AGU PUBLICATIONS

Geophysical Research Letters

RESEARCH LETTER

10.1002/2015GL063192

Key Points:
- Greenland Ice Sheet meltwater volumes are increasing notably at high elevations
- Low strain rates limit the likelihood of moulin formation at high elevations
- High-elevation meltwater will reach an already wet bed at lower elevations

Supporting Information:
- Text S1 and Figures S1 and S2

Correspondence to:
K. Poinar,
kpoinar@apl.washington.edu

Citation:
Poinar, K., I. Joughin, S. B. Das, M. D. Behn, J. T. M. Lenaerts, and M. R. van den Broeke (2015), Limits to future expansion of surface-melt-enhanced ice flow into the interior of western Greenland, *Geophys. Res. Lett.*, *42*, 1800–1807, doi:10.1002/2015GL063192.

Received 23 JAN 2015
Accepted 20 FEB 2015
Accepted article online 24 FEB 2015
Published online 24 MAR 2015

Limits to future expansion of surface-melt-enhanced ice flow into the interior of western Greenland

Kristin Poinar[1], Ian Joughin[1], Sarah B. Das[2], Mark D. Behn[2], Jan T. M. Lenaerts[3], and Michiel R. van den Broeke[3]

[1]Polar Science Center, Applied Physics Laboratory, University of Washington, Seattle, Washington, USA, [2]Department of Marine Geology and Geophysics, Woods Hole Oceanographic Institution, Woods Hole, Massachusetts, USA, [3]Institute for Marine and Atmospheric Research, Utrecht University, Utrecht, Netherlands

Abstract Moulins are important conduits for surface meltwater to reach the bed of the Greenland Ice Sheet. It has been proposed that in a warming climate, newly formed moulins associated with the inland migration of supraglacial lakes could introduce surface melt to new regions of the bed, introducing or enhancing sliding there. By examining surface strain rates, we found that the upper limit to where crevasses, and therefore moulins, are likely to form is ~1600 m. This is also roughly the elevation above which lakes do not drain completely. Thus, meltwater above this elevation will largely flow tens of kilometers through surface streams into existing moulins downstream. Furthermore, results from a thermal ice sheet model indicate that the ~1600 m crevassing limit is well below the wet-frozen basal transition (~2000 m). Together, these data sets suggest that new supraglacial lakes will have a limited effect on the inland expansion of melt-induced seasonal acceleration.

1. Introduction

A major concern at the time of the fourth Intergovernmental Panel on Climate Change report [*Lemke et al.*, 2007] was that surface meltwater could lubricate the base of the Greenland Ice Sheet, enhancing sliding and increasing the seaward flux of ice. Initial observations indicated that surface-melt-induced basal lubrication could accelerate ice sheet flow seasonally by a factor of 2 or more in slow-moving regions (<100 m/yr) [*Zwally et al.*, 2002]. Since then, it has been shown that the modulation of ice flow by basal lubrication is a time-evolving, nonlinear process, whereby the subglacial drainage system adapts to sustained periods of melt so that its sensitivity to melt declines over the melt season [*Schoof*, 2010; *Bartholomew et al.*, 2011a]. Thus, while the input of water initially leads to an ice-flow speedup, increasing the volume of water also increases drainage efficiency, stabilizing ice motion, and giving rise to the observed seasonal cycle in ice motion [*Sundal et al.*, 2009; *Palmer et al.*, 2011; *Bartholomew et al.*, 2011b]. As a result, the time-averaged velocity may be independent of [*van de Wal et al.*, 2008] or even negatively correlated to [*Sundal et al.*, 2011] surface melt volume. Surface melt also appears to have little effect on the acceleration of fast-moving outlet glaciers [*Joughin et al.*, 2008].

As the climate warms, melting is increasing at higher elevations (farther inland) on the ice sheet surface [*van de Wal et al.*, 2008; *Hoffman et al.*, 2011; *Howat et al.*, 2013; *Fitzpatrick et al.*, 2014]. An outstanding question is whether this surface meltwater can access the bed through moulins, and if so, whether the seasonal melt-induced speedup that presently occurs at lower elevations will be observed at new inland locations [*Sundal et al.*, 2009; *Palmer et al.*, 2011; *Bartholomew et al.*, 2011b; *Howat et al.*, 2013]. In particular, if surface melt reaches the bed in areas where the bed is frozen, its latent heat could warm the basal ice to the melting point and more permanently increase its seaward velocity [*Parizek and Alley*, 2004; *Bamber et al.*, 2007; *Sundal et al.*, 2009; *Howat et al.*, 2013].

Moulins often form from hydrofracture under or near supraglacial lakes. Hydrofracture occurs when the driving stress associated with the differential pressure between water in a crevasse and the surrounding lower density ice exceeds the fracture toughness of ice, forcing the crevasse tip downward; this can continue to the bed if the water supply (e.g., supraglacial lakes) is sufficient [*Alley et al.*, 2005; *Krawczynski et al.*, 2009]. Hydrofracture can occur directly beneath a lake or where an overflow stream from a lake reaches a crevasse [*Das et al.*, 2008; *Tedesco et al.*, 2013]. At lower elevations, moulins tend to form near the lakes they drain [*Joughin et al.*, 2013]. We refer to drainage that occurs beneath a lake or through a moulin a few kilometers away as "local" drainage. By contrast, "nonlocal" drainage occurs when melt travels >~10 km downstream before entering a moulin.

