Century-scale discharge stagnation and reactivation of the Ross ice streams, West Antarctica

C. Hulbe and M. Fahnestock

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[1] Flow features on the surface of the Ross Ice Shelf, West Antarctica, record two episodes of ice stream stagnation and reactivation within the last 1000 years. We document these events using maps of streaklines emerging from individual ice streams made using visible band imagery, together with numerical models of ice shelf flow. Forward model experiments demonstrate that only a limited set of discharge scenarios could have produced the current streakline configuration. According to our analysis, Whillans Ice Stream ceased rapid flow about 850 calendar years ago and restarted about 400 years later and MacAyeal Ice Stream either stopped or slowed significantly between 800 and 700 years ago, restarting about 150 years later. Until now, ice-stream scenarios emphasized runaway retreat or stagnation on millennial timescales. Here we identify a new scenario: century-scale stagnation and reactivation cycles, as well as lateral communication with adjacent ice streams through thickness changes on lightly grounded ice plains. This introduces uncertainty into predictions for future sea-level withdrawals by the West Antarctic Ice Sheet, which are based in part on recent slowing of Whillans Ice Stream and the stagnant condition of Kamb Ice Stream.


1. Introduction

[2] The West Antarctic Ice Sheet (WAIS) has the potential for rapid and significant change in grounding line position due to its marine character and fast-flowing ice streams [Alley and Whillans, 1991]. Such change is of interest in part because it would affect sea level immediately. The Ross Ice Shelf (RIS) grounding line has retreated more than 1000 km, from the continental shelf edge to its present location, in the time since the Last Glacial Maximum (LGM) [Conway et al., 1999; Shipp et al., 1999]. Of interest now is whether or not that retreat will continue, and if so, at what rate.

[3] Timescales for grounding line migration and variations in ice-stream discharge in the Ross sea sector of the WAIS are not well known. Observational data are rare but some general statements may be made. Ice streams were present when the grounding line retreated from the continental shelf edge [Mosola and Anderson, 2003]; thus it seems likely that ice streams have persisted since that time. The broad pattern of retreat since the LGM is early retreat to Roosevelt Island in the east and then a gradual “swinging gate” retreat to the present location with the island as a hinge point [Conway et al., 1999].

[4] The temporal character of grounding line retreat since the LGM is also of interest. Seismically imaged sedimentary wedges on the sea floor seaward of the RIS front have been interpreted as grounding line features deposited at sites where a retreating grounding line paused and held a steady position before resuming retreat [Mosola and Anderson, 2003]. Deformed ice–internal stratigraphy upstream of the present-day Kamb Ice Stream (KIS, formerly named C) grounding line suggests that the grounding line of this currently quiescent outlet was upstream of its present location within the last few hundred years [Catania et al., 2006]. The implication of these observations is that grounding line retreat since the LGM has been episodic, with transient stable stands and local readvance events. This behavior must have been tied to variability in the volume of ice discharged by ice streams into the shelf.

[5] The relationship between ice stream discharge and grounding line position is complicated. In simplest terms, the location of the grounding line depends on sea level, bed elevation, and ice thickness (or the volume of ice discharge). All else being equal, an increase in volume flux would lead to grounding line advance while a decrease would lead to grounding line retreat. Changes in strain rates that may accompany changes in volume flux modify this simple relationship. Acceleration of the flow accompanied by increased longitudinal strain rates across an ice stream to ice shelf transition would thin the ice and perhaps yield grounding line retreat, as in the classic marine ice sheet instability hypothesis [Thomas and Bentley, 1978; Hulbe and Payne, 2001]. Complications may arise owing to the
development of sticky spots in the discharge path [Hulbe and Fahnestock, 2004].

Evidence for both past and ongoing variation in ice stream discharge is found in many locations in the Ross Sea sector of the WAIS, for example, the stagnation of KIS about 150 years ago [Retzlaff and Bentley, 1993], the stagnation of Siple Ice Stream (a distributary of Kamb) 420 ± 60 years ago (Ben Smith, personal communication 2005), and recent slowing at the downstream end of Whillans Ice Stream (WIS, formerly B) [Joughin and Tulaczyk, 2002]. Together, these observations suggest a negative trend in the volume of ice being discharged from the WAIS. One conclusion that may be drawn from the recent trend is that the ice sheet is nearing the end of a long downwasting phase and the future contribution to sea level will be negative. However, these are as yet short-lived events in a system that has persisted for tens of thousands of years. Quantification of variability in ice stream discharge will help place the recent events in a useful predictive context.

Numerical models of the coupled ice sheet and ice shelf system have been used to study variability in ice sheet discharge. Unfortunately, models of the whole ice sheet system are in general too coarse to resolve many details of grounding line migration and often do not include realistic transitions from grounded to floating ice [Hulbe and Payne, 2001]. Models designed to study ice stream discharge cycles (slowing, stagnation, and reactivation) produce millennial or longer timescales [MacAyeal, 1992; Payne, 1995; Parizek et al., 2003]. Quantitative interpretations of present-day WAIS conditions [Bougamont et al., 2003; Joughin and Tulaczyk, 2002; Vogel et al., 2005] generally favor ice stream deceleration and relatively long timescales for reactivation. The prediction by Vogel et al. [2005] that the now-quiescent KIS may soon reactivate, based on borehole observations of subglacial conditions, is an exception to conventional views.

