Flow Dynamics of a Soft-Bedded Glacier in Southeast Iceland During Basal Sliding Events

Thesis

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Abstract

The purpose of this study is to determine how glacier motion and stresses vary spatially and temporally in order to clarify weaknesses in current understanding of soft-bedded glacier motion using data collected from Breiðamerkurjökull, Iceland. The dynamics of ice motion are the most substantial source of uncertainty in current models of future ice sheet mass-loss and resulting sea level rise. Currently, there is a general lack of quantitative understanding of how glacial basal conditions, such as the hydrology and till rheology at the bed, control ice motion. This study focuses on the examination of high spatial and temporal resolution surface velocities retrieved from a 12-station GPS grid in the melt seasons of 2009 and 2010 to evaluate the variation of glacial motion and strain rates over time on Breiðamerkurjökull. The first specific objective is to identify any short-term velocity variations. The second is to use the surface motion data to calculate strain rates and other components of the force budget. The third objective is to explain the variations in velocity and force budget components while taking into account glaciomorphic features of the bed. Results reveal five distinct periods of increased surface motion, termed sliding events, corresponding to periods of rainfall and/or increased temperatures during the 2009 and 2010 melt seasons. Along-flow strain rates show extension upglacier and compression downglacier during sliding events. The force budget solution indicates that upglacier, basal drag decreases substantially during speed-up events and cannot resist the local driving stress, most likely indicating pressurization
of a distributed subglacial drainage system. The excess driving stress is then transferred
downglacier, through gradients in longitudinal stress, to a more efficiently draining
terminus where water pressures are lower and basal drag is sufficient to support the
excess stress. The results demonstrate that the till at the terminus accommodates the
excess stress, possibly through extensive grain bridging and dilatant hardening or by a
relocation of stress to bedrock bumps during sliding events. This buttressing role of the
till-bedded margin in resisting increased upglacier sliding, likely over bedrock, is novel
and counter to the prevailing view of soft beds, with implications for simulating the
evolution of past and current ice masses.
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1. Introduction

Glacier motion is largely controlled by basal conditions and processes. If the bed is composed of till, basal motion can occur through both sliding along the ice-till interface and sediment deformation. These two processes are highly sensitive to the subglacial hydrologic conditions. An increase in subglacial water pressure can increase the rate of basal motion by decoupling the ice from the bed, a process called cavitation, and weaken basal sediments. Understanding these processes are essential for assessing the stability of soft-bedded glaciers and ice sheets. Studies of the West Antarctic ice streams demonstrate that high ice flow speeds are caused by weak basal till (Alley et al., 1986; Kamb, 1991; Tulaczyk et al., 2000b). The mass balance of the West Antarctic Ice Sheet, which could greatly contribute to future global sea level rise, depends primarily on these fast moving ice streams (Payne et al., 2004; Tulacyzk et al., 2000b). Similarly, short-term velocity variations on glaciers in Alaska, such as Black Rapids Glacier, have been attributed to high water pressures and changing basal stress conditions (Amundson et al., 2006; Truffer et al., 2000, 2001). Further, recent geophysical evidence suggests that soft sediments underlie Greenland’s largest, rapidly retreating Jakobshavn Isbrae (Block and Bell, 2011). Therefore, it is important to determine a definitive relationship among subglacial hydrology, till rheology, and basal motion.

Numerous past studies have documented enhanced sliding due to meltwater input, mainly from surface melting and precipitation events (e.g., Anderson et al., 2004; Mair et
al, 2001; Sugiyama and Gudmundsson, 2004; Willis, 1995). A common interpretation is that such increases in ice speed are caused by increased pressurization of subglacial drainage systems from melt or rainwater input. Increased pressurization causes the glacier to separate from the bed, resulting in less basal friction and increased ice flow. Nevertheless, the mechanics underlying basal sliding are still poorly understood and observations have not been consistent (Fountain and Walder, 1998; Howat et al., 2008). While various sliding laws (i.e. laws linking basal motion, hydrologic conditions and bed characteristics) have been proposed, there is currently no well-constrained sliding law for modeling future sliding events, particularly over timescales of individual hydrologic events (Fountain and Walder, 1998; Harper et al., 2007; Hooke, 1998).

The initial basal sliding law was based on regelation and plastic flow of the ice over control volumes on an inclined plane (Weertman, 1957). From this, a variety of sliding theories were established using a plane with a continuous range of different sized obstacles or with a sinusoidal bed (Kamb, 1970; Nye, 1969; Nye, 1970). These early theories established that the motion of basal ice due to regelation and creep is directly related to the basal shear stress to a power and inversely related to the fourth power of roughness (Weertman, 1957). However, the role of subglacial water pressure in controlling the distribution of stress on the bed is not included in these early attempts and, consequently, intra-annual variations cannot be explained by these relationships alone (Willis, 1995). More recent sliding theories take into account cavity formation as a contributor to basal motion. As water pressure increases to a critical fraction of the overburden pressure, cavities form on the lee sides of obstacles and the area of glacier in contact with the bed decreases. This results in decreased basal drag. Lliboutry (1968) was
the first to incorporate the role of subglacial effective pressure (the difference between ice overburden pressure and subglacial water pressure) into a basal sliding law. The current standard sliding law is of similar form and demonstrates that the motion of basal ice is directly related to the basal shear stress and inversely proportional to the effective pressure and a roughness constant, each through a nonlinear power function (Bindschadler, 1983; Willis, 1995).

To validate this sliding law, surface motion has been measured together with subglacial water pressure for numerous glaciers on timescales ranging from hours to months. Seasonal-scale results show ice flow generally increasing with water pressure and the speed peaking with pressure (Hooke et al., 1989; Iken and Bindschadler, 1986; Jansson, 1995). Despite this correlation, studies focusing on shorter timescales associated with sliding generated by intense rainfall, melting, or lake drainage, found an inconsistent temporal relationship between velocity and basal water pressure variations (Fudge et al., 2009; Harper et al., 2002, 2005, 2007; Iken, 1981; Iken and Truffer, 1997; Kamb et al., 1994). For example, Kamb and others (1994) discovered that velocity variations in nondiurnal speed-up events on timescales of several days on Columbia Glacier correlated well with water storage in the glacier instead of maximum pressure. Based on this hypothesis, increased ice motion should be accompanied by uplift (vertical ice motion) from increased subglacial cavity formation and water storage, which has been observed in several studies (Anderson et al., 2004; Iken and Bindschadler, 1986; Sugiyama and Gudmundsson, 2004). However, to calculate bed separation, vertical strain and bed-parallel motion must be subtracted from total uplift. These difficulties have inhibited the establishment of a clear and consistent relationship between bed separation and sliding
speed. A second important conclusion from Kamb and others (1994) is that the influence of water pressure on basal sliding is averaged over a large area of the bed while fluctuations in water pressure can occur within much smaller local subareas due to changes in the conduit system. Therefore, measurements from a single borehole may not be representative of the total water pressure and cannot predict basal sliding. Additionally, Iken (1981) emphasized that sliding velocity is not solely a function of subglacial water pressure or the size of the cavities but that it depends on both variables. The results of this study suggest that the maximum sliding speed occurs, not at the highest pressure, but at the beginning of cavity growth, indicating that the evolution of the subglacial drainage system plays a major role in basal sliding (Schoof, 2010).