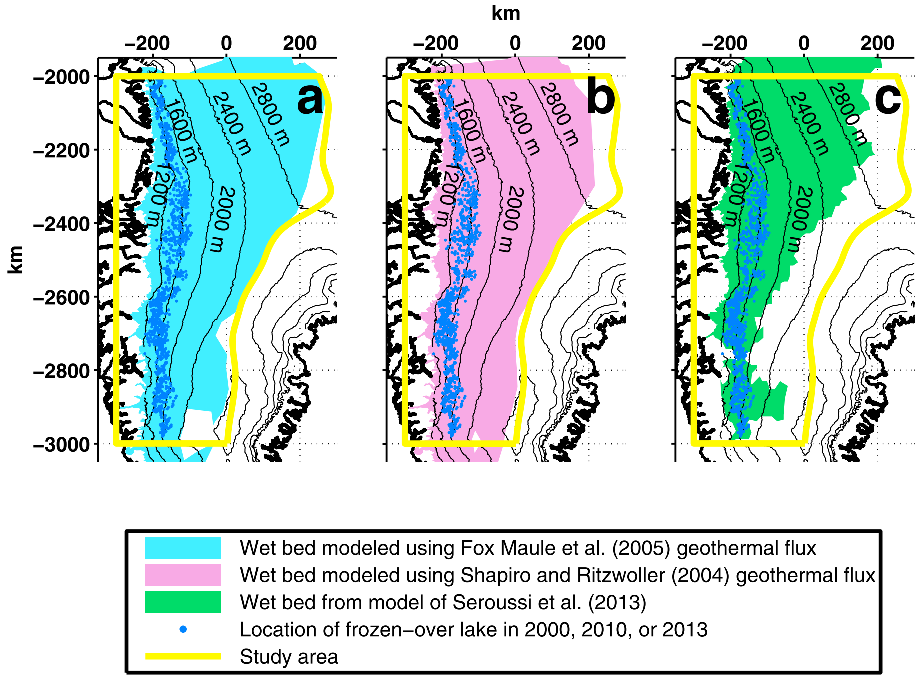

Figure 1. Locations of frozen-over lakes identified using satellite data from 2000, 2010, and 2013 (blue dots) overlain on wet-bedded areas (shaded) modeled using geothermal flux from (a) *Fox Maule et al.* [2005] and (b) *Shapiro and Ritzwoller* [2004]. (c) Wet-bedded areas modeled by *Seroussi et al.* [2013]. The yellow box shows the study area, which extends east to the divide. Our model (Figures 1a and 1b) was not run east of the divide; the white areas west of the divide indicate frozen beds. The black lines indicate the surface elevation contours from the Greenland Ice Mapping Project (GIMP) digital elevation model (DEM) [*Howat et al.*, 2014].

An ongoing concern is the extent to which high-elevation surface melt can drain locally [e.g., *Selmes et al.*, 2013]. This has important implications for ice dynamics because it determines where surface melt lubricates the bed. In this study, we apply numerical models and remote sensing to investigate the extent to which the inland propagation of surface melt is likely to form new moulins that could influence the flow and stability of the Greenland Ice Sheet.

2. Extent of Wet Bed

We first assess the vulnerability of the thermal state of the ice sheet bed to future incursions of surface meltwater. The basal velocity of an ice sheet is influenced by (1) the extent to which the bed is melted or frozen, and (2) the efficiency of the basal drainage system. To assess the former, we performed a series of thermal model calculations (see the supporting information) to identify those areas of the bed that are frozen and those that are melted within a 420,000 km^2 study area from the coast to the divide in western Greenland (Figure 1). Due to the uncertainty in geothermal flux, we varied this parameter in our models. Figures 1a and 1b show the results of our model forced with geothermal flux data sets from *Fox Maule et al.* [2005] and *Shapiro and Ritzwoller* [2004], respectively. Figure 1c shows the results of a model by *Seroussi et al.* [2013], which uses a modified version of *Shapiro and Ritzwoller*'s [2004] geothermal flux. The three panels are similar and also broadly agree with other modeling results (see the supporting information). A key result of these calculations is that little area below 2000 m elevation within our study area has a frozen bed. Thus, we use the 2000 m contour as a conservative estimate of the melted-frozen basal boundary; i.e., the true boundary is likely somewhat farther inland.

3. Expansion of Surface Melt

Next, we analyze the ongoing inland migration of surface melt. Figure 2a shows the annual snow and ice melt versus elevation within our study area. Melt at all elevations has increased from the climatological average

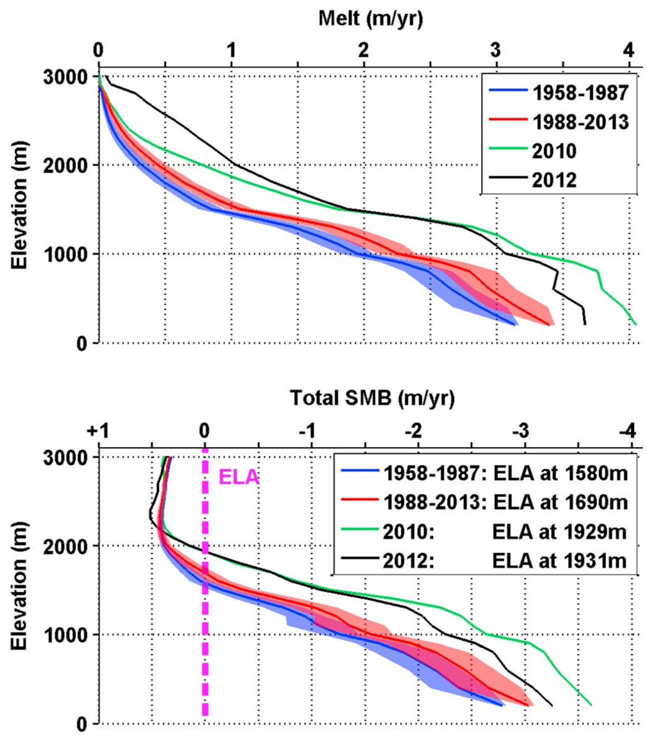

Figure 2. Average annual (a) surface melt and (b) surface mass balance from RACMO2.3 within the study region as a function of elevation. The blue lines show the climatological average over 1958–1987 with the 95% confidence interval for the mean over this period (shaded). The red shows the same over the 26 year period from 1988 to 2013. See the supporting information for details. The exceptional melt years 2010 and 2012 are shown in green and black, respectively.

