EARTH SURFACE PROCESSES AND LANDFORMS
Earth Surf. Process. Landforms **36**, 1230–1239 (2011)

Published online 31 March 2011 in Wiley Online Library
(wileyonlinelibrary.com) DOI: 10.1002/esp.2148

Temporal englacial water content variability associated with a rapidly retreating glacier

Jane K. Hart,[1] Kathryn C. Rose[1]* and Kirk Martinez[2]
[1] Schools of Geography, University of Southampton, Southampton, UK
[2] Electronics and Computer Science, University of Southampton, Southampton, UK

Received 18 November 2010; Revised 1 February 2011; Accepted 7 February 2011

*Correspondence to: Kathryn C. Rose, British Antarctic Survey, Madingley Road, Cambridge, CB3 0ET, UK. E-mail: kase@soton.ac.uk

ESPL
Earth Surface Processes and Landforms

ABSTRACT: This study uses a combination of evidence from ground penetrating radar, borehole, video, and wireless probe data to assess temporal changes in englacial water content associated with Briksdalsbreen, a rapidly retreating Norwegian glacier. Over a 13 day period in 2006, ice radar-wave velocity varied between 0·135 m/ns (± 0·009) and 0·159 m/ns (± 0·003), and water content from 7·8% (+2·6, −2·8) to 2·5% (+0·9, −1·1) [derived from the Looyenga (*Physica* **31**(3): 401–406, 1965) formula]. It is suggested that during warm precipitation free days, void spaces within the glacier become filled with water, resulting in low radar-wave velocity. This stored water then drained during cold, high precipitation days, allowing the radar-wave velocity to rise. These changes in englacial storage were caused by the enhanced crevassing generated by the newly floating ice margin, and were associated with accelerated glacier retreat. Copyright © 2011 John Wiley & Sons, Ltd.

KEYWORDS: englacial hydrology; GPR; glacier water storage; Briksdalsbreen; western Norway

Introduction

It is important to understand glacier water storage, the morphology and distribution of englacial drainage networks and the mechanisms by which these networks may route water into the sub-glacial environment. Understanding the relationships between these components is fundamental to the study of glacier dynamics (Iken and Bindschadler, 1986; Willis, 1995; Fountain and Walder, 1998; Clarke, 2005). In turn, this is particularly important for predictions concerning the stability of the Antarctic (e.g. Hughes, 1996; Vaughan and Sponge, 2004) and Greenland (e.g. Krabill *et al.*, 2002; Velicogna and Wahr, 2006; Nick *et al.*, 2009) ice sheets, in association with current temperature rises.

Glaciers are logistically difficult to study, but they can be 'imaged' by the combined techniques of borehole video cameras (e.g. Harper and Humphrey, 1995), studies of borehole behavior (e.g. Gordon *et al.*, 2001; Harper *et al.*, 2001), *in situ* experiments (e.g. Fischer and Clarke, 2001) and geophysical techniques [e.g. ground penetrating radar (GPR), Murray *et al.*, 2000; seismic surveys, Smith and Murray, 2009; electromagnetic methods, Kulessa, 2007]. In the last 10 years, investigations studying the nature of the englacial hydrological system have revealed a range of new information, highlighting the important role englacial processes may play within the glacier system.

Jansson *et al.* (2003) have argued that glaciers have the potential to store large volumes of water, but that determining the precise location and morphology of these water bodies is more complex. GPR studies have investigated the spatial distribution of water within glaciers (Lingle and Fatland, 2003; Pettersson *et al.*, 2003; Hubbard *et al.*, 2003; Bradford and Harper, 2005; Catania *et al.*, 2008; Kulessa *et al.*, 2008; Endres *et al.*, 2009). Murray *et al.* (2000) showed that the morphology of englacial storage spaces at the temperate glacier Falljökull, comprised a series of connected centimeter to decimeter sized voids. Fountain *et al.* (2005) stressed the importance of crevasses in englacial water storage. Gulley and Benn (2007) have shown that englacial conduits may subsequently develop along these crevasses via a mechanism of cut-and-closure. This process may be an important component in rapid water transfer from the glacier surface to its bed, via the englacial drainage system (e.g. Catania *et al.*, 2008; Das *et al.*, 2008).

In addition, it has been suggested that increases in the water content of glacier ice may affect its behaviour resulting in enhanced creep rates. Duval (1977) noted that the strain rate of ice increased by approximately 3·8 times for every 1% increase in water content (between 0·001 and 0·8%). Consequently, changes in the quantity and distribution of water within the glacier body may affect ice dynamics not only in terms of creep velocities, but also with regard to the opening and closure of englacial drainage pathways or void spaces. In turn, this may influence water routing to the glacier bed, providing an additional control on sub-glacial behaviour. This highlights the importance of constraining both temporal and spatial variations in glacier water content.

Numerous researchers have used GPR to calculate the water content of ice (e.g. Macheret *et al.*, 1993; Moore *et al.*, 1999) and demonstrate that water content varies with glacier depth (Macheret and Glazovsky, 2000; Murray *et al.*, 2000; Bradford and Harper, 2005). Very few studies have looked at temporal changes in water content. Macheret and Glazovsky (2000) examined Fridtjovbreen twice during a year and Irvine-Fynn *et al.* (2006) investigated Stagnation Glacier three times during a year. Jacobel and Raymond (1984) undertook a study of Variegated Glacier three times a day for 16 days; whilst Gades (1998) investigated Black Rapids Glacier at three fixed locations at half-hour intervals for 50 days, during a spring speed up event. The latter showed that the observed speed up of the glacier was probably not driven by local changes in meltwater input at the centre of the glacier, but instead from meltwater inputs originating at the glacier's margins. Kulessa *et al.* (2008) examined Grubengletscher every 10 minutes for one summer day, and showed that automated GPR techniques can monitor hydrological changes at a high temporal resolution. In particular, they argued that future work should combine surface radar monitoring with direct sub-glacial observations.

In light of this, the aim of this study was to combine GPR survey data with sub-glacial observations, derived from borehole and wireless multi-sensor probe data (Hart *et al.*, 2006). These data are then interpreted in combination with meteorological patterns in order to investigate temporal variations in englacial water storage and pathways over a 13 day period at a rapidly retreating glacier and how they relate to changes in external forcing.

Field Site

The study was undertaken at Briksdalsbreen, southern Norway (Figure 1). This is an outlet glacier of the Jostedalsbreen icecap, resting on a Precambrian gneiss bedrock. The glacier advanced 600 m between 1955–1996 (notably 390 m after 1987), initially into its pro-glacial lake, and then over a birch forest (Norwegian Water Resources and Energy Directorate (NVE); Winkler, 1996; Nesje, 2005; Laumann and Nesje, 2009a, 2009b). The 1987–1996 advance has been attributed to increased precipitation resulting from a positive phase of the North Atlantic Oscillation (NAO) (Nesje and Dahl, 2003). However, between 1996–2006 the glacier retreated approximately 300 m, with an additional 133 m of retreat recorded in eight months (August 2006–April 2007). This was associated with a negative phase (reduced precipitation) of the NAO (Winkler *et al.*, 2009).

