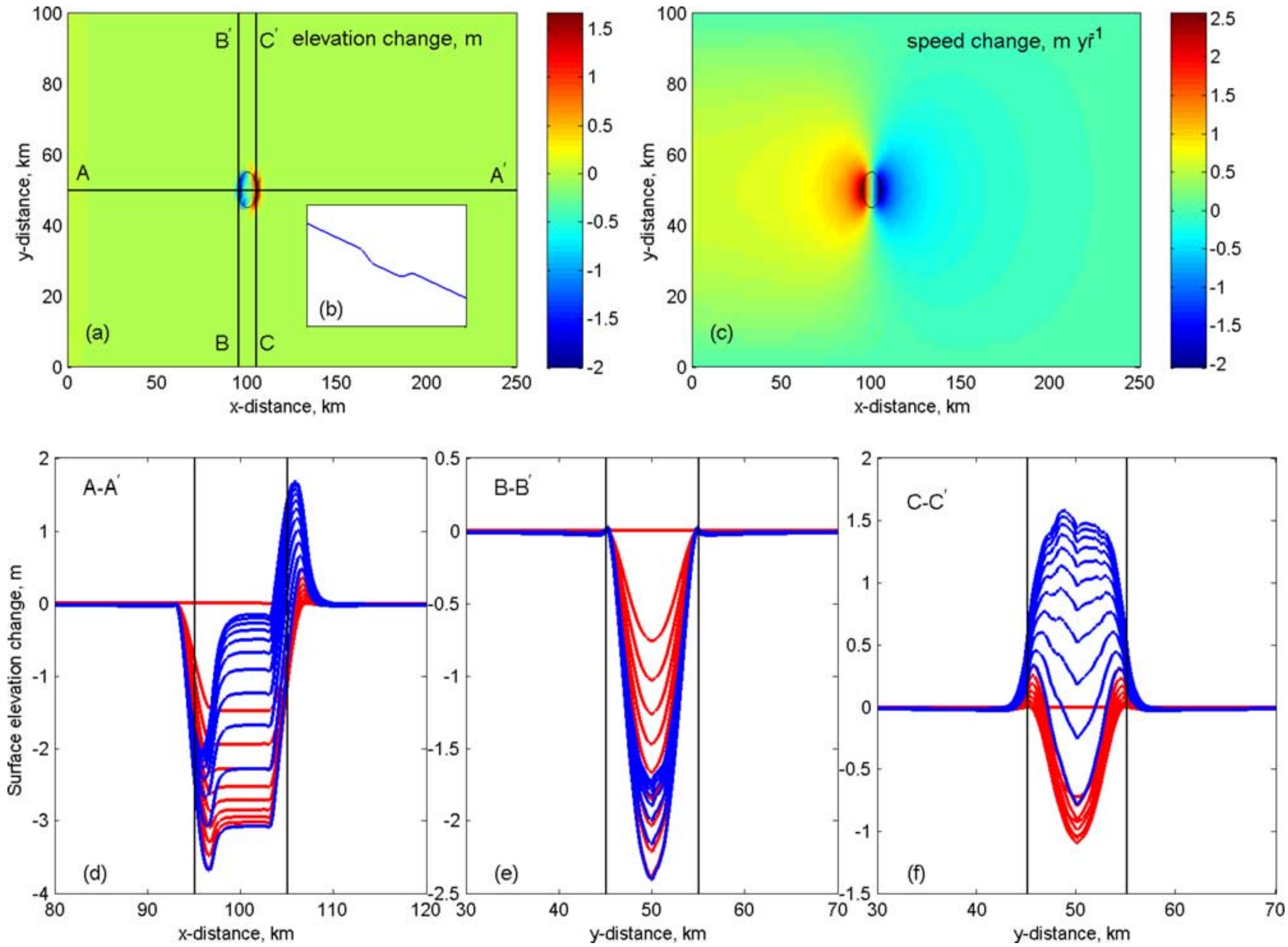

Figure 2. Experiment A: Ice-stream response to lowering ice base in the subdomain Γ_c, (a) ΔS (m) after 10 years; (b) ice base profile at the end of the run; (c) ΔV (m yr^{-1}) after 2 years; (d) ΔS (m) along the A-A′ cross section; (e) ΔS (m) along the B-B′ cross section; (f) ΔS (m) along the C-C′ cross section. Cross-sections in Figures 2d–2f are shown every half of a year, red curves are the first 5 years, blue curves are the second 5 years. Vertical lines outline the extent of Γ_c.

B is a polynomial function of x and y such that the center of Γ_c experiences a drop of 10 m and the edges of Γ_c experiences a drop that is smoothed to 0 at the edge of Γ_c (Figure 2b).

[13] Results of Experiment A, are presented in Figure 2. In this experiment ice speed experiences little variations. Maximum values of ΔV are $\sim$0.9% (3.5 m yr^{-1}) of the initial velocity magnitude. As Figures 2d–2f show, the ice surface mimics the ice base during first years of lake drainage, but with a smaller rate of change. As the experiment proceeds, the surface lowering rate decreases with time. After two years of lake drainage, a dipole-like structure starts to develop, with zones of reduced elevation upstream and increased elevation downstream of the subdomain Γ_c, respectively. This pattern continues to develop over the 10-year duration of the experiment. This structure in surface elevation change develops in response to changes in slope at the upstream and downstream boundaries of Γ_c. At the upstream end, where the ice bed has an additional negative slope due to the