Surface melt-driven seasonal behaviour (englacial and subglacial) from a soft-bedded temperate glacier recorded by in situ wireless probes

Jane K. Hart,†* Kirk Martinez,2 Philip J. Basford,2 Alexander I. Clayton,1† Graeme M. Bragg,2 and David S. Young2

1 Geography and Environment, University of Southampton, Southampton, UK
2 Electronics and Computer Science, University of Southampton, Southampton, UK

Received 21 April 2017; Revised 7 March 2019; Accepted 11 March 2019

*Correspondence to: Jane K. Hart, Geography and Environment, University of Southampton, Southampton, SO17 1BJ, UK. E-mail: jhart@soton.ac.uk
†Current address: Winchester College, Winchester, UK

This is an open access article under the terms of the Creative Commons Attribution License, which permits use, distribution and reproduction in any medium, provided the original work is properly cited.

ABSTRACT: We investigate the spatial and temporal englacial and subglacial processes associated with a temperate glacier resting on a deformable bed using the unique Glacsweb wireless in situ probes (embedded in the ice and the till) combined with other techniques [including ground penetrating radar (GPR) and borehole analysis]. During the melt season (spring, summer and autumn), high surface melt leads to high water pressures in the englacial and subglacial environment. Winter is characterized by no surface melting on most days ('base') apart from a series of positive degree days. Once winter begins, a diurnal water pressure cycle is established in the ice and at the ice/sediment interface, with direct meltwater inputs from the positive degree days and a secondary slower englacial pathway with a five day lag. This direct surface melt also drives water pressure changes in the till. Till deformation occurred throughout the year, with the winter rate approximately 60% that of the melt season. We were able to show the bed comprised patches of till with different strengths, and were able to estimate their size, relative percentage and temporal stability. We show that the melt season is characterized by a high pressure distributed system, and winter by a low pressure channelized system. We contrast this with studies from Greenland (overlying rigid bedrock), where the opposite was found. We argue our results are typical of soft bedded glaciers with low englacial water content, and suggest this type of glacier can rapidly respond to surface-driven melt. Based on theoretical and field results we suggest that the subglacial hydrology comprises a melt season distributed system dominated by wide anastomosing broad flat channels and thin water sheets, which may become more channelized in winter, and more responsive to changes in meltwater inputs. © 2019 The Authors. Earth Surface Processes and Landforms published by John Wiley & Sons Ltd.

KEYWORDS: subglacial processes; in situ experiments; till; GPR; subglacial hydrology

Introduction

Glacier response to climate change is controlled by the two interactive processes of subglacial sediment deformation and hydrology (Boulton and Jones, 1979; Fountain and Walder, 1998). Surface melting generates melt water, which may travel through the glacier to its base, where it can regulate glacier velocity via sliding and/or till deformation (Iken et al., 1993; Willis, 1995; Hubbard and Nienow, 1997; Zwally et al., 2002; Anderson et al., 2004; Schoof, 2010; Sugiyama et al., 2011; Bartholomew et al., 2012). Although there have been an increasing number of studies of these processes due to a series of new technologies, including in situ glacier monitoring (Fischer and Clarke, 2001; Murray and Porter, 2001; Hart et al., 2009), geophysics (Smith, 2006; King et al., 2009; Matsuoka et al., 2010), borehole videos (Fountain et al., 2005; Hubbard et al., 2008) and differential global positioning system (dGPS) (Wiens et al., 2008; Waechter et al., 2015), instrumented data are still rare because of the logistical difficulties in accessing the glacial environment.

Most models of the subglacial environment have assumed a rigid bedrock (Hubbard et al., 1995; Bartholomew et al., 2010; Andrews et al., 2014). However, many of the fast flowing ice streams of Antarctica are resting on unconsolidated sediments (soft bedded) which is important in controlling its stability (Schroeder et al., 2013; Joughin et al., 2014; DeConto and Pollard, 2016), as well as large parts of the Quaternary Ice sheets of Europe and North America (Boulton and Hindmarsh, 1987; Hicock and Dreimanis, 1992; Hart et al., 2011).

When a glacier rests on unconsolidated material, subglacial deformation is very common, with 20–85% of glacier motion occurring within the subglacial sediment layer (till) rather than the ice (Boulton et al., 2001), and deformation in this layer can be modelled as a shear zone (Hart and Boulton, 1991).
Pore-water pressure is an important factor in determining subglacial processes and the resultant sedimentation (Boulton and Jones, 1979; Alley et al., 1986; Clarke, 1987; Brown et al., 1987; Hart and Boulton, 1991; Hicock, 1992; Hicock and Dreimanis, 1992; Iverson, 2010).

The glacier bed will comprise a mosaic of different strengths, comprising weak ‘slippery’ spots and stronger ‘sticky’ spots (Alley, 1993; Piotrowski et al., 2004; Stokes et al., 2007, 2009; Trommelen et al., 2014) and these subglacial processes can change both temporally and spatially (Smith and Murray, 2009; Hart et al., 2011).

Unconsolidated beds will also have a different type of subglacial hydrology. Instead of a channelized system (Röthlisberger, 1972; Nienow et al., 1998), the subglacial water will flow through a combination of a braided/swampy/canal system alongside a thin macroporous layer (Weertman, 1964; Hooke et al., 1990; Kamb, 1991; Hock and Hooke, 1993; Walder and Fowler, 1994; Creyts and Schoof, 2009; Schroeder et al., 2013).

