The annual glaciohydrology cycle in the ablation zone of the Greenland ice sheet: Part 2. Observed and modeled ice flow

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ABSTRACT. Ice velocities observed in 2005/06 at three GPS stations along the Sermeq Avannarleq flowline, West Greenland, are used to characterize an observed annual velocity cycle. We attempt to reproduce this annual ice velocity cycle using a 1-D ice-flow model with longitudinal stresses coupled to a 1-D hydrology model that governs an empirical basal sliding rule. Seasonal basal sliding velocity is parameterized as a perturbation of prescribed winter sliding velocity that is proportional to the rate of change of glacier water storage. The coupled model reproduces the broad features of the annual basal sliding cycle observed along this flowline, namely a summer speed-up event followed by a fall slowdown event. We also evaluate the hypothesis that the observed annual velocity cycle is due to the annual calving cycle at the terminus. We demonstrate that the ice acceleration due to a catastrophic calving event takes an order of magnitude longer to reach CU/ETH (‘Swiss’) Camp (46 km upstream of the terminus) than is observed. The seasonal acceleration observed at Swiss Camp is therefore unlikely to be the result of velocity perturbations propagated upstream via longitudinal coupling. Instead we interpret this velocity cycle to reflect the local history of glacier water balance.

1. INTRODUCTION

Ice discharge from marine outlet glaciers is a function of deformational and basal sliding velocities. It has been suggested that relatively small increases in surface ablation may result in disproportionately large increases in ice discharge via basal sliding (Zwally and others, 2002; Bartholomew and others, 2010). Recent interferometric synthetic aperture radar (InSAR) observations have confirmed that an annual velocity cycle is spatially widespread in the marginal ice of West Greenland. This annual velocity cycle is most likely due to seasonal changes in basal sliding velocity (Joughin and others, 2008). Recently, however, some studies have speculated that projected increases in surface water production will likely result in a net decrease in basal sliding velocity due to a transition from relatively inefficient to efficient subglacial drainage (Schoof, 2010; Sundal and others, 2011). This motivates the need to quantitatively address the physical relation between glacier hydrology and basal sliding velocity. We therefore seek a computationally efficient means of reproducing the observed spatial and temporal patterns of basal sliding so that we can ultimately explore the likely response of outlet glacier dynamics to climate change scenarios. Other observations and models indicate that the ice discharge from Greenland’s marine-terminating glaciers is highly sensitive to calving-front perturbations, which are subsequently propagated upstream via longitudinal coupling (Holland and others, 2008; Joughin and others, 2008; Nick and others, 2009). We also explore this alternative mechanism as a cause for the observed annual velocity cycle.

The Sermeq Avannarleq ablation zone in West Greenland exhibits an annual ice velocity cycle similar to that of an alpine glacier. The critical features of this cycle are a summer speed-up event followed by a fall slowdown event (Colgan and others, 2011a). A qualitative comparison of this annual ice velocity cycle to a modeled annual glacier water storage cycle suggests that enhanced (suppressed) basal sliding generally occurs during periods of positive (negative) rates of change of glacier water storage (Colgan and others, 2011a). This notion is consistent with alpine glacier studies that suggest that changes in basal sliding velocity are due to changes in the rate of change of glacier water storage (dS/dt or the difference between rates of glacier water input and output; Kamb and others, 1994; Anderson and others, 2004; Bartholomaus and others, 2008, 2011). In this conceptual model, three general basal sliding states exist: (1) when local meltwater input exceeds the transmission ability of the subglacial hydraulic system (i.e. dS/dt>0 or increasing glacier water storage); (2) when the transmission ability of the subglacial hydraulic system exceeds the local input of meltwater (i.e. dS/dt<0 or decreasing glacier water storage); and (3) when meltwater input and subglacial transmission ability are in approximate equilibrium. In alpine settings, peak basal sliding velocity can be expected when dS/dt reaches a maximum (although there is some evidence that peak sliding velocity exhibits a slight phase-lag behind peak dS/dt values; Bartholomaus and others, 2008). Following this maximum, both dS/dt and basal sliding velocity decrease, and dS/dt becomes negative during the later part of the melt season when efficient conduits can transmit more water than...
is delivered to the englacial and subglacial system through surface melt.

Our goal is to reproduce the annual ice velocity cycle observed in the Sermeq Avannarleq ablation zone with coupled ice-flow and hydrology models. In this paper, we couple a one-dimensional (1-D) (depth-integrated) ice-flow model to a 1-D (depth-integrated) glacier hydrology model (Colgan and others, 2011a) via a semi-empirical sliding rule. Our goal is not to reproduce specific observed intra- or interannual variations in ice velocity. Rather we attempt to reproduce an annual glaciohydrology cycle in dynamic equilibrium that reproduces the critical features of the observed cycle and may thus serve as a basis for future work investigating the influence of interannual variations in surface ablation on annual ice displacement. The 530 km Sermeq Avannarleq flowline runs up-glacier from its tidewater terminus (km 0 at 69.37°N, 50.28°W) to the main ice divide of the Greenland ice sheet (km 530 at 71.54°N, 37.81°W; Fig. 1). This flowline lies within 2 km of three Greenland Climate Network (GC-Net; Steffen and Box, 2001) automatic weather stations (AWSs): JAR2 (km 14.0 at 69.42°N, 50.08°W), JAR1 (km 32.5 at 69.50°N, 49.70°W), and CU/ETH (‘Swiss’) Camp (km 46.0 at 69.56°N, 49.34°W) (all positions reported for 2008). In a companion paper to this study (Colgan and others, 2011a) we suggested that in the Sermeq Avannarleq ablation zone: (1) englacial water table elevation, which may be taken as a proxy for glacier water storage, oscillates around levels that are relatively close to flotation throughout the year; and (2) observed periods of enhanced (suppressed) basal sliding qualitatively correspond to modeled periods of increasing (decreasing) glacier water storage. In this paper, we propose a semi-empirical and site-specific sliding rule that relates variations in the modeled rate of change of glacier water storage to observed variations in basal sliding velocity.

2. METHODS

2.1. Observed annual ice surface velocity cycle
We characterize the annual ice surface velocity cycle at three locations (JAR2, JAR1 and Swiss Camp) along the terminal ~55 km of the Sermeq Avannarleq flowline using differential GPS observations of 10 day mean ice surface velocity in 2005 and 2006 (Larson and others, 2001; Zwally and others, 2002). This characterization provides a representative annual ice velocity cycle against which the accuracy of the modeled annual ice velocity cycle can be assessed. At all sites, these observations reveal that the ice moves at winter velocity until the beginning of a summer speed-up event in which ice velocities increase above winter velocity (Fig. 2). The summer speed-up event is followed by a fall slowdown event in which ice velocities decrease below winter velocities. Using the positive degree-days (PDDs) observed at each station as a proxy for melt intensity (Ohmura, 2001; Steffen and Box, 2001), the onset of the speed-up approximately coincides with the onset of summer
melt and reaches a maximum approximately halfway through the melt season, while the slowdown occurs after the cessation of melt.

