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Assessing the catastrophic break-up of Briksdalsbreen, Norway, associated with rapid climate change

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Abstract: Recent research has raised concerns about the potential influence of rapid climate change on the stability of major ice sheets. The behaviour of glaciers is determined largely by the processes and conditions operating at their base. Technological advances have allowed these factors to be examined and their contribution to ice flow constrained. This study investigated the rapid disintegration of an aquatic based Norwegian glacier, through the study of boreholes, video, ground-penetrating radar, differential global positioning system, bathymetry and Glacsweb wireless probes. Briksdalsbreen retreated dramatically between 2000 and 2007, with *c*. 56×10^5 m^3 of ice lost from the glacier tongue, equivalent to a rate of 70 m a^{-1}. This was due to the combined effect of higher summer temperatures, decreased precipitation (resulting from a negative phase of the North Atlantic Oscillation) and increased fracturing of the glacier tongue. The enlargement of a proglacial lake played a key role in Brikdalsbreen's rapid retreat, allowing calving events and promoting crevassing and fluctuating water contents at the glacier margin. We suggest that hydro-fracturing was the dominant mechanism responsible for generating more crevasses each year, which facilitated the development of an efficient englacial drainage system. This fed increasing quantities of water to the bed, where it was stored in subglacial cavities and transferred through a distributed ('slow') drainage system. However, despite this increase in subglacial water content, ice velocities remained constant during the break-up. Comparisons are made between the processes observed at Briksdalsbreen and those associated with the acceleration and rapid retreat of Greenland's tidewater glaciers.

The response of glaciers to climate change is complex, and numerical models have failed to predict the rapid ice loss observed (Alley *et al.* 2005; Solomon *et al.* 2007; Vaughan & Arthern 2007). This presents a serious limitation to the prediction of global sea-level changes. Uncertainty within models of ice sheet dynamics is largely attributed to a lack of understanding of internal glacier dynamics that complicate the relationship between climate and key glaciological variables (Howat *et al.* 2007; Nick *et al.* 2009). Mechanisms for rapid break-up of calving glaciers have been attributed to increased air temperatures and a corresponding rise in glacier surface melt that results in elevated englacial and subglacial melt water inputs. In turn, this may result in enhanced basal lubrication (Zwally *et al.* 2002), or hydro-fracturing of water-filled crevasses (Sohn *et al.* 1998) accompanied by a release of back-stresses (Thomas 2004), which may lead to faster flow, thinning and rapid retreat.

Southern Norway has one of the best records of glacier limits since the 'Little Ice Age' maximum (AD 1748) (Grove 1988; Bogen *et al.* 1989; Bickerton & Matthews 1993), from dated moraine studies (Andersen & Sollid 1971; Nesje *et al.* 1991; Matthews 2005) and marginal monitoring from the beginning of the twentieth century (Rekstad 1904; Kjøllmoen 2007). During the 1990s the maritime glaciers of the Jostedalsbreen ice cap in southern Norway advanced rapidly, but since 2000 have been undergoing dramatic retreat. Numerous researchers have argued that the recent readvance was due to increased winter precipitation and snow accumulation (Liestøl 1967; Hurrell 1995; Winkler *et al.* 1997; Nesje *et al.* 2000). Nesje & Dahl (2003) and Chinn *et al.* (2005) have suggested that this is caused by a positive phase of the North Atlantic Oscillation (NAO). Consequently, the subsequent dramatic retreat has been attributed to a combination of decreased winter precipitation (a negative phase of the NAO) and increased summer temperatures (Laumann & Nesje 2009; Winkler *et al.* 2009).

Our study was based at Briksdalsbreen, one of the outlet glaciers (with an aquatic margin) of Jostedalsbreen, which advanced *c*. 400 m (1987–1996), before retreating over 400 m (1996–2007). The rapid retreat of this glacier can be compared with the current dramatic break-up of aquatic Greenland outlet glaciers (Joughin *et al.* 2004; Krabill *et al.* 2004; Howat *et al.* 2005; Nick *et al.* 2009) as both are associated with hard rock beds and ice that is channelled through confined valleys.

The work carried out at Briksdalsbreen was centred around the Glacsweb System (Martinez *et al.* 2004). This comprised the development and deployment of a series of autonomous multi-sensor wireless probes, established within the first glacier-based environmental sensor network (Hart & Martinez 2006; Hart *et al.* 2006). These instruments were designed to monitor the physical properties of, and processes occurring in, both englacial and subglacial environments. A number of additional techniques were used in support of this system. For example, the use of ground-penetrating radar (GPR) provided a second measure of ice depth and allowed us to image both the internal ice structure and subglacial topography across the study grid. These additional techniques allowed us to examine the nature of the glacier as a whole and provided both a multi-sensor and multi-instrument approach to the investigation.

Previous papers have focused on detailed analysis of the probe sensor readings and their relevance to seasonal or short-term glacier activity. Here we present an overview of Briksdalsbreen's frontal variations and its recent catastrophic break-up. We draw upon a range of techniques and data sources to assess the glacier's limits from 1900 and investigate changes in its internal structure since 2003, during the height of the glacier's retreat over the course of the Glacsweb study period (2003–2006). GPR, borehole video, and Glacsweb wireless subglacial probes were used alongside differential global positioning system (dGPS) surveying and sedimentological techniques, to quantify the amount of retreat experienced by the glacier and investigate englacial and subglacial changes associated with the rapid retreat.

Briksdalsbreen fluctuations

Briksdalsbreen is an outlet glacier of the Jostedalsbreen ice cap, the largest in mainland Europe (Fig. 1). Since the Little Ice Age maximum, Briksdalsbreen has shown net retreat up a steep-sided valley, but experienced small readvances during 1910 and 1925, followed by rapid retreat after 1940 (Fig. 2). The present-day proglacial lake (Briksdalsvatnet) first formed in 1939 and reached its maximum size in the early 1950s (Liestøl 1967). The glacier subsequently readvanced from 1955 to 1996, but was relatively stable between 1996 and 2000. It has most recently experienced renewed retreat, at a rate of *c.* 70 m a^{-1} (Kjøllmoen 2007). In November 2006, a final 100 m section of ice collapsed into the lake, resulting in the glacier margin resting on a rock step at the base of the ice fall and above the lake. Since then, the retreat has slowed to 21 m a^{-1} (Kjøllmoen 2007; L. Andreassen, pers. comm. 2009).

The glacier limits between 1996 and 2000 can be clearly seen in the landscape, marked by a sequence of annual push moraines at the margin and an erosional bedrock trimline on the valley sides (Figs 1b, c, e and 2b). Between 2001 and 2005 a subglacial surface, comprising lineations and flutes, was exposed (Winkler & Nesje 1999; Hart 2006; Rose & Hart 2008).

The surface of the glacier was debris-free (Fig. 1b) and marked by crevasses, whose frequency and magnitude increased each year (2003–2006). In addition, during the 2006 field season, the glacier toe was buoyant and, unlike previous years, frequent calving events were observed.

Methods

The locations of the glacier margin and moraines were mapped each year with a Topcon dGPS system using a kinematic survey. Lake bathymetry was surveyed with a 0.25 Hz echosounder mounted on an oar-powered inflatable boat, along an approximate 10 m × 15 m grid mapped using dGPS. Strong currents in the lake and the presence of icebergs led to unavoidable irregularities in the grid shape and prevented sampling close to the ice margin.

Sites on the glacier were chosen for detailed subglacial observations. Site A, used in 2003, became unsafe and so the study area was moved to Site B from 2004 (Fig. 1b). Table 1 highlights that Site A in 2003 and Site B in 2005 were at a similar distance from the margin. Site B in 2004 was furthest from the margin, and Site B in 2006 the closest. At each site a series of boreholes were drilled with a Kärcher HDS1000DE jet wash system (Fig. 3). Once the boreholes were made, the internal structure and bed of the glacier were examined using a custom-made CCD video camera that used IR (900 nm) illumination in 2004 and a white LED illumination colour camera in 2005. Glacsweb wireless probes (16 cm long, axial ratio 2.9:1; Fig. 1d) were also installed in some of the boreholes. Probe micro-sensors measured water pressure, probe deformation, conductivity, tilt and probe temperature, although only the water pressure results are discussed below. Data were collected six times a day, every 4 h, and then transferred daily via radio communications to a base station located at the glacier surface. The base station relayed this information once a day to a reference station 2.5 km away, where it was uploaded onto a web server. The base station was also equipped with a weather station (measuring temperature, wind speed and direction, incoming solar radiation and precipitation) and dGPS capabilities (Hart *et al.* 2006).

The 3D aspects of the glacier (depth) were determined annually from the surface dGPS heights, measured borehole depths and GPR surveys. The last was also used to determine the nature of the bed and water content of the ice. The GPR survey was undertaken using a Sensors and Software Pulse Ekko 100 with a 1000 V transmitter system. Each year a common offset survey was performed using a 50 MHz antenna, with a 2 m antenna spacing and 0.5 m sampling interval along a grid (Fig. 3). A common midpoint survey (CMP) was also performed using the 50 MHz antenna. Details of the GPR analysis are given in the Appendix.

