Ice dynamic response to two modes of surface lake drainage on the Greenland ice sheet

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Received 26 March 2013
Accepted for publication 27 June 2013
Published 16 July 2013
Online at stacks.iop.org/ERL/8/034007

Abstract

Supraglacial lake drainage on the Greenland ice sheet opens surface-to-bed connections, reduces basal friction, and temporarily increases ice flow velocities by up to an order of magnitude. Existing field-based observations of lake drainages and their impact on ice dynamics are limited, and focus on one specific draining mechanism. Here, we report and analyse global positioning system measurements of ice velocity and elevation made at five locations surrounding two lakes that drained by different mechanisms and produced different dynamic responses. For the lake that drained slowly (>24 h) by overtopping its basin, delivering water via a channel to a pre-existing moulin, speedup and uplift were less than half those associated with a lake that drained rapidly (~2 h) through hydrofracturing and the creation of new moulins in the lake bottom. Our results suggest that the mode and associated rate of lake drainage govern the impact on ice dynamics.

Keywords: Greenland, ice dynamics, supraglacial lakes

1. Introduction

The ablation zone of the Greenland ice sheet (GrIS) accelerates each summer due to basal lubrication from surface meltwater that penetrates the ~1-km-thick ice (e.g., Zwally et al 2002, van de Wal et al 2008, Bartholomew et al 2010, Hoffman et al 2011). Basal sliding appears to be controlled by the rate of water delivery to the bed and the capacity of the subglacial drainage system to accommodate it; rapid water delivery overwhelms the hydrologic system, leading to high subglacial water pressures, reduced basal friction and enhanced sliding, whereas slow delivery can be accommodated by gradual enlargement of the system, thereby lowering water pressures, increasing friction and reducing sliding (Bartholomaus et al 2008, Schoof 2010, Pimentel and Flowers 2011).

The cumulation of hundreds of supraglacial lake drainage events on the GrIS each summer (e.g., Selmes et al 2011, Liang et al 2012, Howat et al 2013) affects the seasonal speedup of the ice sheet in two key ways. First, by facilitating hydrofracturing (i.e., the propagation of water-filled cracks to the base of the ice sheet; Weertman 1973, van der Veen 2007, Krawczynski et al 2009), lakes that drain rapidly may temporarily increase surface velocity five- to ten-fold as a direct result of the fracture opening (Doyle et al 2013), and by overwhelming the capacity of the subglacial hydrologic system once the fracture reaches the bed, thereby reducing basal friction (Das et al 2008, Pimentel and Flowers 2011).
Second, by opening connections between the surface and the bed, surface water may continue to be delivered to the base of the ice sheet via moulins (Catania and Neumann 2010) where it may continue to reduce friction and enhance sliding through the remainder of the melt season. In this study, we define rapidly draining lakes as those that drain in the order of a few hours.

Although many lakes drain rapidly by hydrofracturing (e.g., Das et al 2008, Doyle et al 2013), others appear to drain more slowly by feeding supraglacial streams that, in turn, flow into moulins (Catania et al 2008, Hoffman et al 2011, Selmes et al 2013). We define slowly draining lakes as those draining in less than two days but more than a few hours. The speed at which lakes drain, and therefore the rate at which water is delivered to the ice sheet bed, may be important not only for short-term ice dynamics, but also for velocities measured over longer-term (i.e. summer) timescales (Palmer et al 2011). Using radar velocity data at a high spatial resolution Joughin et al (2013), reveal a complex spatio-temporal pattern of 11-day ice velocity, with speedups associated with both fast and slow lake drainage events. However, the relative impact of slowly draining lakes on ice dynamics is yet to be isolated.

Here we report and analyse data collected in the summer of 2011 from five differential global positioning system (GPS) stations situated around two supraglacial lakes in the Paakitsoq region, West Greenland (figure 1), that drained within two days of one another through different mechanisms. The smaller of the two lakes (Lake Half Moon, ‘LHM’, 69.573N, −49.805E, maximum recorded depth, surface area and volume of ∼1.6 m, 60 000 m² and 200 000 m³ respectively) drained slowly (>24 h) via an overspill channel to an existing moulin when the water level rose high enough to breach the lowest point of the lake basin (‘overspill’ drainage). Drainage of the larger, deeper lake (Lake Ponting, ‘LP’, 69.589N, −49.783E, maximum recorded depth, surface area and volume of ∼5.2 m, 480 000 m² and 1 500 000 m³ respectively) was fast (∼2 h) and occurred through its bottom, following hydrofracturing (‘bottom’ drainage). We analyse and compare the impacts of the two different lake drainage modes on the dynamics of a ∼16 km² area of the ice sheet surrounding the two lakes (figure 1).