Flow features in the surface of the RIS provide an opportunity to extend the high-resolution observational record of ice stream and grounding line variability beyond the instrumental era [Fahnestock et al., 2000]. Because the flow of the shelf must adjust to the volume flux of ice discharged into it, physical tracers within the ice form a time-integrated record of past events [MacAyeal et al., 1988]. The RIS record extends back about 1000 years, the approximate age of ice now at the shelf front. Hulbe and Fahnestock [2004] studied features associated with the lightly grounded, low-slope ice plains across which active ice streams make the transition from a grounded to floating condition, and concluded that grounding line migration is episodic and driven by thermal processes near the base of the fast-flowing ice. Here we direct our attention to recon-

Figure 1. Composite MODIS image of the Ross Ice Shelf.
Flow Features in the Ice Shelf Surface

Flow features observed on the surfaces of floating ice shelves chronicle kinematic and dynamic events over the lifetimes of the features [MacAyeal et al., 1988]. Three broad classes of flow feature are observed throughout the RIS (Figures 1 and 2): subtle ridges in surface elevation that in gross scale trend downstream (streaklines); trains of shear-margin crevasses that advect downstream from their sites of formation; and transverse crevasses of multiple possible origins. Streaklines originate in the grounded ice sheet, where ice flows over a basal disturbance such as a sticky spot, roches moutonnee, or other obstacle at the bed [Gudmundsson et al., 1998]. Throughout the WAIS, the farthest upstream streaklines are observed where ice enters a fast-flowing ice stream or tributary [Merry and Whillans, 1993]. Streaklines within ice streams form simple geometries that can be traced over hundreds of kilometers. Trains of shear margin crevasses form at lateral stick-slip margins. Trains may be actively forming at the boundaries of the shelf, or may be relict ice-stream margins advecting through the shelf. The RIS stands out among Antarctic ice shelves in that flow feature geometries are complex, indicating significant variability in ice shelf flow [Casassa et al., 1991; Fahnestock et al., 2000; Jezek, 1984]. Here we will focus on streaklines and relict ice-stream margins as tools to map the provenance of ice within the shelf and to infer past variations in ice stream discharge into the shelf.

The subtle surface morphology present on the RIS can be observed, to a limited extent, in many types of visible-band satellite imagery [Bindschadler, 1993; Casassa et al., 1991; Williams and Ferrigno, 1988]. Enhancements made using multiple images can greatly improve our ability to see such features [Scambos and Fahnestock, 1999; Fahnestock et al., 2000]. Here we use a multiple-image composite from 250 m pixel size red-light (band 1) images from the MODIS sensor on NASA’s Terra platform. The MODIS imagery was obtained from the NASA Goddard Space Flight Center Earth Sciences Distributed Active Archive Center. At any location, the composite includes a few to ten images spanning a few-week period in November and December 2001, using multiple local solar times from 1000 GMT to 2200 GMT. This range of image acquisition times provides a range of sun azimuths in the images used. The images were individually remapped into a polar stereographic grid at a pixel size of 250 m, hand masked for clouds using polygonal areas, and high-pass filtered with a 25 km radius filter to remove the effects of larger-scale slope variations and allow contrast enhancement for smaller spatial-scale slope variations. Subtle topographic ridges, with a short dimension on the order of a kilometer, are visible in the composite image. The range of solar azimuths (illumination directions) necessitates some caution in interpreting slope directions of features seen in the composite but the possible confusion is balanced by the greatly enhanced ability to track features as coherent structures for hundreds of kilometers across the ice shelf. A high-resolution composite image of similar construction but with additional filtering to remove effects of image
2.1. Ice Provenance

Mosaic of Antarctica Center data archive (http://nsidc.org). The composite image edges is available via the U.S. National Snow and Ice Data Center data archive (http://nsidc.org). The composite image in Figure 1 is an extract from that data product, the MODIS Mosaic of Antarctica [Haran et al., 2005].

Figure 3. Development of distorted streaklines around Crary Ice Rise due to a stagnation event on WIS following the experiment I scenario (Table 3). (a) Steady state trajectories with grounded, but not stagnant, CIR. Dark grey swaths indicate soft ice zones specified in the model by adjusting the parameter \( \alpha \). (b) Nine hundred years before end of model run: ice at CIR is stagnant, producing localized shear deformation around the ice rise. Arrows illustrate the stress pattern dominating streakline deformation. The streaklines depicted are in the WIS outflow only. (c) Five hundred years before end of model run: WIS discharge has ceased and ice in the WIS/MIS bay spreads laterally, creating a starter fold. (d) Three hundred years before end of model run: WIS discharge has resumed. The increased volume flux enhances the starter fold by compressing it longitudinally and causing it to extend in the right lateral direction. (e) One hundred years before end of model run: large folds have developed around CIR. (f) End of model run: provenance map of ice in the vicinity of CIR.

2.1. Ice Provenance

Relict margins and streaklines can be traced back upstream to their emergence into the ice shelf, and in some cases upstream into the grounded ice sheet. We have traced major streaklines and shear margin tracks upstream to create a provenance map for ice now in the RIS (Figure 2). The most dramatic feature of the provenance map is the missing ice in the WIS discharge, downstream and to the north of Crary Ice Rise. The implication is that WIS ceased discharging ice into the shelf for several hundred of the last 1000 years and that after having stagnated, the stream reactivated.

2.2. Streakline Distortions

Upon entering the ice shelf, streaklines and relict shear margins become passive tracers that persist for hundreds of kilometers [Casassa and Whillans, 1994; Fahnestock et al., 2000; Jeziak, 1984]. Each streakline is the locus of all fluid particles (or parcels of ice) that have passed through a common upstream location [Acheson, 1990]. The trajectory of any individual parcel is determined by the time-transient velocity field it experiences, so each streakline reveals an integrated record of ice shelf flow over its lifetime. This is analogous to a plume formed by the release of dye at a fixed point in a moving fluid.