It has been argued that enhanced temperature rise in the Arctic (‘Arctic Amplification’ – Serreze and Francis, 2006; Solomon et al., 2007; Cohen et al., 2014) will lead to increased surface melting, resulting in increased basal sliding, which brings more ice into lower altitudes, which accelerates further melting (Zwally et al., 2002; Joughin et al., 2008; Shepherd et al., 2009). However, there is an alternative model that suggests that increases in melt do not necessarily lead to faster glacier velocities (Sundal et al., 2011; Sole et al., 2013; Tedstone et al., 2015). These researchers argue that in early summer, meltwater causes increased storage and higher water pressures. Once channelization occurs, this increases drainage capacity and decreases water pressure, and so the subglacial hydrological system can accommodate increased melt.

There have been fewer studies of the effect on increased surface melting on the behaviour of glaciers with unconsolidated beds. In this study we present an instrumented data set from an Icelandic temperate soft bedded glacier, to investigate water pathways through the ice and the till, to understand a multi-year seasonal response to surface melt and to determine to what degree do water pressure variations moderate ice flow.

Field Site and Previous Studies

The study was undertaken at Skálfellsjökull, Iceland (Figure 1a), an outlet glacier of the Vatnajökull icecap which rests on Upper Tertiary grey basalts. The area of the glacier is approximately 100 km² with a length of 25 km (Sigurðsson, 1998). The study site was located on the glacier at an elevation of 792 m above sea level (a.s.l.), where the ice was flat and...
crevasse free. The subglacial meltwater in this area emerges 3 km away at the southern part of the glacier (known as the Sulfartungajökull tongue, Figure 1b).

The site was examined for five years 2008–2012 and previous studies (Hart et al., 2015, 2018) have shown the following:

a. A fine-grained till (mean grain size 53 μm) covered the majority of the glacier bed.
b. There was evidence of subglacial deformation in the foreland, comprising flutes and push moraines.
c. The data from boreholes (26–86 m in depth) were similar each year and showed that the majority (15 out of 19) remained filled with water during and after drilling. Video evidence showed only one borehole with englacial drainage, two with subglacial cavities, and one with fast flowing water (subglacial channel).
d. Repeat ground-penetrating radar (GPR) surveys (combined with measured glacier depth and video recordings) showed that the glacier had a low water content (0.54% ± 0.26), and that the majority of the glacier had a mean radar velocity of 0.177 ± 0.005 m ns⁻¹, with a thin 1 m debris-rich basal ice layer with a radar velocity of 0.158 ± 0.003 m ns⁻¹.
e. The GPR results from 2008 and 2011 were used to reconstruct the nature of the bed from basal reflection strength.

It was concluded that the glacier has little englacial storage, the ice was impermeable and drainage pathways were concentrated in fractures and moulins. GPR results combined with modelling showed that the subglacial hydrological systems comprised a series of subglacial braided water bodies flowing parallel with the glacier margin.

Methodology

The Glacsweb system

A key component of the data collection is the Glacsweb environmental sensor network (Hart and Martinez, 2006) which comprises sensor nodes (probes and geophones), base stations and a sensor network server in the UK, linked together by radio (Figure 2). Sensors within a probe (0.16 m long, axial ratio 2.9:1), measured water pressure, force, resistance, tilt and temperature within the ice or till. These data were recorded every 15 minutes in 2012 and every hour 2008–2010, and transmitted (via 151 MHz radio) to a base station located on the glacier surface. Probe data, along with dGPS and meteorological data, were sent once a day to a mains powered computer (2.5 km away), where it was forwarded to a web server (via a GPRS link) in the UK (Table I). Details of the 2008–2010 system are discussed in Martinez et al. (2009) and the 2012 system in Martinez et al. (2012). This system can only transmit through 70 m of ice, however dGPS measurements of annual surface velocity (discussed in more detail later) show a positive relationship between velocity and depth ($r^2 = 0.98$), indicating that the ice at the study site is representative of glacier motion as whole (Figure 3).

Probes

These were deployed in the summers of 2008 and 2012, in a series of boreholes (57–69 m in depth), which were drilled with a Kärcher HDS1000DE jet wash system (Figure 1c). The depths of the boreholes were measured to calculate glacier thickness. The glacier and till were examined using a custom made digital infrared light-emitting diode (LED) illuminated colour video camera, via the borehole.

In order to insert probes into the till, the boreholes were drilled to the base of the glacier and the presence of till checked with the video camera. If till was present it was hydraulically excavated (Blake et al., 1992) by maintaining the

![Figure 2. Glacsweb sensor network system. [Colour figure can be viewed at wileyonlinelibrary.com]](image)

| Sensor Type          | Technical Specifications | Range          |
|----------------------|--------------------------|----------------|
| Temperature sensor   | DS1631                   | 0.0625°C to -10°C to +85°C |
| Pressure transducer  | 24PCFG66G                | 1.12 kPa to 0–1724 kPa |
| Strain gauges        | Strain gauge             | 1.25 N to 0–213 kN |
| Conductivity         | Resistance bridge        | 0.005 μS to 0–100 μS |
| Tilt sensors 1 and 2| LIS1L02DQ                | ± 2 g           |

Table I. Sensors in the Glacsweb probes

© 2019 The Authors. Earth Surface Processes and Landforms published by John Wiley & Sons Ltd. Earth Surf. Process. Landforms, Vol. 44, 1769–1782 (2019)
jet at the bottom of the borehole for an extended period of time. The probes were then lowered into this space, enabling the till to subsequently close in around them. The measured depth of the probes within the till is not known, but is approximated at 0.1–0.2 m beneath the glacier base, estimated from video footage of the till excavation prior to deployment.