2. Data and methods

Water levels in the two lakes were measured every five minutes by pressure transducers (HOBO®), after correcting for elevation and for barometric pressure fluctuations, measured by a third pressure transducer located less than 1 km from the lakes. Further details of the approach are given by Tedesco et al (2012). Water levels were converted to volumes using empirically derived depth–volume curves from surface topography data. For this, we used the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) Global Digital Elevation Model (GDEM), which has a nominal grid size of 30 m (http://asterweb.jpl.nasa.gov/gdem.asp). This was smoothed with a 6 × 6 cell median filter to remove small-scale noise then re-sampled to a 100 m resolution using bilinear interpolation (Banwell et al 2012). Knowing the topography of each lake also allows us to use the water level data to derive surface areas. Cumulative water volume curves were differenced to calculate net water discharge flowing to or from each lake at a 5 min temporal resolution. Uncertainty in calculated lake volumes due to error within the GDEM data was assessed by applying 1000 sets of Gaussian noise with a standard deviation of 13.8 m (MacFerrin 2011) to the raw GDEM data, then smoothing and interpolating the resultant DEM as above. The root mean square errors of sequential volume estimates in the cumulative volume curves were used to derive discharge errors.

The five GPS receivers were installed within 2 km of the two lakes, approximately along flowlines through LHM. Proceeding upglacier, receivers named R1 and R2 were located downstream of LHM and R3–R5 were positioned upglacier of the lake (figure 1). Antennas were mounted on aluminum poles drilled into the ice with a portable steam drill to a depth of 6 m. GPS data were logged every 15 s, and positions were determined by carrier-phase differential processing using TRACK software (Chen 1998) with the base station KAGA (25 km to the south) as a reference and final International GNSS Service satellite orbits. Each 15 s GPS time series was re-sampled to a 6 min interval, and a 1-h moving average was then applied to reduce the sidereal noise, following the methods of Hoffman et al (2011). The short filter width of 1 h used here provides a pre-drainage velocity standard deviation ranging between 25 and 35 m yr⁻¹ (depending on the rover). Though noisier, a short filter allows for higher precision of the timing and magnitude of the high signal drainage events. The position data were used to
generate velocity time series averaged over 1 h time windows and posted at 6 min intervals. The recorded peak speed during the drainage of the two lakes are at least four times greater than the pre-drainage speed standard deviation in the case of both drainage mechanisms.

#### 3. Results

##### 3.1. Overspill drainage of Lake Half Moon

Water pressure records indicate that LHM filled in less than three days, reaching a maximum depth of 1.6 m at our sensor at 22:10 (UTC) on 16 June 2011 (figure 2(a)) and a maximum estimated volume of $\sim 200,000$ m$^3$ (figure 2(b)). Thereafter, the lake level began to decline slowly and at a declining rate, with an average net discharge of $1.4 \pm 5.2$ m$^3$ s$^{-1}$ and a peak net discharge of $2.9 \pm 6.5$ m$^3$ s$^{-1}$. The water pressure record ends 22 h and 15 min later at 20:25 on 17 June 2011 when the lake level dropped below our sensor. From daily visual inspection of the lake, we know the lake level continued to drop very slowly until at least 26 June 2011 when we left the site. Water draining from LHM flowed in an existing channel incised into the ice surface to a moulin located $\sim 700$ m down glacier (figure 1). Initially, drainage was relatively rapid due to the observed removal of the previous winter’s snow from the channel via a series of slush flows. Drainage then declined at an exponential rate as the hydraulic head between the lake and the moulin dropped.

In association with the slow drainage of LHM, the ice velocity at our GPS stations increased from baseline values of $\sim 90–100$ m yr$^{-1}$ to a maximum of $\sim 420$ m yr$^{-1}$ (figure 3(a)), with flow trajectories remaining largely unaltered (figure 4) and velocities increasing in an east–west (downglacier) rather than a north–south (transverse) direction (figures 5(a) and (b)). The onset of the speedup began with the upglacier station R5 and proceeded downglacier. All stations started to accelerate before the lake level began to fall, with the acceleration of R5 beginning $\sim 3$ h before the onset of the lake level drop. Water was, however, observed to be entering the moulin at
this time. The amplitude of the acceleration was highest for station R5 (a four-to-five-fold increase on the pre-drainage velocity), and decreased down glacier. After the level of LHM stabilized on 17 June and before drainage of LP began late on 19 June, velocities were \(~160–170~\text{m yr}^{-1}\), over 50% above pre-drainage velocities (figures 3(a), 5(a) and (b)). A gradual and uniform increase in elevation (\(\sim 0.1\text{ m over 45 h}\)) was recorded at all stations, beginning with initial speedup and continuing after completion of the drainage (figure 4(a)). We named this draining mechanism ‘overspill draining mechanism’ and a sketch is reported in figure 6.

