An increase in crevasse extent, West Greenland: Hydrologic implications


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1. Introduction

[2] Recent studies have suggested that the Greenland Ice Sheet possesses a basal sliding mechanism similar to that of an alpine glacier [e.g., Bartholomaus et al., 2008], whereby enhanced basal sliding occurs when surface meltwater input exceeds subglacial transmission capacity and pressurizes the subglacial hydrology system [Shepherd et al., 2009; Bartholomew et al., 2010; Colgan et al., 2011]. Enhanced melt season basal sliding constitutes a relatively larger fraction of annual displacement in land-terminating glaciers than in marine-terminating glaciers [Loughin et al., 2008]. While many studies suggest that relatively small increases in future surface meltwater production may result in disproportionately large increases in basal sliding velocity [Zwally et al., 2002; Shepherd et al., 2009; Bartholomew et al., 2010; Schoof, 2010], other studies suggest this will result in a transition to more efficient subglacial drainage and a net decrease in basal sliding velocity [van de Wal et al., 2008; Schoof, 2010; Sundal et al., 2011].

[3] Meltwater is transferred from the supraglacial system to the subglacial system via either moulins or crevasses (Figure 1). Moulins are near-vertical conduits through the ice that have been demonstrated to rapidly transmit any fluctuations in surface meltwater production to the subglacial system [Shepherd et al., 2009]. Moulins concentrate surface meltwater production from relatively large areas for “point” delivery to the subglacial system. Crevasses collect surface meltwater from comparatively small areas to drain via a more “distributed” englacial network. Thus, the presence of crevasses influences both supra- and englacial water routing. Crevase propagation can precondition ice for moulin formation at crevasse field boundaries [Holmlund, 1988; Phillips et al., 2011]. While crevasses and moulins occur throughout a common elevation band referred to as the “runoff zone”, substantial supraglacial river/moulin systems and lakes typically do not occur within crevassed areas in West Greenland [Thomsen et al., 1988; Phillips et al., 2011].

[4] We find that changes in marginal ice geometry (i.e., thinning and steepening), due to a combination of enhanced surface ablation and the recent acceleration of nearby Jakobshavn Isbrae (JI), have increased the relative fraction of crevasse-type drainage areas in the Sermeq Avannarleq (SA) ablation zone in West Greenland. We provide a first-order demonstration that the characteristic transfer time for supraglacial meltwater to reach the subglacial system is approximately two orders of magnitude faster for moulin-type drainage than for crevasse-type drainage. Thus, we suggest that moulin-type drainage is more efficient in propagating surface meltwater fluctuations (diurnal and otherwise) to the subglacial system than crevasse-type drainage (which attenuates meltwater fluctuations). As enhanced basal sliding is associated with meltwater “pulses”, rather than sustained meltwater input, a net transition from moulin to crevasse-type drainage may ultimately modify the basal sliding response of land-terminating portions of the ice sheet that are not presently crevassed to surface meltwater input.

2. Methods

[5] The SA ablation zone is comprised of predominantly land-terminating ice, with a relatively small tidewater glacier that calves into a sidearm of Jakobshavn Fjord. In 1985 the Geological Survey of Greenland conducted an intensive survey of the SA ablation zone that included the acquisition of high-quality panchromatic aerial photographs on 10 and
Crevassed area polygons. By 10 and 100 m buffers around each respective set of only the most "crevasse extent," and crevasses >10 m wide (this includes >2 m wide (given the pixel size this is equivalent to 60% over the same period (from 83 ± 26 km² to 99 ± 28 km²; not shown). The increase in total crevasse extent, as well as severely crevassed extent, may be generalized as an expansion of existing crevasse fields. While a qualitative assessment suggests a corresponding decrease in the area occupied by supraglacial rivers, we do not assess this decrease in a quantitative fashion here. We simply note that crevasses and prominent supraglacial hydrologic features (e.g., several km lake and river systems) are typically mutually exclusive [cf. Thomsen et al., 1988]. Therefore, the net increase in ice sheet area drained by crevase systems between 1985 and 2009 implies a net decrease in ice sheet area drained by river/moulin systems (cf. Figures 2c and 2d).