We approximate the annual ice surface velocity cycle using two Gaussian curves. These are overlaid on mean winter velocity, with one (positive) curve representing the summer speed-up event and the other (negative) curve representing the fall slowdown event. The use of a bi-Gaussian function allows the amplitude, width and timing of both the summer and fall events to be parameterized independently. Thus, we characterize surface velocity, \( u_s \), as a function of day of year, \( j \), according to:

\[
    u_s = u_w + (u_{\text{max}} - u_w) \exp \left( \frac{j - j_{\text{max}}}{d_{\text{max}}} \right)^2 \\
    - [u_w - (u_{\text{max}} - u_{\text{min}})] \exp \left( \frac{j - j_{\text{min}}}{d_{\text{min}}} \right)^2
\]

where \( u_w \) is the mean winter velocity and \( u_{\text{max}} \) and \( u_{\text{min}} \) are the summer maximum and fall minimum velocities, respectively. The remaining four parameters govern the timing and shape of the speed-up and slowdown curves: \( j_{\text{max}} \) (\( j_{\text{min}} \)) is the day of year of summer maximum (fall minimum) velocity, while \( d_{\text{max}} \) (\( d_{\text{min}} \)) is the duration of the summer (fall) velocity anomaly. All parameters were prescribed manually by visual inspection (Table 1). For \( j_{\text{max}} \) and \( j_{\text{min}} \) we assess an estimated uncertainty equivalent to the temporal resolution of the velocity data (±10 days). This bi-Gaussian characterization was fitted to the aggregated 2005 and 2006 velocity data at each station (Fig. 2).

2.2. Ice-flow model

We apply a longitudinally coupled 1-D (depth-integrated) ice-flow model to the Sermeq Avannarleq flowline. This model solves for the rate of change in ice thickness \( \partial H/\partial t \) according to mass conservation:

\[
    \frac{\partial H}{\partial t} = b - \frac{\partial Q}{\partial x}
\]

where \( b \) is the annual mass balance, \( Q \) is the ice discharge per unit width and \( \partial Q/\partial x \) is the horizontal divergence of ice discharge. To generate dynamic equilibrium ice geometry and velocity fields, the ice-flow model was subjected to a 1000 year spin-up that was initialized with present-day ice geometry and a ‘cooler’ climate with no hydrology cycle. Following the 1000 year spin-up, an annual basal sliding cycle is introduced via the coupled 1-D (depth-integrated) hydrology model. We use a semi-empirical three-phase sliding rule (described in Section 2.3) to convert variations in the rate of change of glacier water storage calculated by the hydrology model into variations in basal sliding velocity.

2.2.1. Annual balance

The annual mass balance of a given ice column is the sum of the annual surface accumulation, \( c_a \), surface ablation, \( a_s \), basal accumulation, \( c_b \), and basal ablation, \( a_b \):

\[
    b = c_a + F_a + c_b + a_b
\]

where \( F \) is the hydrologic system entry fraction based on the ratio of annual surface accumulation to annual surface ablation (Pfeffer and others, 1991; Colgan and others, 2011a). Annual surface accumulation is prescribed as the observed mean annual value over the period 1991–2000 (Burgess and others, 2010). Annual accumulation increases from ~0.25 m at the Sermeq Avannarleq terminus to a maximum (~0.5 m) at ~100 km upstream and decreases again to ~0.25 m at the main flow divide (530 km upstream). In the ablation zone, annual surface ablation, \( a_s \), is taken to be a function of elevation, based on previous observations:

\[
    a_s = \gamma(z_e - z_{ela}) - a_{ela}
\]

where \( \gamma \) is the present-day ablation gradient with elevation \((\Delta a_s/\Delta z_e) \) taken as 0.00372; Fausto and others, 2009), \( z_e \) is the ice surface elevation, \( z_{ela} \) is the equilibrium-line altitude (ELA) and \( a_{ela} \) is the annual surface ablation at the ELA (taken as 0.4 m). The observed regional ELA was ~1125 m over the period 1995–99 (Steffen and Box, 2001) and ~1250 m over the period 1996–2006 (Fausto and others, 2009). We prescribe an ELA of 1125 m, as it is more likely to be consistent with the steady-state surface mass-balance forcing prior to the highly transient post-1990 period. Annual surface ablation is distributed over an annual cycle to yield surface ablation rate \( a_s \) using a sine function to represent the melt season solar insolation history (cf. fig. 4 in Colgan and others, 2011a).

In the ice-flow model, we assume that annual basal accumulation is negligible \((c_b \approx 0 \text{ ma}^{-1})\) and that submarine basal ablation is only significant beneath the floating tidewater tongue (e.g. Rignot and others, 2010). We use the relative magnitudes of ice and englacial water pressures,
\(P_i\) and \(P_w\), respectively, to determine which flowline nodes are grounded \((P_i > P_w)\) or floating \((P_i < P_w)\) in a given time-step. During spin-up and dynamic equilibrium we prescribe a constant submarine ablation rate of \(a_b = 10\, \text{m}^2\, \text{a}^{-1}\) to all floating nodes. This prescribed rate is less than the contemporary submarine ablation rate at Sermeq Avannarleq, which is estimated to exceed 25 m a\(^{-1}\) (Rignot and others, 2010). The ice-flow model, however, does not reproduce a floating tongue when contemporary submarine basal ablation rates are imposed for the duration of spin-up. As our intent is to reproduce the dynamic equilibrium ice geometry that precedes the current rapid transient state, we depart from the present-day submarine basal ablation rate.