To quantify changes in the volume of the glacier tongue during the study period (2003–2007), we needed to reconstruct the glacier shape (surface profile and ice thickness) each year. To do this we carried out a topographical survey of the glacier and foreland, and a bathymetric survey of the lake (Fig. 4). This was combined with GPR analysis and borehole ground truthing to determine both glacier thickness and till depth. To reconstruct englacial and subglacial hydrological processes, we combined borehole analyses, video data, GPR and wireless probe data. Finally, repeat dGPS measurements of control points were used to determine spatial and temporal variations in surface velocities.

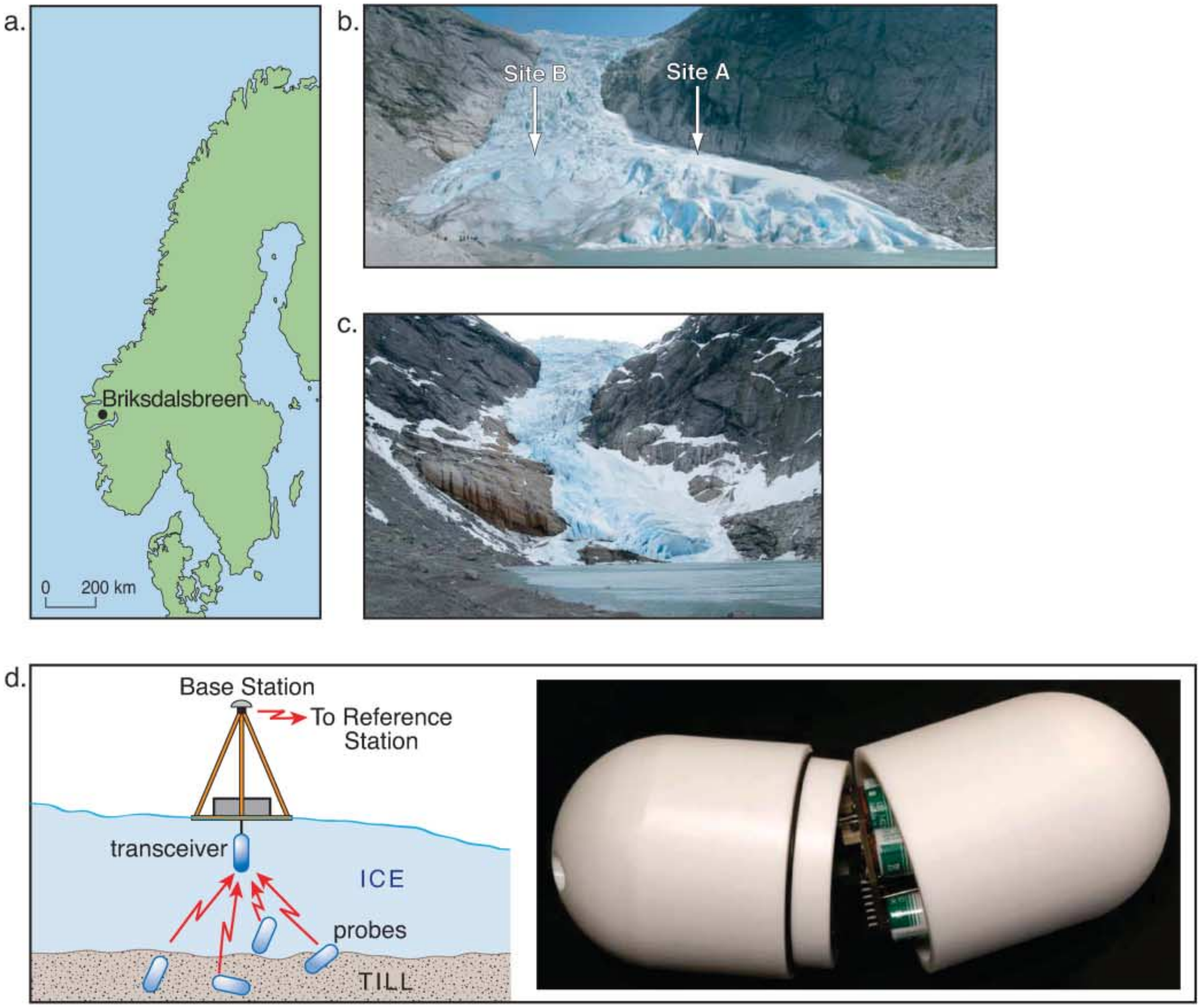
Results

Foreland geomorphology and lake bathymetry

The results of the moraine and bathymetric surveys are shown in Figure 4a and the latter is compared with previous bathymetric surveys undertaken in 1982 by Duck & McManus (1985) (Fig. 4b). The surveys were used to draw a series of cross-sections and a long section of the lake (Fig. 4c). The lake reached a maximum measured depth of 21 m in 2006, but would have been slightly deeper at the ice front, where it was too dangerous to take further readings. In cross-section, the lake displays an asymmetric profile, with a much steeper slope on the southern flank. In addition, there is a ridge *c.* 1–2 m high located in line with the 2004 margin. This is also seen on the long section (Fig. 4c, 240 m along the profile) and represents a subaqueous moraine. On land, a moraine between 1 and 7 m high was formed each year from 2000 to 2006 (Fig. 4a).

Borehole analysis and video data

The analysis of the video footage obtained from the borehole camera indicated significant englacial and subglacial hydrological changes during the period of observation. The average depth of water in boreholes connected to the ice–bed interface de-



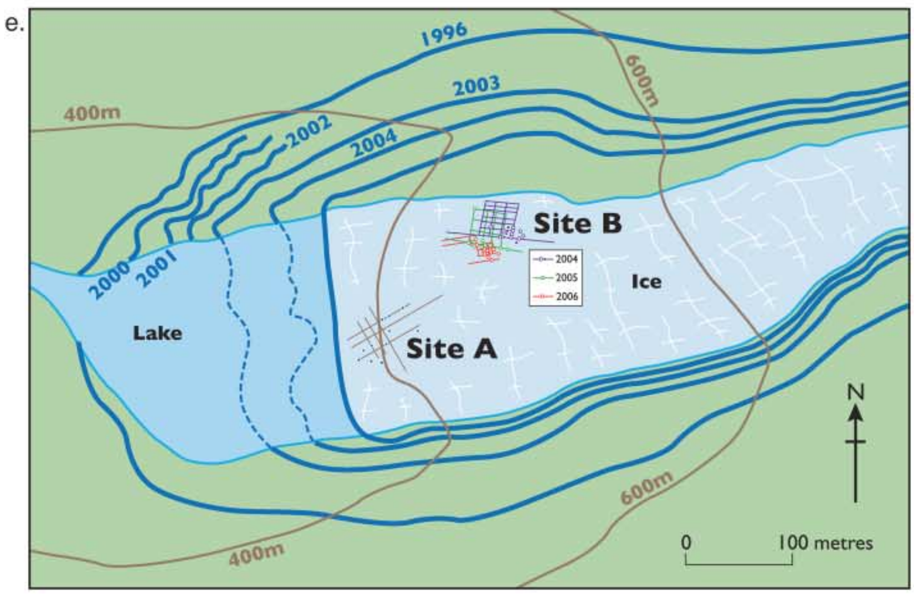


Fig. 1. (**a**) Location in Norway; (**b**) photograph of Briksdalsbreen 2003, with location of study sites shown (Site A, 2003; Site B, 2004–2006); (**c**) photograph of Briksdalsbreen in 2007, resting on a bedrock step at the base of the ice fall; (**d**) Glacsweb probe and sensor network established on the glacier; (**e**) map of Briksdalsbreen with study sites shown (ice limit 2005).