(defined as 1958–1987) in the period of 1988–2013. The difference between the means of the two periods is positive at 95% confidence at all elevations from 400–2600 m (see the supporting information).

The increase in surface melt has raised the equilibrium line altitude (ELA) substantially within our study area (Figure 2b); the ELA now (1988–2013) averages 1690 m elevation, which is 110 m higher and 12 km farther inland than in 1958–1987 (1580 m). In the extreme melt years of 2010 and 2012, the ELA temporarily rose an additional 240 m (31 km farther inland). These results corroborate shorter-term regional observations [e.g., *van de Wal et al.*, 2012]. As the ELA rises, so too does the elevation of the bare ice zone, where supraglacial lakes are able to coalesce and potentially form moulins [*Howat et al.*, 2013].

4. Variations in Supraglacial Lakes and Streams With Elevation

The site of origin of a parcel of surface meltwater is often not coincident with the site where it enters the basal hydrologic system. Thus, although some surface melt reaches the bed locally, some does not, with the relative amounts and transport distances determined by its specific routing through the surface hydrological network of surface lakes and streams. To examine these characteristics, we used Landsat images (pixel size of 30 m) and satellite-based radar images (pixel size of 20 m) to study supraglacial lakes, moulins, and streams in our study area (see the supporting information). In particular, we identified all lakes with "lake ice," still visible on their surfaces in the summers of 2000, 2010, and 2013, which had the greatest number of cloud-free Landsat images available. A frozen ice cover indicates that a lake contained water that overwintered [*Darnell et al.*, 2013]; i.e., it did not drain completely in the previous summer. Figure 1 shows the locations of these "frozen-over" lakes. Their distribution generally spans 1100–1800 m elevation and typically ends just above the ELA.

Unlike more comprehensive mappings [e.g., *Lampkin*, 2011; *Selmes et al.*, 2011, 2013], we did not attempt a thorough analysis of lake size, depth, or other characteristics. Instead, we qualitatively studied how the size and drainage nature of supraglacial lakes varie with elevation. To illustrate the results of this analysis, in Figure 3, we show representative 10×10 km regions at four elevations spanning 1000–1900 m within our study area. At 1000 m elevation, summertime images (Figure 3a) show lakes that are deep blue in color and generally hundreds to thousands of meters in diameter. Wintertime images (Figure 3b) show moulins in or near most drained lake basins, fed by local streams. At 1300 m elevation (Figure 3c), the lakes are slightly larger [*Sundal et al.*, 2009; *Fitzpatrick et al.*, 2014], and some have ice covers that are visible in the wintertime radar data (Figure 3d) as bright or dark patches, depending on whether the ice cover is floating or frozen through, respectively [*Jeffries et al.*, 1994; *Surdu et al.*, 2014]. Evidence from ice-penetrating radar further suggests that these lakes overwinter under snow and lake ice [*Koenig et al.*, 2014]. At 1600 m elevation (Figure 3e), most lakes are ice covered and can exceed 5 km^2 in size. Wintertime images (Figure 3f) also show the presence of floating and frozen-through lake ice here. At 1900 m and above (Figures 3g and 3h), ice cover