Table 1 shows the details about the probe sensors. Water pressure was measured in metres water equivalent (m WE) and calculated as a percentage of glacier thickness. The glacier thickness was determined from the GPR and GPS results. The water depths were calibrated against the measured water depths in the borehole immediately after probe deployment. The conductivity of the material outside the probe was measured across two bolts mounted in the case. This data was used to determine the nature of the material surrounding the probe. The sensors were calibrated in the laboratory. High values indicate wet sediments, and low values indicate dry or frozen material. Probe temperature values were also calibrated in the laboratory using a commercial temperature reference instrument.

Strain gauges measure relative compression and extension of the probe case in two perpendicular planes (case stress) expressed as the force applied to the probes per unit area (in kilopascals). The applied force was calibrated using an Instron 5560 tension/compression machine attached to a nitrogen cooled chamber, where the average chamber temperature was 1.3°C. The chamber was pre-cooled with liquid nitrogen then left to settle around 0°C to be representative of expected basal temperatures.

The Glacsweb probes measure tilt with two dual axis 180° microelectromechanical system (MEMS) accelerometers. Values of 0° x-tilt and y-tilt represent the probe standing vertically, these were calibrated in the laboratory, and dip was calculated by trigonometry. The probe is normally almost vertical during deployment and inclines towards the horizontal as the glacier moves over the till. We define this decrease in dip as synthetic, and movement in the opposite direction (increases in dip) as antithetic.

Ground penetrating radar (GPR)

Common offset (CO) surveys were carried out over the survey grids in 2008, 2011 and 2012 (Figure 1c). The data was used to calculate radar velocity and water content (reported earlier), and the nature of the bed from the strength of the basal reflection (reported in 2008 and 2011). Here we calculate the bed reflection strength from the 2012 data and compare it with the 2008 and 2011 results to show how basal reflection strength varies in space and time. The surveys used a Sensors and Software Pulse Esko 100 with a 1000 V transmitter system and 50 MHz antennas. A 2 m antenna spacing and 0.5 m sampling interval were used for the CO survey. A sledge was constructed to hold the antennae at the correct distance apart and allow the system to move along the grid transects more readily.

A series of standard processing steps were applied to the CO survey data using the software package ReflexW (Moorman and Michel, 2000; Murray et al., 2000). Low frequency noise was eliminated (de-wow filter) and a SEC (spreading and exponential compensation) gain was applied to compensate for signal loss with depth. Next a diffraction stack migration and topographic correction were applied. The ice-bed interface was identified manually in radar echograms from a clear strong reflector. It was then possible to calculate the radar-wave velocity in the whole ice column (v) by comparing the return time of the glacier bed in radar echograms (t) with measured borehole glacier depths (h) according to the equation:

\[ v = \frac{2h}{t} \] (1)

Subsequently, the GPR data were analysed in order to calculate basal reflection to reconstruct the nature of the bed (Gades et al., 2000; Winebrenner et al., 2003; MacGregor et al., 2007; Matsuoka et al., 2010; Jacobel et al., 2009, 2010). In summary, the power of electromagnetic energy returned from the subglacial interface is determined by three factors: the dielectric properties of the reflector (the basal reflectivity) (R), losses due to geometric spreading (which are related to glacier depth), and losses due to dielectric attenuation within the ice (L). We calculate a mean attenuation value (see Hart et al., 2015, for details), and apply this as the best-fit line to the data, as returned power verses depth. The distance from the fitted line is a measure of relative basal reflectivity for each point (Jacobel et al., 2009, 2010). These values were plotted as a histogram and three peaks are identified and these were used to quantify bed strength (see figure 6 in Hart et al., 2015).

Weather data

These were obtained from a meteorological station (Davis Vantage Pro) (temperature, rainfall, wind speed and humidity) sited on the reference station and, during periods of mechanical failure, from a transfer function applied to data from the neighbouring Icelandic meteorological station at Höfn. Daily melt was calculated by the positive degree day algorithm (Braithwaite, 1995; Hock, 2003) using degree day factors for Sáttujökull, Iceland (Johannesson et al., 1995), 5.6 mm d⁻¹°C⁻¹ for snow and 7.7 mm d⁻¹°C⁻¹ for ice. Albedo was calculated from the MODIS data, using the threshold between ice and snow to be 0.45, on a 30 m × 30 m grid ASTER digital elevation model (DEM).

Differential global positioning system (dGPS)

A kinematic dGPS was used to map the glacier margin, boreholes, base station and radar grids. In order to measure surface velocities, a Topcon dGPS was used (2008–2012), with four additional dual frequency Leica dGPS stations installed on the glacier (GPS 1–4) in the study area, with a local base station on the moraine (2012–2013) (Figure 1c). These measured at 15 second sampling rate continuously during the melt season and two hours a day during the winter. The GPS data was processed using TRACK (v. 1.24), the kinematic software package developed by Massachusetts Institute of Technology (MIT) (http://www.unavco.org/, http://geoweb.mit.edu/~lah/track_example/).

Data and Results

Weather results

We define the seasons based on the melt rate/air temperature. The melt season comprises spring, summer and autumn (approximate day of year (DOY) 121–289) when there is constant melt on the glacier surface. During winter, there is no surface melt on the majority of days (negative degree days) apart from a series of warmer days when temperatures rise above zero (positive degree days).
Probe data results

The overall results from the probe data collection are shown in Table II and the data from probes with a water pressure record (2008/2010 probes) are shown in Figure 4 which we will discuss in detail. In 2008, six probes were deployed, and four continued to send data after one day. Of these, one probe (Probe 26) was in a borehole that was not drilled to the glacier bed and so was known to be in the ice. The other three were deployed into the till as discussed earlier (Probe 21, Probe 24 and Probe 25). The 2008 probes sent back data for two years with 141,380 readings in total. All the sensors worked apart from the tilt sensors. However in 2009, the software was de-bugged, which enabled the tilt to be read once a day at mid-day.