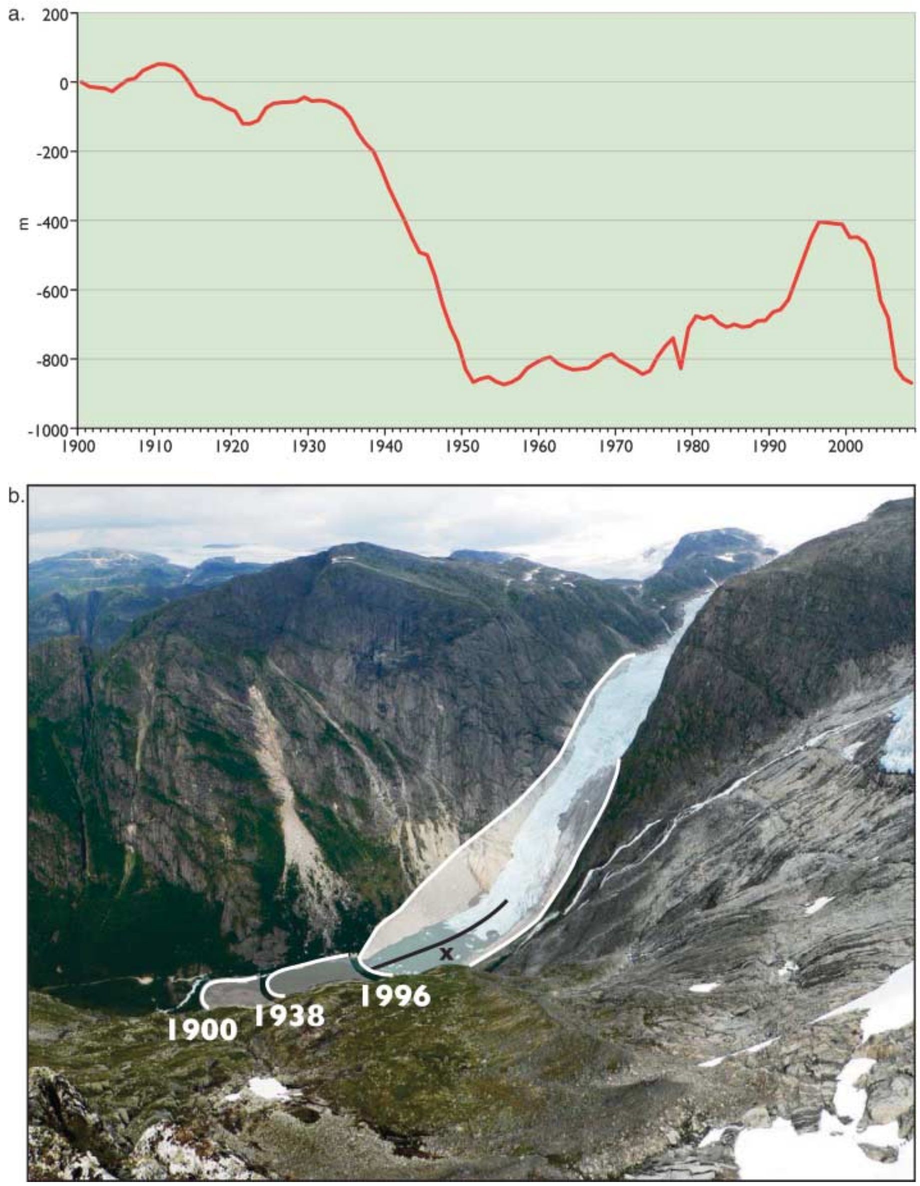


Fig. 2. (**a**) Changes in ice limits since 1900 in metres (data Kjøllmoen 2007); (**b**) photograph (taken 2006) annotated to show recent limits and the centre line of the aquatic based glacier tongue (1996–2006) (marked by an x).

Table 1. *Summary of the data 2003–2006*

Year	Distance from ice front or side (m)	Mean ice depth in study area (m)	Common offset survey mean ice radar-wave velocity (m ns^{-1}) with % error	CMP mean radar velocity (m ns^{-1}) and thickness of upper surface layer (m)	CMP mean radar velocity of main glacier (m ns^{-1})	% boreholes with englacial drainage	% boreholes that drained when borehole reached bed	% area with water at bed	% area with 'sticky spots'
2003	150/150	69	0.175 ± 6%	0.126, 6.62	0.173	63	36	27	26
2004	200/130	68	0.169 ± 3.5%	0.124, 11.22	0.165	20	33	6	32
2005	140/120	51	0.181 ± 9%	0.128, 11.34	0.178	58	58	26	33
2006	70/100	30	0.159 (±7%) to 0.135 (±2%)	0.136, 9.72	0.162	100	58	19	12

creased each year from 63% of ice thickness (43 m) in 2003 to 35% of ice thickness (11 m) in 2006. At the same time, the depth of water became less variable, with the standard error (standard deviation as a percentage of the mean) changing from 78% in 2003 to 9% in 2006 (Tables 1 and 2). In addition, the number of boreholes that drained when the drill reached the bed increased.

Video evidence at Site A in 2003 revealed a wide (30 m),

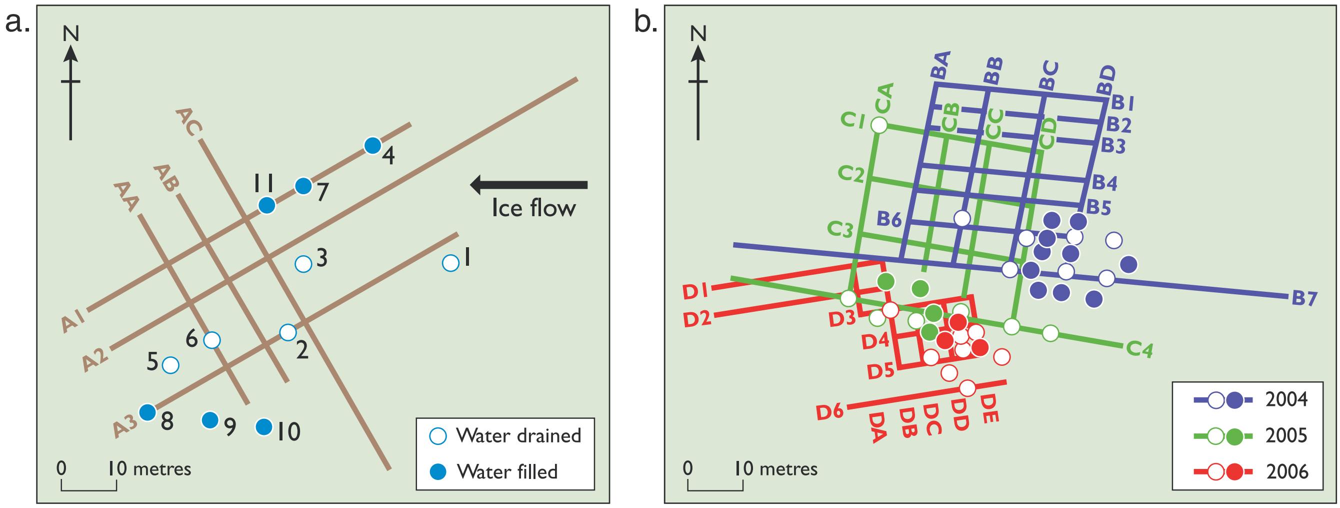


Fig. 3. (**a**) GPR grids 2003 (Site A); (**b**) GPR grids 2004, 2005 and 2006 (Site B) with boreholes shown. Open circles, water-drained holes; filled circles, water-filled boreholes. Ice flow from right to left; location on the glacier is shown in Figure 1e.

shallow (1.6 m), low-pressure, and fast-flowing (>0.1 m s^{-1}) subglacial channel. At Site B, in 2004, a smaller 5 m × 1.5 m, low-pressure, fast-flowing (>0.1 m s^{-1}) subglacial channel was observed. However, between 2005 and 2006, flowing water was not observed in the video data. Instead, the percentage of boreholes observed to have englacial crevasses or voids increased between years and in 2006 water-filled subglacial cavities were observed.

Glacsweb probe water pressure results

Water pressure, measured in metres water equivalent (mW.E.), was recorded continuously throughout the year by the Glacsweb probes in both englacial (probe 4 (2004–2005)) and subglacial (probes 8 (2004–2005), 10 and 12 (2005–2006)) environments (Fig. 5). This was generally low during the autumn and winter, before rising in the spring and summer. The probes that rested within subglacial till (probes 8, 10 and 12) showed a two-stage water pressure rise in the spring (Rose *et al.* 2009). The probe within the ice (probe 4) showed a peak in water pressure at the same time as the beginning of the second water rise in the till. When comparing the subglacial data from 2004–2005 (probe 8) with those from 2005–2006 (probes 12 and 10), it can be seen that water pressure was much lower in the second year (Fig. 5).

GPR results

The annual GPR common offset survey data were combined with known borehole depths to locate the glacier bed (Fig. 6) and calculate bulk radar velocity (Table 1; Appendix). In 2004, the bulk radar velocity was just over 0.16 m ns^{-1}, which is the theoretical value for temperate ice (Davis & Annan 1989). In 2003 and 2005 values were much higher, whereas in 2006 the velocity varied over the field season, ranging between 0.135 and 0.159 m ns^{-1}. The common offset surveys also revealed the presence of additional reflections, in places, beneath the glacier bed (Fig. 6c).

The changes in radar velocity in ice with depth can be calculated using semblance analysis of the CMP velocity survey data. The results identify two sections of the glacier with different radar velocities (Table 1). Each year, there was an upper surface section, *c.* 10 m deep, with a lower mean radar velocity of 0.128 m ns^{-1}. In contrast, the main body of the glacier recorded a higher value, close to that determined from the common offset survey (mean 0.169 m ns^{-1}). This multi-layer characteristic of glacier ice has been widely reported in the literature. For example, at Falljökull, Iceland, Murray *et al.* (2000) found an upper, well-drained, dry layer above the piezometric surface. In contrast, Macheret & Glazovsky (2000) found in the temperate ice at Fridtjovbreen and Hansbreen, Svalbard, that the upper parts of the glacier had the highest water contents. They suggested that this was due to high surface melt rates and 'macro inclusions', such as water-filled cavities and veins.

Next, we used both CMP and bulk radar velocity data to estimate the water content (W) of the glacier, using Looyenga's (1965) formula for dielectric mixtures of air and water inclusions (Macheret *et al.* 1993; Frolov & Macheret 1999) (Table 3; Appendix). The CMP radar velocity data can be used to calculate the water content of the different layers within the glacier, whereas the bulk radar velocity data provide an overall value for the glacier body (Table 3). In terms of the layers, if it is assumed that in the upper part of the glacier (where ice radar velocities are lower) any voids are filled with water, then the two-component model can be used to calculate water content. This results in a water content of *c.* 10% in the upper 10 m of the ice. For the lower part of the glacier, where radar velocities are higher (>0.166 m ns^{-1}), it is assumed that the water content is zero and thus the air component can be calculated. Where velocities are less than 0.166 m ns^{-1}, the air content is assumed to be zero and the water content is calculated. In this lower layer, this results in water contents between 0 and 1.2% and air contents of 0–12%. The values produced using the bulk radar velocities reflect an average of the different layers, and in 2006, the bulk radar velocities varied on different days, in accordance with changes in water content (Table 3).