Since 2004, an area on the northern side of the glacier has been investigated (Figure 1c). This was chosen because it had the fewest crevasses and was not exposed to rockfalls from the adjacent icefall. As the glacier retreated, the ice thinned. In the study area shown in Figure 1c, the average depth of ice decreased from 68 m in 2004, to 30 m in 2006. At the start of the 2006 field season (July), areas adjacent to the study site had become buoyant, and regular calving events were observed. This investigation will focus on data from 2006.

Methods

Boreholes and probe deployment

In 2006, 12 boreholes were drilled with a Kärcher HDS1000DE hot water drill and videos were taken with a custom made CCD camera using infra-red (900 nm) illumination (Table I). The depths of the boreholes were measured with the drill hose and camera cable. It was concluded that borehole deviation from vertical was minimal because the camera remained in the centre of the hole as it was lowered to the borehole base. The boreholes were examined for evidence of englacial and sub-glacial drainage, as well as the presence of till at the ice-bed interface.

Two wireless probes, inserted into the till during the summer of 2005, were still functioning during the summer of 2006. Their sensors record sub-glacial water pressure, case strain, water content of the till, temperature and tilt angle (details are described in Hart *et al.*, 2006). However, only water pressure is discussed in the context of this paper. Previous research had indicated that between 2005/2006 the probes moved at a rate of approximately 40% of surface velocity (Hart *et al.*, 2011). As a result, the predicted positions of the probes in the summer of 2006 could be calculated and are shown in Figure 1d. Air temperature and rainfall data were collected from the glacier. In addition, the glacier was repeatedly surveyed by a TOPCON differential global positioning system (dGPS) (real time kinematic mode), which enabled surface melt to be estimated.

GPR survey

A Sensors and Software Pulse Ekko 100, with a 1000 V transmitter system, was used to survey the glacier in a number of ways. Initially, a common offset (CO) survey was performed on a transect grid (Figure 1d), using 50 MHz antennae, with a 2 m antennae spacing, 0·5 m sampling interval and a stacking of 16. A custom built sledge held the antennae at the correct distance apart and facilitated movement along each transect. These measurements were carried out on two separate days. On 27 July transects A, B, C, D27 and E were surveyed and on 29 July transects D29 and F were surveyed. (Note, the same transect, D, was surveyed on both the 27 and 29 July and, as such, the data from each day is labelled D27 and D29, respectively.) The locations of the transects were recorded using a TOPCON dGPS.

In addition, a static survey using the 50 MHz antennae was undertaken at two specific locations on the transects – Site A and Site B (Figure 1d). At both sites the antennae were fixed to the glacier surface and recordings were made every 30 minutes during the working day for four days (Site A: 1, 2, 4 and 5 August; Site B: 6–9 August 2006). The coupling between the antennae and the ice remained constant throughout the study period. The GPR was not affected by rainfall events as most supra-glacial water was carried in well-defined streams located away from the study area and/or drained into crevasses.

GPR data processing

The GPR data were analysed using the software package ReflexW. For the initial analysis of the CO surveys, low frequency noise was eliminated using a de-wow filter and a SEC gain function was applied. Once the radar-wave velocity was established (see later), the data were migrated using a one–dimensional (1D) Kirchhoff migration, before a topographic correction was applied. Because the glacier was relatively thin, the glacier bed could be clearly seen in the radar echo grams (Figure 2).

The radar velocity of ice (v) can be calculated given the known glacier depths (d), established from the boreholes (where they intersect with the radar grid; Table I; Figure 1d) and the two way radar travel time (t) where:

$$v = 2d/t \tag{1}$$

The percentage error of calculated radar-wave velocities was determined as the standard deviation of the mean (in percentage).