### 2.2.2. Ice discharge

We include depth-averaged longitudinal coupling stress, \(\tau_{xx}\), as a perturbation to the gravitational driving stress derived from the shallow-ice approximation (Van der Veen, 1987; Marshall and others, 2005):

\[
\tau = -\rho g H \frac{\partial z}{\partial x} + 2 \frac{\partial}{\partial x} \left( H \tau_{xx} \right)
\]

where \(\tau\) is total driving stress. Depth-averaged longitudinal coupling stress \(\tau_{xx}\) is calculated following the approach outlined by Van der Veen (1987). This formulation derives longitudinal coupling stress by solving a cubic equation describing equilibrium forces independently at each node, based on ice geometry and prescribed basal sliding velocity, \(u_b\):

\[
0 = \tau_{xx} \left\{ 2 \frac{\partial z}{\partial x} \frac{\partial H}{\partial x} - \frac{\partial z}{\partial x} \frac{\partial H}{\partial x} + H \frac{\partial^2 z}{\partial x^2} - \frac{1}{2} \right\}
+ \tau_{xx} \left\{ \frac{2}{3} \frac{\partial H}{\partial x} \frac{\partial z}{\partial x} + \frac{1}{2} H \frac{\partial^2 z}{\partial x^2} - \frac{1}{6} \left( \frac{\partial z}{\partial x} \right)^2 \right\}
+ \cdots \tau_{xx} \left\{ \frac{1}{3} \frac{\partial H}{\partial x} \frac{\partial z}{\partial x} + \frac{1}{4} H \frac{\partial^2 z}{\partial x^2} - \frac{1}{4} \left( \frac{\partial z}{\partial x} \right)^2 \right\}
+ \tau_{xx} \left( \frac{2}{5} \frac{\partial H}{\partial x} \frac{\partial z}{\partial x} + \frac{1}{4} \frac{\partial H}{\partial x} \frac{\partial z}{\partial x} \right) + \frac{1}{2} \frac{\partial u_b}{\partial z}.
\]

The depth-integrated longitudinally coupled ice velocity due to deformation, \(\tau_{xx}\), may be derived from the equation for horizontal shear rate, \(\partial u_d/\partial z\) (e.g. Marshall and others 2005):

\[
\frac{\partial u_d}{\partial z} = 2A \left( \rho g (z_s - z) \frac{\partial z}{\partial x} \right)^{n-1} \left[ -\rho g (z_s - z) \frac{\partial z}{\partial x} + 2 \frac{\partial}{\partial x} \left( \tau_{xx} \right) \right]
\]

where \(\partial z_s/\partial x\) is ice surface slope along the flowline and \(n\) is an exponent of 3 in the empirical relation between stress and strain rate describing ice rheology (Glen, 1955). Integrating Eqn (7) twice in the vertical and dividing by \(H\) yields

\[
u_b = 2A \left( \frac{\rho g H}{n + 2} \left( \rho g H \frac{\partial z}{\partial x} \right)^{n-1} \left[ -\rho g H \frac{\partial z}{\partial x} + 2 \frac{\partial}{\partial x} \left( \tau_{xx} \right) \right] \right) H.
\]

We calculate the flow-law parameter, \(A\), as a function of both ice temperature, \(T\), and thickness, \(H\), using an Arrhenius-type relation (Huybrechts and others, 1991):

\[
A(T, H) = E(H) A_0(T) \exp \left( \frac{-Q_0(T)}{RT} \right)
\]

where \(A_0\) is a coefficient that depends on ice temperature (taken as 5.47 \times 10^{10} \text{ Pa}^{-3} \text{ a}^{-1} when \(T \geq 263.15\, \text{K}\)) and 1.14 \times 10^{-9} \text{ Pa}^{-3} \text{ a}^{-1} when \(T < 263.15\, \text{K}\), \(Q_0\) is the creep activation energy of ice (taken as 139 kJ mol\(^{-1}\) when \(T \geq 263.15\, \text{K}\), and 60 kJ mol\(^{-1}\) when \(T < 263.15\, \text{K}\), \(R\) is the ideal gas constant (8.314 kJ mol\(^{-1}\) K\(^{-1}\)) and \(T\) is the ice temperature. At each flowline node, we use the steady-state ice temperature at 90% depth, derived from independent thermodynamic modeling of the flowline (personal communication from T. Phillips and others, 2011), to calculate the flow-law parameter. The majority of shear occurs at or below this depth. Thus, along-flowline variations in basal ice temperature result in along-flowline variations in the flow-law parameter.

We enhance the flow-law parameter by a factor \(E\), to account for increased deformation due to the presence of relatively soft Wisconsin basal ice. This enhancement factor linearly transitions from its prescribed value where ice thickness exceeds 650 m, to 1 (i.e. no enhancement) where ice thickness is <550 m. At flowline nodes where ice is >650 m thick, Wisconsin ice is expected to comprise a significant portion of the basal ice (Huybrechts, 1994). We assume that the uncertainty associated with the calculated values of \(A\) is small in comparison to uncertainty associated with the Wisconsin flow enhancement factor. We evaluate dynamic equilibrium ice geometry and velocity fields following spin-up with \(E\) ranging between 2 and 4 (Reeh, 1985; Paterson, 1991), as in situ borehole deformation measurements beneath nearby Jakobshavn Isbræ indicate \(E=1\) (Lüthi and others, 2002). Under the \(E=3\) scenario, calculated flow-law parameter values range between 8.2 \times 10^{-18} and 4.3 \times 10^{-16} \text{ Pa}^{-3} \text{ a}^{-1}. For comparison, the recommended unenhanced \((E=1)\) flow-law parameter at 273 K is 2.1 \times 10^{-16} \text{ Pa}^{-3} \text{ a}^{-1} (Paterson, 1994; Fig. 3).

Local ice discharge, \(Q\), is obtained by multiplying the sum of basal sliding and depth-averaged deformational velocities by ice thickness:

\[
Q = (u_b + \tau_{xx}) H.
\]

Basal sliding velocity is prescribed via a semi-empirical sliding rule described in Section 2.3.

Following the approach taken in previous ice-sheets flowline models (e.g. Van der Veen, 1987; Parizek and others 2005):
Alley, 2004), we neglect lateral velocities and stresses stemming from divergence and convergence. While this assumption is likely valid in interior regions of ice sheets, it is less valid near the ice-sheet margin, where substantial divergence and convergence can occur. We acknowledge in Section 3 that the failure to account for possible lateral effects is a potential factor in the systematic overestimation of ice velocities at JAR1 and underestimation of ice velocities at the terminus. Another inherent shortcoming of any 1-D flowline model is the prescription of the ice-flow trajectory. For example, if the flowline length has been underestimated in the accumulation zone, modeled ice discharge across the equilibrium line will be underestimated and the model will require a decrease in surface ablation rate to maintain the observed ice geometry. In addition, the ice-flow trajectory is unlikely to have been constant through time. Striations on the ice surface mapped by Thomsen and others (1988) suggest that the flowline through JAR1 station likely terminated on land just north of Sermeq Avannarleq in 1985, rather than flowing past JAR2 to the tidewater terminus as shown in Figure 1. The recent acceleration of Jakobshavn Isbræ, immediately south of Sermeq Avannarleq, has caused substantial reorientation of ice flow throughout the Sermeq Avannarleq ablation zone (Colgan and others, 2011b). As the flowline used in this study was derived from a 2005/06 InSAR ice surface velocity field (Joughin and others, 2010), obtained after the onset of the reorganization of ice flow (~1997; Colgan and others, 2011b), it does not accurately reflect the long-term ablation zone flowline trajectory.