3. Results

3.1 Increased Crevasse Extent

Previous research suggests that the land-terminating ice within the study area has experienced widespread thinning in excess of 65 m between 1985 and 2007, with a local maximum of 104 ± 8 m [Motyka et al., 2010]. The cause of this thinning is discussed in the next paragraph. Comparison of a 1985 digital elevation model (DEM), derived from a Geological Survey of Greenland supraglacial topography map [Thomsen et al., 1988], and a 2009 DEM, derived from the ASTER Global DEM (http://asterweb.jpl.nasa.gov/gdem.asp), indicates that mean ice surface slope in the SA ablation zone increased ~0.10° (~7%) over the 24-year period. A two-tailed paired t-test indicates that this increase in mean ice slope from 1.45° (s.d. 0.98°) in 1985 to 1.55° (s.d. 0.91°) in 2009 is significant at the p < 0.001 level (Figure 3). This steepening is an anticipated result of previously observed, and ongoing, low elevation thinning and high elevation thickening of the ice sheet [Parizek and Alley, 2004; Luthcke et al., 2006]. As the lithostatic stress opposing crevasse propagation is proportional to ice thickness and the tensile stress promoting crevasse propagation is proportional to local surface slope [van der Veen, 1998], changes in ice thickness and local surface slope influence crevasse propagation. Ice thinning and steepening both act to enhance crevasse propagation. Application of the crevasse model proposed by van der Veen [1998] suggests that for an initial ice thicknesses of 500 m and an crevasse depth of 20 m (values representative of the SA ablation zone), the combination of a 65 m decrease in ice thickness and a 1° increase in local surface slope could result in a 50% increase in crevasse extent within a given time period (cf. Figures 2c and 2d).

4. Discussion

4.1 Increased Crevasse Extent

24 July 1985 [Thomsen et al., 1988]. These photographs were recently digitized and aggregated into an orthomosaic with the intent of studying changes in the ice geometry of nearby JI (~30 km southeast of SA [Motyka et al., 2010]). We compare this historical orthomosaic with a commercially acquired panchromatic 15 July 2009 WorldView-1 image of the SA ablation zone to assess changes in surface morphology over a 24-year period. As the horizontal accuracies of both the orthomosaic (~2 m resolution) and WorldView-1 image (resampled to 2.2 m resolution) are estimated to be ±10 m, the total positional uncertainty when comparing a given location in the 1985 and 2009 images is taken as ±10 m.

[6] We compared the distribution of supraglacial hydrology features (i.e., rivers and lakes) and crevassed area extent in the 1985 and 2009 images. Supraglacial hydrology features were manually delineated from both imagery datasets. Crevassed area extent was delineated from both images using a Roberts cross edge detector in ENVI. This method involves convolving a gradient operator that enhances areas of the image with sharp changes in pixel value in a given direction (which is characteristic of a crevasse field). After manually digitizing the crevassed area boundaries identified by the Roberts operator results, we performed a manual inspection using the original imagery to verify the accuracy of polygons in representing crevassed regions. We delineated two sets of crevassed area extent polygons: crevasses >2 m wide (given the pixel size this is equivalent to “total” crevasse extent), and crevasses >10 m wide (this includes only the most “severely” crevassed areas). We assess our uncertainty in crevassed area extent as the total area occupied by 10 and 100 m buffers around each respective set of crevassed area polygons.

[7] Inter-annual variations in surface mass balance conditions influence the identification of surface features. Cumulative melt intensity in both 1985 and 2009 was compared using positive degree days (PDDs) observed at the nearby Danish Meteorological Institute (DMI) Ilulissat weather station (WMO 04216/04221; ~35 km east of the ice sheet margin). This comparison suggests that the 1985 melt season was ~12% more intense than the 2009 melt season at the time of image acquisition (416 versus 370 PDD, respectively).

In order to minimize the potential differences in crevasse extent between 1985 and 2009 attributable to differences in snow cover and detection, we use the 1985 snowline position [Thomsen et al., 1988] as the upglacier limit of the comparison area. We note that the 2009 snowline also shares a similar position (Figure 2). We use the 2009 margin as the downglacier limit of the comparison area.
in ice thickness and a 0.10° increase in surface slope increases crevasse depth by ∼30%. Given that crevasse propagation is greatly enhanced by hydrofracture when crevasses are water-filled [van der Veen, 1998], a substantial increase in surface meltwater production since 1985 is also expected to facilitate crevasse propagation. As a crevasse field delineates the local extent of a critical ratio between tensile and lithostatic stresses, widespread thinning and steepening is expected to cause an existing crevasse field to expand outwards.