If ice flow changes over time, the trajectories of individual parcels are diverted, and the streaklines appear to be distorted when compared to the overall flow field at any arbitrary point in time. Streaklines in the eastern RIS do not correspond to present-day flowlines while streaklines in the western RIS do approximate present-day trajectories [Jeziak, 1984; Fahnestock et al., 2000]. This indicates past variability in ice shelf flow and thus in the discharge of ice streams feeding the eastern part of the shelf.

Streaklines are distorted by transverse changes in ice flow. Such changes may result from the emergence of an obstacle in the flow path, for example the formation of an ice rumple or ice rise, or from changes in transverse thickness gradients. In the first case, distortion is due to shear while in the second, distortion is due to transverse stretching. These two forcings produce different distortion geometries. The emergence of an obstacle, a site where basal drag increases and ice flow ceases or slows considerably, may trap a segment of a streakline at the obstacle location. Continued ice flow will displace the upstream continuation of the streakline past the obstacle, creating a shear fold. Such folds will have a distinctive, downstream-pointing, short-wavelength shape. The tightness of such folds is due to the concentration of shear near the margin of the obstacle [Raymond, 1996]. A fold of this type is now forming at the downstream end of Mercer Ice Stream (MIS, formerly A) [Hulbe and Fahnestock, 2004]. Large changes in transverse strain rates are likely to arise near ice stream outlets owing to changes in ice stream discharge and thickness. For example, when flux from an ice stream decreases, ice from adjacent outlets will spread laterally to take the place of the missing ice. A range of fold morphologies may be produced, depending on the magnitude and the longevity of the forcing. Numerical experiments presented later in this paper show that both processes, and combinations of the two, are required to produce the streakline distortions now visible on the RIS (Figure 3).

3. Numerical Simulation of Streakline Evolution

3.1. Model

A map-plane, finite-element model of ice-shelf stress and mass balance, coupled with a lagrangian tracer routine, is used to simulate the deformation of streaklines on the RIS. The model velocity field changes according to changing boundary conditions. The time-transient velocity field is in turn used to move tracers through the model domain. When connected, the sequential positions of tracers emitted from a fixed location within the model domain can be compared with the streaklines observed in satellite imagery of the ice shelf surface. Our goal is to establish the timing of major discharge events, such as the apparent shut-down and reactivation of Whillans Ice Stream.

Ice shelves flow by gravity-driven horizontal spreading that transports ice from the grounding line to the calving front. The location of the grounding line is determined by flotation, where the weight of the ice is balanced by buoy-
ancy. Lightly grounded ice plains broaden the grounding line into a grounding zone at the downstream ends of fast-flowing ice streams. Resistance to ice-shelf flow is provided by lateral shear at bay walls and ice rise margins, by compression upstream of ice rises, and perhaps by basal traction beneath ice plains.

[17] Changes in ice stream discharge affect the flowing shelf in two different ways. Changes in boundary velocity propagate nearly instantaneously through the ice shelf while changes in ice thickness propagate on the advective timescale [MacAyeal and Lange, 1988]. Past volume flux is unknown, so we try many scenarios and evaluate the quality of each set of time-transient boundary conditions by comparison of model products with observed ice shelf features.

[18] The ice shelf model derives from the more complete ice sheet model presented by Hulbe and MacAyeal [1999]. The most essential components of the model, and special features developed for this work, are presented here. The model is used to produce time-transient variations in ice shelf velocity and thickness in response to prescribed changes in ice stream discharge. In order to complete many simulations, we limit the model domain to the present-day ice shelf geometry, with an upstream boundary near the present-day grounding line. This is not entirely unreasonable over the relatively short time span of interest (the last 1000 years) although we will demonstrate that the grounding line position both advances and retreats tens of km over this time period, requiring some adaptation of the ice shelf model.

[19] Ice shelf flow is represented by a set of stress-balance equations, simplified by the usual assumption that horizontal flow is depth-independent [MacAyeal and Thomas, 1982]. The Glen flow law is embodied in an effective viscosity \( \nu_e \) that makes use of a depth-average inverse rate factor \( B \). In a Cartesian coordinate system with horizontal coordinates \( x_i \), the stress balance is expressed,

\[
\begin{align*}
\frac{\partial}{\partial x_i} \left( 2 \nu_e \frac{\partial u}{\partial x_i} \right) &+ \frac{\partial}{\partial x_j} \left( 2 \nu_e \frac{\partial u}{\partial x_j} \right) - \rho g h \frac{\partial x}{\partial x_i} - \frac{\partial}{\partial x_i} &u_i \beta = 0,
\end{align*}
\]

where \( h \) represents the ice thickness, \( u \) represents the horizontal velocity, \( \rho \) represents the depth-average density of the ice, \( g \) represents the acceleration due to gravity, \( s \) represents the surface height, and \( \beta \) is a basal friction parameter that in the case of floating ice is equal to zero and can be set to some nonzero value for ice flowing over an ice plain or rumple [MacAyeal et al., 1995]. Here \( \beta \) is a simple scalar quantity with limited physical meaning beyond its utility in generating basal drag. A no-slip condition is specified at no-flow boundaries such as bay walls, the downstream ends of interstream ridges, and the edges of ice rises. The stress balance equations are solved by iteration on the effective viscosity,

\[
\nu_e = \frac{\alpha B}{2 \left( \left( \frac{\partial u}{\partial x_i} \right)^2 + \left( \frac{\partial u}{\partial x_j} \right)^2 + \frac{1}{3} \left( \frac{\partial u}{\partial x_i} + \frac{\partial u}{\partial x_j} \right)^2 + \frac{1}{3} \frac{\partial u}{\partial x_i} \frac{\partial u}{\partial x_j} \right)^{\frac{3}{2}}},
\]

in which \( \alpha \) is a tuning parameter and the flow-law exponent \( n \) is 3.