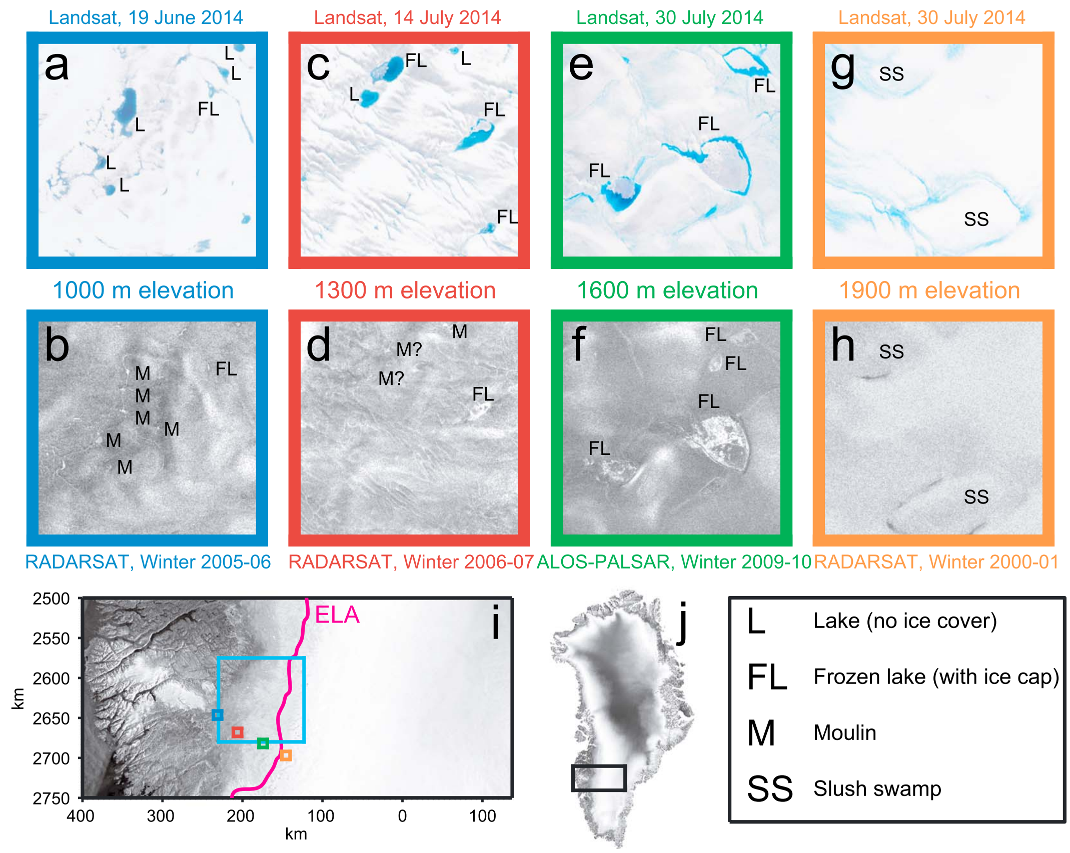

Figure 3. The nature of surface hydrologic features at multiple elevations in southwestern Greenland. (top) Summertime visible-spectrum Landsat images. (bottom) Wintertime radar images of the same scenes. The date, satellite, and approximate elevation [*Howat et al.*, 2014] of each 10 × 10 km scene are shown; elevations increase from right to left. (i) The locations of each (a–h) scene in colors corresponding to the image borders. The cyan square shows the location of Figure 4a; the 1958–2013 average ELA from RACMO2.3 is shown in magenta. Supraglacial features are labeled as follows: L = lake, FL = frozen-over lake, M = moulin, and SS = slush swamp.

persists, but distinct lakes give way to broad topographic lows with fuzzy boundaries. These "slush swamps" [*Bamber et al.*, 2007] occur because the firn absorbs much of the meltwater there, making the borders of water-filled basins more diffuse.

This analysis suggests that high-elevation frozen-over lakes can drain some of their water via overflow streams. Figure 4a shows the 312 major surface streams we identified using satellite imagery over a 6000 km^2 area spanning approximately 1100–1800 m elevation in southwestern Greenland (cyan box in Figure 3i). On average, the streams in this area are 10 km long and descend 60 m elevation, although this will vary in regions with different surface slopes, ice thicknesses, or other characteristics. The streams tend to lengthen at higher elevations, in agreement with stream mapping performed 300 km farther north [*Joughin et al.*, 2013]. For example, streams originating at 1200–1300 and 1700–1800 m elevation are 6 and 40 km long on average, respectively (Figure S2 in the supporting information). The highest-elevation stream termination (moulin) we identified was at 1580 m.

5. Variation of Strain Rate With Elevation

Hydrofracturing from the ice sheet surface to the bed requires both a crack and a sufficient input of meltwater to drive it through the ice sheet [*van der Veen*, 1998; *Krawczynski et al.*, 2009]. If new hydrofractures are to form in inland areas of the ice sheet where increased meltwater volumes are beginning to form supraglacial lakes [*Howat et al.*, 2013], cracks in these regions are required [*Alley et al.*, 2005]. These cracks can either (a) initiate directly beneath the lake water within the basin or (b) exist outside the basin. In the former case, it is sometimes

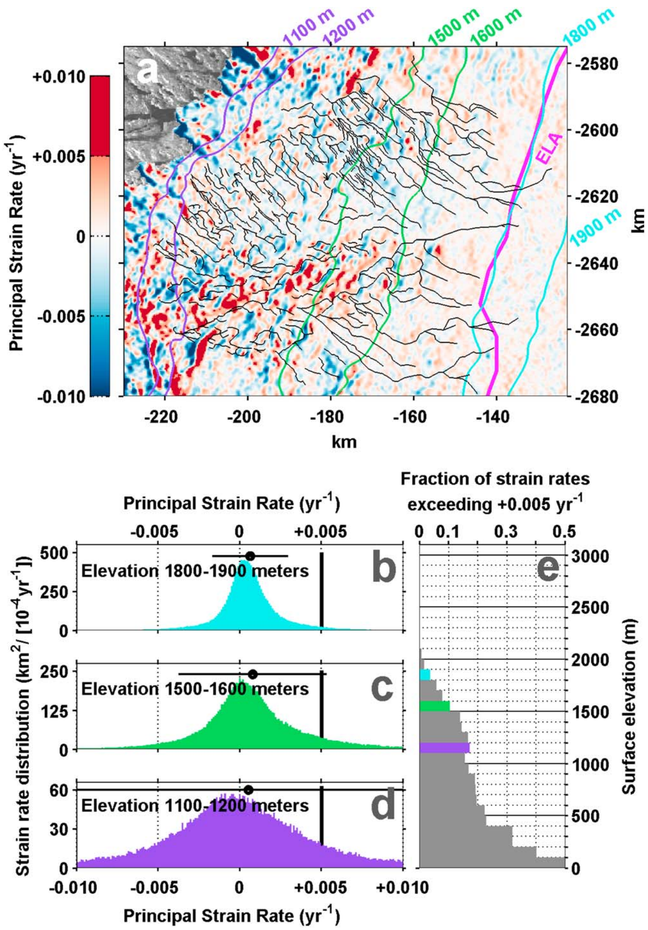

Figure 4. (a) Surface streams (black) identified from 24 June 2004 and 30 July 2014 Landsat images overlain on the principal strain rate field (colored). Positive strain rates indicate extension. Elevation contours from the GIMP DEM [*Howat et al.*, 2014] are labeled, and the 1958–2013 average ELA from RACMO2.3 is shown in magenta. (b–d) Histograms of the strain rates in western Greenland at various elevations: (Figure 4b) 1800–1900 m, (Figure 4c) 1500–1600 m, and (Figure 4d) 1100–1200 m. The approximate strain rate threshold for crevassing, +0.005/year, is shown as a vertical black bar. The mean strain rate (black dot) and one standard deviation (black horizontal line) for each elevation band are also indicated. (e) Bar graph showing the fraction of the entire western Greenland study region (Figure 1) where the tail of the strain rate distribution exceeds the +0.005/year fracture threshold, binned by elevation band. The results derived from the histograms (Figures 4b–4d) are color coded.