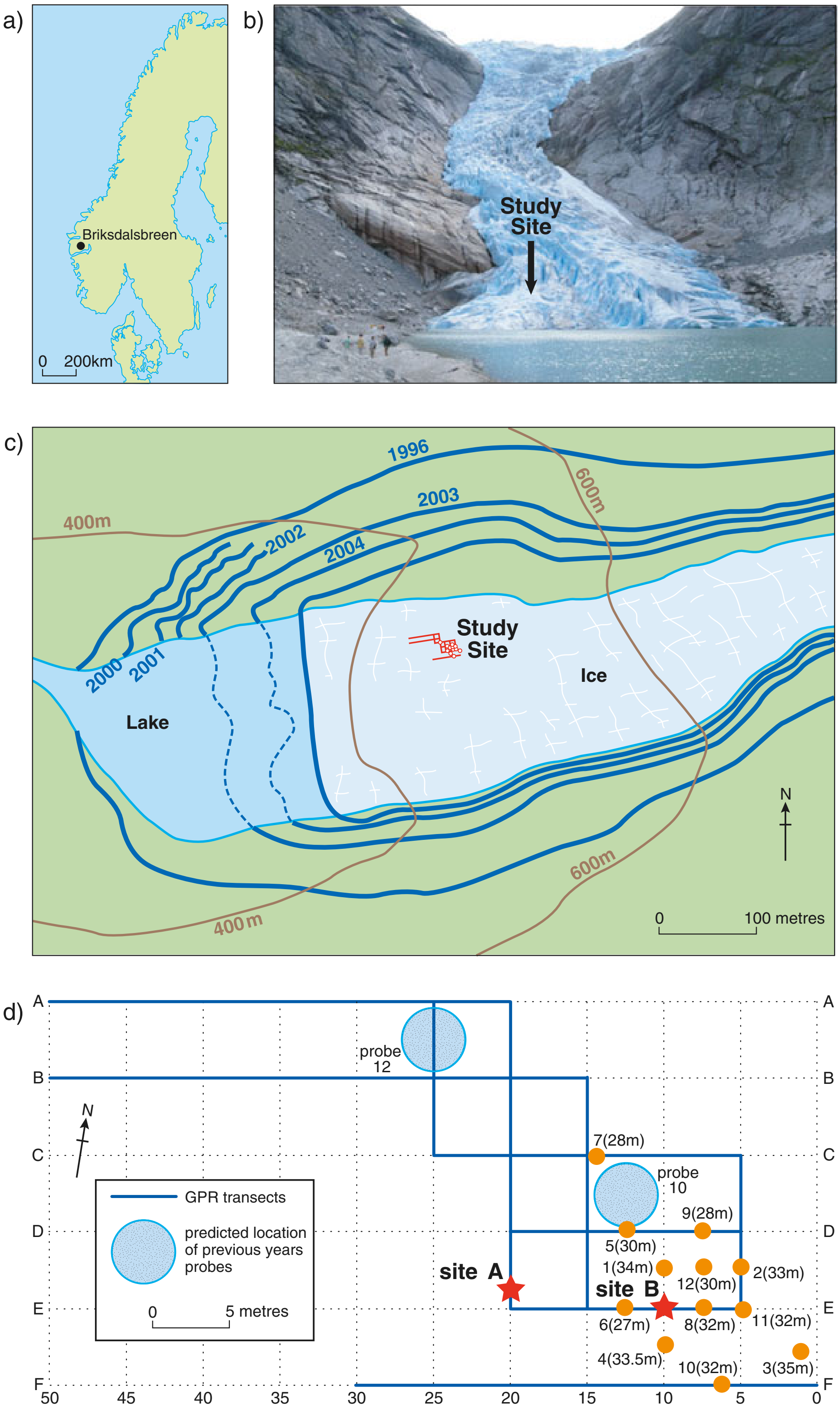

Figure 1. (a) Briksdalsbreen, Norway (61° 39′ 55″ N, 6° 52′ 10″ E); (b) photograph of the glacier; (c) map of ice retreat since 1996 with location of the study site; (d) location of GPR common offset survey transects (thick blue lines), GPR static survey sites A and B (red stars), borehole locations (orange circles) and the 2006 predicted locations of sub-glacial probes deployed in 2005 (large blue circles). This figure is available in colour online at wileyonlinelibrary.com/journal/espl

Glacier water content

It has been shown that the water content (W) of temperate glaciers can be estimated using the formula of Looyenga (1965) for three component dielectric mixtures of ice, air and water inclusions (Sihvola *et al.*, 1985; Macheret *et al.*, 1993; Frolov and Macheret, 1999; Macheret and Glazovsky, 2000). These inclusions or the 'total void space' within temperate ice may constitute both 'macro inclusions' (cavities, crevasses and channels > 1 mm) and 'micro inclusions' (veins, lenses and films at the boundaries of grains < 1 mm) (Macheret and Glazovsky, 2000). The Looyenga model makes no assumptions about

Table I. Borehole details 2006

Borehole number	Measured depth (m)	Depth of other englacial connections from base (m)	Englacial drainage during drilling	Drill pulled away at base by presence of flowing water	Flowing water seen on video	Water depth from base immediately after drilling (m)	Radar velocity	
							27 July (m/ns)	29 July (m/ns)
1	34	9	✓	x	x	13		
2	33	22	✓	✓	x	?		
3	35	9	✓	✓	x	15		
4	33·5	20·5	✓	✓	x	13		
5	30	3	✓	✓	✓	10		0·179
6	27	17	✓	✓	x	7	0·133	
7	27·5	12·5	✓	✓	✓	11	0·127	
8	31·5	24·5,14·5	✓	✓	x	11	0·146	
9	28	21	✓	x	x	8	0·126	0·158
10	32	20	✓	✓	✓	12		0·160
11	32	7	✓	x	x	11	0·144	
12	30	10	✓	✓	x	10		

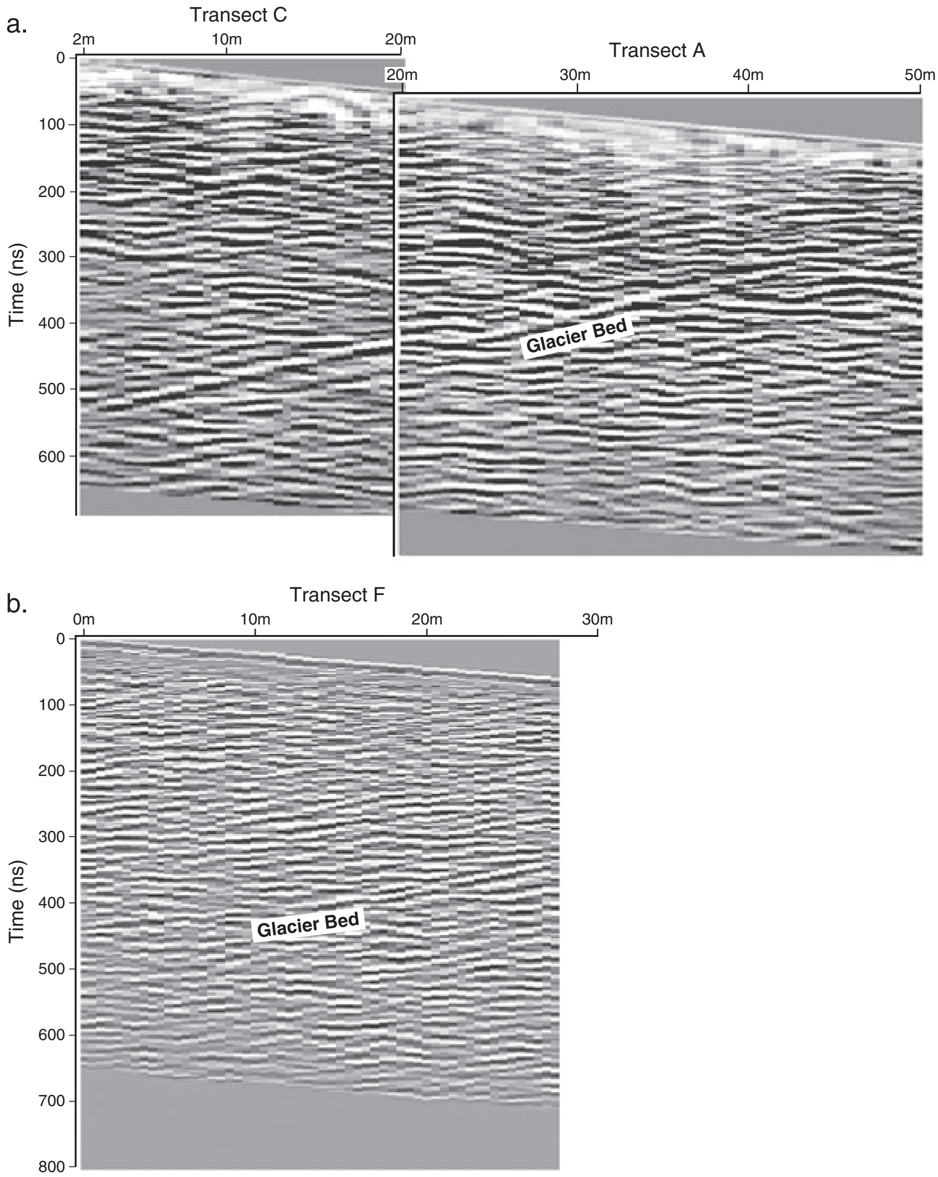

Figure 2. Composite down glacier radargrams: (a) transects C and A; (b) transect F. The following processing was applied to the images (de-wow, the application of a SEC gain function, Kirchhoff migration and a topographic correction).