The scale of results is similar to that reported by Pettersson *et al.* (2004), who showed that the range of water contents determined from GPR using Looyenga's model was 0–9.1%. Our results have revealed a change in water contents between years, with a decrease in water content of the upper layer in 2006, but an increase in the overall water content of the glacier.

We also use GPR data to investigate the nature of the bed (i.e.

differentiate between water, till and bedrock) by calculating basal reflectivity, R, from the strength of the basal reflection power (Gades *et al.* 2000). Pattyn *et al.* (2003) used the three-layer reflectivity model of Born & Wolf (1999) to show the relationship between porosity, layer thickness and subglacial material. These values were similar to those found in other studies, including that by Gades *et al.* (2000), who showed that the saturated till beneath Siple Ice stream had an R value of 0.16–0.32.

The Briksdalsbreen data were filtered using a five-point (2.5 m) moving average to remove the spatial variation caused by single data points. Data were then divided into the three classes of R (water, till and bedrock, as in Table 4), and converted into a percentage of the bed area surveyed (Table 1). A single dominant water body was observed in 2003 and 2004, whereas smaller irregular-shaped water bodies were noted in subsequent years.

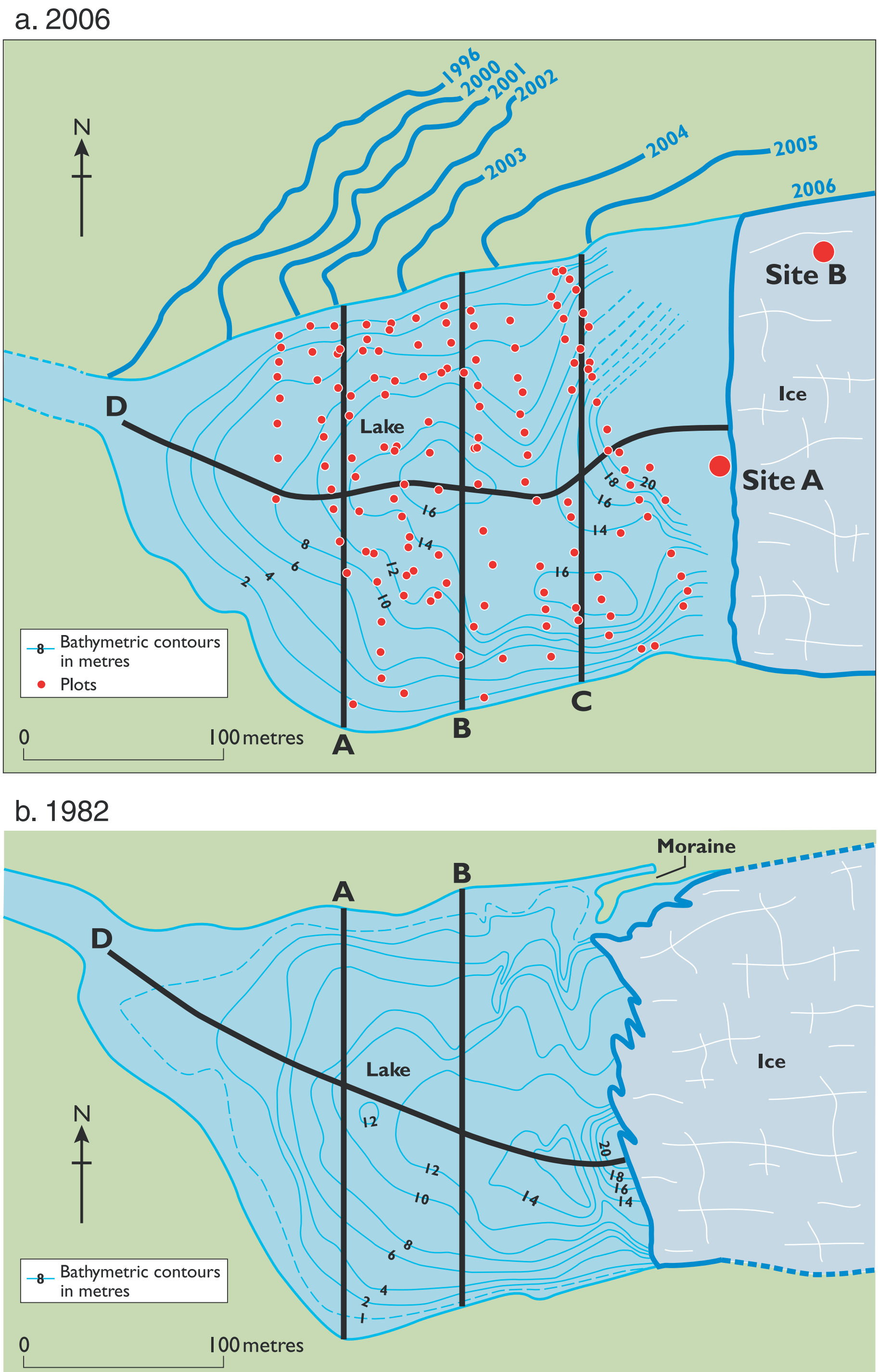


Fig. 4. (**a**) Map of the moraine and bathymetric survey (1996–2006); (**b**) bathymetric survey 1982 (Duck & McManus 1985); (**c**) cross-sections (from north (0 m) to south (230 m)) and long sections (approximately west (0 m) to east (340 m) from both years.

Ice velocity changes

Ice velocity at Site B was measured from 2004 to 2006 using dGPS. The velocity along the central line was measured by Elvehøy (2001) from 1996 to 2000, while the glacier was relatively stable. Figure 7 shows the winter to spring data plotted as a relative distance from the glacier margin. The 2004–2006 data have been corrected for marginal effects, based on Nye

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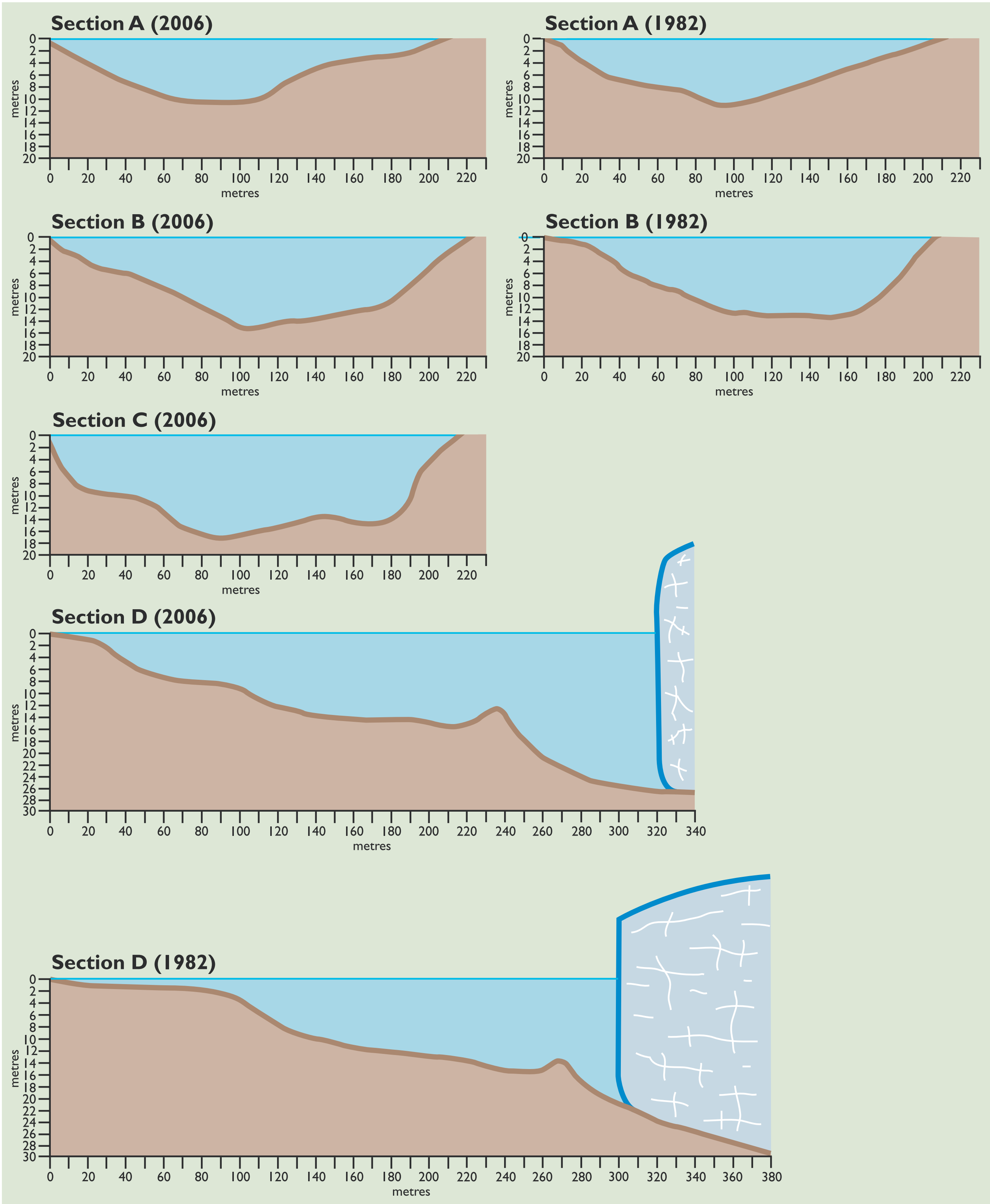