initial drop of the lake's roof, ice starts to flow faster due to increased driving stress. As a result of the local increase in mass transport, ice becomes thinner and a depression is developed. An opposite situation occurs at the downstream end: the initial ice bed change has a positive, downstream slope, that reduces local driving stress, makes the ice flow slower and induces an increase in the surface elevation to develop downstream of Γ_c (Figure 2c). The characteristic pattern of ice-velocity changes is a dipole with increased velocity upstream and decreased velocity downstream. After the lake discharge is complete (at t = 5 years), the ice-stream surface reverses its change and starts to relax toward its initial state and eventually reaches it in $\sim$20 years. It is noteworthy, that the maximum drop of surface elevation is 3.8 m, while the roof of the lake drops by 10 m. This difference serves as a reminder that it is impossible to estimate of water-volume loss from area integrals of the surface elevation change without consideration of ice flow effects.

[14] Select cross-sections of ΔS both along and across the direction of ice flow (Figures 2d–2f) are used to simulate air-borne altimetry observations which sample ice-stream surface elevation along tracks. The experimental results show that analysis of only cross-sections B-B′ and C-C′ does not allow an accurate assessment of the spatial pattern of ΔS. This highlights the fact that limited sampling of ΔS patterns in the altimetry observations can yield misleading or inaccurate estimates of sub-ice-stream lake volume changes.