### 2.3. Three-phase basal sliding rule

The sliding rules employed in glacier models have improved with advances in the conceptualization of basal sliding. Initial sliding rules prescribed basal sliding velocity as proportional to driving stress on the assumption that higher driving stress results in greater till deformation (Weertman, 1957; Kamb, 1970). Observations that subglacial water was capable of enhancing ice velocities by lubricating and pressurizing the subglacial environment led to parameterizations in which basal sliding velocity was taken as proportional to subglacial water pressure (Iken, 1981; Iken and others, 1983), as well as sliding rules that included both effective water pressure ($P - P_w$) and driving stress (Bindschadler, 1983; Iken and Bindschadler, 1986). Recent models have utilized basal sliding rules that are Coulomb friction analogues, whereby basal drag is parameterized to take on a maximum value at low sliding velocities and decrease with decreasing effective pressure and increasing sliding velocity (Schoof, 2005; Gagliardini and others, 2007; Pimentel and Flowers, 2010). To have predictive value, a basal sliding rule should ideally be capable of reproducing observed sliding velocities from first principles of hydrology and force balance with a minimum of free parameters. Here we describe an empirical and site-specific three-phase basal sliding rule that focuses on the hydrologic aspect of basal sliding. We regulate the magnitude and sign of a perturbation to background basal sliding velocities using modeled rates of change of glacier water storage.

The Swiss Camp GPS data indicate that ice velocity is on average ~14% faster during the winter than in the midst of the slowest motion of the year, during the fall slowdown (113 and 99 m a$^{-1}$, respectively; Fig. 2; Table 1). This fall velocity minimum matches very well the ice surface velocity predicted by internal deformation alone. According to the shallow-ice approximation, the ice surface velocity solely due to deformation may be calculated as (Hooke, 2005)

$$\frac{u_d(z)}{n+1} \left( \frac{\partial z}{\partial x} \right) \beta \rho H^{n+1}. \quad (11)$$

Taking ice thickness, $H$, as 950 m, ice surface slope, $\Delta z_i/\Delta x$, as 0.01 and $A$ as $3.33 \times 10^{-16}$ Pa$^{-2}$ a$^{-1}$ (the flow-law parameter for ice at 272 K, which is the pressure-melting point beneath ~1 km of ice, enhanced by a Wisconsin factor of 3), yields an ice surface velocity of 99 m a$^{-1}$.

We interpret this as suggesting that Swiss Camp experiences significant background basal sliding velocity during the winter, which is suppressed during the fall velocity minimum. Thus, at Swiss Camp (and similarly at JAR1 and JAR2), the background basal sliding velocity may be approximated as the difference between observed mean winter ($u_{\text{w}}$) and fall minimum ($u_{\text{min}}$) velocities. We linearly interpolate these basal sliding velocities along the flowline to provide the background basal sliding boundary condition used in spin-up of the ice-flow model (Fig. 4). The year-round persistence of the englacial hydrologic system in the Sermeq Avannarleq ablation zone provides a mechanism capable of maintaining favorable basal sliding conditions year-round (i.e. available liquid water; Catania and Neumann, 2010). After spin-up, we overlay an annual basal sliding velocity cycle on the background basal sliding velocity.

Following theoretical developments in alpine glaciology (Kamb and others, 1994; Anderson and others, 2004; Bartholomaus and others, 2008, 2011), we propose a sliding rule that depends on the sign of the rate of change of glacier water storage to prescribe ‘speed-up’ during periods of increasing glacier water storage and ‘slowdown’ during periods of decreasing glacier water storage. We take rate of change of englacial water table elevation (or head, $\partial h_w/\partial t$) as a surrogate for rate of change of glacier water storage (Colgan and others, 2011a). We formulate a three-phase basal sliding rule that imposes: (1) background basal sliding velocity during the winter when $\partial h_w/\partial t = 0$; (2) enhanced basal sliding during positive rates of change of glacier water storage ($\partial h_w/\partial t > 0$); and (3) suppressed basal sliding during negative rates of change of glacier water storage ($\partial h_w/\partial t < 0$). We accomplish this by conceptualizing $u_b$ as the sum of...
background basal sliding velocity, \( u_{b,o} \), and a perturbation, \( \Delta u_b \):
\[
 u_b = u_{b,o} + \Delta u_b \quad \text{limit: } u_b \geq 0 \quad (12)
\]

The alpine glaciohydrology literature suggests that we may expect basal sliding velocity to scale nonlinearly with the rate of change of glacier water storage (i.e. \( \Delta u_b \propto \partial h_e / \partial t \), where \( m \geq 1 \); Anderson and others, 2004; Bartholomaus and others, 2008). We impose \( m = 3 \) and express the perturbation to the background basal velocity as

\[
\Delta u_b = \left\{ \begin{array}{ll}
 n_c(x) k(x) \left( \frac{\partial h_e}{\partial t} \right)^m \text{sign} \left( \frac{\partial h_e}{\partial t} \right) & \text{if } \frac{\partial h_e}{\partial t} \geq 0.25 \text{ m d}^{-1} \\
 0 & \text{if } \frac{\partial h_e}{\partial t} < 0.25 \text{ m d}^{-1}
\end{array} \right. \quad (13)
\]

where \( n_c(x) \) is the number of subglacial conduits \( m^{-1} \) in the across-flow direction (Colgan and others, 2011a) and \( k(x) \) is a tunable site-specific sliding coefficient. We impose an arbitrary threshold of 0.25 m d\(^{-1}\) (constant in time and space) that rates of change of glacier water storage must exceed in order to either enhance or suppress background basal sliding. We tune the sliding coefficient values (which exceed in order to either enhance or suppress background basal sliding velocity). By parameterizing \( \Delta u_b \) so that it goes to zero when \( |\partial h_e / \partial t| \) falls below a critical threshold (taken as 0.25 m d\(^{-1}\)), Eqn (13) provides the framework for three phases of basal sliding.

### 2.4. Input datasets and boundary conditions

Following Colgan and others (2011a), the ice-flow model is initialized with observed ice surface and bedrock topography (Bamber and others, 2001; Scambos and Haran, 2002; Plummer and others, 2008). During spin-up, the observed surface mass balance was adjusted by decreasing surface ablation in order to reproduce the observed present-day ice geometry. We evaluate dynamic equilibrium ice geometry and velocity fields following spin-ups by decreasing surface ablation by a factor of 25–75%. Previous Greenland ice sheet modeling studies have implemented similar surface mass-balance corrections during spin-up in order to achieve equilibrium present-day ice geometries (i.e. Huybrechts, 1994; Ritz and others, 1997; Parizek and Alley, 2004). This adjustment is typically justified by the notion that the present-day Greenland ice sheet geometry reflects colder ‘glacial’ conditions (i.e. less surface ablation with no change in accumulation). Both the observed surface accumulation (Burgess and others, 2010) and ablation (Fausto and others, 2009) datasets have been validated by in situ observations, including the GC-Net AWSs along the Sermeq Avannarleq flowline (Steffen and Box, 2001; Fig. 1).