The probes were designed so that if the data were not immediately accessed then they were stored for later retrieval. There were some problems with communications between the probes and the base station which unfortunately led to the probes filling their programmable memory (EPROM), resulting in some data gaps (Figure 4).

In 2012, four probes were deployed which all functioned. Two of the probes (Probe 31 and Probe 32) were deployed in boreholes that reached the bed (till probes), and Probe 33 was deployed in a borehole that did not reach the bed (ice probe). Probe 34 was installed on a wire as a ‘relay’ probe in the ice. Although this probe could not move freely, it could record other properties of the ice. The 2012 probes lasted for 15 weeks. There were some sensor failures (particularly the water pressure and conductivity), however 296,235 data points were received from the probes.

After deployment, the probes become embedded into the till as the glacier moves over the initial location of the probes and the boreholes slowly close. For Probe 25, at the beginning of the record (DOY 223–307), water pressure falls were accompanied by a rise in probe temperature and fall in conductivity. This may reflect short periods of connection with the subglacial hydrological system, leading to water pressure decline combined with warmer water entering the till (from its surface source) whilst the borehole was still open and connected to the base. The other probes were incorporated much faster (approximately 15 days). Further discussions of the probe data will not include the days prior to incorporation.

Velocity data

Detailed velocity data was collected 2012–2013 and is shown plotted against surface melt and the continuous tilt data (Figure 5). During the summer, there is no statistical correlation between air temperature and velocity, although the extremes in temperature correlate with the extremes in velocity. However, during the winter there is a different pattern. There is a ‘base’ velocity during the negative degree days, and velocity peaks during the positive degree days (speed-up events). Although we only have detailed velocity data for 2012 to 2013 we can assume that these speed-up events also occurred in previous years during the positive degree days.

GPR results

The GPR results can be used to calculate the strength of the glacier bed reflection. The results from the 2008 and 2011 surveys were reported in Hart et al. (2015), and the same techniques are used to analyse the 2012 data. Using the method of Jacobel et al. (2009, 2010) the reflection values were divided into three classes: low reflection values ($R$) less than $-3$ dB (indicative of
We then calculate the percentage of each of these categories along the survey grids (Table III). Areas with low reflectivity/stiffer till or bedrock covered 19%; areas with high reflectivity/water bodies covered 4%; with the remainder (77%) representing the deforming bed. These percentages were consistent for different years, and where repeat surveys were made over the same area, the bed types remained the same.

We are also able to calculate the size of the different patches (also shown in Table III). Over the three surveys, the mean length of the deforming bed areas was 7.9 m, the length of the low reflectivity areas was 2.2 m, and the length of the high reflectivity areas was 1.9 m. These values were very similar each year and demonstrate the scale of the patchwork elements.

**Discussion**

We now discuss the evidence for spatial and temporal changes in the englacial and subglacial environment, and compare these results with reported *in situ* water pressure data from Greenland.

**The englacial environment**

The data from Probe 26 shows an excellent example of the behaviour of the englacial environment (Figure 4a). During the summer and autumn most sensor readings remained fairly constant. In Probe 26 water pressure remained high except for DOY 265–267 where there was a decrease in water pressure, accompanied by a rise in probe temperature (Figure 4a).

In contrast, during the winter sensor readings were more variable. Beginning DOY 292 the water pressure began to have a diurnal pattern that continued until the end of the record. This pattern commenced once melting ceased at the glacier surface. This comprised a peak at 13:00 followed by a slow decline. After each positive degree day melt event there was an immediate (but) slow rise in water pressure which reached a peak (12% increase) after approximately five days. There was an inverse relationship between air temperature/melt during the melt event and the resultant peak in water pressure five days later ($r^2 = 0.8065$). There was also a drop in conductivity as winter began (DOY 292), and this remained low throughout the winter.

Case stress remained constant (and high) until DOY 79 when there was a dramatic decrease, and after that the case stress was very variable, with the largest changes on the day or the day after a peak in meltwater. DOY 79 was the warmest day since the...
previous November. There was also a rise in probe temperature during the high water pressure events.

The GPR evidence showed a low water content, and borehole videos showed no fractures or conduits intersecting the borehole, which suggests the system was relatively closed. It could be argued that the melt season water pressure stability was because the borehole remained full of water (from the drilling) and was unable to drain (DOY 220–291). However, immediately once winter began (i.e. negative degree days) a diurnal cycle commenced (DOY 292–112), water pressure declined and there developed an immediate but slow response to positive degree day events. This implies that the borehole was connected with the englacial or subglacial system. During the melt season, inputs must be equal or greater than outputs to preserve the water pressure stability. But once surface inputs were stopped, during winter, water pressure decreased and the only water sources must be within the ice. Since there is a diurnal change (and winter diurnal rises in discharge were recorded (Young et al., 2015), these sources must be a combination of the release of storage (Rennermalm et al., 2013; Schoof et al., 2014), as well as melt generated from glacier movement. This ‘new’ water has much lower conductivity (due to dilution) and higher temperatures.