As suggested by theory and numerous observations, the subglacial drainage system generally evolves from a distributed (inefficient) to channelized (efficient) system over the course of the melt season. At the beginning of the melt season, the subglacial drainage system is typically cavity dominated, with the cavities probably poorly connected. As the seasonal snowpack melts, there is a larger water flux into the glacier and to the bed through englacial channels. If there is a large enough pressure perturbation from daily variations in water pressure or from a more sudden event, such as a rainfall event, the cavities will enlarge, become unstable, and eventually create an efficient channelized system (Fountain and Walder, 1998; Kamb, 1987). This evolution has a major impact on sliding because as cavities grow, the area of ice in contact with the bed decreases, allowing for increased sliding speed. In this case, the ice displacement is limited by the geometry of obstacles on the bed. This indicates the need for a sliding law
that incorporates area of separation at the bed as well as effective pressure (Harper et al., 2007; Howat et al., 2008).

Over spatial scales comparable to or larger than the ice thickness, temporal variations in subglacial water pressure will likely cause matching variations in basal drag and consequently in rates of basal motion. The spatial variations in basal motion will ultimately produce deviations in horizontal stress gradients in the overlying ice, affecting a broader area of the glacier than the local basal drag perturbation. Therefore, it is necessary to study glacier dynamics over large areas in order to distinguish between motion due to local influences, such subglacial water pressure or driving stresses, and motion due to non-local influences transmitted by stress gradients (Mair et al., 2001). To fully understand the glacier dynamics, it is necessary to identify sites that are most important to restraining ice flow (van der Veen, 1999). One method for examining factors controlling glacier velocity variations is a force-budget analysis. This method, developed by van der Veen and Whillans (1989) is based on the assumption that, since accelerations in flow speed are small, there is zero net force acting on any section of a glacier. The glacier force budget can therefore be partitioned into driving and resistive stresses, the sum of which must always be zero. Glaciers are forced downslope by a driving stress due to gravity and the weight of the ice. The driving stress is opposed by frictional drag on the bed and shearing on the glacier walls. Horizontal gradients in stresses acting along-flow (longitudinally) and vertically (i.e. stress bridging) may either be driving or resistive. Thus, the basal drag can be calculated if the other stresses can be constrained from surface observations (van der Veen and Whillans, 1989). The force budget method is useful because it does not require direct observations of the bed, obtains an estimate of
drag over a large area of the bed and can be completed wherever measurements of glacier thickness, surface slope and surface velocity are available and ice rheology, through the stiffness parameter of Glen’s Flow Law, is known or can be assumed.

Force budget analyses have been carried out on numerous glaciers, typically using an integrated-thickness approach in which horizontal strain rates are assumed constant with depth and vertical motion is ignored. Van der Veen and Whillans (1993) tested this method on Columbia Glacier, Alaska and determined that basal resistance was spatially variable, which they attributed to the nature of the subglacial drainage system. They also found that the horizontal strain rates were large and spatially variable, indicating stress redistribution across and along flow. Hooke and others (1989) found similar results when looking at 5 velocity events at Storglaciären, Sweden. Hedfors and others (2003) established greater confidence in the force budget method applied to a small valley glacier when they successfully resolved variations in basal conditions that corresponded predictably with known bed topography. A number of studies have built on the results of these earlier force balance analyses with the hypothesis that basal motion is principally controlled by stress redistribution beneath and within glaciers (Kavanaugh and Clarke, 2001; Mair et al., 2001, 2002a; Rippin et al., 2005).

Finally, knowledge of changes in basal and englacial stresses has been acquired from ice flow models constrained by surface velocity measurements. For example, Amundson and others (2006) used observations of surface motion and ice deformation and a simple inverse finite-element flow model to infer mean stress fields in a transverse cross-section of Black Rapids Glacier, Alaska. While they focused on seasonal changes, they did state that observed rapid motion events were probably caused by rapid stress
redistribution towards the margins and are most likely responsible for the larger overall ice displacement in summer compared to winter. Lastly, Sugiyama and others (2010) monitored 3 speed-up events at Gornergletscher, Switzerland. They found that ice motion deviated toward the side margins during events. Also, as the glacier accelerated there was a transverse velocity component generated, which turned flow away from the central flow line. This was due to elevated subglacial water pressure over a limited part of the bed. They confirmed their results with a two-dimensional ice flow model applied to the transverse glacier cross section. Collectively, these analyses reveal that short-term variability in basal drag can result in stress redistribution from weaker to stronger regions of the bed, resulting in rapid changes in glacier speed.

While the results of all of these studies have contributed greatly to our understanding of basal motion, a predictable relation between glacial basal conditions, such as the hydrology and till rheology, and ice motion remains elusive. Current models must assume an arbitrary boundary condition for basal drag, leading to potential inaccuracies in results. It is also assumed that basal drag does not vary temporally, despite previous observations demonstrating varying basal hydrologic conditions and sediment rheology through time (Fountain and Walder 1998; Truffer et al., 2001). Therefore, our study aims to clarify weaknesses in current understanding of soft-bedded glacier motion and sliding as a whole.

The motivation for this study originates from the research done by Howat and others (2008) on Breiðamerkurjökull, Iceland. The principle objective of that study was to find consistent patterns between surface ice motion, stress distribution, and bed separation that would help define the mechanisms controlling short-term variations in ice
speed. They hypothesized that the transient evolution of the subglacial drainage system should be important in controlling the relationship between sliding, effective pressure, and basal shear stress. In addition, along-flow gradients in resistive stress could facilitate redistribution of basal shear stress during increased basal motion. Using surface motion data collected over a 66-day period from five DGPS stations located within 7 km of the front, they observed multiple periods of increased surface motion and uplift corresponding to periods of rainfall or increased temperatures. Results showed that there was an increase in the along-flow longitudinal stress during sliding events, indicated by increasing rates of horizontal compression, implying increased basal drag towards the terminus and increased longitudinal stress transfer from upglacier onto the study area. This is consistent with the idea that the evolution of the subglacial drainage system plays a major role in controlling short-term velocity variations. The terminus is expected to have efficient, tunnelized drainage with a relatively smaller area of bed separation and higher basal drag. On the other hand, upglacier of the study site is believed to have inefficient, distributed drainage for a longer period of the melt season, resulting in a larger area of bed separation and a decrease in basal drag. In this case, the efficiently draining terminus would act as a barrier to the inland ice during increased basal motion (Howat et al., 2008). Observations used in that study, however, were limited to the compressive zone of the glacier, so that this hypothetical evolution of along-flow stress distribution remains unconfirmed.

In order to determine the mechanisms underlying basal motion of Breiðamerkurjökull, it is necessary to collect more spatially variable data, including areas further upglacier than the sites monitored in Howat et al. (2008). In addition to a broader
study area, there are three main goals for this project. The first objective is to observe the glacier surface motion of Breiðamerkurjökull at high spatial and temporal resolution to identify whether it experiences short-term velocity variations over a large area. The second is to use the surface motion measurements to calculate spatial and temporal variations in longitudinal, lateral, and basal stresses using a force budget approach. The third aim is to explain the variations in velocity and force budget components with reference to glaciomorphic features and changes in subglacial hydrology inferred from meteorological and hydrological data.
2. Site Description

Breiðamerkurjökull is an outlet glacier on the southern margin of the Vatnajökull ice cap located in southeast Iceland (Figure 1).