The differential equations describing the evolution of ice thickness (Eqs (2–10)) were discretized in space using first-order finite volume methods with grid spacing \( \Delta x = 500 \text{ m} \). The semi-discrete set of coupled ordinary differential equations at the computational nodes was then solved using ‘ode15s’, the stiff differential equation solver in MATLAB R2008b. During the 1000-year spin-up, the ice-flow model was solved with a time-step, \( \Delta t \), of 10 years. Following spin-up, the ice-flow model was solved concurrently with the hydrology model with a 2 day time-step. Post-spin-up, basal sliding velocity was calculated according to the three-phase sliding rule described above. We apply a second-type (zero flux) Neumann boundary condition at the main ice-flow divide as the upstream boundary condition (\( \partial z / \partial x = 0 \) and \( Q = 0 \) at \( x = 530 \text{ km} \)). When the glacier terminus reaches the downstream boundary of the model domain (km0), the terminal node ice discharge (\( Q_{term} \) or calving flux) is calculated as the difference between the ice discharge of the adjacent upstream node and the annual mass balance (negative in the ablation zone) of the terminal node (\( Q_{term} = Q_{term-1} + b \Delta x \) at \( x = 0 \text{ km} \)). When the glacier terminus does not reach the downstream boundary of the model domain, no calving flux is imposed, as a dynamic equilibrium has been achieved in which ice inflow across the grounding line is balanced by total ablation (both surface and submarine) in the floating ice tongue.

The 1-D (depth-integrated) hydrology model tracks glacier water storage and discharge through time. Glacier water input is prescribed based on observed ablation rates, whereas glacier water output occurs through conduit discharge. Conduit discharge varies in response to the dynamic evolution of conduit radius. When coupled, the ice-flow and hydrology models receive the same surface ablation forcing at each time-step. The ice-flow model updates ice geometry used by the hydrology model each time-step, while the hydrology model updates the subglacial
3. RESULTS

The GPS-observed ice velocities provide insight into the relative importance of basal sliding in the Sermeq Avannarleq ablation zone (Fig. 2). The net annual displacements observed at JAR2, JAR1 and Swiss Camp are about 114, 69 and 115 m, respectively. The fall minimum velocities, taken to represent purely deformational velocities (i.e. no basal sliding), observed at each of these stations are about 80, 59 and 101 m a\(^{-1}\), respectively. Therefore, basal sliding appears to be responsible for about 30, 14 and 12% of the annual net displacement, respectively. These observations suggest that summer speed-up events are responsible for about 13, 4 and 4 m of net displacement at JAR2, JAR1 and Swiss Camp, respectively, equivalent to about 11, 6 and 3% of the annual displacement. These increases in net displacement due to summer speed-up events are partially offset by decreases in net displacement due to fall slowdown events. The fall slowdown events suppress annual displacement by about 1, 2 and 2 m at JAR2, JAR1 and Swiss Camp, respectively, equivalent to about 4, 1 and 2% of the annual displacement. Thus, at Sermeq Avannarleq, year-round (or ‘background’) basal sliding appears to contribute a larger fraction of annual net displacement than seasonal basal sliding.

We compared modeled dynamic equilibrium ice geometry and velocity fields of the ice-flow model with observed ice surface elevation (Scambos and Haran, 2002) and velocity (Joughin and others, 2008) following a 1000 year spin-up. We explored a range of Wisconsin enhancement factors (2–4) and surface ablation perturbations (decreases of 25–75% below contemporary rates). While this parameter space produces a relatively narrow range of ice surface geometries that are similar to the observed profile, it produces a relatively wide range of ice velocities that can differ substantially from the observed profile (Fig. 6). A comparison between modeled mean ice surface elevation and velocity versus observed mean ice surface elevation (688 m) and velocity (116 m a\(^{-1}\)) along the terminal 50 km of the flowline suggests that observations are most accurately reproduced with a 50% decrease in surface ablation below contemporary rates and a Wisconsin enhancement factor of \(E = 3\) (Table 3). This depression in contemporary surface ablation may represent a combination of: (1) an underestimation of historical accumulation rate; (2) the increase in surface ablation that has occurred since the termination of the last glaciation; and (3) error in our delineation of the flowline.

Under all spin-up scenarios, the ice-flow model overestimates ice velocities in the km 25–35 portion of the flowline and underestimates ice velocities in the terminal 5 km. We speculate that both systematic errors are artifacts of the 1-D (flowline) character of the ice-flow model. A 1-D model is expected to overestimate ice discharge (and hence ice velocity) in reaches of divergent ice flow. Thus, the overestimation of ice velocities in the vicinity of JAR1 may suggest that divergent ice flow is occurring in this region. Similarly, a 1-D model is expected to underestimate ice velocities in regions of convergent ice flow. Failure to account for convergent flow at the terminus has likely contributed to the severe underestimation of terminus ice

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**Table 2.** Specific parameterization of the 1-D (depth-integrated) hydrology model (notation follows Colgan and others, 2011a)

<table>
<thead>
<tr>
<th>Variable</th>
<th>Definition</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>(\alpha)</td>
<td>Glacier hydrology length scale</td>
<td>20 km</td>
</tr>
<tr>
<td>(\phi)</td>
<td>Bulk ice porosity</td>
<td>0.01</td>
</tr>
<tr>
<td>(F_r)</td>
<td>Firn meltwater retention fraction</td>
<td>0.5</td>
</tr>
<tr>
<td>(Q_g)</td>
<td>Geothermal flux</td>
<td>57 mW m(^{-2})</td>
</tr>
<tr>
<td>(f)</td>
<td>Conduit friction factor</td>
<td>0.05</td>
</tr>
<tr>
<td>(r_{\text{term}})</td>
<td>Conduit spacing at terminus</td>
<td>0.005 m(^{-1})</td>
</tr>
<tr>
<td>(r_{\text{term}}^\text{max})</td>
<td>Maximum conduit radius at terminus</td>
<td>2 m</td>
</tr>
</tbody>
</table>

---

Fig. 6. (a) Modeled ice surface elevation, \(z_s\), and (b) modeled ice surface velocity, \(u_s\), over a range of Wisconsin enhancement factors, \(E\), and fractional contemporary surface ablation values, \(a_s\). Observed ice surface elevation (Scambos and Haran, 2002) and velocity (Joughin and others, 2008) are shown for comparison. The basal boundary conditions (BC) for ice surface elevation and velocity are observed bedrock elevation and prescribed background basal sliding velocity. Vertical dashed lines denote the locations of JAR2, JAR1 and Swiss Camp (SC).
thickness. Observations suggest the bedrock overdeepening at the terminus contains grounded ice ~600 m thick (Plummer and others, 2008) while the modeled terminus is floating with ice thicknesses of ~300 m.