[10] The general continuity equation for ice suggests that there are two possible causes of ice sheet thinning: a decrease in surface mass balance (b) and an increase in divergence of ice flux [Hooke, 2005]. In situ surface mass balance observations suggest that the maximum thinning observed by Motyka et al. [2010] cannot be attributed to surface mass balance alone, which would invoke \( b \approx -4.7 \) m/a over the study period [cf. Fausto et al., 2009]. An increase in the divergence of ice flux, either in the along or across-flow directions, must therefore be invoked to explain excessive thinning. A comparison of 1985 velocity vectors derived from photogrammetry [Fastook et al., 1995] with 2005/06 velocity vectors derived from InSAR observations [Joughin et al., 2010] suggests that the recent acceleration of JI has increased southbound ice flow at the expense of westbound ice flow in the SA ablation zone (Figure 4). Surface crevasses, which typically form perpendicular to a glacier’s
longitudinal stress field (i.e., across-flow under extensional flow), have rotated by up to 45° throughout the study area in directions consistent with this notion (Figures 4b and 4c). We expect that the majority of this reorganization likely occurred after 1998, rather than in a monotonic fashion, following the 1997 breakup of the JI ice tongue and the consequent doubling of ice speed and dramatic changes in ice geometry [Motyka et al., 2010].

Previous studies indicate that recent changes at JI are part of a widespread change in the low elevation velocity and geometry of the Greenland Ice Sheet [Luthcke et al., 2006; Rignot and Kanagaratnam, 2006]. Around the Greenland Ice Sheet, the “runoff zone” elevation band, which is characterized by meltwater runoff via crevasses and moulins [Phillips et al., 2011], resides below the higher elevation “dark zone”, the elevation band of maximum meltwater accumulation in supraglacial lakes [Greuell, 2000]. Thus, our observations in the SA runoff zone may be part of an ice sheet-wide response in crevasse extent to changing ice velocity and geometry throughout the low elevation runoff zone.

Figure 3. Change in ice surface slope derived from a comparison of 1985 [Thomson et al., 1988] and 2009 (http://asterweb.jpl.nasa.gov/gdem.asp) DEMs at 500 m grid spacing. Values in parentheses indicate the percent area in each class. Inset: Scatter plot of 1985 and 2009 ice surface slopes at each DEM grid point with line $y = x$ shown for reference.

Figure 4. (a) Comparison between 1985 [Fastook et al., 1995] and 2005/06 [Joughin et al., 2010] velocity vectors in the Jakobshavn Isbrae region. Colored contours denote change in absolute velocity. The dashed magenta box denotes the extent of Figures 2a and 2b. The magenta detail box is a highlighted region in which crevasse rotation between (b) 1985 and (c) 2009 is shown. Arrows denote the approximate local crevasse orientation in each year.
4.2. Hydrologic Implications

[12] Meltwater is transferred from the supraglacial system to the subglacial system via either moulins or crevasses. Assuming that changes in englacial water volume due to internal meltwater generation and deformational closure are negligible, the rate of change of water volume \( \frac{dV}{dt} \) in either type of englacial transfer system may be conceptualized as the difference between the rate of surface meltwater input \( (a_s A) \) and the rate of englacial discharge (which we represent as \( V/\tau \)):

\[
\frac{dV}{dt} = a_s A - \frac{V}{\tau}
\]

(1)

where \( a_s \) is the surface ablation rate, \( A \) is the supraglacial catchment area, and \( \tau \) is a characteristic englacial transfer time. This characteristic transfer time represents the mean time required for supraglacial meltwater input to travel from the ice surface to the subglacial system. This formulation assumes that the englacial system behaves as a linear reservoir [Jansson et al., 2003]. The total englacial conduit volume \( V \) may be estimated as:

\[
V \approx H \cdot S_e
\]

(2)

where \( H \) is ice thickness and \( S_e \) is the depth-averaged cross-sectional area of the englacial conduit draining either a crevasse or a moulin. This formulation assumes that water and conduit volumes are equivalent (i.e., discounts open channel flow), and also implicitly takes ice thickness as a proxy for englacial travel distance. This neglects the reality that the englacial conduits draining moulins and crevasses may travel through the ice column at varying angles off-vertical.