![MODIS image of Iceland, Breiðamerkurjökull is circled in red.](image)

This temperate (isothermal) glacier is approximately 12 km wide and has a total area of 910 km² (Björnsson, 1996). Breiðamerkurjökull can be divided into three branches that
are separated by medial moraines, each with different geometry and flow. The central flowline of the western third of the glacier is about 18 km and terminates in a narrow calving front at the proglacial lake Breiðárlón. At the bed, there is subglacial trench that is about 7 km long where bed elevations fall below sea level (Björnsson, 1996). The central branch is about 26 km long and is bounded by medial moraines. Here, the glacier terminates on land, has ice thicknesses of up to 300 m, and has a bed above sea level. Finally, the eastern branch is the longest at 34 km in length. The branch follows a subglacial trough, located 2 to 3 km upglacier from the terminus, which reaches depths of 300 m below sea level (Björnsson, 1996). Here, the glacier terminates at a calving front at Jökulsárlón lagoon. The ice thickness in the trough exceeds 600 m (Howat et al., 2008). Surface elevations range from about 1500 m.a.s.l. upglacier to about sea level at the terminus (Björnsson, 1996).

Breiðamerkurjökull is situated in a region with a maritime, cold-temperate climate (Evans and Twigg, 2002). The glacier has a net mass balance of 4-7 ma⁻¹ in the accumulation area and an ablation rate of about 10 ma⁻¹ at the terminus (Björnsson, 1996). Regional hydrology plays a major role in the mass balance and motion of Breiðamerkurjökull. The glacier receives a large amount of water from rainfall. As of 1996, the recorded maximum rainfall intensity was 230 mm in 24 hours (Björnsson, 1996). Historical documentation provides valuable detailed observations of previous advances and retreats of the glacier. For example, frequent mapping of glacier margins has made it possible to reconstruct terminus profiles through time (Evans and Twigg, 2002). The first accurate map of Breiðamerkurjökull was produced in the early 20th century and the glacier has been the site of numerous studies since due to its ease of
accessibility and range of glaciological and geomorphic features. Most notably, the seminal paper on soft-bedded glacier dynamics, Boulton and Hindmarsh (1987), was based on data from Breiðamerkurjökull. The flow law for soft beds presented in that paper has underpinned most relations used in ice sheet flow models. Past programs have resulted in extensive datasets for till properties, ice thickness, bed topography and speed (Björnsson, 1996, Björnsson et al., 2001; Boulton and Hindmarsh, 1987; Evans and Twigg, 2002; Howat et al., 2008).
3. Methods

3.1 Surface Motion

Surface motion of Breiðamerkurjökull was recorded for the 2009 and 2010 melt seasons using a strain-grid network comprising 12 continuously-recording dual-frequency geodetic GPS (Trimble R7 & 5700) receivers powered by solar panels provided by UNAVCO. In 2009, the units were spread approximately 2 km apart on a broad region of the glacier approximately 2 km from the front (Figure 2).
Figure 2. Velocity map overlay of Breiðamerkurjökull derived from automated surface-feature tracking between repeat ASTER images. Coordinates are in UTM. The locations of the 12 GPS stations for the 2009 field season are marked by white dots. Arrows show flow direction at each station and are scaled to show relative velocities.
The western and central lines (‘A’ and ‘B’ lines, respectively) of the grid were located within the central branch on more slow-moving ice overlying bedrock. The eastern line (‘C’ line) of the grid was within the eastern branch on a fast flowing region underlain by till and was upglacier of the calving front. Ice-penetrating radar data obtained during the deployment revealed ice thicknesses ranging from 200 to 395 m on the A line, 220 to 520 m on the B line, and 460 to 640 m on the C line (K. Matsuoka, unpublished data). These data further reveal that downglacier stations (1 and 2) overlie till whereas the upglacier stations (3 and 4) are positioned over bedrock. The GPS receivers were installed using the methodology of Anderson et al. (2004). This involved mounting the receivers on a tripod platform that was supported by insulated stakes drilled several meters into the ice with a hot steam drill (Figure 3).
This allowed the motion of the GPS antenna to be independent of ablation. Two additional GPS units were placed on the till plain in front of the glacier to serve as a reference for differential processing. Data were collected from March 10 to June 25, 2009. Three units on the C line (C2, C3, and C4) provided no measurements and due to numerous data gaps, position data for the A and B line were available from May 15 to June 19, 2009.

In 2010, a similar strain-grid was installed on Breiðamerkurjökull except that all GPS units were located on the central branch due to heavy crevassing on the eastern part of the glacier and to focus our analysis on a smaller area with more uniform flow direction (Figure 4).
Figure 4. Velocity map overlay of Breiðamerkurjökull derived from automated surface-feature tracking between repeat ASTER images. Coordinates are in UTM. The locations of the 12 GPS stations for the 2010 field season are marked by white dots. Arrows show flow direction at each station and are scaled to show relative velocities.
The units were spread 2 km apart along-flow and 1 km apart across-flow. Again, an additional GPS unit was placed on a till plain at the front for differential processing. Data were collected from March 17 to July 2, 2010. Observations of vertical position of the base station for the beginning of the 2010 season verified that the base station, while on a till plain, did not move substantially.

3.2 GPS Processing

For both years, raw GPS position data were collected at 5 second epochs and post-processed in kinematic mode using the GPS processing software Track, which is part of the Massachusetts Institute of Technology (MIT)’s GPS processing suite GAMIT/GLOBK (http://www-gpgs.mit.edu/~simon/gtgp/). A kinematic data analysis approach minimizes systematic errors and resulting spurious signals (King, 2004). Baselines in this study were short (< 15 km) and so problems associated with processing of long baselines, for example, a lack of common satellites between the rover and base or an increase in relative ionospheric delay, were avoided (King, 2004). A linear combination of the carrier phase observables L1 and L2 was used to remove the first order effects of the ionosphere. Since the difference in vertical position of the stations was small (< 250 m), the error in the positions due to the troposphere is trivial. Track allows parameterization of residual zenith total delay (ZTD) at every measurement epoch. With Track, the ZTD is mapped to the elevation angle of each satellite using the MTT mapping function (Herring, 1992). No additional tropospheric modeling was necessary for the scope of the study. The resulting positions were converted to UTM for analysis of
displacement. Following the methodology of Truffer et al. (2009), errors in position were estimated by locating a period of the time series where there was no significant diurnal or short-term variation in flow. The standard deviation of the linear change in position at this time was taken as the error. The resulting mean standard deviation in the horizontal position was 3 mm and the vertical was 4 mm. In order to remove noise, the data were averaged over 10 minutes and a robust loess algorithm with a 24-hour smoothing window was applied to the time-position data series in Matlab™ (v.7.8.0).

3.3 Meteorological Data

The Iceland Meteorology Office provided daily precipitation data from the Kvísker station, located about 15 km southwest of Breiðamerkurjökull. The Institute for Marine and Atmospheric Research Utrecht, Holland (IMAU) and the Glaciology Group at the Institute of Earth Sciences, University of Iceland provided 30-minute temperature data from a weather station on Breiðamerkurjökull.