Numerous researchers have discussed how water travels through the ice. Flux through veins is negligible as ice has a very low permeability (Fountain and Walder, 1998; Gulley et al., 2009). Raymond and Harrison (1975) reported millimetre scale passages, whilst Hodge (1976), Pohjola (1994) and Harper and Humphrey (1995) reported passages 0.1 m in scale with an estimated flow of 0.01 to 0.1 m s\(^{-1}\). We recorded very few englacial passages; only 5% (one out of 19) of boreholes intersected with a fracture and/or conduit. This was much lower than others

![Figure 4](continued)

![Figure 5](Graph showing the mean daily horizontal surface velocity, tilt and surface melt 2012. [Colour figure can be viewed at wileyonlinelibrary.com])

![Figure 5](Graph showing the mean daily horizontal surface velocity, tilt and surface melt 2012. [Colour figure can be viewed at wileyonlinelibrary.com])
figures reported from mostly temperate rigid bedded glaciers. Fountain and Walder (1998) described from a range of glaciers that approximately 50% of boreholes drained before the bedrock was reached, indicating the presence of englacial passages. In addition, Fountain et al. (2005) reported, from video images of 48 boreholes from Storglaciären, Sweden, that 78% intercepted englacial passages of which most were fractures. However, it is possible that millimetre scale passages or crevasses were present at Skálfafellsjökull which were too small to observe on the video images.

The subglacial environment

Probes 21 and 25 (in the till) showed a similar pattern to each other (Figures 4b and 4c). Water pressure was generally high and either stable or variable. Case stress was intermediate and had an inverse relationship with water pressure. The conductivity at the beginning of the record was low and then increased (indicative of wet sediments). The stable water pressure pattern may reflect a lack of connection with the subglacial drainage system (Murray and Clarke, 1995; Kavanaugh and Clarke, 2001; Nienow et al., 2005). These locations could represent ‘islands’ of high pressure located away from the main drainage systems. In contrast, the variable water pressure pattern may indicate a connection with the subglacial water system, similar to that suggested by Hubbard et al. (1995) (‘variable pressure axis’) and Gooseff et al. (2002).

The style of variable water pressure varied between the seasons. This is clearly illustrated from Probe 25 (Figure 4c). During the melt season 2010 there was variation in water pressure, but there was no relationship between water pressure and surface melt. This was in complete contrast to the winter events. In winter 2009 Probe 25 showed a distinct drop in water pressure immediately after the only two positive degree days (DOY 52 and 80). This also occurred during winter 2009/2010. These later events are seen more clearly in Probe 21 during 2009/2010. In that winter, every time air temperatures rose above 2°C at the Base Station, water pressure dramatically decreased. This was accompanied by a rise in case stress.

We argue that the overall nature of the water pressure record (stable or variable) was due to the location of the probe in relation to the drainage system. For those probes undergoing variable water pressure changes, these changes were not related to melt during the melt season, but were related to melt during the winter. In addition, during the winter, these melt events are accompanied by speed-up events.

We suggest that the continuous change in tilt that was recorded during the melt season reflects deformation within the till. This deformation was continuous throughout the year, although slower in winter. These results are similar to those from Hart et al. (2009) from Briksdalsbreen, where there was also continuous rotation of the probe through the year, with a rate related to glacier surface velocity.

We determined the strength of the bed at the initial probe deployment sites by taking the mean of a 2 m² area around the location to allow for error. They all fell within the intermediate bed reflection range, which we have interpreted as deforming till. Results from the tilt data (Table II) showed that although each probe moved at a different rate, there was a generally similar pattern. During the melt season, there was a continuous change in tilt (average 0.1° per day). During the winter, the tilt change was reduced to 60% of the melt season rate. The winter base velocity was also similarly 63% of the melt season mean velocity. The major difference however, was that in winter there were distinct oscillations in tilt (superimposed on the general trend), which occurred during the positive degree days (Figure 5).

Spatial data on the nature of the bed was provided by the GPR data. This showed the widespread presence of a saturated till bed (77% of glacier bed, typically 8 m in length). This remained relatively constant 2008–2012. We use all this evidence to suggest that there was a deforming bed beneath the study area, which was active throughout the year.

The ice/till interface

Probe 24 had a very different record from the three others discussed in detail (Figure 4d). This probe was initially deployed in the till, and the first part of its record (DOY 202–307) was very similar to that of Probe 21, with high and stable water pressures, intermediate and stable case stress, and increasing conductivity. However, when data returns (DOY 27–66) water pressures were over 100% (mean 108%) (and this is the only time in the record any probe showed this), the case stress remained similar, but the conductivity was much lower. On DOY 67 (unrelated to any melt event) a water pressure diurnal cycle was initiated, and water pressures rose to over 113% (case stress rose slightly to 236.6 N); then after DOY 75 water pressures responded directly to melt events (combined with a diurnal cycle). Conductivity rose after DOY 50, and then experienced a small rise during the extreme high water pressures (DOY 67–75) (3.64–3.68 ohms), and then decreased and had an inverse relationship with water pressure until the end of the data series.

At Briksdalsbreen no diurnal cycles were recorded in probes deployed in the till (Hart et al., 2011). Even the till probes at Skálfafellsjökull with a variable water pressure record (and assumed to be connected with the subglacial water system), were not variable at the diurnal scale. Therefore, we suggest that when Probe 24 began a diurnal cycle it was connected directly with surface water (since diurnal cycles were recorded in the ice probes), i.e. that it was at the ice/sediment interface.