inclusion shape (Barrett *et al.*, 2007) and is expressed as follows:

$$\varepsilon_m^a = \sum f_k \varepsilon_k^a \qquad (2)$$

where ε_m is the permittivity of the mixture, ε_k is the *k*th component with a volume portion $f_{k,}$ and $a = 1/3$. The permittivity of each component is:

$$\varepsilon_k = (c/v)^2 \qquad (3)$$

where *c* is the velocity of light and *v* is the measured radar-wave velocity. For the permittivity of temperate ice (ε_s), Equation 2 can also be expressed as follows:

$$\varepsilon_s = \left(\left(P_i\varepsilon_i^{1/3}\right) + \left(P_w\varepsilon_w^{1/3}\right) + \left(P_a\varepsilon_a^{1/3}\right)\right)^3 \qquad (4)$$

where ε_i is the permittivity of solid dry ice (taken as 3·19), ε_w is the permittivity of water (taken as 86), ε_a is the permittivity of air (taken as one) (values taken from Macheret *et al.*, 1993; Macheret and Glazovsky, 2000). The values of P_i, P_w and P_a are the proportions of ice, water and air where $P_i + P_w + P_a = 1$. If it is assumed that within temperate ice the total void space is water filled, then Equation 4 can be simplified to a two component model and water content ($W = P_w \times 100$) is expressed as follows:

$$W = \left(\varepsilon_s^{1/3} - \varepsilon_i^{1/3}\right) / \left(\varepsilon_w^{1/3} - \varepsilon_i^{1/3}\right) \qquad (5)$$

Glacier water content was calculated from the GPR CO grid survey radar-wave velocity readings collected on the 27 and 29 July. High radar-wave velocities indicate low water contents, whilst low velocities suggest higher ice water contents. The technique of Macheret and Glazovsky (2000) assumes that, on the day with the lowest recorded radar-wave velocity, all the void spaces in the ice will be filled with water. This allows the two component model to be used to calculate both the water content and the ice content from this data set ($P_i = 1 - P_w$). On the other days the three component model can be used and, since P_i is known, Equation 4 can be rewritten as follows:

$$\varepsilon_s = \left(\left(P_i\varepsilon_i^{1/3}\right) + \left(P_w\varepsilon_w^{1/3}\right) + \left((1 - P_i - P_w)\varepsilon_a^{1/3}\right)\right)^3 \qquad (6)$$

This then allows the relative proportions of ice, water and air to be determined on days when the total void space is not water saturated.

Results

Borehole and sub-glacial probe records

The video footage showed that the ice was debris free. Evidence for englacial drainage was observed in all the boreholes (Table I). Video footage displayed a variety of horizontal to vertical fractures within the walls of the boreholes. Those which were seen to have water flowing from them were considered to represent drainage passages. In addition, many boreholes drained during drilling when they intersected these englacial passages and/or voids. A number of boreholes also drained when they connected with the ice-bed interface.

Till and moving water were observed at the base of three boreholes (5, 7 and 10). In another three boreholes (1, 9 and 11), till and stationary water were noted. In the remaining boreholes (2, 3, 4, 6, 8 and 12), stationary water was viewed at the bed, but till was not visible. Unusually, after drilling, all the holes refilled to a mean height of 10·6 m above the bed [standard deviation (s.d.) 1·9 m], equivalent to 308 m above sea level (a.s.l.). This elevation corresponded with the lake surface level. Water levels stayed at this height for the remainder of the field season.

The results from the sub-glacial probes 10 and 12 recorded a slow decline in water pressure over the study period. Both probes indicate relatively high water pressures in the till. Probe 12 water pressure represented 73% of the glacier thickness and probe 10, only 26%. These results are discussed in more detail later.

At the glacier surface, repeat dGPS readings showed an elevation change of −1·88 m between the 29 July and the 10 August, equating to a melt rate of approximately 0·145 m/day.

GPR survey

Ice radar-wave velocities, derived from the CO survey, showed velocity variations between survey days (Table I). Those readings collected on the 27 July gave an average value of 0·135 m/ns (s.d. 0·009); whilst those obtained on the 29 July gave an average value of 0·159 m/ns (s.d. 0·001). The percentage error for the radar-wave velocities collected was 7% and 1% on the 27 and 29 July, respectively. Barrett *et al.* (2007) have argued that radar-wave velocity estimates have an error of approximately 2%. As a result, we have applied an adjusted error bar of 2% to the radar-wave velocities collected on 29 July, to reflect the minimum error estimate of Barrett *et al.* (2007).

In terms of the static GPR surveys, it was noted that the radar traces had a distinctive peak, which represents the ice-bed

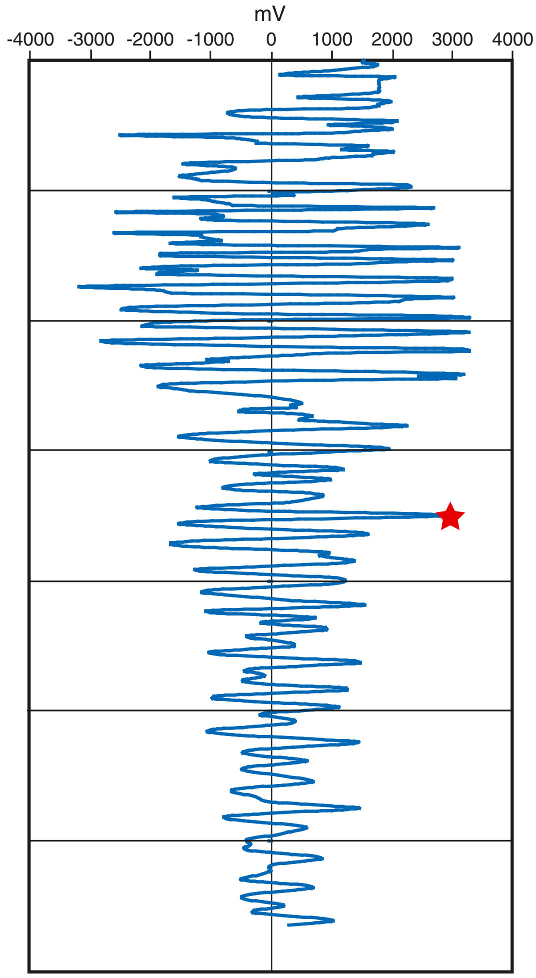

Figure 3. Radar trace (unprocessed) from Site B showing the location of the bed (indicated with a star). This figure is available in colour online at wileyonlinelibrary.com/journal/espl

interface (Figure 3a). The original depth of the glacier was known at Sites A and B from comparisons between the CO survey ice depths and corresponding measured boreholes depths. Thus, assuming a constant surface melt rate derived from dGPS readings, it was possible to calculate ice radar-wave velocity over the static survey study period (Table II).