3.4 Force Budget Calculations

The force budget method was developed by van der Veen and Whillans (1989) and assumes there is zero net force acting on any section of a glacier. The along-flow component of the glacier force budget can therefore be written as:
\[ 0 = \tau_{dx} + \tau_{hx} + \frac{\partial}{\partial x} \int_{h-H}^{h} R_{xx} \, dz + \frac{\partial}{\partial y} \int_{h-H}^{h} R_{xy} \, dz + \frac{\partial}{\partial x} \int_{h-H}^{h} R_{xz} \, dz \]  

(1)

where the driving stress, \( \tau_d \), due to gravity and the weight of the ice forcing the glacier downslope is opposed by frictional drag on the bed, \( \tau_b \), along-flow gradients in longitudinal stress, \( R_{xx} \), across-flow gradients in shear stress, \( R_{xy} \), and vertical gradients in resistive stress, \( R_{xz} \), of the glacier. In this equation, Cartesian coordinates \( x, y \) and \( z \) represent the along-flow, across-flow, and vertical directions, respectively. \( H \) is the ice thickness and \( h \) is surface elevation. In this case, \( \tau_{dx} \), is always positive because the stress acts along the direction of increasing \( x \) and \( \tau_{hx} \) is always negative.

Where previous studies, including Howat et al. (2008), typically assume depth-invariant strain rates, which reduces the integrals in Equation 1 to averages over the ice thickness, we solve the integrals numerically using an approach similar to that of van der Veen and Whillans (1989). This technique takes into account the vertical resistive stress and so the resistive stresses are integrated from a general depth to the surface, allowing for the calculation of strain rates at all depths. Using this method, the force budget is solved at discrete depth intervals, starting from the stress-free surface boundary, to solve for the basal stress boundary. At each depth, the velocities are used to solve the force budget equation, including the shear strain rates, which are then used to calculate the velocities for the next lower measurement point (Figure 5) (van der Veen and Whillans, 1989).
Figure 5. Illustration of the solution scheme for calculating velocities and stresses at depth. Panel A shows the geometry and the input data (surface slope, ice thickness, surface velocities). From the velocities, longitudinal strain rates are calculated (Panel B) and the force balance equation solved for the shear strain rate (Panel C). From the vertical shear, the velocities at the next depth can be calculated (Panel D). Next, longitudinal stretching is calculated (Panel E) and force balance is solved for this layer (Panel F). This procedure is repeated for all depth layers (Panel G). Finally, when the bed is reached, basal drag and stress normal to the bed are calculated (Panel H) (from: van der Veen and Whillans, 1989).
In order to calculate strain rates at depth, a two-dimensional grid, similar to the one in Figure 5 must be established in which the vertical coordinate $s$ is the depth relative to the ice thickness:

$$s = \frac{h - z}{H}$$  \hspace{1cm} (2)

where $h$ represents surface elevation, $H$ is ice thickness, and $z$ is the vertical coordinate.

Using this convention, at the surface $z = h$ and $s = 0$ and at the bed $z = h - H$ and $s = 1$.

First, the driving stress is calculated from the glacier geometry,

$$\tau_{dx} = -\rho g H \frac{\partial h}{\partial x}$$  \hspace{1cm} (3)

where $\rho$ is ice density and $g$ is acceleration due to gravity.

The strain rates are obtained from the gradients in the measured glacier surface velocities ($u$ and $v$ corresponding to along and across-flow, respectively).

$$\dot{\varepsilon}_{xx} = \frac{\partial u}{\partial x}$$  \hspace{1cm} (4)

$$\dot{\varepsilon}_{yy} = \frac{\partial v}{\partial y}$$  \hspace{1cm} (5)

$$\dot{\varepsilon}_{zz} = -\left( \dot{\varepsilon}_{xx} + \dot{\varepsilon}_{yy} \right)$$  \hspace{1cm} (6)

$$\dot{\varepsilon}_{xy} = \frac{1}{2} \left( \frac{\partial u}{\partial y} + \frac{\partial v}{\partial x} \right)$$  \hspace{1cm} (7)
\[
\dot{\varepsilon}^2 = \frac{1}{2}\left(\dot{\varepsilon}_{xx}^2 + \dot{\varepsilon}_{yy}^2 + \dot{\varepsilon}_{zz}^2 + \dot{\varepsilon}_{xy}^2 + \dot{\varepsilon}_{xz}^2 + \dot{\varepsilon}_{yz}^2\right)
\]  

(8)

Applying Glen’s flow law, the deviatoric stresses can be expressed in terms of strain rates,

\[
\tau_{ij} = B \dot{\varepsilon}^n \dot{\varepsilon}_{ij} \quad i, j = x, y
\]  

(9)

where \(B\) is ice viscosity and \(n\) is the flow exponent for ice.

Next, the resistive stresses are expressed in terms of stress deviators,

\[
R_{xx} = 2\tau_{xx} + \tau_{yy} + \tau_{zz}
\]  

(10)

\[
R_{sx} = \tau_{sx}
\]  

(11)

The vertical resistive stress is found from incompressibility and based on the \((x,s)\) coordinate system becomes,

\[
R_{sz}(s) = -\Delta_s R_{sz} - \frac{\partial}{\partial x} \left[ \int_0^s HR_{sz} \, ds \right]
\]  

(12)

with

\[
\Delta_s = \frac{\partial h}{\partial x} - s \frac{\partial H}{\partial x}
\]  

(13)
After partitioning full stress into resistive and lithostatic components and integrating with respect to $s$, the along flow equation for the shear stress at depth is,

$$R_{xx}(s) = s\tau_{ds} - \frac{\partial}{\partial x} \left[ \int_s^0 \! H R_{xx} \, ds \right] + \Delta_s R_{xx}(s) \quad (14)$$

The force balance equation in terms of strain rates, using Equations 2-14, becomes,

$$B\varepsilon_e^{-\frac{1}{n}} \dot{\varepsilon}_{xx} = s\tau_{ds} - \frac{\partial}{\partial x} \left[ \int_s^0 \! 2HB\varepsilon_e^{-\frac{1}{n}} \dot{\varepsilon}_{xx} \, ds \right] + 2\Delta_s B\varepsilon_e^{-\frac{1}{n}} \dot{\varepsilon}_{xx} + \Delta_s R_{xx} - \frac{\partial}{\partial x} \left[ \int_s^0 \! H R_{xx} \, ds \right] \quad (15)$$

The numerical solution is found iteratively from the surface to the bed using finite-difference methods. Force budget solutions were determined at 2-hourly time steps for the central nodes of the two 9-point strain grids from the 2010 data.