Since Probe 24 was initially deployed in the till, and during the early part of its record it behaved very similarly to the other till probes with a stable water pressure record (Probe 21), we suggest the following scenario (Figure 6):

a. Sometimes between DOY 307 and 27 the water pressure in Probe 24 increased to over 100%. We suggest the probe must be in some type of confined area. Although Murray
and Clarke (1995) argue that greater than 100% water pressures in unconnected areas occur due to low water pressures in connected areas, this is unlikely to explain this situation. We saw no change in water pressure associated with the meltwater event around DOY 50. We suggest the confined area could be a stiffer part of the till, a more clast-rich till, or the probe was held against an obstacle (Figure 6i).

b. Although there was no change in water pressure during the melt event around DOY 50, we can assume there was a speed-up event and there was a subsequent rise in conductivity associated with an increase in wetness. A combination of the sediment change, and glacier speed-up event may have destabilized the probe, and so it subsequently moved on DOY 67 into a location at the ice/sediment interface where it became directly connected to the subglacial water drainage system (hence the diurnal signal). Since the probe water pressure was even higher (and the case stress still high), we suggest that probe remained trapped behind an obstacle at the ice/sediment interface (Figure 6ii).

c. On DOY 75, there was another meltwater driven event (with an associated speed-up), which coincided with the dramatic water pressure and conductivity decrease (Figure 4d). We suggest, due to the associated basal sliding, the probe was moved out of its ‘lodged’ location, but was still in a connected area at the ice/sediment interface. It may have resided in a microcavity related to the obstacle (Figure 6iii). The reduction in conductivity could result from the till draining or surface meltwater entering the microcavity and causing dilution.

d. The melt events of DOY 75, DOY 93 and DOY 110, were marked by a decrease in water pressure and conductivity. This was followed by a rise in water pressure which peaked five days later, accompanied by continued decreases in conductivity, so the maximum water pressure and minimum conductivity occurred simultaneously with the peak in water pressure in Probe 26. This is further evidence that the probe is now at the ice/sediment interface since it is behaving in a similar way to Probe 26 in the ice.

Contrasting behaviours during the melt season and winter

There are few instrumented studies of the whole glacier (i.e. ice, till and ice/sediment interface) and even fewer during the winter. Here we contrast the behaviour in the three environments (Figure 7).

During the melt season, melt water input into the glacier was high, and equalled or exceeded output. There were no diurnal water pressure changes in the ice. Water pressure in the till was stable or variable related to the location (whether connected or not), but variable water pressures were not related to melt water inputs. The conductivity results in the till from Probe 21 and Probe 25 show that the till remained saturated throughout the season. This accords with the till data for constant deformation.

Winter behaviour was very different and was characterized by two distinct regimes, normal ‘base’ behaviour and melt events. During the melt events, water travelled via different pathways. The first was by direct drainage system through crevasses and moulins, which transported the surface water generated during the positive degree days to the glacier bed, which resulted in basal sliding and glacier speed-up ‘fast’ direct delivery system discussed by Zwally et al., 2002, Fountain et al., 2005, Das et al., 2008 and Gulley et al., 2009]. This was accompanied by a dramatic fall in water pressure in the till, and change in tilt of the probe.

Some of this water joined the ‘indirect’ system flowing through the smaller interconnected elements in the englacial...
system, and took five days to reach maximum water pressure [similar slow delivery systems have been described by Pohjola, 1993, Murray et al., 2000, Wilson et al., 2013 and Schoof et al., 2014 who described a summer two day lag]. This could be fed from smaller elements such as cracks which can act as englacial storage systems (Schoof et al., 2014). Similar reports of the co-existence of fast and slow drainage have come from Storglaciären, Sweden (Gulley et al., 2009) that had a fast conduit system (Hooke et al., 1988) combined with a slow fracture system (Fountain et al., 2005). We were able to show that during larger melt events, proportionally more of this water was channelled directly to the bed rather than feeding the slow englacial system. This may be because increased surface discharge led to the opening of the larger crevases and moulins. During the negative degree days (‘base’ behaviour), water was generated from storage and/or from the frictional heat from glacier movement to produce diurnal fluctuations in water pressure, whilst the indirect flow fed from the melt events continued to travel through the glacier. There was also slow deformation in the till (60% of summer rate).

We have no data concerning the ice/sediment interface during the melt season, however during the winter we saw all three water flow patterns. This included, constant diurnal flow, and then a drop in water pressure during the melt event associated with the direct flow, and then a rise in water pressure peaking five days later reflecting the indirect flow. This latter behaviour does not appear to affect the till nor glacier movement.

Comparison with Greenland

We can compare our results to those from Greenland (Smeets et al., 2012; Meierbachtol et al., 2013; van der Wal et al., 2015) who also used a wireless probe in the ice to record water pressure. They showed the water pressure readings from two boreholes 5 m apart had almost identical behaviour. The pattern of water pressure was very different to that seen at Skálfellsjökull. In Greenland, during autumn and winter, water pressure was relatively high and stable and slowly rose over the period, although in autumn and early winter single melt events affected water pressure and velocity. With the arrival of spring, water pressures initially rose to over 100% and then there was an overall decline in velocity over the summer, with some small velocity increases due to strong melt events or lake drainage (Hoffman et al., 2011; Das et al., 2008) before slowly rising as the summer ended. In general, water pressure and melt were inversely related. During summer, there was a diurnal water pressure record, with meltwater peaks at 14:00 and water pressure and velocity peaks at 15:00.

Figure 7. Schematic diagram to show water pathways during the melt season and winter. [Colour figure can be viewed at wileyonlinelibrary.com]
Abrupt rise (and then fall) in water pressure and velocity increase (spring event)\(\text{MW} > \text{DC}\).

Low water pressure = channels \(\text{MW} < \text{DC}\).

High water pressure = distributed system \(\text{MW} \approx \text{DC}\).

Abrupt rise (and then fall) in water pressure and velocity increase (speed-up events) \(\text{MW} > \text{DC}\).

Low water pressure (rising between melt events) = more channelized \(\text{MW} < \text{DC}\).

Note: MW, melt water; DC, drainage capacity.

### Relationship to different responses of melt to glacier dynamics

There has recently been a discussion concerning the effect of increased surface melting on glacier dynamics (as discussed in the Introduction), with most researchers from studies in Greenland, suggesting that once a summer subglacial channelized system develops, then additional melt water has little effect (Sundal et al., 2011; Sole et al., 2013; Tedstone et al., 2015).