Generally, however, the ice-flow model achieves a dynamic equilibrium ice geometry and velocity field that closely matches observations throughout the Sermeq Avannarleq ablation zone (i.e. terminal 60 km of the flowline). This suggests that the basal sliding boundary condition imposed during the 1000 year spin-up (i.e. 'background basal sliding'; Fig. 4) captures the essence of the basal sliding profile beneath the terminal ~60 km of the flowline. When we allow the annual hydrologic cycle to operate, producing seasonal variations in basal sliding around this background basal sliding profile, the three-phase basal sliding rule produces a reasonable annual basal sliding velocity cycle throughout the Sermeq Avannarleq ablation zone (Animation 1). While the absolute magnitude and seasonal timing of modeled maximum and minimum velocities do not precisely match observations (discussed in Section 4.1), the coupled model produces an annual basal sliding cycle that captures the essence of the observed annual velocity cycle, namely: (1) prescribed background basal sliding velocity during the winter, (2) a summer speed-up event, followed by (3) a fall slowdown event.

4. DISCUSSION
We developed a representation of basal sliding velocity that depends on the local rate of change of glacier water storage. Incorporating this basal boundary condition in a 1-D ice-flow model captures the essence of the annual velocity cycle of the Sermeq Avannarleq flowline (i.e. summer speed-up and fall slowdown events). The mean annual glaciohydrology cycle reproduced by this coupled model may serve as a basis for future investigations into the transient response of the Sermeq Avannarleq flowline to predicted increases in surface meltwater production. While we do not drive our basal sliding rule with absolute head values, $h_0$, but rather with the rate of change of head, $\partial h_0/\partial t$, we achieve a

<table>
<thead>
<tr>
<th>$E$</th>
<th>$0.25 a_s$</th>
<th>$0.50 a_s$</th>
<th>$0.75 a_s$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\delta z_s$</td>
<td>$\delta u_s$</td>
<td>$\delta z_s$</td>
</tr>
<tr>
<td>2</td>
<td>m</td>
<td>m a$^{-1}$</td>
<td>m</td>
</tr>
<tr>
<td>3</td>
<td>32 (5%)</td>
<td>-11 (-9%)</td>
<td>2 (0%)</td>
</tr>
<tr>
<td>4</td>
<td>48 (7%)</td>
<td>17 (15%)</td>
<td>30 (4%)</td>
</tr>
<tr>
<td>5</td>
<td>59 (9%)</td>
<td>40 (34%)</td>
<td>45 (7%)</td>
</tr>
</tbody>
</table>
satisfactory annual basal sliding cycle (Fig. 7). We interpret this as suggesting that the $\frac{\partial h_e}{\partial t}$ term contains important information that modulates basal sliding. This suggests that sliding rules that are dependent on basal stress and absolute head may overlook important variables that are correlated with $\frac{\partial h_e}{\partial t}$, particularly those related to transient subglacial transmissivity or higher-order features of subglacial hydrologic geometry (e.g. the $n_n(x)$ ‘conduit spacing’ parameter used in this study). Recent observations of hysteresis in the ratio of inferred subglacial water storage to sliding velocity over the course of a melt season support this notion (Howat and others, 2008a). Thus, we suggest that an improvement towards achieving a physically based sliding rule would be to combine an absolute head/basal stress rule that modulates background (i.e., winter) sliding velocities (which are prescribed in this study) with a $\frac{\partial h_e}{\partial t}$-type parameterization that honors the empirical subtlety that sliding reaches a maximum when $\frac{\partial h_e}{\partial t}$ reaches a maximum (rather than when $h_e$ reaches a maximum).

4.1. Modeled speed-up and slowdown events

There are discrepancies in the absolute magnitude and timing between modeled and observed velocity maxima and minima at all three stations (Fig. 8). As our interest lies in producing an annual velocity cycle using an annual hydrologic cycle, we focus the following discussion primarily on temporal discrepancies. We speculate that the relatively damped annual velocity cycle observed at JAR1 is due to a relatively damped annual hydrology cycle. While the hydrology model forces subglacial water to flow over the bedrock high at km 30, in reality subglacial water likely flows around this bedrock high in the y direction. Thus, the hydrology model likely overestimates $\frac{\partial h_e}{\partial t}$ values along the km 25–40 portion of the flowline, for which the scaling parameter $k(x)$ compensates. In regard to the absolute magnitude of the annual cycle, we note that at any location along the flowline an infinite combination of the parameters $k(x)$ and $m$ may be used to scale $\frac{\partial h_e}{\partial t}$ to a desired value. We also note that the modeled basal sliding velocity cycles exhibit abrupt transitions between background sliding and speed-up and slowdown events. These abrupt transitions are due to the imposition of a relatively high threshold of rate of change of glacier water storage to transition between the three sliding phases ($|\frac{\partial h_e}{\partial t}| > 0.25 \text{ m d}^{-1}$). In reality, this threshold likely varies in time and space.

We take uncertainty in the timing discrepancy as equivalent to the temporal resolution of the velocity data (i.e. ±10 days). Modeled maximum velocity precedes observed maximum velocity by ~20 ± 10 days at Swiss Camp and ~18 ± 10 days at JAR1 (Fig. 8). It is more difficult to assess the timing discrepancy of the summer speed-up event at JAR2, where an apparently premature onset of the modeled slowdown event truncates the speed-up event, but the discrepancy appears to be <10 ± 10 days. The relatively early onset of the modeled speed-up event can be attributed to a combination of: (1) incorrect timing of the prescribed meltwater input and (2) the assumption embedded in the hydrology model that meltwater produced at the surface is immediately routed to the glacier interior, where it raises the englacial water table. This second assumption assumes that there is no temporary supraglacial meltwater storage (e.g. within a saturated snowpack or in ponded water) and that supra- and englacial travel times are negligible. In reality, temporary supraglacial meltwater storage and travel time can delay the initial meltwater pulse from reaching the englacial water table for several weeks after the onset of melt (Fountain and Walder, 1998; Flowers and Clarke, 2002; Jansson and others, 2003). This lag would be expected to be greater near the equilibrium line (i.e. at Swiss Camp) than at the terminus (i.e. at JAR2). In addition to temporary supraglacial meltwater storage, englacial transfer time can be expected to range over two orders of magnitude as a consequence of the morphology of the englacial hydrologic system (i.e. moulin- versus crevasse-type drainage; Colgan and others, 2011b; McGrath and others, 2011). A more detailed treatment of supra- and englacial meltwater routing in the hydrology model would likely reduce the discrepancies between modeled and observed velocity maxima.