assumed that when a certain energy threshold or melt volume is reached, a lake can self-generate the crack necessary for drainage [*Georgiou et al.*, 2009; *Johansson et al.*, 2013]. However, whether lakes drain is not well correlated to their volume [*Fitzpatrick et al.*, 2014], implying that some other drainage initiation mechanism is required [*Selmes et al.*, 2013]. For drainage outside the basin (b), overflow channels deliver lake water into a preexisting moulin or a crevasse that, through hydrofracture, becomes a moulin [*Catania et al.*, 2008; *Das et al.*, 2008; *Tedesco et al.*, 2013]. Together, these studies suggest that without an existing crevasse, hydrofracture cannot initiate.

To constrain where on the ice sheet hydrofracture may initiate, we use surface strain rates as a proxy to identify where crevasses are likely to form. A survey of submeter-resolution imagery [*Joughin et al.*, 2013] indicated a general correspondence between areas with principal strain rates above +0.005/year and areas where crevasses are visible. More traditional methods of identifying where ice may fracture are stress thresholds or fracture toughness values [*van der Veen*, 1998], but these are difficult to observe remotely. Thus, while approximate, surface strain rates have the advantage that they are easily obtained from available velocity data. With this in mind, we use principal strain rates exceeding +0.005/year to indicate where crevassing is likely to occur.

We calculate principal strain rates on a 1×1 km grid from the average 2007–2010 wintertime ice sheet surface velocity [*Joughin et al.*, 2010] (see the supporting information) and examine their variation with elevation over our study region (Figure 1). Figures 4b–4e show the distribution of strain rates with elevation. These quantities are inversely correlated: higher elevations have lower magnitude, less variable strain rates. For example, at 1100–1200 m elevation (Figures 3a and 3b), the strain rates span from -0.01 to +0.01/year and exceed the +0.005/year threshold in 17% of the surface area (Figure 4d). At 1500–1600 m elevation (Figures 3c–3f and 4c), the distribution of strain rates is narrower and 10% of the ice sheet area is susceptible to fracture formation. At 1800–1900 m (Figures 3g, 3h, and 4b), only 3.6% of the area likely is susceptible to crevassing.

6. Discussion

As we show with our analyses of the RACMO2.3 output (Figure 2), the ice sheet surface has been melting farther inland (or, equivalently, at higher elevations) over the last several decades. This surface meltwater may potentially reach the bed and seasonally accelerate ice flow at new inland locations, but for this to occur, the water must create new local hydrofractures that access the bed. Here we discuss the limits to inland expansion of hydrofracture and the likely consequences of new high-elevation surface melt reaching the bed.

6.1. Limitation to High-Elevation Hydrofracture

Our results and work by others [*Johansson et al.*, 2013] show that at low elevations (below ~1300 m; Figures 3a–3d), lakes are typically empty by the end of the melt season, draining either rapidly through hydrofractures [*Krawczynski et al.*, 2009] or more slowly through short (local) overflow streams. By contrast, higher-elevation lakes tend to drain incompletely (Figures 3e and 3f) through long overflow streams that carry the melt to distant (nonlocal) moulins (Figure 4a and Figure S1 in the supporting information) [*Johansson et al.*, 2013].

Figure 3 shows a general increase in the area of individual supraglacial lakes with elevation. Thicker ice dampens the surface expression of bedrock topography, passing only longer-wavelength basal undulations to the surface [*Gudmundsson*, 2003; *Lampkin and VanderBerg*, 2011]. Thus, higher-elevation lakes tend to be shallow and broad; accumulated meltwater can therefore sometimes overtop the lake basin. However, such overflow channels may not completely empty the lake; if the inflow of meltwater is slow, the turbulent outflow volume may be insufficient to incise a channel to lake-bottom level.

Water that is left in higher-elevation lakes after the melt season can form an ice cover, which sometimes superimposes onto the ice sheet ice (Figures 3c–3f) and is advected through and out of the lake basin. This process forms bands called lake ogives, which have been observed at several higher-elevation lakes [*Darnell et al.*, 2013]. The small width of lake ogives (~100 m) compared to that of the lakes (~2 km at 1600 m elevation) suggests that lake ogives may contribute only slightly to the water balance of a high-elevation lake.

The lake water that cannot drain into an in-basin hydrofracture or refreeze into lake ogives must eventually overflow the basin. At high elevations, this overflow often forms surface streams that extend tens of kilometers downglacier, sometimes cascading through multiple lakes (as seen between the two central lakes in Figures 3e and 3f and throughout Figure S1 in the supporting information) until the streams encounter a crevasse or moulin (Figures 4a and Figure S2 in the supporting information) and terminate. Such crevasses or moulins are likely to initiate in areas of tensile ice flow located near topographic highs [*Catania et al.*, 2008; *Das et al.*, 2008], yet streams flow toward topographic lows. This hydrologic characteristic is especially apparent at higher elevations, where short, efficient streams that connect lakes directly to nearby moulins are uncommon (Figures 3e–3h), and streams are generally longer (Figure S2 in the supporting information). Our data show that at 1600 and 1900 m elevation, the formation of surface crevasses is infrequent (8% of the ice sheet area) and rare (1.6%), respectively (Figure 4e). Thus, the seeding of moulins should be similarly infrequent and rare at these elevations, as *Selmes et al.* [2013] posited and observed. Our satellite imagery and previous observations [*Howat et al.*, 2013; *Johansson et al.*, 2013; *Selmes et al.*, 2013] indicate that although lakes exist above approximately 1600 m, many do not drain completely to moulins and tend to freeze over winter. *Catania et al.* [2008] also found a scarcity of moulins above the ELA despite the occurrence of lakes there.