Our results suggest that we have a summer distributed system dominated by wide anastomosing broad flat channels and thin water sheets (Hock and Hooke, 1993; Creyts and Schoof, 2009; Schroeder et al., 2013). Water from surface melt rapidly reaches the bed and causes velocity increases during both the melt season and the winter (although the latter is more dramatic). Water is stored within the subglacial system rather than englacially (as we recorded low glacier water content), and the distributed system can rapidly adjust to changes in meltwater production. Similarly, Minchew et al. (2016) have argued that soft bedded glaciers, are more sensitive to meltwater production.

However, Andrews et al. (2014) and Hoffman et al. (2016) have argued that the pattern proposed for the subglacial hydrology for the Greenland ice sheet (discussed earlier) may be more complex. They also stress the importance of a distributed system (alongside a channelized system) during the melt season. They suggest the distributed system beneath a rigid bedded glacier comprises linked cavities, with a low hydraulic conductivity. They suggest that these patches cover approximately 66% of the bed.

### Conclusions

We demonstrate the englacial and subglacial process associated with a soft bedded temperate glacier. This glacier has a fast englacial transfer system which leads to immediate changes in the subglacial hydrological system and glacier velocity in response to surface melt. However, there was an additional slow englacial system which had a five day lag between surface event and subsurface response. This slow system also had an effect at the ice/sediment interface, but was not recorded in the till.

The till and the subglacial hydrological system were coupled, and the bed was composed of a mosaic of different bed strengths (Alley, 1993; Hart and Boulton, 1991; van der Meer et al., 2003; Piotrowski et al., 2004). We are able to show the scale of these patches (e.g. deforming bed equals 77% of total area, typically 8 m in length) which remain relatively...
constant throughout the five year study period. However, these classifications may be too coarse to reflect the smaller scale differences between stable and variable water pressures, and the continuum of subglacial hydrological patterns which determine stabilization during the winter, which varied each year.

We were able to show that the till deforms throughout the year. Water pressures were normally high during the melt season (due to very high meltwater inputs), but were often more variable during the winter, and were often associated with a distinctive water pressure cycle related to surface melt events.

We were able to contrast our results with those from a rigid bedded glacier in Greenland. We argue that the seasonal behaviour at Skåófellsjökull is very different, instead of the typical high pressure winter distributed system and low pressure summer channelized system we show the opposite. We argue that because there it is a soft bedded system, the subglacial hydrology is dominated by a distributed system that may become more channelized in winter. These systems are very responsive to melt water inputs, particularly in winter.

Acknowledgements—The authors would like to thank the Glacsweb Iceland 2008/2011 and 2012 teams for help with probe development and data collection. Thanks also go to Lyn Aspden and Mark Dover and in the Cartographic Unit for figure preparation. This research was funded by Engineering and Physical Sciences Research Council (EP/C511050/1), Leverhulme Trust (F/00 180/AK) and the National Geographic Society (GEFNE45-12) and the GPR and Leica 1200 GPS units were loaned from the NERC Geophysical Equipment Facility.