3.2. Bottom drainage of Lake Ponting

Lake Ponting reached its maximum depth of 5.2 m at our sensor in six and half days at 14:35 on 19 June 2011 (figure 2(a)) and a maximum estimated volume of \(\sim 1500000\text{ m}^3\) (more than six times greater than the volume of LHM, figure 2(b)). In contrast to the relatively slow drainage of LHM, LP subsequently drained completely in 2 h and 10 min (figure 2), with an average net discharge of \(166 \pm 31\text{ m}^3\text{s}^{-1}\) and a peak net discharge of \(586 \pm 19\text{ m}^3\text{s}^{-1}\). Frames from a time-lapse sequence of the drainage are shown in figures 7(a)–(f). The rapid increase in LP’s depth/volume seen at \(\sim 12:00\) on 18 June 2011 was due to the overflow of an upstream lake into LP (figure 7(g)) and therefore the overall enlargement of LP’s catchment (Banwell et al 2012).

Reconnaissance of the LP basin shortly after drainage revealed a recently formed northwest–southeast trending fracture (\(~600\text{ m long},\) ranging from a few centimetres to several metres wide, figure 8(a)) that ran along the former lake bed, centred on what would have been its deepest part. Numerous ice blocks, several metres in size, lay close to the fracture (figure 8(b)) and six moulins were found along it, ranging in size from a few metres to \(~10\text{ m in diameter}\) (figure 8(c)). The blocks had been observed floating on the lake during the initial phase of the drainage, but before we had set up our time lapse camera (figure 7). Overall, the evidence suggests that LP drained through its bottom by hydrofracture, that the ice blocks were plucked from the fracture, floated temporarily due to buoyancy and then became grounded nearby as the lake level dropped, and that the moulins were produced as water flow concentrated in places along the fracture during and immediately after the drainage event. We named this draining mechanism ‘bottom draining mechanism’ (figure 6).

Compared to the relatively slow drainage of LHM, the faster drainage of LP had a larger, more immediate impact on ice velocities and elevation (figure 3(b)) and displayed a more spatially variable ice dynamic response across our GPS station network (figures 4, 5(c) and (d)). In contrast to the velocity response following LHM drainage, downstream stations R1 and R2 responded first to LP drainage, and reached greater peak velocities of \(1500–1600\text{ m yr}^{-1}\) (\(~\text{a ten-fold increase on pre-LP-drainage velocities and \(~\text{a fifteen-fold increase on pre-LHM-drainage velocities, figures 3(b), and 5(d)\. The initial acceleration of station R1 followed the onset of LP drainage by 10 min, and was followed by acceleration of R2 \(~20\text{ min later. The remaining stations accelerated concurrently \(~30\text{ min after R2 accelerated, reaching peak velocities of 270 to 370 m yr}^{-1}\). All stations reached peak velocities around 17:00 on 19 June 2011, fifteen minutes after LP finished draining. Unlike observations during the drainage of LHM, all stations temporarily changed flow direction to varying degrees towards the south during the drainage of LP (figure 4) and velocities showed a strong north–south (transverse) as well as an east–west (downglacier) component (figures 5(c), and (d)). Rapid increases in elevation were recorded at all stations, peaking between 20:00 on 19 June and 01:00 on 20 June 2011 (figure 3(b)). Stations R1 and R2 reached maximum uplift of \(~0.20\text{ m, R3 and R5 reached maximum uplift of }\sim 0.10\text{ m, and peak uplift at R4 }\sim 0.05\text{ m. Post-drainage velocities were on the order of }\sim 300–350\text{ m yr}^{-1}\), nearly twice as high as pre-drainage velocities, which were already nearly twice as high as baseline velocities prior to LHM drainage.