### 3.5 Error Analysis

To assess the impact of measurement uncertainties on the force budget solutions, a perturbation sensitivity analysis was performed for each grid. We assumed position errors were normally distributed with a mean of zero and standard deviation of ± 5 m for thickness measurements and ± 5 mm for position. Thickness and position measurements were perturbed by adding errors drawn randomly from this distribution. The force budget was then calculated using these perturbed values. This was repeated 1000 times for each time step in the Event 5 time series. The resulting distribution for each stress was examined and the standard deviation was taken as the error.
4. Results

4.1 Ice Motion: 2009

Ten-minute horizontal ice surface velocities of the A line and B line stations for the 2009 melt season are shown in Figures 6 and 7, respectively. Overall, the magnitude of horizontal speed increases upglacier from station 1 to station 3 and then decreases slightly at station 4. Three speed-up events were observed during the study period, each lasting approximately 24 to 48 hours. These speed-ups will be referred to as Event 1 (Day 143-144, May 23-24), Event 2 (Day 149-150, May 29-30) and Event 3 (Day 167-169, June 16-18). The timing of the events was synchronous across all stations at the resolution of the data. Uplift occurred at all stations during all three events (Figure 8). The average uplift of all stations during Event 1 was 3 to 4 cm. Station A4 showed the least variation (Figure 8a) throughout the event, while station B3 exhibited the greatest uplift (~ 6 cm) (Figure 8b). Through the second half of the record, all stations except A1 gained elevation relative to the mean at a rate of approximately 3 mm/day. There was a mean temperature of 4.87 °C throughout the time series. The temperature remained above 0 °C and the maximum temperature was 11.21 °C (Figures 6b and 7b).

Event 1 began on May 23 and correlates with a precipitation event reaching its peak on May 24 (Figures 6 and 7). The peak speeds along the A line varied widely reaching 0.30 – 0.70 m/day or about 4 times greater, on average, than the mean speed at
the station. The peak speeds along the B line reached 0.45 – 0.95 m/day and were about 2.5 times greater, on average, than the mean speed at the station.

The onset of Event 2 correlates well with a 2-day sustained increase in temperature and an increase in precipitation. Peak speeds occurred the evening of the same day, reaching 0.30 – 0.50 m/day along the A line and 0.50 – 0.70 m/day along the B line. Event 2 was smaller in magnitude but similar in duration to Event 1.

Days 155 to 160 exhibit a strong diurnal variation in flow speed. Event 2 began on June 16, correlating well to the largest rainfall event (66 mm) of the study period on the same day. The flow speeds peaked early in the day on June 17 and decreased to pre-event level by June 18. Peak speeds along the A line ranged from 0.20 – 0.65 m/day and along the B line from 0.50 – 0.70 m/day.
Figure 6. (a) Time series of horizontal flow speed of the A line stations in 2009, starting with the most downglacier on the top plot and moving upglacier in successive plots. (b) Time series of half-hourly air temperature (line) and daily precipitation (bar). Shaded areas indicate periods of speed-up events.
Figure 7. (a) Time series of horizontal flow speed of the B line stations in 2009, starting with the most downglacier on the top plot and moving upglacier in successive plots. (b) Time series of half-hourly air temperature (line) and daily precipitation (bar). Shaded areas indicate periods of speed-up events.
4.2 Horizontal Strain Rates: 2009

Longitudinal (along-flow) strain rates are calculated using Equation 4. In this case, negative strain rates indicate compression between adjacent stations. In general, strain rates for the A line remained close to zero except during speed-up events. The downglacier and mid-glacier stations (B1/B2 and B2/B3) were compressive throughout the study period while the upglacier stations (B3/B4) were extensive.

During speed-up events, data show increased compression between the most downglacier stations (1 and 2) similar in duration to the increased ice flow speed. The midglacier stations (2 and 3) also had increased compression but smaller in magnitude to
the downglacier stations. Finally, the upglacier stations (3 and 4) experienced periods of greater extension during the speed-up events (Figure 9).

![Figure 9. Horizontal along flow strain rates during Event 1 (left), Event 2 (middle), and Event 3 (right) for the A line.](image)

4.3 Ice Motion: 2010

Ten minute horizontal velocities of the A, B and C lines for the 2010 melt season are shown in Figures 10, 11 and 12, respectively. Similar to 2009, mean ice speeds increase upglacier to station 3 and decrease slightly at station 4 for all lines. In addition, mean speeds increase from west (A line) to east (C line) across the glacier. Two speed-up events were observed during the study period, once in April and once in May. These speed-ups will be referred to as Event 4 (Day 99-101, April 9-11) and Event 5 (Day 138-140, May 18-20).
Event 4 began on April 9 and lasted approximately 48 hours. The increase in horizontal speed correlated well with a two day combined increase in both temperature and rainfall. Rainfall amounts reached 60 mm per day for both days and temperature reached 16.48 °C (Figure 10b). Uplift was present at all stations during Event 4 but varied greatly (Figure 13). Station A1 experienced 19 cm of uplift while the majority of the stations experienced approximately 9 cm of uplift. Horizontal speeds along the A line reached 0.45 – 0.95 m/day or about 7 to 16 times the mean speed of the respective station. The B line showed peak speeds of 0.75 – 0.90 m/day or 6-8 times the mean speed. Finally, horizontal speeds along the C line reached 0.75- 1.00 m/day or 2-9 times the mean speed of the respective station.

Event 5 is shorter in duration than Event 4, beginning on May 18 and lasting approximately 28 hours. Again, horizontal speed increase was accompanied by uplift at all stations. Station A1 had the most uplift (~10 cm) and most other stations experienced approximately 5 cm (Figure 13). The C line showed a wide variety of vertical motion during Event 5. Station C4 experienced approximately 15 cm of uplift, station C3 approximately 6 cm, and C1 and C2 approximately 3 cm (Figure 13c). Peak horizontal speeds along the A line reached 0.55 – 0.75 m/day or 9-12 times the mean speed at the respective station. The B line stations reached speeds of 0.65-1.0 m/day or 6-7 times the mean speed. Peak horizontal speeds along the C line ranged from 0.55 to 1.0 m/day or 2-8 times the mean speed.
Figure 10. (a) Time series of horizontal flow speed of the A line stations in 2010, starting with the most downglacier on the top plot and moving upglacier in successive plots. (b) Time series of half-hourly air temperature (line) and daily precipitation (bar). Shaded areas indicate periods of speed-up events.
Figure 11. (a) Time series of horizontal flow speed of the B line stations in 2010, starting with the most downglacier on the top plot and moving upglacier in successive plots. (b) Time series of half-hourly air temperature (line) and daily precipitation (bar). Shaded areas indicate periods of speed-up events.
Figure 12. (a) Time series of horizontal flow speed of the C line stations in 2010, starting with the most downglacier on the top plot and moving upglacier in successive plots. (b) Time series of half-hourly air temperature (line) and daily precipitation (bar). Shaded areas indicate periods of speed-up events.
4.4 Horizontal Strain Rates: 2010

Event 4 strain rates could not be calculated due to data gaps. Event 5 strain rates are shown in Figure 14. Longitudinal strain rates were calculated in the same manner as the 2009 data. As expected, the results are similar to 2009. During speed-up events, upglacier stations experience extension, while mid-glacier and downglacier stations
experience compression at the onset. In the A line, this compression switches to
extension at about the same time as peak speed (Figure 14).

Figure 14. Horizontal along-flow strain rates during Event 5 for the A (left), B (middle), and C (right) lines.

4.5 Force Budget: Event 5

Force budget calculations were carried out for Event 5, which had sufficient data
to form two 9-unit strain grids referred to as upglacier (Figure 15a) and downglacier
(Figure 15b). An inverse distance-weighted average was performed to complete the B1
time series, using the mean of data from A1 and C1. A detrended comparison of the
position interpolation error shows a standard deviation of 8.5 mm in the horizontal
direction and 11 mm in the vertical.