References
Alley RB. 1993. In search of ice-stream sticky spots. Journal of Geology 39: 447–454.
Alley RB, Blankenship DD, Bentley CR, Rooney S. 1986. Deformation of till beneath ice stream B, West Antarctica. Nature 322(6074): 57–59.
Anderson RS, Anderson SP, MacGregor KR, Waddington ED, O’Neill S, Rihimaki CA, Loso MG. 2004. Strong feedbacks between hydrology and sliding of a small alpine glacier. Journal of Geophysical Research: Earth Surface 109(F3).
Andrews LC, Catania GA, Hoffman MJ, Gulley JD, Lüthi MP, Ryser C, Hawley RL, Neumann TA. 2014. Direct observations of evolving subglacial drainage beneath the Greenland Ice Sheet. Nature 514(7520): 80.
Bartholomew I, Nienow P, Mair D, Hubbard A, King MA, Sole A. 2010. Seasonal evolution of subglacial drainage and acceleration in a Greenland outlet glacier. Nature Geoscience 3(6): 408.
Bartholomew I, Nienow P, Sole A, Mair D, Cowton T, Palmer S, Wadhams J. 2011. Supraglacial forcing of subglacial drainage in the ablation zone of the Greenland Ice sheet. Geophysical Research Letters 38(8).
Bartholomew I, Nienow P, Sole A, Mair D, Cowton T, King MA. 2012. Short-term variability in Greenland Ice Sheet motion forced by time-varying meltwater drainage: Implications for the relationship between subglacial drainage system behavior and ice velocity. Journal of Geophysical Research: Earth Surface 117(F3).
Blake EW, Clarke GKC, Gerrin MC. 1992. Tools for examining subglacial bed deformation. Journal of Glaciology 38: 388–396.
Boulton GS, Hindmarsh RCA. 1987. Sediment deformation beneath glaciers: rheology and geological consequences. Journal of Geophysical Research: Solid Earth 92(B9): 9059–9082.
Boulton GS, Jones AS. 1979. Stability of temperate ice caps and ice sheets resting on beds of deformable sediment. Journal of Glaciology 24: 29–43.
Boulton GS, Dobbie KE, Zatešpin S. 2001. Sediment deformation beneath glaciers and its coupling to the subglacial hydraulic system. Quaternary International 86: 3–28.
Braithwaite RJ. 1995. Positive degree-day factors for ablation on the Greenland Ice-sheet studied by energy-balance modelling. Journal of Glaciology 41: 153–160.
Brown NE, Hallet B, Booth DB. 1987. Rapid soft bed sliding of the Puget Glacier lobe. Journal of Geophysical Research 92(B9): 8985–8997.
Crande GKC. 1987. Subglacial till: a physical framework for its properties and processes. Geophysical Research: Solid Earth 92(B9): 9023–9036.
Cohen J, Screen JA, Furtado JC, Barlow M, Whittleton D, Coumou D, Francis J, Dethloff K, Entekhabi D, Overland J, Jones J. 2014. Recent Arctic amplification and extreme mid-latitude weather. Nature Geoscience 7(9): 627.
Cowton T, Nienow P, Sole A, Wadhams J, Lit G, Bartholomew I, Mair D, Chandler D. 2013. Evolution of drainage system morphology at a land-terminating Greenlandic outlet glacier. Journal of Geophysical Research: Earth Surface 118(1): 29–41.
Creyts TT, Schoof CG. 2009. Drainage through subglacial water sheets. Journal of Geophysical Research: Earth Surface 114(F4).
Das SB, Joughin I, Behn MD, Howat IM, King MA, Lizzaralde D, Bhatia MP. 2008. Fracture propagation to the base of the Greenland Ice Sheet during supraglacial lake drainage. Science 320(5877): 778–781.
DeConto RM, Pollard D. 2016. Contribution of Antarctica to past and future sea-level rise. Nature 531: 591–597.
Fischer UH, Clarke GKC. 2001. Review of subglacial hydro-mechanical coupling: Trapridge glacier, Yukon Territory, Canada. Quaternary International 86(1): 29–43.
Fountain AG, Walder JS. 1998. Water flow through temperate glaciers. Reviews of Geophysics 36(3): 299–328.
Fountain AG, Jacobel RW, Schlichting R, Jansson P. 2005. Fractures as the main pathways of water flow in temperate glaciers. Nature 433(7026): 618–621.
Gades AM, Raymond CF, Conway H, Jacobel RW. 2000. Bed properties of Siple Dome and adjacent ice streams, West Antarctica, inferred from radio-echo sounding measurements. Journal of Glaciology 46(152): 88–94.
Gooseff MN, McKnight DM, Lyons WB, Blum AE. 2002. Weathering reactions and hyporheic exchange controls on stream water chemistry in a glacial meltwater stream in the McMurdo Dry Valleys. Water Resources Research 38(12).
Gulley JD, Benn DJ, Scootan E, Martin J. 2009. Mechanisms of englacial conduit formation and their implications for subglacial recharge. Quaternary Science Reviews 28(19): 1984–1999.
Harper JT, Humphrey NF. 1995. Borehole video analysis of a temperate glacier/englacial and subglacial structure: Implications for glacier flow models. Geology 23(10): 901–904.
Hart JK, Boulton GS. 1991. The interrelationship of glaciotectonic and glaciodepositional processes within the glacial environment. Quaternary Science Reviews 10(4): 335–350.
Hart JK, Martinez K. 2006. Environmental sensor networks: a revolution in the earth system science? Earth-Science Reviews 78(3): 177–191.
Hart JK, Rose KC, Martinez K, Ong R. 2009. Subglacial clast behaviour and its implication for till fabric development: new results derived from wireless subglacial probe experiments. Quaternary Science Reviews 28(7): 597–607.
Hart JK, Rose KC, Martinez K. 2011. Subglacial till behaviour derived from in situ wireless multi-sensor subglacial probes: rheology, hydro-mechanical interactions and till formation. Quaternary Science Reviews 30(1): 234–247.
Hart JK, Rose KC, Clayton A, Martinez K. 2015. Englacial and subglacial water flow at Skåófellsjökull, Iceland derived from ground penetrating radar, in situ Glacsweb probe and borehole water level measurements. Earth Surface Processes and Landforms 40(15): 2071–2083.
Hart JK, Clayton AI, Martinez K, Rolson BA. 2018. Erosional and depositional subglacial streamlining processes at Skåófellsjökull, Iceland: an analogue for a new bedform continuum model. GFF 1–17.
Hicock SR, Hicock SR. 1992. Lobal interactions and rheologic superposition in subglacial till near Bradville, Ontario, Canada. Boreas 21(1): 73–88.
Hicock SR, Dreimanis A. 1992. Deformation till: implications for rapid flow along the south-central margin of the Laurentide Ice Sheet. Canadian Journal of Earth Sciences 29(7): 1565–1579.
Hock R. 2003. Temperature index temperature modelling in mountain areas. Journal of Hydrology 282: 104–115.
Stokes CR, Clark CD, Lian OB, Tulaczyk S. 2007. Ice stream sticky spots: a review of their identification and influence beneath contemporary and palaeo-ice streams. *Earth-Science Reviews* **81**(3): 217–249.

Stokes CR, Clark CD, Storrar R. 2009. Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide Ice Sheet. *Quaternary Science Reviews* **28**(7): 721–738.

Sugiyama S, Skvarca P, Naito N, Enomoto H, Tsubaki S, Tone K, Morioka S, Aniya M. 2011. Ice speed of a calving glacier modulated by small fluctuations in basal water pressure. *Nature Geoscience* **4**: 597–600.

Sundal AV, Shepherd A, Nienow P, Hanna E, Palmer S, Huybrechts P. 2011. Melt-induced speed-up of Greenland ice sheet offset by efficient subglacial drainage. *Nature* **469**(7311): 521.

Tedstone AJ, Nienow PW, Gourmelen N, Dehecq A, Goldberg D, Hanna E. 2015. Decadal slowdown of a land-terminating sector of the Greenland Ice Sheet despite warming. *Nature* **526**(7575): 692.

Waechter A, Copland L, Herdes E. 2015. Modern glacier velocities across the Icefield Ranges, St Elias Mountains, and variability at selected glaciers from 1959 to 2012. *Journal of Glaciology* **61**(228): 624–634.