4. Discussion

Through the opening of local surface-to-bed connections, the drainage of both LHM and LP caused increases in the local ice velocity, at least temporarily, from the relatively slow pre-drainage velocities. The drainage events also induced changes in vertical motion from downward movement associated with bed-parallel flow to upward movement. The evidence suggests, therefore, that in each case the surface water drained to and immediately exceeded the capacity of the subglacial drainage system (e.g. Hoffman et al 2011). We interpret the evidence in terms of hydraulic jacking within basal cavities and increases in basal sliding. Fractures from elastic plates loaded from below typically produce radial fractures on the surface (the higher the stress, the greater the number of radial fractures) (Beltaos 2002, his figure 1). We suggest that water-filled cavities forming beneath the ice sheet uplifted the ice and that the main northwest–southeast fracture which we observed at LP formed along the long-axis of this temporary water body. That velocities remained higher after each drainage event than before, suggests that the hydrologic system remained water filled and operated at higher pressure...
Figure 5. North–south ((a), (c)) and east–west ((b), (d)) velocities (m a$^{-1}$) estimated from GPS measurements collected during the drainage of Lake Half Moon ((a), (b)) and Lake Ponting ((c), (d)). Note that scales on the y-axis are different for the four panels.

Figure 6. Schematic representation of the two types of lake drainage, ‘overspill drainage’ and ‘bottom drainage’ and their different dynamic responses.

than before drainage, likely in response to the continued flow of water into the respective moulins. Peak velocities during both events occurred during the maximum uplift rate, shortly after lake drainage began, rather than when the capacity of the subglacial drainage system had reached a maximum, showing that changes in water storage are more important in driving velocity increases than the magnitudes of inputs (Iken 1981, Bartholomaeus et al 2008, Schoof 2010, Bartholomew et al 2012), similar to observations and modelling of valley glaciers during ‘spring events’ (Iken et al 1983, Mair et al 2003), and the drainage of other lakes on the GrIS (Das et al 2008, Hoffman et al 2011, Pimentel and Flowers 2011, Doyle et al 2013).

The rapid drainage of LP generated a greater, but more spatially variable ice dynamic response across our GPS network than the slower drainage of LHM. During LP drainage, the speedup and uplift at R1 and R2 were about twice as large as those associated with LHM drainage, although the response at the other three stations was more
muted and of similar magnitude to that which occurred during the drainage of LHM (figures 3 and 5). Even though the drainage of LHM occurred first, and therefore the water likely impinged on a lower capacity subglacial hydrologic system than existed when LP drained, the rapid drainage of LP still produced a larger ice dynamic impact.

Another difference between the two drainage events concerns the spatial pattern of displacement at each GPS receiver. The slow LHM drainage caused minor (<20°) adjustments in flow directions that were maintained subsequently through the rest of the GPS record (figures 4, 5(a) and (b)). We interpret this in terms of lasting changes to the spatial distribution of basal friction caused initially by the first arrival of surface water to the bed, but maintained thereafter by continued input of water via the moulin. In contrast, the initial displacements of all receivers during the fast LP drainage show large deviations from their pre-event trajectories to the south, away from LP, with R1 and R2 moving in a transverse direction by ∼0.1 m over the course of a few hours (figures 4, 5(c) and (d)). A substantial fraction of this motion is subsequently recovered with northward motion. This is consistent with observations of Doyle et al. (2013) who interpreted this type of motion in terms of fracture opening and closing during and immediately after rapid lake drainage. Thus, a substantial component of the initial increased ice velocities associated with the rapid drainage of LP is temporary, lateral displacement and not associated with increased longitudinal downglacier displacement. Thereafter, displacement is predominantly downglacier along trajectories that are slightly modified compared with those that existed previously (figure 4). As with the LHM drainage, we interpret this in terms of enduring changes to the spatial distribution of basal friction caused by the water from the lake drainage and continued input via the moulins.

Although the dynamic response to the slow LHM drainage was relatively uniform across the GPS network, there was some anomalous behaviour. The receivers began to accelerate prior to the initial drop in LHM water level. However, observations in the field showed that water was already entering the moulin prior to LHM attaining its peak water level at 22:10 (UTC) on 16 June 2011. We suggest that the receivers responded to the arrival of surface water at the bed at around 19:00 but that lake levels continued to rise up to 22:10 as water inputs to the lake exceeded the capacity of the snow-filled surface channel between the lake and the moulin to discharge the water.