[13] Assuming that englacial conduit volume evolves to achieve a quasi-steady-state diurnal cycle (i.e., \( \int (dV/dt) dt \approx 0 \) over the diurnal cycle), the characteristic englacial transfer time may be approximated as:

\[
\tau \approx \left( \frac{H \cdot S_e}{A \cdot a_s} \right)
\]

(3)

This quasi-steady-state assumption is reasonable, given that empirical observations on alpine glaciers suggest that the 1/\( e \) timescale for conduit closure beneath \( \sim 500 \text{ m} \) of ice is expected to be \( \sim 5 \text{ hr} \) when water pressures are low [Bartholomaus et al., 2008]. Furthermore, assuming high temporal resolution GPS observations near “K-transect” \( (250 \text{ km south of SA}) \) are broadly representative of the Greenland Ice Sheet, the “summer speedup” event is actually comprised of a series of diurnal events [Shepherd et al., 2009]. Additionally, at SA, there is observational evidence of the persistence of englacial hydrologic features throughout the year [Catania and Neumann, 2010] and model inferences that englacial hydraulic head fluctuations annually within a relatively small range [Colgan et al., 2011].

[14] To provide a first-order demonstration of the fundamental difference in englacial transfer time between crevasse-type and moulin-type drainage systems, we evaluate equation (3) for both systems using order-of-magnitude parameter estimates. For moulins, we assume that englacial conduits are generally circular in shape, with cross-sectional area approximated as \( \pi \tau^2 \) (where \( \tau \) is depth-averaged englacial conduit radius). For crevasses, we do not impose a circular cross-sectional conduit geometry, but rather approximate \( S_e \) as \( w l \) (where \( l \) is crevasse length and \( w \) is depth-averaged width of the crevasse and underlying distributed englacial system). In both scenarios we take mean daily ablation rate as \( 4 \text{ cm/d} \) and ice thickness as \( 500 \text{ m} \). In the moulin scenario we assume that 1 km\(^2\) of ice sheet area drains into a single moulin with \( \tau = 1 \text{ m} \). In the crevasse scenario we assume that the same ice sheet area is drained by multiple crevasses, with each crevasse draining an area of \( A = dl \) (where \( d \) is a mean crevasse spacing of 100 m) and assume \( \tau = 0.1 \text{ m} \). Using these specified values, the mean englacial transfer times for moulin-type and crevasse-type drainage are \( \sim 1 \text{ hr} \) and \( \sim 12 \text{ days} \), respectively. As \( H/a_s \) may be regarded as constant in a given region, \( S_e/A \) controls this \( \sim 200 \)-fold difference in \( \tau \). This suggests that drainage systems with relatively large \( S_e/A \) (i.e., crevasse-type) will attenuate surface meltwater fluctuations in comparison to those with relatively small \( S_e/A \) drainage (i.e., moulin-type). Decreasing the assumed \( \tau \) by an order-of-magnitude yields \( \tau \approx 1.25 \text{ days} \) (i.e., crevasse-type drainage still remains an order-of-magnitude slower than moulin-type drainage), while increasing the assumed \( \tau \) further increases the discrepancy between crevasse-type and moulin-type drainages.

[15] The attenuation of meltwater fluctuations can also be demonstrated by analytically solving equation (1). This reveals an inverse relation between englacial transfer time and englacial discharge variability. To facilitate an analytical solution, we approximate surface ablation rate \( (a_s) \) as a sinusoidal function of the form:

\[
\dot{a}_s(t) = a_s \left[ \frac{1}{2} \left( 1 - \sin \left( \frac{2 \pi t}{T} \right) \right) \right]
\]

(4)

where \( t \) is time and \( T \) is period of the surface ablation cycle (24 hr). Using this surface ablation forcing, the analytical solution of equation (1) (after many cycles of forcing (i.e., loss of memory of the initial water volume) may be expressed as:

\[
V(t) = \frac{a_s A \tau}{2} \left[ 1 + \left( \frac{1 - 4 \pi^2 \tau^2 / T^2}{\sqrt{1 + 4 \pi^2 \tau^2 / T^2}} \cos \left( \frac{2 \pi t}{T} - \varphi \right) \right) \right]
\]