[20] The inverse rate factor used here is based on an estimated ice temperature and is modified via the spatially variable multiplier \( \alpha \) in equation (2). First, a depth-varying temperature field is created using a parabolic function with endpoints at a spatially variable surface temperature \( T_s \) and a spatially uniform basal temperature \( T_b \). The temperature field is then used to compute a depth-average rate factor. Finally, the product \( \alpha B \) is tuned by comparison with Ross Ice Shelf Geophysical and Glaciological Survey (RIGGS) velocity measurements [Bindschadler, 1993; Thomas et al., 1984].

[21] The temperature shape function is

\[
T(\zeta) = T_s + (T_b - T_s)\theta(\zeta),
\]

where \( \zeta \) is a normalized vertical coordinate ranging from 0 at the base to 1 at the upper surface and

\[
\theta = (1 - \zeta)^2.
\]

The parabolic shape mimics temperature profiles observed in large ice shelves. This is similar to the approach of Humbert et al. [2005], although those authors use a colder (clear-sky only) \( T_b \) than is used in the present work. \( T(\zeta) \) could also be estimated with a linear shape function for \( T(\zeta) \) or by specifying \( T(\zeta) \) at inflow boundaries and including energy balance in the model equations. We have experience with all three approaches in various settings [Hulbe and MacAyeal, 1999; Hulbe et al., 2005; Scambos et al., 2000]. Here we use the parabolic shape function because it is straightforward and yields flow fields with downstream strain rates that more closely resemble observed present-day strain rates than does the simpler linear function. A more complete treatment of energy balance is unwarranted without an ice sheet model upstream of the shelf.

[22] The inverse rate factor can be thought of as an ice “stiffness.” Its temperature dependence is expressed,

\[
B = a \left( \exp \left( \frac{-Q}{RT} \right) \right) \frac{1}{\zeta},
\]

where \( T^* \) represents the homologous temperature, \( R \) represents the gas constant, and the rate-factor constant \( a \) and activation energy for creep flow \( Q \) have temperature-dependent values,

\[
a = \begin{cases}
7.23 \times 10^{-12} \text{Pa}^{-3} \text{s}^{-1} & \text{if } T^* < 263 \text{ K} \\
3.47 \times 10^4 & \text{if } T^* \geq 263 \text{ K}
\end{cases}
\]

\[
Q = \begin{cases}
6.0 \times 10^4 \text{ J mol}^{-1} & \text{if } T^* < 263 \text{ K} \\
13.9 \times 10^4 & \text{if } T^* \geq 263 \text{ K},
\end{cases}
\]

following Payne [1995]. \( T(\zeta) \) and in turn \( B \) are computed using mean annual surface temperature from the Comiso [1994] analysis of data from the Nimbus 7 infrared radiometer. A uniform \( a \) of 1.4 in equation (2) yields a good fit to the RIGGS velocity field through the central portion of the ice shelf, indicating that the depth-average of our simple \( T(\zeta) \) is too warm.
reduce what the model would otherwise simulate as a very fast discharge through that region. Two values are assigned to $\beta$, a background $\beta_0 = 0.1 \times 10^8$ Pa s m$^{-1}$ across most of the ice plain and $1 \times 10^8$ Pa s m$^{-1}$ beneath the narrow region upstream of CIR. The basal shear stress takes up, on average, 50% of the driving stress across the ice plain. Other modeling studies, which fail to account for shear margins or the ice plain, or to adequately resolve the narrows between Crary and the TAM, fail to match measured ice speed in this area [MacAyeal et al., 1996].

[26] The basal traction is also manipulated in transient experiments where ice runs aground owing to thickening. We have little insight into the selection of this value and use the background $\beta_0$ defined above. $\beta$ is also used to simulate the emergence of ice rises. Ice speed is nearly zero on the large present-day Crary and Steershead ice rises but this has not always been the case. Temperature profiles measured in CIR indicate a melted base about 1000 years ago [Bindschadler et al., 1990] and the prominent crevasse track downstream of Steershead is about 350 years old [Fahnestock et al., 2000]. Here we simulate the emergence of ice rises via a linear increase from $\beta_0$ to $10 \times 10^8$ Pa s m$^{-1}$, at which point the model nodes are set to a no-flow condition. Nonzero $\beta$ over the site of an ice rise increases the magnitude of flow diversion around the obstruction (Figure 3).

[27] Ice shelf thickness varies in response to changing ice stream discharge and basal traction under grounded ice. Mass conservation is expressed, 

$$\frac{\partial h}{\partial t} = \dot{a} + b - \nabla \cdot (uh), \tag{6}$$

in which $t$ represents time, $\dot{a}$ and $\dot{b}$ are upper and lower surface accumulation rates and $u$ represents the vector-valued horizontal velocity. The upper surface ice accumulation rate is specified according to Vaughan et al. [1999]. Bottom accumulation, which varies in magnitude and in sign and will certainly change as regions ground and go afloat over time, is not easy to prescribe. Correct application of $\dot{b}$ may affect details of ice grounding but in our experience has a minor effect on the outcome of any particular experiment.