The stream paths we identified do indicate that though rare, higher-elevation moulin formation does occur. Of the 134 stream termini (resulting from convergence of 312 total streams) we identified, 25 occurred above 1500 m. Thus, the spatial density of moulins in the 1600 km^2 area at 1500–1600 m (0.02/km^2) is a tenth or less of that observed at lower elevations [*Zwally et al.*, 2002; *Phillips et al.*, 2011].

In summary, relatively few high-elevation lakes drain their water to the bed locally. Instead, most high-elevation meltwater (originating above ~1600 m) appears to drain through surface streams into nonlocal, lower-elevation moulins (Figure 4a). Thus, we hypothesize that as climate change increases meltwater volumes at higher elevations (Figure 2), surface streams are likely to lengthen and transport that melt downglacier to where moulins are more likely to form. Consequently, in the coming decades, the effects of an increasing amount of high-elevation meltwater on ice sheet sliding velocity generally will be concentrated at lower elevations.

As the ice sheet evolves over the next hundreds to thousands of years, regions that are currently unlikely to hydrofracture (above ~1600 m) may thin enough to reach elevations (below ~1300 m) where hydrofracture is more common, as explored by *Leeson et al.* [2014]. This will bring water to these inland areas of the bed and may eventually activate sliding in new areas, forming a feedback for ice sheet destabilization. However, this enhanced basal motion will be a consequence of a large drawdown of Greenland's ice—not its cause.

6.2. Effect of New Meltwater at the Bed

The results of our thermal model suggest that the boundary between frozen and melted basal areas likely lies at or above the 2000 m surface elevation contour (Figure 1), although this result is somewhat sensitive to the geothermal flux. If we take this elevation as a rough boundary between frozen and wet beds in western Greenland, then a wet bed underlies the areas where lakes currently form and where they may form in the next several decades. Our data indicate that at elevations above 2000 m, hydrofracture to the bed will be extremely rare (Figure 4e), so that surface melt thawing a frozen bed is unlikely.

Although our data suggest that hydrofracture becomes less frequent with elevation, there is evidence of surface meltwater affecting the flow of high-elevation ice through longitudinal coupling. Observations of speedup at a site at 1840 m elevation, where there was no evidence of moulins or surface uplift, suggest that melt reached the bed tens of kilometers downstream [*Doyle et al.*, 2014]. The speedup was greatest in 2012, when melt was well above the current climatic mean (Figure 2); this response may thus be representative of the mean behavior several decades in the future. Nonetheless, the consequent seasonal (~8%) and annual (~2%) high-elevation speedups were relatively small [*Doyle et al.*, 2014].

7. Conclusions

Our results indicate that the likelihood of crevassing and, consequently, lake drainage via local spillover into crevasses decreases substantially with elevation. Thus, despite the observed inland migration of the ELA and surface melt in western Greenland, creation of new hydrofractures at high elevations is unlikely. Instead, most high-elevation meltwater likely will flow downhill via an extended network of surface streams and drain through existing moulins at lower elevations. Strain rate data indicate that relatively little water should reach the bed at elevations above ~1600 m, and this should diminish almost completely by 2000 m elevation. Furthermore, results from numerical models indicate that the bed in western Greenland is generally melted below ~2000 m surface elevation. Thus, it is unlikely that surface melt will reach the bed in regions where it could thaw the bed and produce large changes in ice sheet velocity. Instead, new water that reaches the bed is likely to have a modest effect similar to that already observed [e.g., *Bartholomew et al.*, 2011b; *Doyle et al.*, 2014]. Overall, in the next several decades, the inland migration of melt is unlikely to produce large changes in flow due to increased basal lubrication or the thawing of a frozen bed.

Acknowledgments
National Science Foundation grants supported K.P. (CReSIS; ANT-0424589), I.J. (ARC-1023382), S.B.D., and M.D.B. (ARC-1023364). The Polar Program of the Netherlands Organisation for Scientific Research supported J.T.M.L. and M.R.v.d.B. Velocity data are available through NASA MEaSUREs at http://nsidc.org/data/NSIDC-0478. Radar imagery will be available shortly at the NSIDC. Landsat imagery is publicly available at http://landsat.usgs.gov. All other data in this study are available at the University of Washington Library at http://hdl.handle.net/1773/27336. We thank Nick Selmes and an anonymous reviewer for their constructive comments.

The Editor thanks Nick Selmes and an anonymous reviewer for their assistance in evaluating this paper.

References

Alley, R. B., T. K. Dupont, B. R. Parizek, and S. Anandakrishnan (2005), Access of surface meltwater to beds of sub-freezing glaciers: Preliminary insights, *Ann. Glaciol.*, *40*(1), 8–14, doi:10.3189/172756405781813483.

Bamber, J. L., R. B. Alley, and I. Joughin (2007), Rapid response of modern day ice sheets to external forcing, *Earth Planet. Sci. Lett.*, *257*(1–2), 1–13, doi:10.1016/j.epsl.2007.03.005.