The discrepancy between modeled and observed velocity minima is more variable; the model produces a significantly earlier onset at JAR2 (~65 ± 10 days), an extreme delay at Swiss Camp (~110 ± 10 days) and a reasonable match at JAR1 (15 ± 10 days; Fig. 8). We believe that this reflects the difficulty in capturing the behavior of subglacial conduits in an along-flowline 1-D hydrology model. The timing of the fall slowdown event depends on the timing of the negative (or
decreasing) glacier water storage (\( \partial h_m / \partial t \)) phase. This phase is initiated by the upstream propagation of a nickpoint in englacial water table elevation that corresponds to the opening of efficient subglacial conduits that in turn lower glacier water storage (Animation 1; e.g. Kessler and Anderson, 2004). A two-dimensional (2-D) (xy) hydrological model, which allows water to flow both parallel and perpendicular to the ice dynamic flowline, would be inherently more realistic in propagating changes in englacial water table elevation upstream from various outlets at the ice margin. Because the timing of the slowdown event is controlled by the evolution of conduit sizes, the fall slowdown event is more dependent on an accurate representation of subglacial conduit dynamics than is the spring speed-up event. The spring speed-up event initiates as long as conduits have collapsed to their minimum radii (i.e. 'closed'), which is a relatively simple geometry to capture given the lengthy winter period over which this geometry is achieved.

**4.2. Upstream limit of seasonal basal sliding**

The observed dates of summer maximum \( (j_{\text{max}} \text{; red}) \) and fall minimum \( (j_{\text{min}} \text{; blue}) \) ice surface velocity at the GC-Net stations versus distance upstream. Vertical whiskers denote ±10 days uncertainty in \( j_{\text{max}} \) and \( j_{\text{min}} \) at each station. Solid lines denote least-squares linear fit. Dashed lines denote 95% confidence bounds for the linear fit. The grey shading denotes the 95% confidence envelope for the convergence of \( j_{\text{max}} \) and \( j_{\text{min}} \). The black dot denotes day 223 and upstream distance km 69.5.

that this occurs at ~69.5 km upstream on day of year ~223. This distance upstream corresponds to the ~1350 m ice elevation contour, which is slightly above the regional ELA over the period 1996–2006 (~1250 m; Fausto and others, 2009). The lower limit of this convergence position, delineated by the 95% confidence envelope, is ~55 km upstream. GPS velocity observations at Up50 (km 98.5 at 69.75° N, 48.14° W in 2008), which show no evidence of an annual velocity cycle (Colgan and others, 2009), provide an upstream limit for the annual velocity cycle.

**4.3. Importance of longitudinal coupling**

Although in situ GPS data indicate that Swiss Camp exhibits an annual velocity cycle, these data cannot indicate whether this is due to: (1) local meltwater-induced acceleration (Zwally and others, 2002; Bartholomew and others, 2010) or (2) an annual velocity cycle originating downstream of Swiss Camp that is propagated upstream via longitudinal coupling. This downstream annual velocity cycle may be due to either lower-elevation meltwater-induced acceleration (Price and others, 2008) or the annual tidewater calving cycle (Howat and others, 2008b; Joughin and others, 2008). A previous study has suggested that a 10–20% velocity increase at Swiss Camp can be achieved by a roughly 100% velocity increase initiated ~12 km downstream from Swiss Camp that is propagated upstream through longitudinal coupling (Price and others, 2008). This previous study used a 2-D (cross-sectional) ice-flow model in which basal sliding was parameterized to occur through deformation of a fluid layer several meters thick underlying the ice. This basal fluid layer was assumed to have an effective viscosity of ~2.8 × 10^8 Pa a. For comparison, typical flow-law parameter values (i.e. \( A=10^{-10} \text{Pa}^{-1} \text{a}^{-1} \)) represent an effective viscosity of ~2.5 × 10^8 Pa a when \( H=500 \text{m}, \beta z_r/\partial x=0.01 \text{ and } n=3 \). In addition to a vertical structure that is conducive to transmitting longitudinal coupling, the magnitude of the assumed downstream seasonal velocity perturbation (i.e. 100% or doubling) greatly exceeds that recorded by JAR1 station 13.5 km downstream of Swiss Camp.

In the present study, the 1-D (depth-integrated) ice-flow model suggests that the absolute longitudinal coupling stress, \( \left| \sigma_{xz} \right| \text{ (Eqn (6))} \) is only significant (defined here as >10% of total driving stress) along the terminal ~6 km of the flowline (Fig. 10). Upstream of the icefall at ~6 km, where observed ice thickness decreases to <300 m, the absolute longitudinal...
coupling stresses seldom exceed 10% of total driving stress.
In addition, any perturbation to the tidewater tongue would also experience rapid radial diffusion (in the x-y plane) with distance inland. This inference of minimal inland coupling stresses fits the theoretical notion that coupling stresses are typically important only in the terminal few km of ice-sheet flowlines (where ice thickness becomes small). In these terminal regions the magnitude of coupling stresses can be equivalent to, or exceed, driving stresses (Van der Veen, 1987). Where longitudinal coupling stresses are insignificant along the Sermeq Avannarleq flowline, the forces governing ice flow can be assumed to be local in nature.