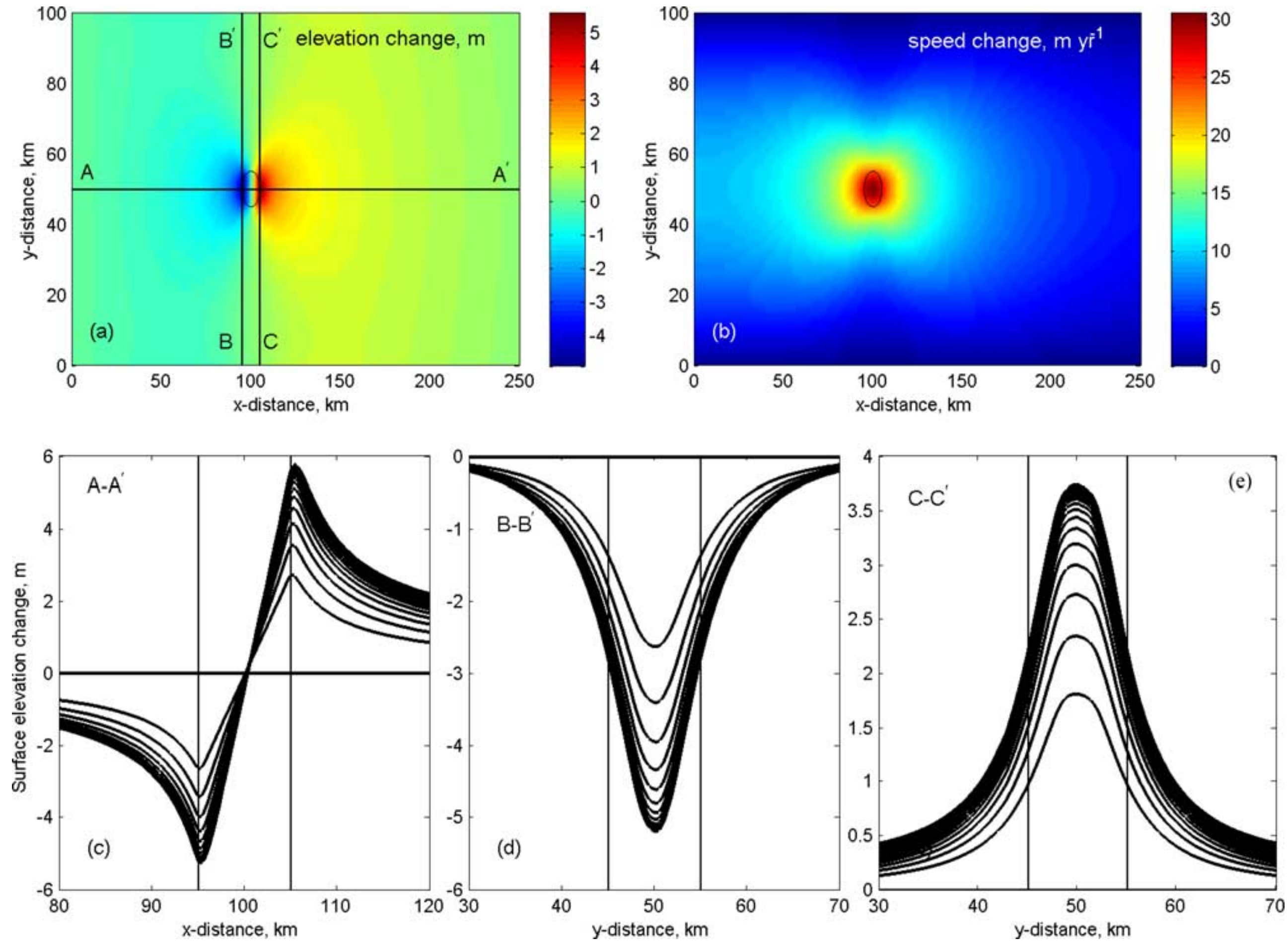

Figure 3. Experiment B: Ice-stream response to a sudden reduction of basal resistance in the subdomain Γ_c, (a) ΔS (m) after 10 years; (b) ΔV (m yr^{-1}) after 2 years; (c) ΔS along the A-A′ cross section in Figure 3a; (d) ΔS along the B-B′ cross section; (e) ΔS along the C-C′ cross section. In Figures 3c–3e cross sections are shown every half year and vertical lines outline the extent of Γ_c.

3.2. Experiment B: Reduction of Basal Resistance

[15] Reduction in basal resistance is simulated by changing the basal resistance parameter T within the subdomain Γ_c from an initial value of 10 kPa at $t = 0$ to 0 kPa for $0 < t \leq 10$ years. The initial condition is the steady-state configuration of the ice stream with the uniform basal resistance parameter $T = 10$ kPa.