### 3.2. Experiments

[28] Our numerical model experiments simulate 1600 years of ice shelf flow with the goal of identifying and placing time constraints on the causative events for large streakline deformations observed on the RIS (Figures 1 and 2). The timing of each event is approximate. We use the term “years ago” to mean years before the end of the modeled time interval, which corresponds to the range of dates over which images for the composite were acquired, November and December 2001. We consider a successful experiment to be one in which the general form and downstream locations of streakline fold packages compare well with the observed streaklines. The lateral extents of fold packages are more difficult to reproduce than the downstream placement of folds because lateral extent depends on the behavior of adjacent ice streams in concert, as well as correct estimation of past volume flux and basal traction on the ice plain. Repeated experiments with slight

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**Table 1. Chi-Square Test of Modeled Modern Ice Speed Using Spatially Variable, as in Text, and Homogeneous Values of $\alpha$**

<table>
<thead>
<tr>
<th>$\alpha$</th>
<th>$\chi^2_{\text{CIR}}$</th>
<th>$\chi^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>As in text</td>
<td>8220</td>
<td>1393</td>
</tr>
<tr>
<td>1</td>
<td>120960</td>
<td>8878</td>
</tr>
<tr>
<td>1.6</td>
<td>5806</td>
<td>2152</td>
</tr>
</tbody>
</table>

$^*$Comparison is with the RIGGS data set [Thomas et al., 1984]. The subscript CIR indicates Crary Ice Rise region only; $n = 143$ and 46 for the whole-shelf and CIR region comparisons, respectively.

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[23] The ice shelf model infrastructure can correctly reproduce many features of present-day RIS flow, even with a uniform $\dot{B}$ and no notion of basal traction $\dot{b}u$ on ice plains [MacAyeal et al., 1996]. However, modifications to both parameters are necessary to accurately simulate regional details of importance to this study, in particular the pattern of flow around Crary Ice Rise (CIR) [Hulbe and Fahnestock, 2004]. First, to promote ice flow through the narrows between CIR and the Transantarctic Mountains (TAM), ice softness must be enhanced in shear margins along the coasts of both features. This is reasonable, given the large degree of crevassing along ice rise margins and the advection of glacier shear margins downstream along the TAM front [Bindschadler, 1993; Thomas et al., 1984]. In the model, ice is softened by setting $\alpha < 1$. A reasonable agreement with observed velocity around CIR is achieved with $\alpha = 0.7$ in narrow (2 to 3 km) bands along the TAM coast from 170 W to 180 W longitude and along the right lateral side of CIR, and with $\alpha = 0.1$ in a band about 15 km outboard of the southern side of CIR. A narrow shear margin with $\alpha = 0.9$ is specified around the margins of Roosevelt Island in order to match speeds in the narrows between the island and adjacent coast.

[24] The quality of the simulated modern flow field may be evaluated both qualitatively and quantitatively. Qualitatively, we are interested in the best representation of specific attributes of the flow field that are important to the intended use of the model. Here it is important to achieve an appropriately large flux of ice around CIR so that the ice does not thicken unrealistically in diagnostic simulations of the ice shelf with variable boundary conditions. It is less important to correctly reproduce ice speed at the front of the shelf, which is distant from our area of interest. Correctly reproducing ice flow everywhere in the shelf requires spatial variation in $\alpha$ throughout the ice shelf [Rommelaere and MacAyeal, 1996] that could not be justified in the diagnostic model runs. Thus we rely on a relatively simple spatial variation in $\alpha$. The goodness of the present-day model tuning of $\alpha$ can be evaluated using a chi square test to compare ice speed measured during the RIGGS campaign [Thomas et al., 1984] with simulated ice speed [MacAyeal et al., 1996]. Tuning $\alpha$ as described above improves the fit around CIR but degrades the overall model fit to the RIGGS observations (Table 1). This is the case because the uniform midshelf $\alpha$ promotes too-large strain rates as the ice flows toward the shelf front. We accept this trade-off for reasons described above.

[25] Second, basal traction is applied between CIR and the downstream end of the ridge between WIS and KIS to
Table 2. Ice Stream Discharge Boundary Conditions

<table>
<thead>
<tr>
<th>Boundary Inflow</th>
<th>Speed</th>
<th>Volume</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Fast-Flowing KIS Initialization (Experiment I; Figure 4)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mercer IS</td>
<td>400</td>
<td>7.5</td>
</tr>
<tr>
<td>Whillans IS</td>
<td>550</td>
<td>39</td>
</tr>
<tr>
<td>Kamb IS (Duckfoot active)</td>
<td>500</td>
<td>41</td>
</tr>
<tr>
<td>Bindschadler IS</td>
<td>300</td>
<td>13.6</td>
</tr>
<tr>
<td>Siple IS</td>
<td>300</td>
<td>5.0</td>
</tr>
<tr>
<td>MacAyeal IS</td>
<td>350</td>
<td>17</td>
</tr>
<tr>
<td>Echelmeyer IS</td>
<td>140</td>
<td>2.2</td>
</tr>
<tr>
<td>Prestrud Inlet</td>
<td>200</td>
<td>1.5</td>
</tr>
<tr>
<td>Scott and Amundsen Glaciers</td>
<td>170</td>
<td>2.7</td>
</tr>
<tr>
<td>Beardo Glacier</td>
<td>470</td>
<td>4.2</td>
</tr>
<tr>
<td>Nimrod Glacier</td>
<td>250</td>
<td>2.4</td>
</tr>
<tr>
<td>Byrd Glacier</td>
<td>600</td>
<td>11.5</td>
</tr>
<tr>
<td>Mulock Glacier</td>
<td>290</td>
<td>2.1</td>
</tr>
<tr>
<td><strong>Slower KIS Initialization (Experiment II; Figure 5)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Kamb IS (Duckfoot active)</td>
<td>300</td>
<td>24.6</td>
</tr>
<tr>
<td><strong>Miscellaneous Changes</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Bindschadler IS</td>
<td>500</td>
<td>23</td>
</tr>
<tr>
<td>MacAyeal IS</td>
<td>550</td>
<td>27</td>
</tr>
<tr>
<td>Mercer IS</td>
<td>800</td>
<td>15</td>
</tr>
</tbody>
</table>

*Boundary speeds are in m yr⁻¹ and volume fluxes are km³ yr⁻¹ [see MacAyeal and Thomas, 1986; Shabtaie and Bentley, 1987].