The inability of a seasonal terminus perturbation (cf. Howat and others, 2008b; Joughin and others, 2008) to propagate upstream to produce the ~55% seasonal velocity acceleration observed at Swiss Camp (from $u_a = 113 \text{ m a}^{-1}$ to $u_{max} = 175 \text{ m a}^{-1}$) can also be demonstrated by a simple numerical simulation. In this fully transient simulation, we spin-up the ice-flow model for 1000 years under the $E=3$ and 50% of contemporary $a_s$ scenario. To isolate potential changes in ice dynamics, we disable hydrological coupling and prescribe temporally constant background basal sliding along the flowline. At 1 year after spin-up, we impose a first-type (specified head) Dirichlet boundary condition of $H=0 \text{ m}$ at km 8. This boundary condition, which instantly removes the terminal several km of the flowline, is meant to represent a catastrophic terminus perturbation. Despite imposing an unprecedented terminus perturbation that is an order of magnitude larger than the annual advance/retreat cycle, longitudinal coupling stresses inland of JAR2 (km 14) fail to exhibit any significant change in the 4 years following the perturbation (Animation 2). The ice acceleration is propagated up to Swiss Camp (km 46) not by longitudinal stresses but by changes in ice geometry (i.e. surface slope steepening) over ~2.75 years (Fig. 11). The modeled upstream propagation rate (~17 km a$^{-1}$) is over an order of magnitude slower than the upstream propagation rate of the summer speed-up event inferred from the GPS observations (~1.2 km d$^{-1}$). In addition, upon reaching Swiss Camp the modeled velocity perturbation is only ~5% of the modeled dynamic equilibrium winter velocity at Swiss Camp, which is an order of magnitude less than the observed ~55% seasonal acceleration. Thus, it is unlikely that terminus perturbations associated with the annual retreat/advance cycle (Howat and others, 2008b; Joughin and others, 2008) are propagated upstream by longitudinal coupling (Price and others, 2008) to produce the seasonal velocity variations observed inland of JAR2 (km 14). Instead, we interpret the upstream propagation rate (~17 km a$^{-1}$) is over an order of magnitude slower than the upstream propagation rate of the summer speed-up event inferred from the GPS observations (~1.2 km d$^{-1}$). In addition, upon reaching Swiss Camp the modeled velocity perturbation is only ~5% of the modeled dynamic equilibrium winter velocity at Swiss Camp, which is an order of magnitude less than the observed ~55% seasonal acceleration. Thus, it is unlikely that terminus perturbations associated with the annual retreat/advance cycle (Howat and others, 2008b; Joughin and others, 2008) are propagated upstream by longitudinal coupling (Price and others, 2008) to produce the seasonal velocity variations observed inland of JAR2 (km 14). Instead, we interpret the
annual velocity cycle at upstream sites to reflect local mismatches between glacier water inputs and outputs (i.e., local melt-induced acceleration; Zwally and others, 2002; Kessler and Anderson, 2004).

5. SUMMARY REMARKS
We examined the annual glaciohydrology cycle in the ablation zone of the Sermeq Avannarleq flowline. We coupled a 1-D (depth-integrated) ice-flow model to a previously published 1-D (depth-integrated) hydrology model via a semi-empirical and site-specific basal sliding rule. This sliding rule imposes seasonal perturbations to the background basal sliding velocity that are dependent on rate of change of glacier water storage. Following a 1000 year spin-up, the ice-flow model produces dynamic equilibrium ice geometry and velocity fields that compare well with observations. After spin-up, the coupled model reproduces the broad features observed in the annual basal sliding cycle in the terminal 60 km of the flowline, namely: (1) prescribed background basal sliding during the winter; (2) a summer speed-up event; and (3) a fall slowdown event and return to winter velocities. While we have put forth a plausible rule that connects sliding velocity to the state of the glacier hydrologic system, there remains a significant challenge to develop a sliding rule that is more firmly based upon first principles (i.e., no free parameters and based on physical mechanisms by which basal water produces sliding). This requires both a proper characterization of all components of the glacier hydrologic system, from the spatial-temporal pattern of melt generation to the complete evolution of englacial and subglacial transport and storage, and a physics-based connection between the state of the hydrologic system and the basal sliding.

GPS observations of ice surface velocity at three stations during 2005 and 2006 suggest the annual velocity cycle propagates as far upstream as ~70 km. We examined the relative magnitude of driving and coupling stresses and performed a simple simulation to assess the possible contribution of an upstream propagation of a terminus perturbation through longitudinal coupling to the annual velocity cycle. We find that an extreme terminus perturbation is unlikely to influence velocities upstream of JAR2 (km 14) on the seasonal timescale. Thus, we suggest the annual ice velocity cycle along the majority of the flowline (km 14–70) is instead driven by variations in background basal sliding during the winter; (2) a summer speed-up event; and (3) a fall slowdown event and return to winter velocities. While we have put forth a plausible rule that connects sliding velocity to the state of the glacier hydrologic system, there remains a significant challenge to develop a sliding rule that is more firmly based upon first principles (i.e., no free parameters and based on physical mechanisms by which basal water produces sliding). This requires both a proper characterization of all components of the glacier hydrologic system, from the spatial-temporal pattern of melt generation to the complete evolution of englacial and subglacial transport and storage, and a physics-based connection between the state of the hydrologic system and the basal sliding.

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REFERENCES


**APPENDIX**

**Variable notation**

\( \gamma \) Ablation gradient

\( \rho_i \) Density of ice (kg m\(^{-3}\))

\( \tau \) Total driving stress (Pa)

\( \tau_{xx} \) Longitudinal stress (Pa)

\( A \) Flow-law parameter (Pa\(^{-3}\) a\(^{-1}\))

\( E \) Wisconsin enhancement factor

\( F \) Englacial hydrology system entry fraction

\( H \) Ice thickness (m)

\( K \) Sliding rule coefficient (m\(^{1/3}\) d\(^{2/3}\))

\( N \) Glen law exponent

\( P_i \) Ice pressure (Pa)

\( P_{aw} \) Water pressure (Pa)

\( Q \) Ice discharge per unit width (m\(^2\) a\(^{-1}\))

\( Q_c \) Creep activation energy of ice (kJ mol\(^{-1}\))

\( R \) Ideal gas constant (J mol\(^{-1}\) K\(^{-1}\))

\( T \) Ice temperature (K)

\( a_b \) Annual basal ablation (m)

\( a_s \) Annual surface ablation (m)

\( a \) Annual mass balance (m)

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*Note: The above text is a modified version of the original content for clarity and to improve readability.*
Annual basal accumulation (m)  
cs Annual surface accumulation (m)  
dmax Duration of summer speed-up event (days)  
dmin Duration of fall slowdown event (days)  
g Gravitational acceleration (m s⁻²)  
h₀ Englacial water table elevation (or head) (m)  
j Day of year (days)  
jmax Day of year of summer maximum velocity (days)  
jen Day of year of fall minimum velocity (days)  
m Sliding rule exponent  
n₀ Subglacial conduits per meter (m⁻¹)  
t Given time (years)  
u₀ Basal sliding velocity (m a⁻¹)  
u₀o Background basal sliding velocity (m a⁻¹)  
u₀d Deformational velocity (m a⁻¹)  
u₀max Ice surface summer maximum velocity (m a⁻¹)  
u₀min Ice surface fall minimum velocity (m a⁻¹)  
u₀s Ice surface velocity (m a⁻¹)  
u₀w Ice surface winter velocity (m a⁻¹)  
x Given distance upstream (m)  
z Given elevation (m)  
z₀ Bedrock elevation (m)  
zₛ Ice surface elevation (m)

All heads and elevations are in reference to sea level as datum.

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