[16] Results of this experiment are shown in Figure 3. A dipole with lower surface elevation upstream, and a higher surface elevation downstream of Γ_c develops in response to reduction of the basal resistance within Γ_c. The ice flowing into Γ_c experiences less friction, flows faster (Figure 3b) and increases mass transport, causing thinning and $\Delta S < 0$ on the upstream side of Γ_c. At the downstream side of Γ_c, the situation is the opposite: bed resistance is stronger, the ice flows slower and mass transport is reduced. This results in ice thickening, which produces $\Delta S > 0$.

[17] Figures 3c–3e show surface elevation changes along various lines during the 10-year model simulation. Cross-sections taken along ice flow (Figure 3c) show development of the dipole structure described above. Cross-sections taken across ice flow show development of the surface-elevation deflation (Figure 3d) upstream, and of the surface-elevation inflation (Figure 3e) downstream of Γ_c. Magnitudes of the surface elevation changes strongly depend on a magnitude of the basal resistance reduction. To assess sensitivity of the surface elevation to the magnitude of basal resistance reduction, we have performed a set of experiments with various background basal resistances –30, 10 (present experiment), 1, 0.1 and 0.05 kPa, respectively. The corresponding maxima of surface elevation changes are 12.4, 5.6, 2.3, 0.8 and 0.02 m, respectively.

[18] Increase of the ice velocity magnitude is produced both immediately over the area with reduced basal resistance as well as over a much larger area both upstream and downstream of the subdomain Γ_c (Figure 3b). The maximum ice-flow increase, ΔV, is produced over the subdomain Γ_c, and is more then 50 m yr^{-1} (~15%) of the initial velocity magnitude (340 m yr^{-1}).

3.3. Experiment C: Increase of Basal Resistance

[19] Experiment C simulates a circumstance opposite to Experiment B - a sudden increase in basal traction in a limited area - to emphasize the fact that ΔS of one sign observed in a limited region can be generated by either basal resistance change scenario. Real-world analogs for this simplified simulation can include melt-water refreezing to the ice base thereby hardening the underlying subglacial till.