(5)

where \( V(t) \) is the total englacial water volume and \( \varphi \) is a phase angle equal to:

\[
\varphi = \tan^{-1} \left( \frac{T}{2 \pi \tau} \right)
\]

(6)

Conduit discharge is obtained by dividing equation (5) by \( \tau \). [16] Thus, it becomes evident that the amplitude of conduit discharge is inversely proportional to characteristic englacial transfer time. For small \( \tau \) (i.e., \( \leq T/2\pi \)), as expected in moulin-type drainage, variations in conduit discharge precisely follow those of surface ablation forcing and there is little attenuation of the diurnal signal. For large \( \tau \) (i.e., \( \geq T/2\pi \)), as expected in crevasse-type drainage, englacial discharge exhibits smaller amplitude variations. As basal sliding requires meltwater inputs to overwhelm the transmission capacity of the subglacial system, it is reasonable to expect efficient moulin-type drainage (i.e., meltwater “pulses”) to enhance basal sliding more than inefficient crevasse-type drainage (i.e., sustained meltwater input). In addition to “englacial hydraulic adjustment” [van de Wal et al., 2008], decreased
basal sliding could provide a complementary explanation for observations of decreasing mean annual velocities of land-terminating ice along K-transect.

[17] We note that the potential change from moulin to crevasse-type drainage is a one-time transition limited to areas of the low elevation ice sheet runoff zone that are not presently crevassed; ice sheet areas that are already crevassed would not be expected to be sensitive to this mechanism. As enhanced melt season basal sliding contributes to a larger fraction of annual displacement in land-terminating glaciers than marine-terminating glaciers [Joughin et al., 2008], potential decreases in basal sliding velocity would exert a greater effect on land-terminating glaciers than on marine-terminating glaciers.

4.3. Other Implications

[18] While a net increase in crevasse extent may result in a net decrease in basal sliding sensitivity, a net increase in crevasse extent may enhance ice flow through other mechanisms. As cryo–hydrologic warming is highly sensitive to the mean spacing of englacial hydrologic pathways [Phillips et al., 2010], a net increase in crevasse extent is expected to expose an increased area of the ice sheet to closely-spaced hydrologic pathways that facilitate cryo-hydrologic warming. These pathways would be capable of raising ice temperatures, thereby enhancing deformational ice velocities, in newly crevassed regions of the runoff zone [van der Veen et al., 2011]. A positive feedback could be possible if the subsequent change in the ice velocity field further expands crevassed areas. Thus, predicting the implications of a change in crevasse extent on ice flow is complicated by counteracting processes. In addition to ice flow, the presence of crevasses has been shown to more than double the absorption of solar radiation (and hence enhance surface ablation) in comparison to ice sheet areas without crevasses [Pfeffer and Bretherton, 1987]. Thus, a spatially widespread increase in crevasse extent (i.e., an increase in extent characterized by the outward expansion of multiple existing crevasse fields) is expected to result in a widespread increase in summer ablation.

5. Summary Remarks

[19] We suggest that the recent acceleration of Jakobshavn Isbrae has resulted in an increase in crevasse extent in the Sermeq Avannaq Fjord alication zone, due to a combination of ice thinning and steepening surface slope. We argue that the resultant change from moulin-type to crevasse-type drainage is expected to result in a net dampening of basal sliding sensitivity to surface meltwater input. This dampening would be most significant in portions of the ice sheet that are land-terminating and not presently crevassed. Due to the absence of an ice sheet-wide time series of crevasse extent, the potential prevalence of this transition outside our study area is not known at present. We expect, however, that our observations at Sermeq Avannaq are representative of a widespread response to low elevation thinning and outlet glacier acceleration throughout the low elevation runoff zone of the Greenland Ice Sheet. We acknowledge that any potential reduction of ice discharge stemming from decreased basal sliding may be offset by other counteracting processes. In particular, increased crevasse extent may enhance surface ablation through increased absorption of solar radiation, and increase deformational ice velocities by exposing an increased ice sheet area to closely-spaced hydrologic pathways that facilitate cryo-hydrologic warming.

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