Water content

Using the technique of Macheret and Glazovsky (2000), glacier water content was calculated to be 7·8 % (+2·6, −2·8) on 27 July (low radar-wave velocities; Equation 5). On the 29 July (high radar-wave velocities; Equation 6) water content was 2·7 % (+0·9, −1·1) and air content 5·1% (+1·7, −1·7). The uncertainties were based on the radar-wave velocity percentage error discussed earlier. This technique was also used to calculate the water and air content at Sites A and B, during the static survey study period (Table II). Values ranged from a water content of 2·5% (+0·9, −1·1) and air content of 5·4% (+1·7, −1·7), to a water content of 5·3% (+1·0, −1·0) and air content of 2·5% (+1·6, −1·8), averaged over both sites.

Briksdalsbreen's water content is within the range of normal values for glacier ice. Petterson *et al.* (2003) and Macheret and Glazovsky (2000) have reported the water content of temperate glacier ice to range between 0 and 9%, although values are normally less than 2%. Of a glacier's total void space Macheret and Glazovsky (2000) suggest that low water contents reflect water stored in micro inclusions, as the water content measured directly from temperate ice cores ranges from 0 to 1·4% (Raymond and Harrison, 1975). In contrast, higher water contents will reflect water storage in both micro and macro inclusions.

In summary, the GPR surveys (CO and static) showed that water content was variable over the 10 measured survey days. For six days the water content was low (average 2·7%), for two days the water content was intermediate (average 3·7%) and for two days the water content was high (average 6·7%). High water contents were recorded at the start of the survey period (27 July), but values subsequently decreased and continued to fluctuate during the remainder of the study.

Discussion

Comparison of glacier water content with the meteorological record

Using the techniques discussed earlier, it is possible to show variations in the ice radar-wave velocity and thus, infer changes in englacial water content over the study period (Table II; Figure 4d). The temporal variation in water content in the ice must reflect the changing hydrology of the glacier. This variation will result from a combination of changing meltwater inputs, storage and outputs. To investigate the relationship between these three components, the englacial conditions recorded can be compared with meteorological patterns and the sub-glacial water variability documented by the boreholes and wireless probes over the same time period (Figure 4).

Weather conditions on the 27 July, when glacier water content was high, were warm and precipitation free and had been preceded by a further nine precipitation free days. From the 28 July to the 2 August there was rainfall on five of the six days, during which time the glacier water content remained low. Then, between the 3 and 6 August, when little rainfall was recorded, the water content rose. This pattern continued whereby water content declined during periods of rainfall (e.g. 7 August), but rose again when rainfall began to decrease.

Sub-glacial water pressure data from the probes simply show a general reduction in pressure throughout the study period that does not correlate with temperature or precipitation (Figure 4a). Similarly, the height of water in the boreholes remained relatively constant. This would suggest that changes in englacial water content were not caused by changes in sub-glacial hydrology, nor did these changes influence sub-glacial drainage in that region of the bed. Instead, we argue that the differences in glacier water content are due to changes in englacial storage.

Interpretation

Numerous researchers (e.g. Macheret *et al.*, 1993; Moore *et al.*, 1999; Murray *et al.*, 2000) have argued that high water content values in ice are due to the presence of water-filled void spaces (both macro and micro inclusions). Our data indicate that during the beginning of the study period, the void spaces within the glacier were filled with water. It is suggested that the nine day period of warm, precipitation free, weather (prior to the GPR survey) caused high rates of continuous water input from surface melting. This resulted in the exceptionally high water content values recorded by the CO survey on the 27 July. Then, on the 28 July, the day was cooler, with a rainfall event, and water inputs were much lower (i.e. surface melting decreased). This enabled the glacier to drain slightly, reducing its water content by the 29 July.

This pattern continued over the rest of the study period. During relatively cool periods with precipitation, the water content in the ice was low. In contrast, during warm, precipitation free periods, the void spaces within the ice became filled with water. This continued until cooler, wetter conditions began to dominate the pattern of weather, from the

Table II. Ice radar-wave velocities and water content calculated over the study period using the Looyenga method (explained in the text)

Date	Ice radar-wave velocity (m/ns)	Water content	Air content
27 July	0·135 (± 0·009)	7·8 (+2·6, −2·8)	0
29 July	0·159 (± 0·003)	2·7 (+0·9, −1·1)	5·1 (+1·7,−1·7)
1 August	0·159 (± 0·003)	2·7 (+0·9, −1·1)	5·1 (+1·7,−1·7)
2 August	0·160 (± 0·003)	2·5 (+0·9, −1·1)	5·4 (+1·7,−1·7)
4 August	0·152 (± 0·003)	4 (+0·9, −1·1)	3·8 (+1·7,−1·7)
5 August	0·155 (± 0·003)	3·4 (+0·9, −1·1)	4·4 (+1·7,−1·7)
6 August	0·145 (± 0·007)	5·3 (+1·0, −1·0)	2·5 (+1·6,−1·8)
7 August	0·160 (± 0·003)	2·5 (+0·9, −1·1)	5·4 (+1·7,−1·7)
8 August	0·159 (± 0·003)	2·7 (+0·9, −1·1)	5·1 (+1·7,−1·7)
9 August	0·157 (± 0·003)	3·0 (+0·9, −1·1)	4·8 (+1·7,−1·7)