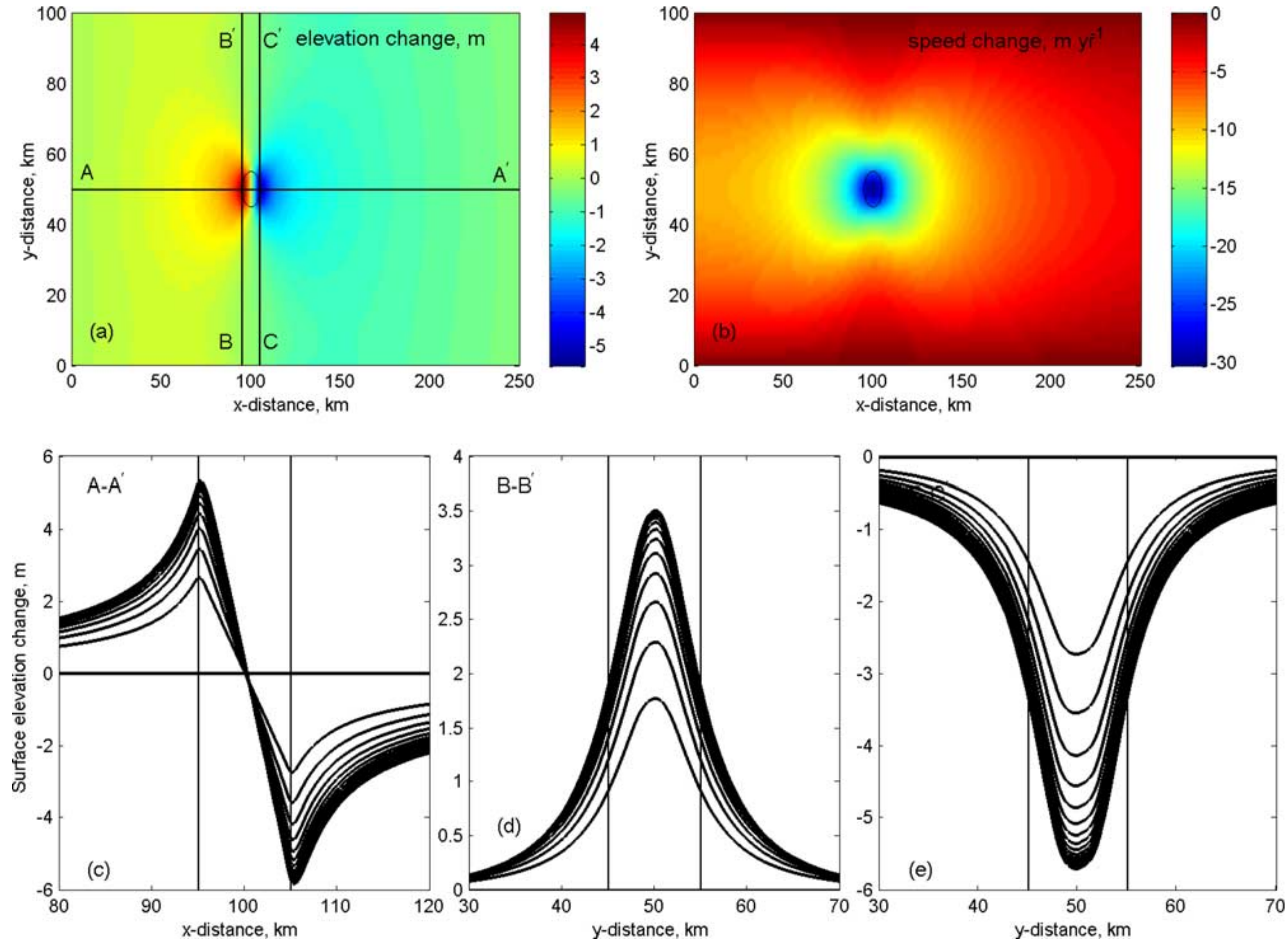

Figure 4. Experiment C: Ice-stream response to doubled basal resistance in the subdomain Γ_c. (a) surface elevation change (m) after 10 years; (b) ΔV (m yr^{-1}) after 2 years; (c) ΔS along the A-A′ cross section in Figure 4; (d) ΔS along the B-B′ cross section; (e) ΔS along the C-C′ cross section. In Figures 4c–4e cross sections are shown every half year and vertical lines outline the extent of Γ_c.

[20] In this experiment, the basal resistance parameter T in the subdomain Γ_c is changed from an initial value of 10 kPa at $t = 0$ to 20 kPa for $t > 0$. As Figures 4a and 4c–4e show, a dipolar structure in ΔS develops in response to such a basal resistance variation. It is similar to that of Experiment B but with opposite polarity: an uplifting zone upstream and lowering zone downstream of Γ_c. As in the experiment with reduced basal resistance, significant change (reduction) in ice velocity magnitude is observed over a large area (Figure 4b), with maximum change associated with the subdomain Γ_c (where ΔV is ~50 m yr^{-1}(15%)).

[21] Results of a "real world" lake drainage experiment - lowering of the lake roof followed by increase of its basal resistance (combination of Experiments A and C) are presented in Figure S2 of the auxiliary material. Surface elevation response to the combined forcing is complex and does not allow for making any conclusions about magnitudes of either the sub-ice-stream lake volume change or basal resistance change.

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.

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R. A. Bindschadler, NASA Goddard Space Flight Center, Greenbelt, MD 20771, USA.

D. R. MacAyeal, Department of the Geophysical Sciences, University of Chicago, Chicago, IL 60637, USA.

O. V. Sergienko, Biospheric and Hydrospheric Research Lab, Oak Ridge Associated Universities at NASA Goddard Space Flight Center, Code 614, Building 33, Room A109, Greenbelt, MD 20771, USA. (olga@neptune.gsfc.nasa.gov)