Variations in discharge event timing indicate that we can correctly identify the timing for creation of well-simulated deformation features within about 50 years.

Each experiment starts from a steady state initialization of h and then runs forward in time for 1600 years with changing boundary conditions (Tables 2 and 3). Steady state for any initialization is taken to be a change in h of ≤1 × 10⁻⁶ m a⁻¹ at any node in the model domain. The first 600 years of each transient experiment are used to transform the Crary region from a location of grounded but fast-flowing ice into an ice rise. While it seems unlikely that the system we study was near steady state at any time since the last glacial maximum, let alone 1600 years ago, such an initialization is needed if we are to evaluate the effects of boundary perturbations on streakline geometry.

Past boundary fluxes are not known, and indeed, one objective of the present work is to infer large-scale changes in those boundary conditions over time. We have experimented with several initializations, the primary difference being the flux of ice from Kamb, Bindschadler, and MacAyeal ice streams. Prior to its stagnation, KIS was a fast-flowing outlet of the ice sheet [Ng and Conway, 2004]. The results presented here are not very sensitive to the choice of initialization. One exception is boundary flux from Bindschadler Ice Stream (BIS, formerly D) and MacAyeal Ice Stream (MacIS, formerly E). In order to prevent excessive grounding of ice between Roosevelt Island and the grounding line, fluxes from these two ice streams are set lower than the present flux through much of the model time (Table 2).

Ice around CIR is grounded at the start of all of our simulations. This is due to the relatively high bed elevation in this area [Hulbe and Fahnestock, 2004]. In this context, ice rise formation is related to changing basal traction at the location of the ice rise, not grounding of floating ice.

Transient boundary conditions include changes in ice stream discharge and changes in basal traction beneath grounded portions of the ice shelf (Table 3). Flux changes are made in a step-wise manner. Studies of internal layers and buried crevasses on Kamb Ice Stream support the notion that its shut-down transpired quickly [Rettloff and Bentley, 1993; Smith et al., 2002] so this approach is reasonable. Basal traction on the ice plain is also treated in a simple way, as discussed above. Margin jumps are represented by setting influx to 0 at affected boundary segments.

3.2.1. Whillans and Mercer Events

Our experiments place the timing for stagnation of the WIS outlet at about 850 years ago with reactivation 400 years later. The stagnation event is responsible for the formation of the complicated suite of folds around and downstream of CIR (Figures 3 and 4). When Whillans shuts down, ice from Mercer Ice Stream and nearby TAM glaciers spreads laterally, creating a large fold in streaklines upstream of CIR. When Whillans reactivates, the starter fold is enhanced by compression upstream of CIR and shearing around its margins. Alternatives, such as reduced, but not zero, flux, even over longer time periods, do not reproduce the observed streakline features.

The large streakline distortions downstream of CIR, first recognized by meteorologist Jorge Carrasco in weather satellite images [Casassa et al., 1991] are a result of the WIS stagnation event, not a surge from Mercer or TAM glaciers, as has been suggested. Indeed, the effect of a surge from those sources would be to relax folds because it would increase downstream stretching (Figure 5). It is interesting

Table 3. Changes to Boundary Conditions Used in Two Numerical Model Experiments

<table>
<thead>
<tr>
<th>Event</th>
<th>Timing, Years Ago</th>
<th>Magnitude</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>I: WIS and MacIS Stagnation Events (Figure 4)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CIR β</td>
<td>1600 to 1000</td>
<td>β to 10 × 10⁸</td>
</tr>
<tr>
<td>CIR stagnates</td>
<td>1000</td>
<td>u = 0</td>
</tr>
<tr>
<td>WIS stagnates</td>
<td>850</td>
<td>u = 0</td>
</tr>
<tr>
<td>MacIS stagnates</td>
<td>800</td>
<td>u = 0</td>
</tr>
<tr>
<td>Duckfoot margin jump</td>
<td>670</td>
<td>u = 0</td>
</tr>
<tr>
<td>MacIS reactivates</td>
<td>650</td>
<td>u = 350</td>
</tr>
<tr>
<td>Steershead northern β</td>
<td>530 to 500</td>
<td>β to 5 × 10⁴</td>
</tr>
<tr>
<td>Siple Ice Stream stagnates</td>
<td>460</td>
<td>u = 0</td>
</tr>
<tr>
<td>WIS reactivates</td>
<td>450</td>
<td>u = 550</td>
</tr>
<tr>
<td>MacIS speed increases</td>
<td>350</td>
<td>u = 550</td>
</tr>
<tr>
<td>BIS speed increases</td>
<td>350</td>
<td>u = 500</td>
</tr>
<tr>
<td>Steershead β to 5</td>
<td>250 to 200</td>
<td>u = 10 × 10⁸</td>
</tr>
<tr>
<td>Steershead stagnates</td>
<td>200</td>
<td>u = 0</td>
</tr>
<tr>
<td>KIS trunk shuts down</td>
<td>150</td>
<td>u = 0</td>
</tr>
<tr>
<td><strong>II: MIS Surge (No WIS Cycle) Conventional KIS Events (Figure 5)</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CIR β</td>
<td>1600 to 1000</td>
<td>β to 10 × 10⁸</td>
</tr>
<tr>
<td>CIR stagnates</td>
<td>1000</td>
<td>u = 0</td>
</tr>
<tr>
<td>MIS speed increases</td>
<td>850</td>
<td>u = 800</td>
</tr>
<tr>
<td>MacIS stagnates</td>
<td>800</td>
<td>u = 0</td>
</tr>
<tr>
<td>Scott + Amundsen speed increases</td>
<td>850</td>
<td>u = 340</td>
</tr>
<tr>
<td>MacIS reactivates</td>
<td>650</td>
<td>u = 350</td>
</tr>
<tr>
<td>KIS speed increases</td>
<td>550</td>
<td>u = 340</td>
</tr>
<tr>
<td>Duckfoot margin jump</td>
<td>460</td>
<td>u = 0</td>
</tr>
<tr>
<td>MIS speed increases</td>
<td>450</td>
<td>u = 400</td>
</tr>
<tr>
<td>Scott + Amundsen speed increases</td>
<td>450</td>
<td>u = 170</td>
</tr>
<tr>
<td>Steershead stagnates</td>
<td>360</td>
<td>u = 0</td>
</tr>
<tr>
<td>MacIS speed increases</td>
<td>350</td>
<td>u = 550</td>
</tr>
<tr>
<td>BIS speed increases</td>
<td>350</td>
<td>u = 500</td>
</tr>
<tr>
<td>KIS trunk shuts down</td>
<td>150</td>
<td>u = 0</td>
</tr>
</tbody>
</table>