Table 2. *Annual variations in borehole water depth measurements*

2003		2004		2005		2006	
Borehole	Water depth above bed (m)	Borehole	Water depth above bed (m)	Borehole	Water depth above bed (m)	Borehole	Water depth above bed (m)
03/1	0	04/1	36–74	05/3	22–50	06/1	13
03/2	0	04/3	47–67	05/4	35	06/4	13
03/3	0	04/4	7–16	05/5	1–9	06/5	10
03/4	86	04/5	0–2	05/6	0–50	06/6	7
03/5	0	04/7	0–74	05/8	13–14	06/7	11
03/6	65	04/8	0–52	05/9	39	06/8	11
03/7	82	04/9	11–65	05/10	12	06/9	8
03/8	56	04/10	0–24	05/11	44	06/10	12
03/9	57–62	04/12	20–41	05/12	0	06/11	11
03/10	63	04/13	0–37			06/12	10
03/11	56–74	04/14	0–1				
		04/15	2–12				
		04/16	21–67				
Mean	43.18	Mean	25.1	Mean	23.2	Mean	10.6
SD	35.46	SD	19.4	SD	16.1	SD	1.9
% error	82.1	% error	77.3	% error	69.4	% error	18.9

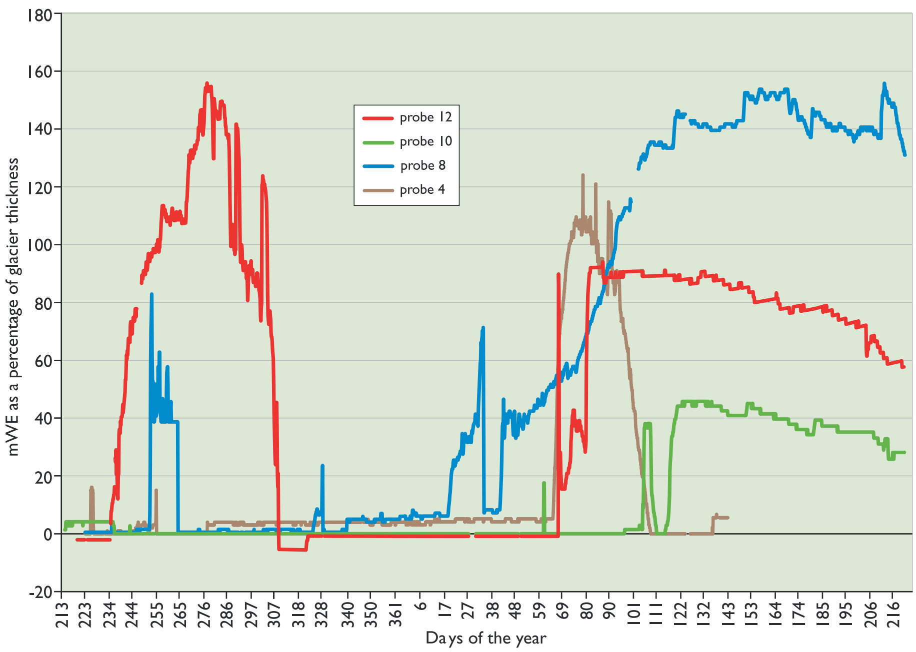

Fig. 5. Water pressure data obtained from the Glacsweb wireless probes: 2004–2005 englacial probe 4 and subglacial probe 8; 2005–2006, subglacial probes 10 and 12.

(1965). The 1997 data were slightly higher, as they contain some summer values, but otherwise velocities remain constant over the study period.

Discussion

We initially discuss the reconstruction of a 3D model of the glacierorder to calculate ice volume loss over the retreat period. The nature of sedimentation into the lake during the advance and retreat stages is reviewed. Then we examine the englacial and subglacial hydrological processes, and combine these results to characterize the anatomy of the glacier during rapid retreat.

Three-dimensional glacier model and ice volume reconstruction

Using the morphology (GPS), bathymetry, borehole and GPR data, it was possible to reconstruct a 3D model of the glacier tongue. This is illustrated by the long profile and three cross-sections shown in Figure 8. It can be seen on the long profile that there is a large subglacial bedrock obstacle at the beginning of the profile. The 'additional reflections' beneath the glacier bed shown by the common offset survey (Fig. 6c) were interpreted to represent this feature. During ice retreat, the presence of this bedrock obstacle was revealed (Fig. 1c), confirming this prior evaluation and providing evidence of the till–bedrock interface

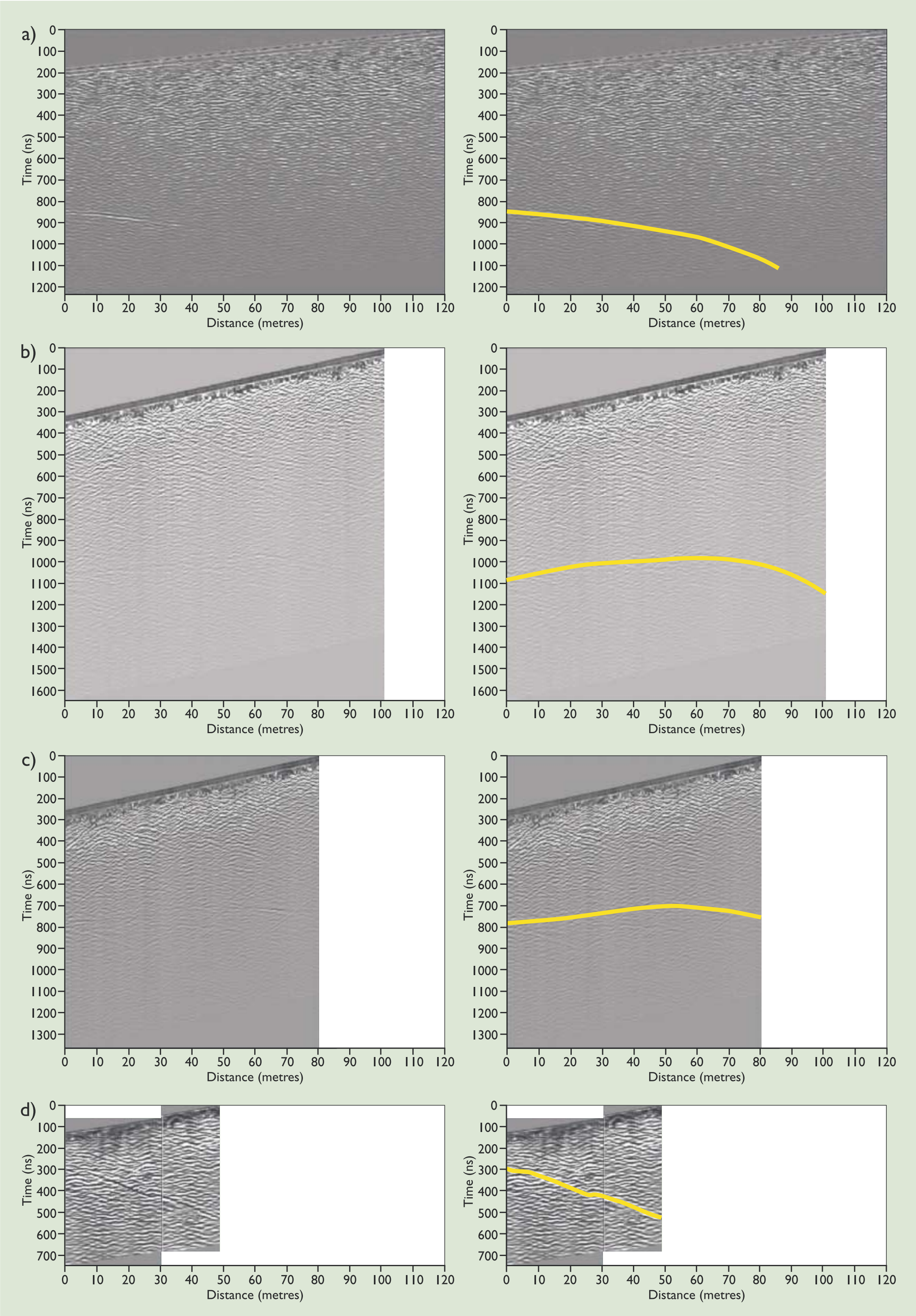

Fig. 6. GPR images with glacier bed shown in the right-hand image: (**a**) Site A, 2003, survey line A2; (**b**) Site B, 2004, survey line B7; (**c**) Site B, 2005, survey line C4; (**d**) Site B, 2006, survey line D6. Survey lines drawn west (left) (0 m) to east (right); ice flow direction right to left. Location of survey lines is shown in Figure 3. In (**c**) it is possible to see an additional reflection *c*. 100 ns two-way travel time beneath the glacier bed from 20 to 80 m along the profile, which is interpreted to be the bedrock–till interface.