Driving stress is nearly constant with time such that variations in longitudinal and
lateral stress must be balanced by variations in basal drag. It is therefore important to
determine the direction in which each stress is acting. Driving stress is always along-flow while basal drag is always acting in the direction opposite of the driving stress or against-flow. The longitudinal stress changes direction in time and space, as seen in Figure 15a.

An ice viscosity parameter ($B$ in Section 3.4) of 100 kPa yr$^{1/3}$ is typically assumed for temperate ice. However, assuming no sliding at the slowest and least variable part of the record, the surface speed and driving stress can be used to calculate $B$. At Breiðamerkurjökull, this resulted in a $B$ of $\sim$230 kPa yr$^{1/3}$. This would represent a lower bound, as sliding greater than zero would reduce the deformation rate, requiring stiffer ice. This unexpectedly stiff value of ice may be due to the grain size of ice at the terminus of Breiðamerkurjökull, which can be on the order of several cm at the surface (Duval et al., 1983). To characterize the uncertainty in $B$, we solve the force budget solutions using $B$ values of 150, 200, and 250 kPa yr$^{1/3}$ and present the range of solutions. This increasing $B$ results in increased magnitude of variations in calculated stress, but does not affect the temporal pattern of the variations (Figure 15). In Figure 14, the solid lines are results for a $B$ value of 200 kPa yr$^{1/3}$ while the colored regions represent the bounds of 150 kPa yr$^{1/3}$ and 250 kPa yr$^{1/3}$.

Initially, the longitudinal stress upglacier is acting in the along-flow direction (i.e. acting as a driving, as opposed to resistive, stress) and averages 11 kPa until the evening of May 18$^{th}$ (day 138) and the onset of Event 5. During the speed-up, the longitudinal stress switches direction and increases against-flow (i.e. becomes resistive), reaching 40 kPa or about 32% of the driving stress at its peak in approximately 12 hours. At the end of the speed-up, another 12 hours later, it returns to acting along-flow. The lateral stress upglacier is near 0 kPa for the majority of the time series, increasing for 12 hours during
the speed-up. Lateral drag reaches a maximum of 32 kPa, or about 25% of the driving stress. These increases in lateral and longitudinal resistive stresses are balanced by a ~45% reduction in basal drag, from approximately equal to the driving stress, to 60 kPa. Following the speed-up event, the basal drag returns to near the driving stress (Figure 15a).

Downglacier, the longitudinal stress does not change direction and is acting in the along-flow direction (i.e. as a driving stress) for the entire time series. It averages 12 kPa, or 10% of the driving stress, until the morning of May 19th (day 139). The increase and peak in longitudinal stress downglacier lags the increase and peak in longitudinal stress upglacier by 10 hours (Figure 15). Longitudinal stress peaks at 47 kPa (37% of the driving stress) after approximately 5 hours and decreases to 18 kPa in about 10 hours. The lateral stress remains close to 0 kPa and increases to only 12 kPa (10% of the driving stress) during Event 5. This increase in along-flow longitudinal stress is balanced by a 12% increase in basal drag during the event, from 135 to 152 kPa or ~120% of the driving stress (Figure 15b). After reaching a peak within 5 hours of the onset of speedup, the basal drag returned to a slightly higher level than before the event. The increase in basal drag was only about 7 hours in length, 12 hours shorter than variations in stresses upglacier. The vertical stress was small both upglacier and downglacier, remaining less than or equal to 1 kPa over the time series.
Figure 15. Plots of longitudinal, lateral, and basal stresses calculated using a force budget. Force budget solutions in terms of resistive stresses are shown for upglacier and downglacier on plots (a) and (b), respectively. The absolute value represents magnitude while the sign represents direction in which the stress is acting (positive for against-flow and negative for along-flow).
4.6 Error Analysis

The mean error for each stress during Event 5, downglacier and upglacier, is shown in Table 1. The smallest error, both upglacier and downglacier, was that of the driving stress at 0.8 kPa (~0.5% of the maximum driving stress). The error in basal drag was also relatively small at approximately 15.8 kPa (~10% of the maximum basal drag). Longitudinal and lateral stress errors were larger at approximately 14 kPa (~33% of the maximum) and 4.1 kPa (~22% of the maximum), respectively. The errors are less than 65% of the range in the respective stress, indicating that the variation in stresses during the Event 5 time series cannot be explained by error alone and confirms the pattern of stress redistribution during the sliding event.

Table 1. Mean errors (kPa) of stresses during Event 5.

<table>
<thead>
<tr>
<th>Stress</th>
<th>Mean Error (kPa) Downglacier</th>
<th>% of Maximum Stress</th>
<th>Mean Error (kPa) Upglacier</th>
<th>% of Maximum Stress</th>
</tr>
</thead>
<tbody>
<tr>
<td>Driving</td>
<td>0.8</td>
<td>0.6</td>
<td>0.7</td>
<td>0.5</td>
</tr>
<tr>
<td>Basal</td>
<td>15.7</td>
<td>10.3</td>
<td>15.9</td>
<td>11.3</td>
</tr>
<tr>
<td>Longitudinal</td>
<td>13.2</td>
<td>28.1</td>
<td>15.9</td>
<td>39.0</td>
</tr>
<tr>
<td>Lateral</td>
<td>3.9</td>
<td>30.5</td>
<td>4.4</td>
<td>13.7</td>
</tr>
</tbody>
</table>

In addition to the mean error of each stress, it is worthy to note the change in errors over the time series (Figure 16). The errors in driving and lateral stresses do not vary much over the time series. However, the errors in basal and longitudinal stresses noticeably decrease at the beginning of Event 5 around day 139.
Figure 16. Time series of the errors in stresses during sliding Event 5.
5. Discussion

5.1 Glacier force budget variations

The time series of surface motion provides insight into the along-flow variations in stresses during sliding events. For all events, we find an increase in the rate of compression downglacier and extension upglacier, suggesting a synchronous increase in basal drag downglacier and a decrease upglacier during sliding events (Figure 17).

![Graph showing along-flow profile with arrows representing strain rates.](image)

Figure 17. Along-flow profile (data courtesy of K. Matsuoka, University of Washington and Norwegian Polar Institute) of the A line and arrows representing the apparent longitudinal strain rates during speed-up events. GPS stations are represented by black dots.

Our force budget results support the initial expectations from surface strain rate observations. The Event 5 velocity increase correlates well with a precipitation event and
also an increase in temperature. Force budget solutions show that, during this event, there were substantial variations in longitudinal stress that were balanced by variations in basal drag. Longitudinal stress increases along-flow (i.e. adds to the driving stress) downglacier while longitudinal stress increases against-flow (i.e. increases resistance to driving stress) upglacier. Since lateral and vertical horizontal stress gradients did not vary substantially, these changes in longitudinal stress were accommodated by increasing basal drag down glacier and decreasing basal drag upglacier. This indicates that during speed-up events, driving stresses are transferred longitudinally from the interior toward the terminus such that the basal drag over the marginal till wedge exceeds the local driving stress. Thus, from a force-budget perspective, the terminus region acts as a barrier to interior motion during sliding events.