Bartholomew, I. D., P. Nienow, A. Sole, D. Mair, T. Cowton, S. Palmer, and J. Wadham (2011a), Supraglacial forcing of subglacial drainage in the ablation zone of the Greenland ice sheet, *Geophys. Res. Lett.*, *38*, L08502, doi:10.1029/2011GL047063.

Bartholomew, I. D., P. Nienow, A. Sole, D. Mair, T. Cowton, M. A. King, and S. Palmer (2011b), Seasonal variations in Greenland Ice Sheet motion: Inland extent and behavior at higher elevations, *Earth Planet. Sci. Lett.*, *307*(3–4), 271–278, doi:10.1016/j.epsl.2011.04.014.

Catania, G. A., T. A. Neumann, and S. F. Price (2008), Characterizing englacial drainage in the ablation zone of the Greenland ice sheet, *J. Glaciol.*, *54*(187), 567–578, doi:10.3189/002214308786570854.

Darnell, K. N., J. M. Amundson, L. M. Cathles, and D. R. MacAyeal (2013), The morphology of supraglacial lake ogives, *J. Glaciol.*, *59*(215), 533–544, doi:10.3189/2013JoG12J098.

Das, S. B., I. Joughin, M. D. Behn, I. M. Howat, M. A. King, D. Lizarralde, and M. P. Bhatia (2008), Fracture propagation to the base of the Greenland Ice Sheet during supraglacial lake drainage, *Science*, *320*(5877), 778–781, doi:10.1126/science.1153360.

Doyle, S. H., A. Hubbard, A. Fitzpatrick, D. van As, A. B. Mikkelsen, R. Pettersson, and B. Hubbard (2014), Persistent flow acceleration within the interior of the Greenland ice sheet, *Geophys. Res. Lett.*, *41*, 899–905, doi:10.1002/2013GL058933.

Fitzpatrick, A., A. Hubbard, J. E. Box, D. J. Quincey, D. Van As, A. P. B. Mikkelsen, S. H. Doyle, C. F. Dow, B. Hasholt, and G. A. Jones (2014), A decade (2002–2012) of supraglacial lake volume estimates across Russell Glacier, West Greenland, *Cryosphere*, *8*(1), 107–121, doi:10.5194/tc-8-107-2014.

Fox Maule, C., M. E. Purucker, N. Olsen, and K. Mosegaard (2005), Heat flux anomalies in Antarctica revealed by satellite magnetic data, *Science*, *309*(5733), 464–467, doi:10.1126/science.1106888.

Georgiou, S., A. Shepherd, M. McMillan, and P. Nienow (2009), Seasonal evolution of supraglacial lake volume from ASTER imagery, *Ann. Glaciol.*, *50*(52), 95–100, doi:10.3189/172756409789624328.

Gudmundsson, G. H. (2003), Transmission of basal variability to a glacier surface, *J. Geophys. Res.*, *108*(B5), 2253, doi:10.1029/2002JB002107.

Hoffman, M. J., G. A. Catania, T. A. Neumann, L. C. Andrews, and J. A. Rumrill (2011), Links between acceleration, melting, and supraglacial lake drainage of the western Greenland Ice Sheet, *J. Geophys. Res.*, *116*, F04035, doi:10.1029/2010JF001934.

Howat, I. M., S. de la Peña, J. H. van Angelen, J. T. M. Lenaerts, and M. R. van den Broeke (2013), Expansion of meltwater lakes on the Greenland Ice Sheet, *Cryosphere*, *7*(1), 201–204, doi:10.5194/tc-7-201-2013.

Howat, I. M., A. Negrete, and B. E. Smith (2014), The Greenland Ice Mapping Project (GIMP) land classification and surface elevation data sets, *Cryosphere*, *8*(4), 1509–1518, doi:10.5194/tc-8-1509-2014.

Jeffries, M. O., K. Morris, W. F. Weeks, and H. Wakabayashi (1994), Structural and stratigraphie features and ERS 1 synthetic aperture radar backscatter characteristics of ice growing on shallow lakes in NW Alaska, winter 1991–1992, *J. Geophys. Res.*, *99*(C11), 22,459–22,471, doi:10.1029/94JC01479.

Johansson, A. M., P. Jansson, and I. A. Brown (2013), Spatial and temporal variations in lakes on the Greenland Ice Sheet, *J. Hydrol.*, *476*, 314–320, doi:10.1016/j.jhydrol.2012.10.045.

Joughin, I., S. B. Das, M. A. King, B. E. Smith, I. M. Howat, and T. Moon (2008), Seasonal speedup along the western flank of the Greenland Ice Sheet, *Science*, *320*(5877), 781–783, doi:10.1126/science.1153288.

Joughin, I., B. E. Smith, I. M. Howat, T. Scambos, and T. Moon (2010), Greenland flow variability from ice-sheet-wide velocity mapping, *J. Glaciol.*, *56*(197), 415–430.

Joughin, I., S. B. Das, G. E. Flowers, M. D. Behn, R. B. Alley, M. A. King, B. E. Smith, J. Bamber, M. R. van den Broeke, and J. H. van Angelen (2013), Influence of supraglacial lakes and ice-sheet geometry on seasonal ice-flow variability, *Cryosphere*, *7*, 1185–1192, doi:10.5194/tc-7-1185-2013.

Koenig, L. S., et al. (2014), Wintertime storage of water in buried supraglacial lakes across the Greenland Ice Sheet, *Cryosphere Discuss.*, *8*(4), 3999–4031, doi:10.5194/tcd-8-3999-2014.