Table 3. *Water content (%) calculated from GPR analysis*

Year	Upper layer (from CMP)		Lower layer (from CMP)		Whole glacier (from common offset survey and measured borehole depths)	
	Water content	Air content	Water content	Air content	Water content	Air content
2003	10.5	0	0	6	0	9
2004	11.2	0	0.6	0	0.6	2.5
2005	9.9	0	0	12	0	15.5
2006	7.5	0	1.2	0	7.8–1.4	0–6.4

Table 4. *Basal reflection values (from Pattyn et al. 2003)*

R value	Equivalent porosity	Subglacial material
>0.5	>0.6	Water body
0.18–0.4	0.2–0.3	Saturated deforming till
<0.18	<0.2	Frozen till, rigid till or bedrock

within the GPR dataset. Consequently, assuming a till radar velocity of 0.079 m ns^{-1} (Murray *et al.* 1997), we were able to calculate the depth of till across the glacier bed, which ranged from 0 to 12 m (mean 6.7 m).

On the lee side of the obstacle, the glacier base and till depth remained consistent between 2004 and 2006 when the area was surveyed. On the cross-sections the till thickness calculated is also consistent with observations. From this we can reconstruct ice volume loss since 1996 (Table 5). Although ice loss was slow at first, by 2004 over 50% of the ice volume of the glacier tongue was lost. The 3D glacier and till model was also used to calculate the glacial sedimentation rate, assuming the till was deposited since 1996. This was 2.34×10^4 m^3 a^{-1} (0.3 m over a unit square metre).

Comparison of lake bathymetry before and after the recent advance

Duck & McManus (1985) also carried out a bathymetric survey of Briksdalsbreen's proglacial lake in 1979 and 1982, during which time the ice had advanced 30 m. The 1982 margin was in a similar location to the 2003 margin. By comparing the 1982 and 2006 bathymetric surveys we can see how the lake has changed in response to glacier advance and retreat (Fig. 4). Table 6 shows a comparison of surface area and lake volume from the three surveys.

Many aspects of the lake's morphology remained similar between the advance and retreat stages. The maximum depth was 21 m at the glacier margin and the lake width was 200 m. Both cross-sectional profiles had a steeper southern flank, and long sections show an ice-marginal sub-aqueous moraine. The later survey also demonstrates that the water was shallower in 2006. The fact that there was only one moraine in both surveys implies that the others must have been either covered by sediment or eroded by a combination of waves and iceberg scour.

Duck & McManus (1985) argued that between 1979 and 1982 there was a 10.2% reduction in water volume as a result of glacier advance. They suggested that the additional ice present in 1982 had displaced 2×10^4 m^3 of water, and the remaining 1.2×10^4 m^3 of volume lost was attributed to sediment infill (4×10^3 m^3 a^{-1} or 0.24 m per unit square metre). The lake's water volume in 2006 represents 76.7% more than the 1982 level.

However, if the 2006 water volume is calculated over the same area as in 1982, this gives a value of 8.62×10^4 m^3, which represents a 69.4% reduction in comparison with 1982. This suggests a sediment infill of 1.96×10^5 m^3. Assuming that the sediment infill has formed since 1996 (last glacial maximum advance), then this reflects 1.96×10^4 m^3 a^{-1} of deposition into that part of the lake, which is equivalent to 0.44 m per unit

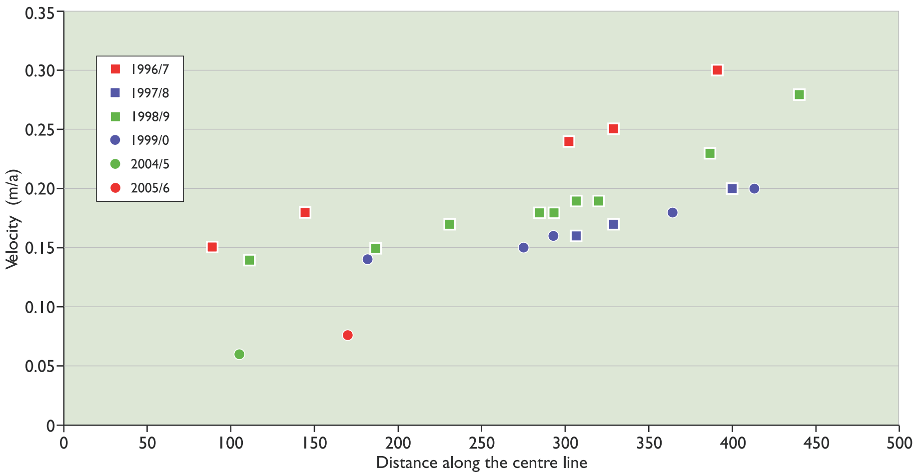

Fig. 7. Winter ice velocities along the centre line (marked in Fig. 2b) from 1996–1997 to 2005–2006.

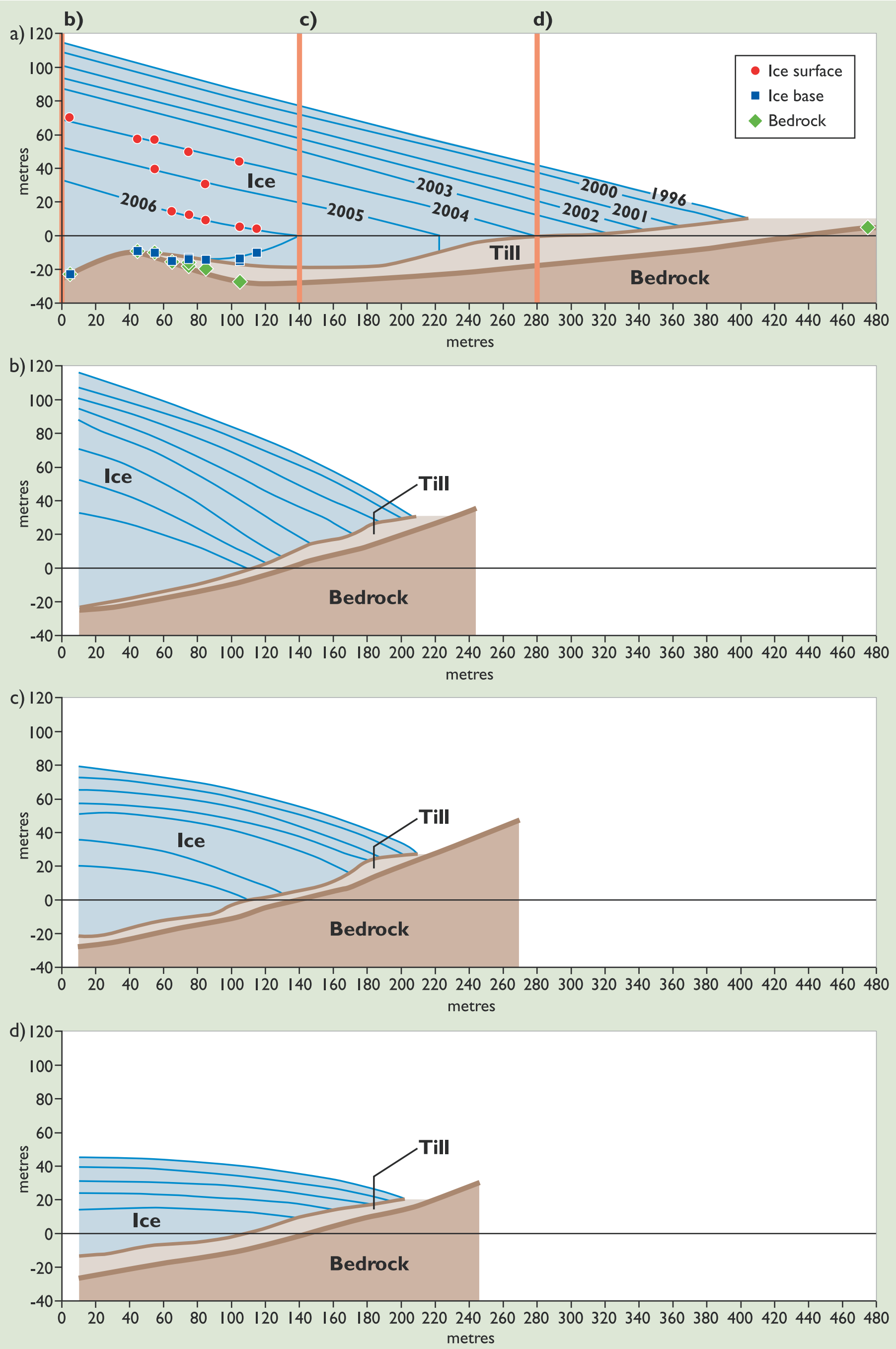

Fig. 8. Reconstructed glacier and till profiles as the glacier retreated, measured using GPS, GPR and bathymetry: (**a**) long section along a line x (Fig. 2b) (ice flow direction left to right) with years shown, and known glacier surface, glacier base and bedrock surface shown; (**b**–**d**) half cross-sections perpendicular to centre line (x in Fig. 2b) at the places shown in (**a**). Ice flow direction into the page; 0 m height represents the 2006 lake level.