*Boundary speeds are in m YR⁻¹; β is Pa m s⁻¹.

This is an event timed differently than suggested by the literature.
to note that a core of relatively old ice is preserved within fold limbs in the wake of CIR.

3.2.2. Kamb and Steershead Events

The recent stagnation of the Kamb Ice Stream outlet is suggested to have occurred in three stages, with inward margin jumps preceding stream-wide stagnation [Catania et al., 2006]. The larger of two recent margin jumps resulted in the stagnation of an area on the right-lateral side of the ice stream outlet often called the “Duckfoot” (a dashed line in Figure 2). Stagnation of the Duckfoot is a likely source for a starter fold that evolved into the large folds now observed around Steershead. Unlike the WIS event, where an obstruction (CIR) predates fold initialization, in the case of the Duckfoot (and then KIS) stagnation [Fahnestock et al., 2000], observations suggest that Steershead formation is closely timed with the ice stream event. Indeed, the deformed streaklines around Steershead should record important information regarding the timing of those events.

Simulations using timing reported in the literature [Catania et al., 2005; Fahnestock et al., 2000] for events associated with KIS stagnation fail to reproduce the observed degree of streakline folding in the Steershead region. We have more success with an early Duckfoot margin jump and delayed stagnation of ice on the basal high beneath Steershead (Table 3 and Figures 4 and 5). There are two obvious causes for the difficulty in reproducing the observed streakline distortions around the Steershead. First, the Duckfoot stagnation 350 years ago [Catania et al., 2005] is too recent for a starter fold to form, advect downstream, and become trapped and distorted around Steershead. The problem is exacerbated by the shutdown of the entire ice stream, which reduces the downstream advection and distortion of the fold. Second, discharge from WIS limits the left-lateral expansion of folds as they deform around Steershead. Bathymetry between Steershead and the downstream end of Siple Dome may be a third contributor to the modest streakline distortions in our experiments. The relatively shallow bed in this region discourages northward redirection of ice flow, limiting the size of the starter fold. Streakline geometries improve when we impose an upstream propagation of the increase in basal traction under the Steershead. It is of course possible that flawed knowledge of bed geometry through this region also contributes to the problems described here. More work is required to adequately reproduce this feature, hopefully with the result of improving our understanding of the recent KIS stagnation.

3.2.3. MacAyeal and Bindschadler Events

An unanticipated outcome of the present work is the recognition of a stagnation and reactivation event on MacAyeal Ice Stream. Muted streakline folds along the southern side of Roosevelt Island indicate lateral expansion of ice discharging from the adjacent Bindschadler Ice Stream during MacIS quiescence and fold enhancement by shear around Roosevelt Island after reactivation of MacIS. The outlet stagnation event began about 800 years after the Duckfoot event.

Figure 4. Streakline geometries at the end of model experiment I in Table 3, in which WIS and MacIS experience stagnation events. Steershead becomes an ice rumple and stagnates late in the KIS stagnation cycle. This is the most successful simulation to date, capturing the major streakline deformations observed in the RIS. Heavy grey lines along the boundary of the model domain indicate influx gates.

Figure 5. Streakline geometries at the end model of experiment II in Table 3, in which MIS and adjacent Scott and Amundsen Glaciers surge. This fails to reproduce folds observed both upstream and downstream of Crary Ice Rise, as discussed in the text. Steershead stagnates at a time suggested by other authors. KIS discharge increases 100 years before the Duckfoot margin jump in an unsuccessful attempt to displace WIS ice eastward. MacIS experiences a stagnation event. Heavy grey lines along the boundary of the model domain indicate influx gates.
ago, with reactivation about 150 years later. The shorter quiescent phase compared to the WIS event may be related to differences in potential gradients in the two systems.

Siple Ice Stream (SIS) is a former distributary of KIS. Internal layers observed via ice-penetrating radar have been used to place the age of SIS stagnation between 400 and 450 years ago [Conway et al., 2002]. There is no obvious record of this event in the RIS streaklines and indeed, no distortion is produced by the deactivation of this outlet in our model experiments.

4. Discussion

4.1. Ice Streams Stop and Start

Flow features preserved in the Ross Ice Shelf reveal that the Ross ice streams stagnate and reactivate on century timescales. Whillans Ice Stream stagnated about 850 years ago and reactivated 400 years later. MacIS experienced a similar cycle 800 to 650 years ago. It is reasonable to suppose that these, and other, ice streams have experienced similar cycles in the past. Indeed, the sea floor record of episodic grounding line retreat can be read as evidence of just such events [Anderson et al., 2002].