There are several hypotheses that may explain the redistribution of stress during speed-up events. The first stems from the standard model of the evolution and distribution of the subglacial drainage system from inefficient to efficient (Fountain and Walder, 1998; Mair et al., 2002b). Since channel development is driven by water input from the surface, the development of the subglacial drainage system should start downglacier and progress upglacier throughout the melt season. This hypothesis has been supported by observations using dye-trace studies (Nienow et al., 1998). Thus, efficient, channelized drainage will likely develop earlier closer to the terminus where there is more surface melt and throughflow of water. In addition, thinner ice at the terminus results in less overburden pressure and slower creep-closure rates of the tunnels. We would expect the area of separation between ice and bed to be lower in a channelized drainage system relative to a distributed system, resulting in higher basal drags for a
given water pressure. Conversely, the upglacier region of the study area is expected to maintain an inefficient drainage system for a greater part of the melt season due to higher overburden pressures and less water flux. This drainage configuration is incapable of efficiently draining larger water inputs reaching the bed, leading to higher basal water pressures, increased bed separation, and decreased basal drag.

In the case of Breiðamerkurjökull, it seems that basal drag upglacier decreases substantially below the local driving stress during speed-up events. The excess driving stress is then transferred downglacier, through gradients in longitudinal stress, to a more efficiently-draining terminus where the water pressure and/or ice-bed separation area are lower and basal drag is sufficient to support the excess stress.

A second hypothesis for the observed stress distribution during sliding is the influence of varying bed characteristics. Ice-penetrating radar surveys conducted in 2009 reveal that the till wedge upon which the glacier terminates pinches out onto bedrock inland in the vicinity of the middle part (i.e. Stations 2 and 3) of our study area. It is therefore possible that the observed stress transfer reflects the change from bedrock to till approaching the terminus.

The flow over granular material such as till is caused by basal sliding, subglacial sediment deformation, or a combination of the two (Alley, 1989). The processes that control the amount of sliding and sediment deformation contributing to basal motion depend strongly on the mechanical and hydrological coupling at the ice-bed interface. Many studies have indicated a strong correlation between subglacial water pressure and both sliding (Ilken and Bindschadler, 1986; Kamb and Engelhardt, 1987; Hooke et al., 1989) and sediment deformation (Boulton and Hindmarsh, 1987; Iverson et al., 1995).
Yet, the complex relationships between variations in basal drag, sliding, and sediment deformation remain poorly understood.

Till rheology is an important component of soft bed glacier flow modeling. It is crucial to understand at what point failure of the till will occur and what the rate of deformation following this failure will be. Early hypotheses suggested a linear-viscous rheology in which the strain rate increases proportionally with applied stress (Boulton and Hindmarsh, 1987). However, more recent evidence suggests a Coulomb-plastic rheology in which no deformation occurs up to a yield stress, above which deformation will increase non-linearly (Hooke et al., 1997; Iverson et al., 1998, 1998; Kavarnaugh and Clarke, 2006; Truffer et al., 2000; Tulaczyk et al., 2000a). A major implication of this theory is that the subglacial till cannot accommodate any additional stress above the failure strength. The additional stress must then be transferred to other parts of the bed, such as the glacier margins or ‘sticky spots’ on the bed (Amundson et al., 2006; Truffer et al., 2001). A similar mechanism may operate at Breiðamerkurjökull such that during sliding events, the weak till beneath the terminus fails and stresses are transferred to large bedrock bumps, several of which are visible on the recently-deglaciated proglacial plain.

Alternatively, the till itself may be strong enough to resist the additional driving stress. As with any granular material, till grains dilate and align under increasing shear. Chains, or bridges, of contacting grains spanning the entire layer of deforming till may form, increasing the till strength. In order for till deformation to occur, grain bridges or networks must fail through grain breakage (Hooke, 2005). Iverson et al. (1998) determined that the shearing of till beneath glaciers that are subject to daily changes in water pressure is not steady but periodic. This periodic compression and decompression
will cause a cumulative decrease in till porosity, leading to dilatation of the grain bridges upon shearing. While dilation weakens dry granular materials by reducing friction between grains, pore expansion in water-saturated till can cause reductions in pore-fluid pressure that increases normal stresses at grain contacts, increasing friction and strength (Iverson et al., 1998; Moore and Iverson, 2002). This process is called dilatant hardening (Iverson et al., 1998) and has been observed by Tulaczyk and others (2000) beneath Whillans Ice Stream (WIS). They completed undrained tests on till beneath WIS and found that, for a given shear stress, strain rate decreases for an increasing pore pressure (Tulaczyk et al., 2000a). At Breiðamerkurjökull, dilatant hardening could result in till that is strong and can balance the excess longitudinal stress in addition to the driving stress. In order for dilatant hardening to occur, the till at Breiðamerkurjökull must experience frequent water pressure fluctuations, suggesting that it is well-drained.

Most recently, a model has been proposed by Piotrowski and others (2004), suggesting that soft beds are actually a “mosaic” of deforming and stable spots. They call attention to the fact that the glacier system is extremely complex with parameters controlling basal mechanics that vary in both time and space. While till will deform if its yield strength is exceeded by the basal shear stress, its shear strength can change depending on water input and drainage. For example, thick, permeable sediments can efficiently drain pore water and so are often stronger than fine-grained sediments where water pressure can rise more easily, facilitating deformation. Therefore, varying basal water pressures will lead to varying degrees of deformation within the till. This model agrees well with the plastic rheology hypothesis because it discounts widespread, pervasive deformation. Using this “mosaic” model, basal ice motion occurs through
sliding (decoupling at the ice-bed interface) and ploughing (stress support and glacier stabilization) (Evans et al., 2006; Piotrowski et al., 2004). The “mosaic” model may also explain the variation in basal drag and longitudinal stress during the sliding event at Breiðamerkurjökull. Here, the till bed would behave similar to bedrock, with the stable spots acting similar to obstacles in hard-bed sliding theory. Therefore, the subglacial hydrology would likely have a greater impact on basal sliding than the transition from hard bed to soft bed.

5.2 Implications

Whether the observed transfer of stress during sliding events is the result of variations in the subglacial drainage system, basal character, or a combination of both, the key outcome is that the terminus acts as a barrier to ice flow. This has important implications for understanding past and current behavior of ice masses. For example, the bed characteristics at Breiðamerkurjökull are similar to the Laurentide Ice Sheet (LIS) but on a smaller scale. Much of the southern margin of the LIS was underlain by till, inciting a debate over the extent of till deformation beneath the LIS and the role it played in the relatively rapid retreat of the ice sheet. Boulton and Jones (1979) and Alley (1991) believe that there was a deep, pervasively deforming till layer that contributed greatly to the ice motion at the southern margin of the LIS. Conversely, others such as Piotrowski et al. (2001) argue that the major movement mechanism of past ice sheets was sliding rather than bed deformation and that the glacial debris found at the southern margin of the LIS was transported englacially rather than within the deforming bed. The results from our
study agree with those of Piotrowski et al. (2001) and support the idea that there was no continuous, active deformation of viscously behaving basal till since the till at the terminus of Breiðamerkurjökull sufficiently resisted increased interior ice motion. Understanding the mechanics of ice motion and the sediment and bed conditions beneath modern glaciers, such as Breiðamerkurjökull, can greatly improve our knowledge of the behavior of ice sheets, past and present.