Table 5. *Ice volume of the 'tongue' (see Fig. 2b for location)*

Year	Ice volume (m^3)	% loss	Ice volume loss per year (m^3)
1996	5.60×10^6		
2000	5.24×10^6	6.28	3.52×10^5
2001	4.70×10^6	16.01	5.44×10^5
2002	4.11×10^6	26.55	5.90×10^5
2003	3.59×10^6	35.80	5.17×10^5
2004	2.61×10^6	53.42	9.86×10^5
2005	1.89×10^6	66.16	7.13×10^5
2006	3.56×10^5	93.64	1.54×10^6
2007	0	100	3.56×10^5

Table 6. *Details of the lake during the recent advance (1979 and 1982) (Duck & McManus 1985) and retreat (2006) of Briksdalsbreen*

	1979	1982	2006
Surface area (ha)	4.71	4.53	73.47
Lake volume (m^3)	3.14×10^5	2.82×10^5	4.98×10^5
Change in water volume (m^3)	–	-3.2×10^4	$+2.16 \times 10^5$

square metre. These results suggest that the sedimentation rate during the retreat was much greater than during the advance. In addition, the amount of sedimentation into the lake is slightly greater (31%) than that on land.

Englacial changes

Data from an *in situ* englacial Glacsweb probe showed that water was able to efficiently drain through the ice, and water pressure did not rise until the spring event. Each year the number of water-filled surface crevasses and boreholes with evidence for englacial drainage (crevasses and voids) increased. Given that high bulk radar velocity values indicate air pockets within the glacier whereas low radar velocity values indicate high water contents, the decrease in radar velocities between years, and variable velocities in 2006, indicated a rise in the number of potential water storage sites within the glacier.

This was supported by the increase in calculated water contents (formula of Looyenga 1965) in 2006, which suggested that the significant number of englacial voids and crevasses noted between 2003 and 2005 had indeed become water filled.

Jansson *et al.* (2003) and Lingle & Fatland (2003) have argued that large volumes of water can be stored within glaciers, but its location is unknown. Much of this water is stored in a vein system of connected, centimetre- to decimetre-sized, voids (Murray *et al.* 2000) or crevasses (Nienow *et al.* 1998; Fountain *et al.* 2005), which can readily divert water from the glacier surface to its bed (Zwally *et al.* 2002; Das *et al.* 2008; Benn *et al.* 2009). Given the general increase in crevasses and voids observed within boreholes between years, we envisage that the englacial drainage system at Briksdalsbreen developed a similar morphology (Murray *et al.* 2000; Fountain *et al.* 2005).

However, although the connectivity of the englacial drainage system may have developed, the overall increase in bulk glacier water content recorded suggests that numerous sites of water storage were also present. The variable velocities recorded in 2006 may in fact reflect a pattern of water transfer, temporary storage, and then further transfer, as drainage routes throughout the glacier body proceeded to open and close in response to water inputs. Indeed, Fountain *et al.* (2005) observed just such a pattern of fracture opening and closing within the englacial drainage system at Storglaciären, Sweden; and Benn *et al.* (2009) and Gulley *et al.* (2009) noted that with the onset of the summer melt season a mechanism of hydro-fracture propagated the development of englacial pathways from surface to bed on parts of the Greenland ice sheet. Alternatively, the increase in water content may represent ponding along low-gradient englacial channels (Stuart *et al.* 2003).

Furthermore, semblance analysis of the CMP data indicated that there were two main layers within Briksdalsbreen. We suggest that the upper, low-velocity, area of the glacier comprised shallow water-filled surface crevasses, voids and a system of veins that were water saturated from surface melt. These features were generally of limited spatial extent and formed a dominantly closed system, where water could be stored. In contrast, beneath this surface layer, the main body of the glacier, which showed higher velocities, supported a more efficient drainage network. It comprised a series of voids, veins and crevasses that were interconnected, allowing water to easily flow through the glacier. This englacial drainage may have been enhanced through a process of hydro-fracture (Boon & Sharp 2003; Benn *et al.* 2009), as the number of initially water-filled crevasses increased each year.

Subglacial changes

Each year there was more water at the glacier bed. This was manifest in the percentage of boreholes that drained when the drill reached the bed, indicating that they had intersected some form of subglacial drainage or cavity. Borehole water levels also became stable, reflecting a consistent hydraulic head supplied from the subglacial environment. The increase in water at the glacier bed may also suggest enhanced transfer of water from the glacier surface. This is supported by the more numerous observations of englacial crevassing. The increasing buoyancy of the glacier tongue and proximity to the lake in 2006 may also increase the amount of water accessing the glacier bed from the lake and thus increase subglacial water content.

The GPR basal reflectivity (R) data also agreed with the borehole results, as the spatial coverage of areas of high reflectivity, representing water bodies, increased as the glacier retreated. Areas of low reflectivity correspond to low-porosity till or bedrock, which can be conceptualized as zones of high basal drag or 'sticky spots' (Alley 1993; Fischer *et al.* 1999; Kavanaugh & Clark 2001; Mair *et al.* 2003). In addition, data from *in situ* Glacsweb probes showed that water pressures were very high during the summer, but decreased as a percentage of overburden pressure each year. This may indicate increased drainage through the till (e.g. via Darcian or pipe flow) (Boulton & Hindmarsh 1987).

In 2003, basal reflectivity indicated the presence of a large discrete subglacial channel (Tables 1 and 4), whose morphology was later exposed as the glacier retreated. In 2004, R values indicated that a smaller subglacial channel existed. Video observations provided evidence for a small amount of moving water flow in both of these wide, shallow, low-pressure channels. These are interpreted as discrete R-channels, similar to those described in previous studies (Seaberg *et al.* 1988; Hooke *et al.* 1990; Hock & Hooke 1993; Cutler 1998).

However, in 2005 and 2006, there was a different style of subglacial water body. This was represented by the irregular-shaped bodies shown in basal reflectivity values (Tables 1 and 4), which covered a relatively large proportion of the glacier bed. These bodies may represent drainage at the ice–bed interface in the form of 'microcavities' (Kamb 1991), a braided canal network (Walder & Fowler 1994) or a linked cavity system (Lliboutry 1976; Kamb 1987; Nienow *et al.* 1998; Willis *et al.* 2009). We suggest that these two styles of subglacial water bodies reflect both 'fast' (discrete *R*-channels in 2003 and 2004) and 'slow' (irregular shapes in 2005 and 2006) connected water flow at the ice–bed interface (Fountain & Walder 1998).

Unusually, despite an increase in water at the glacier bed, an increase in glacier velocity was not observed (Fig. 7). However, the enhanced efficiency of the englacial drainage system in response to elevated melt water inputs or greater transfer of melt water from the surface may in fact have been the cause for such restrained annual velocities (van de Wal *et al.* 2008).

Synthesis

The retreat of Briksdalsbreen was very rapid. Approximately 70 m a^{-1} or 56×10^5 m^3 of ice was lost from the glacier tongue from 2000 to 2007. This result agrees with that of Nick *et al.* (2009) and Winkler *et al.* (2009), who argued that the break-up of the tongues of outlet glaciers is immediate and not necessarily related to mass-balance changes.

Numerous researchers have argued that once a glacier has an aquatic margin, this will transform its dynamic behaviour (Chinn 1996; Benn *et al.* 2007). At Briksdalsbreen, once the proglacial lake formed in 1940, the retreat rates were very rapid (Fig. 2). At an aquatic margin, buoyancy effects and variations in water levels, velocity and stress, lead to increased fracturing in this zone and thus enhanced calving (Holdsworth 1973; Theakstone 1989; Motyka *et al.* 2003; Joughin *et al.* 2004). Additional water within the glacier can also allow the propagation of crevasses to greater depths (Nye 1957; Benn *et al.* 2007, 2009). This can then lead to increased flow velocities, calving rates, and consequently glacier break-up.

We suggest that there were three zones associated with the Briksdalsbreen aquatic margin: marginal, intermediate and distal (Fig. 9). In the extreme marginal zone (Site B, 2006), there were many crevasses and fluctuating water contents within the glacier. There was also a high percentage of interconnected areas, as well as subglacial cavities, at the ice–sediment interface. In the intermediate zone (Site B, 2005), both crevasses and voids store water. In the subglacial environment there is a less well-connected drainage system and water pressures in the till are higher. Furthest from the glacier margin (Site A, 2003; Site B, 2004), the aquatic effects are not significant. Englacial storage is dominated by voids and subglacial conditions by 'fast' water flow.