Krawczynski, M. J., M. D. Behn, S. B. Das, and I. Joughin (2009), Constraints on the lake volume required for hydrofracture through ice sheets, *Geophys. Res. Lett.*, *36*, L10501, doi:10.1029/2008GL036765.

Lampkin, D. J. (2011), Supraglacial lake spatial structure in western Greenland during the 2007 ablation season, *J. Geophys. Res.*, *116*, F04001, doi:10.1029/2010JF001725.

Lampkin, D. J., and J. VanderBerg (2011), A preliminary investigation of the influence of basal and surface topography on supraglacial lake distribution near Jakobshavn Isbrae, western Greenland, *Hydrol. Processes*, *25*(21), 3347–3355, doi:10.1002/hyp.8170.

Leeson, A. A., A. Shepherd, K. Briggs, I. Howat, X. Fettweis, M. Morlighem, and E. Rignot (2014), Supraglacial lakes on the Greenland ice sheet advance inland under warming climate, *Nat. Clim. Change*, *5*(1), 51–55, doi:10.1038/nclimate2463.

Lemke, P., et al. (2007), Observations: Changes in snow, ice and frozen ground, in *Climate Change 2007: The Physical Science Basis, Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by S. Solomon et al., Cambridge Univ. Press, Cambridge, U. K., and New York.

Palmer, S., A. Shepherd, P. Nienow, and I. Joughin (2011), Seasonal speedup of the Greenland Ice Sheet linked to routing of surface water, *Earth Planet. Sci. Lett.*, *302*(3–4), 423–428, doi:10.1016/j.epsl.2010.12.037.

Parizek, B. R., and R. B. Alley (2004), Implications of increased Greenland surface melt under global-warming scenarios: Ice-sheet simulations, *Quat. Sci. Rev.*, *23*, 1013–1027, doi:10.1016/j.quascirev.2003.12.024.

Phillips, T., S. Leyk, H. Rajaram, W. Colgan, W. Abdalati, D. McGrath, and K. Steffen (2011), Modeling moulin distribution on Sermeq Avannarleq glacier using ASTER and WorldView imagery and fuzzy set theory, *Remote Sens. Environ.*, *115*(9), 2292–2301, doi:10.1016/j.rse.2011.04.029.

Schoof, C. (2010), Ice-sheet acceleration driven by melt supply variability, *Nature*, *468*(7325), 803–806, doi:10.1038/nature09618.

Selmes, N., T. Murray, and T. D. James (2011), Fast draining lakes on the Greenland Ice Sheet, *Geophys. Res. Lett.*, *38*, L15501, doi:10.1029/2011GL047872.

Selmes, N., T. Murray, and T. D. James (2013), Characterizing supraglacial lake drainage and freezing on the Greenland Ice Sheet, *Cryosphere Discuss.*, *7*(1), 475–505, doi:10.5194/tcd-7-475-2013.

Seroussi, H., M. Morlighem, E. Rignot, and A. Khazendar (2013), Dependence of century-scale projections of the Greenland ice sheet on its thermal regime, *J. Glaciol.*, *59*(218), doi:10.3189/2013JoG13J054.

Shapiro, N. M., and M. H. Ritzwoller (2004), Inferring surface heat flux distributions guided by a global seismic model: Particular application to Antarctica, *Earth Planet. Sci. Lett.*, *223*, 213–224.

Sundal, A. V., A. Shepherd, P. Nienow, E. Hanna, S. Palmer, and P. Huybrechts (2009), Evolution of supra-glacial lakes across the Greenland Ice Sheet, *Remote Sens. Environ.*, *113*(10), 2164–2171, doi:10.1016/j.rse.2009.05.018.

Sundal, A. V., A. Shepherd, P. Nienow, E. Hanna, S. Palmer, and P. Huybrechts (2011), Melt-induced speed-up of Greenland Ice Sheet offset by efficient subglacial drainage, *Nature*, *469*(7331), 521–524, doi:10.1038/nature09740.

Surdu, C. M., C. R. Duguay, L. C. Brown, and D. Fernández Prieto (2014), Response of ice cover on shallow lakes of the North Slope of Alaska to contemporary climate conditions (1950–2011): Radar remote sensing and numerical modeling data analysis, *Cryosphere*, *8*, 167–180, doi:10.5194/tc-8-167-2014.

Tedesco, M., I. C. Willis, M. J. Hoffman, A. F. Banwell, P. Alexander, and N. S. Arnold (2013), Ice dynamic response to two modes of surface lake drainage on the Greenland ice sheet, *Environ. Res. Lett.*, *8*(3), 034007, doi:10.1088/1748-9326/8/3/034007.

van de Wal, R. S. W., W. Boot, M. R. van den Broeke, C. J. P. P. Smeets, C. H. Reijmer, J. J. A. Donker, and J. Oerlemans (2008), Large and rapid melt-induced velocity changes in the ablation zone of the Greenland Ice Sheet, *Science*, *321*(5885), 111–113, doi:10.1126/science.1158540.

van de Wal, R. S. W., W. Boot, C. J. P. P. Smeets, H. Snellen, M. R. van den Broeke, and J. Oerlemans (2012), Twenty-one years of mass balance observations along the K-transect, West Greenland, *Earth Syst. Sci. Data*, *4*, 31–35, doi:10.5194/essd-4-31-2012.

van der Veen, C. J. (1998), Fracture mechanics approach to penetration of surface crevasses on glaciers, *Cold Reg. Sci. Technol.*, *27*(1), 31–47.

Zwally, H. J., W. Abdalati, T. Herring, K. Larson, J. Saba, and K. Steffen (2002), Surface melt-induced acceleration of Greenland Ice-Sheet flow, *Science*, *297*(5579), 218–222, doi:10.1126/science.1072708.