That the acceleration began with the station furthest upglacier (R5) and proceeded downglacier past LHM is also puzzling. We considered the possibility that the receivers responded to a lake drainage event higher up on the ice sheet prior to the acceleration at R5, but an analysis of daily MODIS imagery shows no evidence for this. We suggest that despite its greater distance from the LHM moulin, R5 may have accelerated prior to R4 and R3, with R2 and R1
studies have identified that rapid drainage is a small fraction of all drainages. In a study of supraglacial lakes detectable in MODIS imagery across the entire GrIS over five years Selmes et al (2011), found that only 13% of drainages occurred over less than two days, with the southwest and northeast regions of GrIS having higher rates of fast lake drainage than the rest of the ice sheet. Slow drainage, by contrast, accounted for 34% of lake drainage events over the same period (Selmes et al 2013). Similarly Liang et al (2012), found that less than 20% of lakes drain faster than 0.5 km² d⁻¹ in a MODIS-based study in western Greenland. In an analysis of a network of nine GPS stations in western Greenland Hoffman et al (2011), were able to identify speedups associated with only 17% of lake drainages identified in the region from satellite imagery. Thus, the hydrofracture-induced rapid drainage mechanism appears to be relatively rare Selmes et al 2011, Liang et al 2012 consistent with the observation that lake drainage associated speedups account for <5% of all summer ice motion (Hoffman et al 2011).

Slow draining lakes are unable to provide a rapid delivery of water to the ice sheet bed. Because the capacity of the subglacial hydrologic system adapts to steady inputs, slow drainage events theoretically will induce less sliding than pulsed inputs (Bartholomaus et al 2008, Schoof 2010, Bartholomew et al 2012), highlighting the importance of distinguishing between these two modes of surface lake drainage. Some lakes have different drainage speeds, and therefore presumably different modes, from year-to-year, and predicting which mode of drainage a particular lake will be subject to appears difficult. However, deeper lakes and those whose basins intersect extensional stress regimes in the ice would theoretically be more likely to experience hydrofracture and bottom drainage (e.g., Krawczynski et al 2009).

5. Conclusions

Our measurements indicate that the impact of surface lake drainage on GrIS dynamics depends on the drainage mechanism (figure 6): either (i) relatively slow (<24 h) overspill drainage via an existing channel to an existing moulin; or (ii) relatively rapid (~2 h) bottom drainage via hydrofracture and the creation of new moulins. The overspill draining lake resulted in less speedup (four-to-five-fold versus fifteen-fold compared to pre-drainage speeds) and uplift (0.1 m versus 0.2 m) than the bottom draining lake, despite the fact that the overspill drainage occurred first, and therefore likely impinged on a lower capacity subglacial hydrologic system than existed after it drained. Due to the muted dynamic response associated with the overspill drainage mechanism, we caution against extrapolation of the dynamic response of draining by the previously studied bottom draining mechanism (e.g., Das et al 2008, Doyle et al 2013) to all lake drainages on the GrIS. Our observations also indicate spatially variable ice dynamic responses, presumably due to differences in flow coupling in both the longitudinal and lateral directions. Care should be taken in interpreting point measurements of ice velocity associated with lake drainage, for example, by GPS receivers.
As evidence suggests that fast drainage events account for a relatively small percentage of total lake drainage events on the GrIS (e.g., Selmes et al. 2011, Liang et al. 2012), we suggest that the potential contribution of the slower drainage mechanism to seasonally averaged ice velocities may actually be higher than that associated with the rapid drainage mechanism. However, as both drainage mechanisms act to open up surface-to-bed connections (i.e. moulins), which can remain open for the rest of the melt season, both drainage mechanisms have the potential to enable diurnally varying meltwater inflow to access the bed for the remainder of the melt season (Banwell et al. 2013). It is this variability of water delivery to the bed, rather than the absolute magnitude, which is thought to have more of a significant influence on sliding speeds; thus the greater the number of open moulins the higher the potential for diurnal variations in velocities and overall summer enhancement of flow (Schoof 2010, Bartholomew et al. 2012).

Acknowledgments

This study was supported by the National Science Foundation (NSF-ARC 0909388), the NASA Cryosphere Program, the Natural Environment Research Council (Grant LCG/133, CASE Studentship with GEUS), the Earth System Modeling program of the Office of Biological and Environmental Research within the US Department of Energy’s Office of Science, St Catharine’s College (Cambridge), the Scandinavian Studies Fund and the B B Roberts Fund. GPS systems were provided by UNAVCO. Finally we thank Douglas MacAyeal for valuable discussions.

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