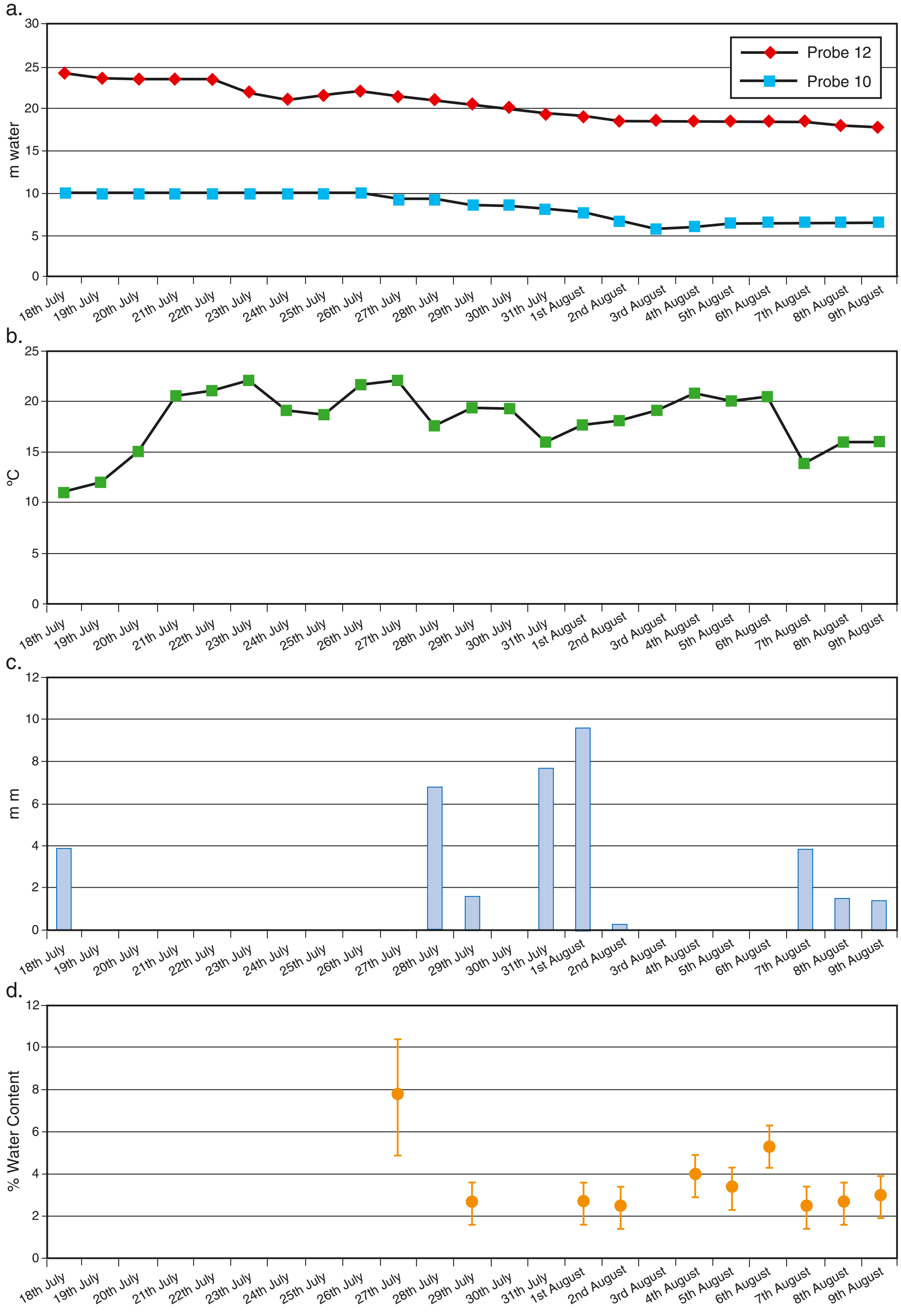

Figure 4. Results over the study period: (a) water pressure (measured in metres of water equivalent) from the Glacsweb wireless probe data; (b) daily mean air temperature; (c) daily total precipitation at the Base Station; (d) water content (for errors see Table II). This figure is available in colour online at wileyonlinelibrary.com/journal/espl

7 August. Presumably, in keeping with the earlier pattern, these conditions would have resulted in a continued decline in englacial water content.

Borehole water levels suggest that the sub-glacial hydraulic head was the same as the lake level. This suggests that there was connection between the sub-glacial water system and the lake. Video data showed stationary water at the base of the majority of boreholes. Thus, it is suggested that the ice-bed interface was dominated by a 'slow', distributed, sub-glacial water system (Fountain and Walder, 1998). This may have comprised a system of either 'microcavities' (Kamb, 1991), a braided canal network (Walder and Fowler, 1994) or a linked cavity system (Lliboutry, 1976; Kamb, 1987; Nienow *et al.*, 1998; Willis *et al.*, 2009).

It is interesting to note that the periods of decline in englacial water content were not apparent within the sub-glacial probe pressure records. The stored water may have been discharged to another region of the bed where the probes were not located. Unfortunately a greater spatial array of functioning sub-glacial probes was not available at this time to clarify this further. Alternatively, given that the sub-glacial environment showed increasing connectivity with the lake as the glacier continued to retreat, any water draining from the glacier body may have been readily accommodated and transferred to the lake by the sub-glacial drainage network. This pattern might also indicate a steady, continuous, transfer of water to the glacier bed, rather than sudden water release events.

Implications

There have been few reports of large temporal changes in ice radar-wave velocity in the literature. However, examples of short-term englacial storage can be found from studies of outburst floods from ice-dammed lakes. Anderson *et al.* (2003) showed that 30% of the lake volume at Kennicott Glacier was stored in englacial void space. Similarly, Huss *et al.* (2007) calculated water storage as 38% (2004) and 54% (2005) of the lake volume at Gornersee Glacier. They suggested that some of this water was stored in sub-glacial cavities (which produced surface uplift), whilst the rest was stored in the englacial system. They calculated the integrated void ratio (ratio between the volume of englacial voids and the total ice volume) as 0·1–10%. This range corresponds with the 7·8% total void space determined in this study.

These investigations demonstrate the importance of glacier water storage in both sub-glacial cavities and englacial voids. In recent work, Harper *et al.* (2010) have shown that a complex pattern of connectivity may exist between sub-glacial and englacial water stores. They demonstrate that significant englacial water storage may be driven from the glacier bed. Their study outlines new evidence for the extensive propagation of basal crevasses into the glacier body, in some cases reaching 20 m from the glacier surface. In these examples, water is derived from the sub-glacial environment, but large volumes of water are then stored within the glacier body in these basal crevasses. Harper *et al.* (2010) suggest these crevasses can develop rapidly and may provide storage areas during periods of sub-glacial flooding (Tsai and Rice, 2010). They may also be associated with aquatic margins (Sohn *et al.*, 1998).

A key aspect of this study is that, due to the rapid retreat of Briksdalsbreen, the 2006 study area had become part of a grounded-floating ice transition zone (Hughes, 1975). Numerous researchers have shown how a combination of buoyancy effects, changing water levels, complex velocity patterns, and stress changes lead to increased fracturing in such zones (Holdsworth, 1973; Fastook and Schmidt, 1982; Theakstone, 1989; Warren *et al.*, 2001; Benn *et al.*, 2007). It was evident that in 2006 Briksdalsbreen's margin had become more buoyant and numerous calving events were now observed during the field season. The sub-glacial hydraulic head was also at lake level, indicating that connectivity between the two environments had been established.

We suggest that a series of interconnected crevasses (both surface and basal) formed at Briksdalsbreen. The formation of these crevasses was promoted by the increased buoyancy of the glacier tongue, which had in turn resulted from the rapid retreat and breakup of the glacier. These crevasses provided zones for water storage, which became filled during high meltwater input days. During periods of low meltwater production, water was able to drain away into the sub-glacial system at a constant rate. This resulted in little hydrological change in the sub-glacial environment, but significant changes in short-term englacial storage.

Conclusion

The location and behaviour of water within englacial and sub-glacial environments is vital to our understanding of glacier dynamics and their response to climate change. This study showed that the water content in the ice varied dramatically from 7·8% to 2·7%. We suggest that this occurred because water was stored and/or released in englacial void space. During warm, precipitation free, periods, enhanced melt caused these void spaces to become filled with water; but during cooler, high precipitation, periods stores were able to drain.