The results of our study demonstrate the potential for a surge-like mechanism on temperate glaciers like Breiðamerkurjökull. With the terminus resisting flow, there is the opportunity for increasing water storage and low basal drag upglacier until the strength of the margin is exceeded. At this point, the abundance of water could overwhelm the efficient drainage system and the glacier could experience a surge in ice flow, with the potential for significant advance of the terminus.

The flow dynamics of Breiðamerkurjökull during sliding events may have implications for geomorphology and erosion. Basal drag is due to friction at the base of the glacier and so if basal drag decreases upglacier during sliding events, erosion would likely be reduced upglacier as well. Similarly, if basal drag increases at the terminus during sliding events, then we would expect to see greater erosion downglacier with the abrasion of bedrock bumps or the deformation and breaking of till. This could ultimately affect the way in which the till deforms. As demonstrated in Iverson and others (1996), when grain size is similar throughout the till, stresses are supported by networks that are focused on a small number of contacts, which optimizes the potential for crushing. Once finer grains from crushing are introduced and assuming they are not washed away by water drainage, they distribute stresses over more contacts, cushioning the larger grains.
This till composition favors failure by grain sliding rather than by crushing, which could affect the way in which the till reacts to daily changes in water pressure and increasing shear stress.

In addition, the presence of a till wedge downglacier and the lack of sliding at the ice-till interface introduces the potential for the evolution of a push moraine such as those found at Taku Glacier, Alaska (Kuriger et al., 2006) and modeled by Leysinger Vieli and Gudmundsson (2010). In the study by Leysinger Vieli and Gudmundsson (2010), they completed a series of model experiments of a glacier advancing over deforming till of varying stiffness. They determined the interaction of the ice and till based solely on shearing through the till column and horizontal compression towards the terminus. They found that if the till is sufficiently soft, it is extruded from underneath the glacier towards the terminus, forming a push moraine. Movement of a glacier over soft till causes shearing within the till that can result in spatially variable rates of basal motion. In all of the modeling experiments, the highest ratio of basal motion to surface velocity was found in the terminus area and so basal motion measured at the terminus, such as those by Boulton and Hindmarsh (1987) at Breiðamerkurjökull, is not representative of the glacier as a whole. A sediment bulge at the front of the glacier terminus would most likely strengthen the resistance at the glacier terminus during sliding events.

The results of this study highlight the importance of horizontal stress redistribution during short-term motion events. In a study by Leysinger Vieli and Gudmundsson (2004), they conclude that, on timescales longer than a few years, it is not necessary to include the effects of horizontal stresses when calculating the reaction of alpine glaciers to climatic changes. Instead, it is more important to determine the mass

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balance distribution. While this may be true on longer timescales, our study shows that on shorter timescales such as hours and days, the details of ice dynamics cannot be ignored. Consequently, if much of the mass motion and erosion occurs during these sliding events, not including them could alias predictions of both mass balance and erosion in model results.

Lastly, with growing concern over the future of the Greenland Ice Sheet and its contribution to sea level rise, recent studies have focused on the behavior of Greenland outlet glaciers (Bartholomew et al., 2010, 2011; Schoof, 2010; Sundal, 2011). All of these studies indicate that the evolution of the subglacial drainage system, similar to alpine glaciers, plays a major role in ice flow. Schoof (2010) and Sundal (2011) demonstrate that variability in water input is more likely to increase ice velocity rather than an increase in mean melt supply. An increase in mean melt supply will allow for a more rapid change to an efficient drainage system, decreasing ice speed. The variability of meltwater input inferred from increases in precipitation and temperature during the melt season at Breiðamerkurjökull results in significant increases in ice flow. While we are unable to confirm the spatial and temporal distribution of the subglacial drainage system, our results corroborate the concept that meltwater pulses largely affect total seasonal flow speed. Therefore, flow models should focus on capturing short-term drainage variability and channelization processes in order to predict future behavior of the margins of the Greenland Ice Sheet.
6. Conclusions

Observations of surface motion and meteorological data on Breiðamerkurjökull reveal multiple speed-up events during the 2009 and 2010 melt seasons corresponding with periods of increased rainfall and/or melting. To elucidate the processes controlling short-term variations in ice speed, numerical force budget solutions were determined for two 9-unit GPS grids (upglacier and downglacier) during the final speed-up of the 2010 time series (Event 5). Our results confirm the hypothesis of Howat and others (2008) that the longitudinal resistive stresses play an important role in the redistribution of basal drag during increased basal motion. Longitudinal stress increased against-flow upglacier and along-flow towards the terminus, indicating a transfer of stress from upglacier to downglacier. Basal drag decreased to about 46% of the driving stress upglacier, synchronous with the velocity increase, while downglacier it increased slightly after the peak in velocity. Our force budget results provide empirical evidence that basal motion is controlled by large-scale redistribution of glacier stresses during sliding events.

One explanation for this stress redistribution is that the spatial and temporal variation in longitudinal resistive stresses and basal shear stress reflects the distribution of the subglacial drainage system. The upglacier area is expected to have an inefficient, distributed drainage system resulting in greater pressurization of the subglacial drainage system and a decrease in basal drag. On the other hand, downglacier is expected to have
efficient, channelized drainage, leading to greater basal drag. During high water pressure, there was no significant transfer of stress laterally towards the glacier margins or bedrock outcrops, suggesting that pervasive till deformation did not occur. The results demonstrate that the till at the terminus accommodates excess stress transferred from upglacier. This could be due to extensive grain bridging and dilatant hardening or the stress may be transferred to bedrock bumps during sliding events.

Our results confirm that the transient evolution of the subglacial drainage system is important in controlling the relationship between sliding, effective pressure, and basal shear stress and, second, that along-flow gradients in resistive stress facilitate redistribution of basal shear stress during increased basal motion. This research provides additional evidence that at short-term timescales related to a single precipitation and/or melting event, current steady state sliding laws cannot accurately predict the relationship between basal motion and water pressure. The results of this study have important implications for understanding past and current behavior of ice masses. For example, the Laurentide Ice Sheet had similar bed characteristics to Breiðamerkurjökull except on a larger scale, further supporting the hypothesis that till at the southern margin of the Laurentide Ice Sheet was transported englacially as opposed to by deformation. Findings at Breiðamerkurjökull indicate that erosion near the terminus during sliding events is likely greater than upglacier where basal drag is low, which could have consequences on till behavior during future sliding events. Geomorphologic implications arise from model experiments by Leysinger Vieli and Gudmundsson (2010) that indicate the potential for the development of a push moraine at the terminus of Breiðamerkurjökull. Finally, more recent studies demonstrate that, similar to alpine glaciers, the evolution of the subglacial
drainage system plays an important role in the ice flow of Greenland outlet glaciers. Therefore, the flow dynamics witnessed at Breiðamerkurjökull suggest that horizontal stress redistribution may also control basal motion of Greenland outlet glaciers. As a result, flow models should focus on capturing short-term drainage variability and channelization processes in order to predict future behavior of the margins of the Greenland Ice Sheet.
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