While the details of timing and magnitude of the events interpreted here may be modified by additional experiments, the fundamental conclusions are robust. The observed features are not easy to reproduce; they are a distinctive fingerprint of a particular set of events. Large, abrupt change in transverse normal strain is always the causative agent in the formation of large streakline folds. Features created by such events may be enhanced by subsequent advection and rotation or muted by downstream stretching.

The notion of rapid reactivation of stagnant ice streams and past stagnation/reactivation cycles is new. It bears directly on interpretation of ongoing events such as the slow-down of WIS and on projections of future change in the WAIS. Clearly, models intended to produce useful projections of ice sheet contribution to sea level rise on human timescales must be able to reproduce century-scale cycles in ice stream discharge.

4.2. Grounding Line Location

Recent past positions of the Ross grounding line are not known in any detail but have been inferred from several lines of evidence. Jezek [1984] mapped the spatial pattern of basal reflectance downstream of Crary Ice Rise and estimated grounding line retreat past the ice rise within the last 1000 years on the basis of the pattern of reflection strength within the ice shelf. Catania et al. [2006] imaged downwarped internal stratigraphy in the low-relief ice terrain on the downstream end of the ridge between KIS and WIS and interpreted them as evidence of a grounding line well inland of its present location within the last few hundred years. Together, these observations suggest 100s of km of grounding line retreat and readvance on century timescales. The stagnant KIS grounding line was observed to be retreating at a rate of about 30 m yr$^{-1}$ in the early 1980s [Thomas et al., 1988].

The ice stream discharge cycles simulated in our numerical model experiments yield large-scale changes in grounding line position on century timescales (Figure 6). In brief, when an ice stream stagnates, ice across the ice plain thins and goes afloat, producing widespread grounding line retreat. When an ice stream reactivates, ice across the ice plain regrounds, producing widespread grounding line advance. Of particular interest is the rapid change in KIS grounding line position resulting from stagnation and reactivation of WIS. Regression and transgression of the grounding line across the ice plain proceeds at a mean rate of about 50 km per century, much faster than the recent retreat rate of 30 m yr$^{-1}$ [Thomas et al., 1988]. Those authors suggested that the low-relief ice plain could allow grounding line retreat rates up to 20 times faster than the observed rate, a range that embraces the rates generated in our simulations. It is possible that the upstream incursion of the KIS grounding line suggested by Catania et al. [2006] is a result of WIS stagnation. Our models do not, at present, extend far enough upstream to verify this suggestion.

4.3. Grounding Zone Dynamics

The experiments presented here suggest that a discharge event on one ice stream may be transmitted quickly to the downstream end of an adjacent stream across a broad, low-slope ice plain. In the present study, as discharge from WIS ceases, ice downstream of Mercer, Whillans, and Kamb ice streams goes afloat. The regions reground when WIS reactivates. It is possible that this communication via ice thickness across the ice plain may play an important role in the phasing of discharge cycles, in addition to timing of grounding line retreat. If so, the forcing mechanisms would be local changes to surface slope (and the potential gradients directing basal water flow) and basal energy balance caused by regrounding.

Consider the sequence of events leading to the recent stagnation of the KIS outlet. Five hundred years ago, a few decades prior to the reactivation of WIS, the KIS grounding line is suggested to have been about 100 km upstream of its present location [Catania et al., 2006]. One hundred and fifty years later, the grounding line moved to its present location [Thomas et al., 1988]. Those observations suggest 100s of km of grounding line retreat and readvance on century timescales. The stagnant KIS grounding line was observed to be retreating at a rate of about 30 m yr$^{-1}$ in the early 1980s [Thomas et al., 1988].

Our numerical model experiments lead to several important interpretations of the distorted streaklines observed in the surface of the RIS. The complex of folds around CIR involving ice from WIS, MIS, and several TAM glaciers can be reproduced only by a large-scale stagnation event on WIS. Our experiments date that event from 850 to 450 years ago. A more subdued set of folds on the left-
lateral side of Roosevelt Island are reproduced by a stagnation event on MacIS. Our experiments date the MacIS event from 800 to 650 years ago. The stagnation of SIS, a small KIS distributary on the north flank of Siple Dome is not recorded in the observed streaklines. [47] Streakline folds downstream of the present-day KIS grounding line are more difficult to reproduce than are folds associated with other ice stream outlets. The difficulty arises from the short time between stagnation of the Duckfoot, the event that initiates folding, and stagnation of the entire KIS outlet. Folds with the correct basic geometry are created but do not advect far enough downstream. Lateral compression due to WIS ice pushing into the area after reactivation of that ice stream exacerbates the problem in our numerical model experiments. Increasing the age of the Duckfoot margin jump and decreasing the age of Steershead stagnation improves simulation of the folds around Steershead but this is not supported by published dates for these events. [48] Our work also leads to several interesting possibilities regarding the dynamics of ice stream outlets. First, the grounding line may experience far more rapid regression and transgression cycles than previously appreciated. Second, changes in discharge from an ice stream may affect the flow of an adjacent stream by altering local thickness and surface elevation gradients. Interactions between the downstream reaches of adjacent ice streams may be as important to ice stream discharge variability as the more widely considered production and routing of basal meltwater in their upstream reaches.

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References

Figure 6. Changes in grounded area over course of a model experiment I in which WIS and MacIS experience stagnation events (Table 3 and Figure 4). Dots represent grounded nodes in the FEM domain. Contours are sea floor elevation within the model domain with an interval of 100 m. The heavy contour is at −500 m below sea level. The heavy, light grey lines in the 0 a plot are boundaries of present-day ice plains.


M. Fahnestock, Institute for the Study of Earth, Oceans, and Space, University of New Hampshire, Durham, NH 03824, USA. (mark.fahnestock@unh.edu)

C. Hulbe, Department of Geology, Portland State University, Portland, OR 97207-0751, USA. (chulbe@pdx.edu)