We suggest that these unusual characteristics at Briksdalsbreen, in August 2006, occurred because the study area became part of a grounded-floating ice transition zone, which developed as a result of the rapid retreat of the glacier at this site. The dramatic crevassing observed at the glacier tongue increased the total void space available and thus, the water content of the glacier. By April 2007, these processes ceased as the glacier had retreated a further 133 m, whereby it was no longer calving into the lake, but had become grounded on its gneiss bedrock.

These findings demonstrate the importance of studying both temporal and spatial variations in glacier water content. We suggest that investigations should continue to study the quantity and distribution of stored water, as well as the mechanisms by which it may be propagated to the glacier bed, in order to improve our understanding of englacial–sub-glacial interactions and their effect on glacier dynamics.

Acknowledgements—The authors would like to thank the Glacsweb 2006 team for help with data collection (Al Riddoch, Ahmed Elsaify, Gang Zou, Paritosh Padhy, Celine Ragault, David Vaughan-Hirsch, Matthew Westoby) and Inge and Gro Melkevoll for help with logistics. Thanks also go to Bob Smith in the Cartographic Unit for figure preparation. This research was funded by the Royal Society, the Paul Instrument Fund and ESPRC, and supported by a GPR loan from the NERC Geophysical Equipment Pool.

References

Anderson SP, Walder JS, Anderson RS, Kraal ER, Cunico M, Fountain AG, Trabant DC. 2003. Integrated hydrologic and hydrochemical observations of Hidden Creek Lake jökulhlaups, Kennicott Glacier, Alaska. *Journal of Geophysical Research* **108**: F1, 6003. DOI. 10.1029/2002JF000004

Barrett BE, Murray T, Clark R. 2007. Errors in radar CMP velocity estimates due to survey geometry, and their implication for ice water content estimation. *Journal of Environmental and Engineering Geophysics* **12**: 101–111.

Benn DI, Warren CR, Mottram RH. 2007. Calving processes and the dynamics of calving glaciers. *Earth-Science Reviews* **82**: 143–179.

Bradford JH, Harper JT. 2005. Wavefield migration as a tool for estimating spatially continuous radar velocity and water content in glaciers. *Geophysical Research Letters* **32**: L08502. DOI. 10.1029/2004GL021770

Catania G, Neumann T, Price S. 2008. Characterizing englacial drainage in the ablation zone of the Greenland Ice Sheet. *Journal of Glaciology* **54**(187): 567–578.

Clarke GKC. 2005. Subglacial processes. *Annual Review of Earth and Planetary Sciences* **33**: 247–276.

Das SB, Joughin I, Behn MD, Howat IM, King MA, Lizarralde D, Bhatia MP. 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

Duval P. 1977. The role of water content on the creep rate of polycrystalline ice. *IAHS* **118**: 29–33.

Endres AL, Murray T, Booth AD, West JL. 2009. A new framework for estimating englacial water content and pore geometry using combined radar and seismic wave velocities. *Geophysical Research Letters* **36**: L04501.

Fastook, JL, Schmidt WF. 1982. Finite-element analysis of calving ice fronts. *Annals of Glaciology* **3**: 103–106.

Fischer UH, Clarke GKC. 2001. Review of subglacial hydro-mechanical coupling: Trapridge Glacier, Yukon Territory, Canada. *Quaternary International* **86**: 29–44.

Fountain AG, Walder JS. 1998. Water flow through temperate glaciers. *Reviews of Geophyics* **36**: 299 (97RG03579).

Fountain AG, Schlicting RB, Jacobel RW, Jansson P. 2005. Fractures as main pathways of water flow in temperate glaciers. *Nature* **433**: 618–621.

Frolov AD, Macheret YY. 1999. On the dielectric properties of dry and wet snow. *Hydrological Processes* **13**: 1755–1760.

Gades AM. 1998. *Temporal and Spatial Variations of Basal Conditions Beneath Glaciers and Ice Sheets Inferred from Radio Echo Measurements*, PhD Thesis, University of Washington, Seattle, WA, 192 pp.

Gordon S, Sharp M, Hubbard B, Willis I, Smart C, Copland L, Harbor J, Ketterling B. 2001. Borehole drainage and its implications for the investigation of glacier hydrology: experiences from Haut Glacier d'Arolla, Switzerland. *Hydrological Processes* **15**: 797–813.

Gulley J, Benn DI. 2007. Structural control of englacial drainage systems in Himalayan debris-covered glaciers. *Journal of Glaciology* **53**: 399–412.

Harper JT, Humphrey NF. 1995. Borehole video analysis of a temperate glacier's englacial and subglacial structure: implications for glacier flow models. *Geology* **23**: 901–904.

Harper JT, Humphrey NF, Pfeffer WT, Huzurbazar SV, Bahr DB, Welch BC. 2001. Spatial variability in the flow of a valley glacier: deformation of a large array of boreholes. *Journal of Geophysical Research* **106**: 8547–8563.

Harper JT, Bradford JH, Humphrey NF, Meierbachtol TW. 2010. Vertical extension of the subglacial drainage system into basal crevasses. *Nature* **467**: 579–582. DOI. 10.1038/nature09398

Hart JK, Martinez K, Ong R, Riddoch A, Rose KC, Padhy PA. 2006. A wireless multi-sensor subglacial probe: design and preliminary results. *Journal of Glaciology* **52**(178): 389–397.

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* **3**: 234–247.

Holdsworth G. 1973. Ice calving into the proglacial Generator Lake, Baffin Island, N.W.T., Canada. *Journal of Glaciology* **12**: 235–250.

Hubbard BP, Hubbard A, Mader H, Tison J-L, Grust K, Nienow PW. 2003. Spatial variability in the water content and rheology of temperate glaciers: Glacier de Tsanfleuron, Switzerland. *Annals of Glaciology* **37**: 1–6.

Hughes T. 1975. The West Antarctic Ice Sheet: instability, disintegration and initiation of ice ages. *Reviews of Geophysics and Space Physics* **13**: 502.

Hughes T. 1996. Can ice sheets trigger abrupt climatic change? *Arctic and Alpine Research* **28**: 448–465. DOI. 10.2307/1551856

Huss M, Bauder A, Werder M, Funk M, Hock R. 2007. Glacier-dammed lake outburst events of Gornersee, Switzerland. *Journal of Glaciology* **53**: 189–198.

Iken A, Bindschadler RA. 1986. Combined measurements of subglacial water pressure and surface velocity Findelengletscher, Switzerland: conclusions about drainage system and sliding mechanism. *Journal of Glaciology* **32**: 101–119.

Irvine-Fynn TDL, Moorman BJ, Williams JLM, Walter FSA. 2006. Seasonal changes in ground-penetrating radar signature observed at a polythermal glacier, Bylot Island, Canada. *Earth Surface Processes and Landforms* **31**: 892–909.

Jacobel R, Raymond C. 1984. Radio echo-sounding studies of englacial water movement in Variegated Glacier, Alaska. *Journal of Glaciology* **30**: 22–29.

Jansson P, Hock R, Schneider T. 2003. The concept of glacier storage – a review. *Journal of Hydrology* **282**: 116–129.

Kamb B. 1987. Glacier surge mechanism based on linked cavity configuration of the basal water conduit system. *Journal of Geophysical Research* **92**: 9083–9100.

Kamb B. 1991. Rheological nonlinearity and flow instability in the deforming-bed mechanism of ice stream motion. *Journal of Geophysical Research* **96**: 16585– 16595.

Krabill W, Abdalati W, Frederick E, Manizade S, Martin C, Sonntag J, Swift R, Thomas R, Wright W, Yungel J. 2002. Greenland Ice Sheet: high-elevation balance and peripheral thinning. *Science* **289**: 428.

Kulessa B. 2007. A critical review of the low-frequency electrical properties of ice sheets and glaciers. *Journal of Environmental & Engineering Geophysics* **12**: 23–36. DOI. 10.2113/JEEG12.1.23

Kulessa B, Booth AD, Hobbs A, Hubbard AL. 2008. Automated monitoring of subglacial hydrological processes with ground-penetrating radar (GPR) at high temporal resolution: scope and potential pitfalls. *Geophysical Research Letters* **35**: L24502. DOI. 10.1029/2008GL035855

Laumann T, Nesje A. 2009a. A simple method of simulating the future frontal position of Briksdalsbreen, western Norway. *The Holocene* **19**: 221–228.

Laumann T, Nesje A. 2009b. The impact of climate change on future frontal variations of Briksdalsbreen. *Journal of Glaciology* **55**: 789–796.

Lingle CS, Fatland DR. 2003. Does englacial water storage drive temperate glacier surges? *Annals Glaciology* **36**: 14–20.

Lliboutry L. 1976. Physical processes in temperate glaciers. *Journal of Glaciology* **16**: 151–158.

Looyenga H. 1965. Dielectric constant of heterogeneous mixtures. *Physica* **31**(3): 401–406.

Macheret YY, Glazovsky AF. 2000. Estimation of absolute water content in Spitsbergen glaciers from radar sounding data. *Polar Research* **19**: 205–216.

Macheret YY, Moskalevsky MY, Vasilenko EV. 1993. Velocity of radio waves in glaciers as an indicator of their hydrothermal state, structure and regime. *Journal of Glaciology* **39**: 373–384.

Moore JC, Pälli A, Ludwig F, Blatter H, Jania J, Gadek B, Glowacki P, Mochnacki D, Isaksson E. 1999. High-resolution hydrothermal structure of Hansbreen, Spitsbergen, mapped by ground-penetrating radar. *Journal of Glaciology* **45**: 524–532.

Murray T, Stuart GW, Fry M, Gamble NH, Crabtree MD. 2000. Englacial water distribution in a temperate glacier from surface and bore hole radar velocity analysis. *Journal of Glaciology* **46**(154): 389–398.

Nesje A. 2005. Briksdalsbreen in western Norway: AD 1900–2004 frontal fluctuations as a combined effect of variations in winter precipitation and summer temperature. *The Holocene* **15**: 1245–1252.

Nesje A, Dahl SO. 2003. The "Little Ice Age" – only temperature? *Holocene* **13**: 171–177.

Nick FM, Vieli A, Howat IM, Joughin I. 2009. Large-scale changes in Greenland outlet glacier dynamics triggered at the terminus. *Nature Geoscience* **2**: 110–114. DOI: 10.1038/NGEO394

Nienow P, Sharp M, Willis I. 1998. Seasonal changes in the morphology of the subglacial drainage system, Haut Glacier d'Arolla, Switzerland. *Earth Surface Processes and Landforms* **23**: 825–843.

Norwegian Water Resources and Energy Directorate (NVE). http://www.nve.no/en/About-NVE/ [31 January 2011]

Pettersson R, Jansson P, Holmlund P. 2003. Thinning of the cold surface layer on Storglaciaren, observed by repeated ground penetrating radar surveys. *Journal of Geophysical Research* **108**(F1): 6004. DOI. 10.1029/2003JF000024

Raymond CF, Harrison WD. 1975. Some observations on the behavior of the liquid and gas phases in temperate glacier ice. *Journal of Glaciology* **14**(71): 213–233.

Sihvola A, Nyfors E, Tiuri M. 1985. Mixing formulae and experimental results for the dielectric constant of snow. *Journal of Glaciology* **31**: 163–170.

Smith AM, Murray T. 2009. Bedform topography and basal conditions beneath a fast-flowing West Antarctic ice stream. *Quaternary Science Reviews* **28**: 584–596.

Sohn HG, Jezek KC, van der Veen CJ. 1998. Jakobshavn Glacier, west Greenland: 30 years of spaceborne observations. *Geophysical Research Letters* **25**: 2699–2702.

Theakstone WH. 1989. Further catastrophic break-up of a calving glacier: observations at Austerdalsisen, Svartisen, Norway, 1983–87. *Geografiska Annaler. Series A, Physical Geography* **71**: 245–253.

Tsai VC, Rice JR. 2010. A model for turbulent hydraulic fracture and application to crack propagation at glacier beds. *Journal of Geophysical Research* ***115***: F03007. DOI. 10.1029/2009JF001474

Vaughan DG, Sponge JR. 2004. Risk estimation of collapse of the West Antarctic Ice Sheet. *Climatic Change* **52**: 65–91.

Velicogna I, Wahr J. 2006. Acceleration of Greenland ice mass loss in spring 2004. *Nature* **443**: 329–331.

Walder JS, Fowler A. 1994. Channelized subglacial drainage over a deforming bed. *Journal of Glaciology* **40**: 3–15.

Warren C, Benn D, Winchester V, Harrison S. 2001. Buoyancy-driven lacustrine calving, Glacier Nef, Chilean Patagonia. *Journal of Glaciology* **47**(156): 135–146.

Willis I. 1995. Intra-annual variations in glacier motion: a review. *Progress in Physical Geography* **19**: 61–106.

Willis I, Lawson W, Owens I, Jacobel B, Autridge J. 2009. Subglacial drainage system structure and morphology of Brewster Glacier, New Zealand. *Hydrological Processes* **23**: 384–396.

Winkler S. 1996. Front variations of outlet glaciers from Jostedalsbreen, western Norway, during the twentieth century. *Norges Geologiske Undersøkelse Bulletin* **431**: 33–47.

Winkler S, Elvehøy H, Nesje A. 2009. Glacier fluctuations of Jostedalsbreen, western Norway, during the past 20 years: the sensitive response of maritime mountain glaciers. *The Holocene* **19**: 395–414.