Implications and comparison with Greenland

Recent research in Greenland and the Arctic has provided new evidence of englacial drainage in cold (Catania *et al.* 2008; Das *et al.* 2008; Catania & Neumann 2010; Parizek *et al.* 2010) and polythermal (Boon & Sharp 2003; Bingham *et al.* 2008) ice masses. Therefore the englacial processes commonly observed at temperate glaciers, such as Briksdalsbreen, may have more relevance to assessing the causes of the enhanced ice flow and retreat of Greenland's outlet glaciers. In fact, we can see that a number of the englacial and subglacial processes discussed from Briksdalsbreen's retreat are similar to several of the patterns observed in Greenland's tidewater outlet glaciers (e.g. Sohn *et al.* 1998; Thomas 2004; Das *et al.* 2008). Specifically, both have shown thinning, increased calving and associated terminus retreat (Nick *et al.* 2009), as well as an increase in water content and water-filled voids (e.g. Benn *et al.* 2009). Similar to the study by Nick *et al.* (2009), this investigation also indicates that enhanced hydro-fracturing and subsequent calving was the dominant mechanism for Briksdalsbreen's retreat. Although Funk & Röthlisberger (1989) have argued that, for a given water depth, calving rates are lower for terrestrial aquatic glaciers than tidewater glaciers, Venteris (1999) showed that the salinity of water is not important to this process.

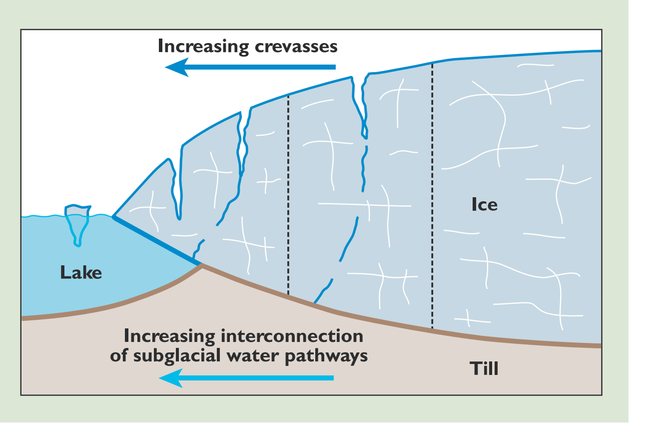

Fig. 9. Schematic diagram illustrating the physical conditions in three glacial zones at Briksdalsbreen (delineated by vertical dotted line): marginal zone (left); intermediate zone (centre); distal zone (right).

In addition, a notable outcome of this study is that Briksdalsbreen displayed stable velocity throughout its retreat. This contradicts the pattern of acceleration associated with rapid retreat displayed by the majority of outlet glaciers in Greenland (Joughin *et al.* 2004; Howat *et al.* 2005; Pritchard *et al.* 2009). This may reflect differences between mechanisms of freshwater and tidewater calving or local climatic influences, for example. However, an exception lies in NW Greenland, where (similar to Briksdalsbreen) glaciers have shown little change in flow, despite observed thinning (Rignot & Kanagaratnam 2006). This pattern highlights the spatial variability in glacier response to increases in discharge. Rignot & Kanagaratnam (2006) suggested that such stable flow velocities may indicate 'that the glaciers were already flowing above balance velocity conditions' when ice front discharge was initially calculated. In contrast, van de Wal *et al.* (2008) suggested that stable annual-scale velocities may be the result of the development of efficient englacial drainage. This system can then successfully and continuously transfer surface water to the glacier bed, preventing sudden large melt water inputs and associated speed-up events (Catania & Neumann 2010). The latter would appear to be applicable to the pattern of drainage development observed at Briksdalsbreen.

Where glacier velocities are increasing, any future loss of ice in Greenland may be more rapid and catastrophic than that recently observed at Briksdalsbreen, and currently predicted for the whole ice sheet. However, some investigations have shown evidence of feedback mechanisms, which act to stabilize glacier activity in response to enhanced subglacial melt water inputs, causing any acceleration in ice velocities to be temporary rather than continuous (Howat *et al.* 2007; Das *et al.* 2008; Nick *et al.* 2009). Similarly, other researchers have suggested that longitudinal coupling of ice flow represents a better explanation for acceleration (Price *et al.* 2008) than melt water inputs (Zwally *et al.* 2002).

At Briksdalsbreen, retreat was halted as the glacier withdrew away from the lake onto a previously hidden bedrock step at the base of the ice fall. Since this time, the glacier has remained relatively stable. In this instance, a localized topographic feature acted as a feedback mechanism for re-stabilization. Similarly, most of the basal troughs through which many of Greenland's tidewater outlet glaciers flow do not extend far inland and it is thought that this will prevent any runaway destabilization (Nick *et al.* 2009). However, Jakobshavn Isbrae's basal topography is an exception to this, where rapid retreat may continue inland, and it is these types of unique variables (e.g. topography, geometry) that make predicting glacier response more complex.

Consequently, it is difficult to assess whether or not all of Greenland's outlet glaciers will be able to quickly re-stabilize after a period of rapid retreat (Howat *et al.* 2007; Nick *et al.* 2009), as indicated at Briksdalsbreen; or if the pervasive dynamic thinning observed across Greenland (Pritchard *et al.* 2009) will ensure a future of continued collapse. In addition, whereas an aquatic (freshwater) margin may initiate specific responses in glacier dynamics (Chinn 1996; Benn *et al.* 2007), a different set of mechanisms may be more applicable to Greenland's terrestrial (and marine) outlet glaciers. Despite such variation, it seems that as dynamic thinning and melt water production continue to increase with rising atmospheric and oceanic temperatures, mechanisms of glacier disintegration, such as those observed at Briksdalsbreen, are likely to be observed with greater frequency.

Conclusion

Using combined topographic, GPR and bathymetric surveys, we were able to reconstruct, in three dimensions, the dramatic retreat of Briksdalsbreen since 1996. In addition, we were able to use GPR, borehole drilling, borehole video and the Glacsweb wireless subglacial probes to 'image' the englacial and subglacial environments associated with this retreat.

Overall, we would argue that the rapid break-up of Briksdalsbreen was due to increased fracturing, which generated crevasses. These promoted an efficient englacial drainage system, which supplied water to the glacier body and, ultimately, the bed. This water was stored in crevasses and voids, and transferred through interconnected subglacial drainage networks at the ice–bed interface and within the till. However, the predicted increases in velocities associated with rapid break-up (owing to increased lubrication) were not observed. Instead, our data suggest that hydro-fracturing of water-filled crevasses, accompanied by a possible release of back-stresses, was the dominant mechanism of glacier collapse.

The investigations discussed here highlight the need to better understand the mechanisms associated with glacier retreat in response to climate change, if we are to improve modelled predictions of future sea-level rise (Nick *et al.* 2009). Our observations at Briksdalsbreen provide evidence of rapid collapse as a result of mechanisms of increased crevassing, and englacial and subglacial water storage and transfer. This is similar to several of the mechanisms of glacier retreat observed in Greenland. Consequently, such processes should be taken into consideration when monitoring the complex behaviour of Greenland's outlet glaciers.

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Appendix: GPR analysis

Comparison of common offset surveys with known depth

The following processes were applied to the common offset surveys using the software package ReflexW: the elimination of low-frequency noise (de-wow filter) and the application of an SEC (spreading and exponential compensation) gain to compensate for signal loss with depth. Analysing the two-way radar travel time (t), we were able to reconstruct the location of the bed (see Fig. 6) and compare this with known borehole depths (d). This allowed the calculation of the bulk radar velocity of ice (v) (and percentage errors), where

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

Barrett *et al.* (2007) have suggested that a typical error on these values is 2%.

Common midpoint velocity survey

Semblance analysis was carried out on the CMP data to calculate the radar velocity at different depths within the glacier from the following relationship (Yilmaz 1987; Eisen *et al.* 2002; Moorman & Michel 2000):

$$v_{\mathrm{rms}} = \sqrt{x^2/(t^2 - t_0^2)} \tag{2}$$

where v_{rms} is the root mean square radar velocity, x is the antenna separation and t_0 is the two-way zero offset time.

Water content

The radar velocity data can be used to estimate the water content (W) of glaciers using Looyenga's (1965) formula for two- and three-component dielectric mixtures of air and water inclusions (Sihvola *et al.* 1985; Macheret & Glazovsky 2000):

$$\varepsilon_{\mathrm{m}}^{a} = \sum f_k \varepsilon_k^{a} \tag{3}$$

where ε_{m} is the permittivity of the mixture, ε_k is the kth component with a volume portion f_k, and $a = 1/3$. The permittivity of the mixture (i.e. temperate ice ε_{s}) is

$$\varepsilon_{\mathrm{s}} = (c/v)^2 \tag{4}$$

where c is the velocity of light and v is the measured radar-wave velocity. Equation (3) can also be expressed as follows:

$$\varepsilon_{\mathrm{s}} = [\varepsilon_{\mathrm{i}}^{1/3}(1 - P) + W\varepsilon_{\mathrm{w}}^{1/3} + P - W]^3 \tag{5}$$

where ε_{i} is the permittivity of solid dry ice (taken as 3.19), ε_{w} is the permittivity of water (taken as 86), W is the water content and P is the total fractional water and air content. If it is assumed that within temperate ice all the cavities are water filled, then equation (5) can be simplified to the two-component model:

$$W = \left(\varepsilon_{\mathrm{s}}^{1/3} - \varepsilon_{\mathrm{i}}^{1/3}\right) / \left(\varepsilon_{\mathrm{w}}^{1/3} - \varepsilon_{\mathrm{i}}^{1/3}\right). \tag{6}$$

Although Endres *et al.* (2009) have suggested that Looyenga's model slightly underestimates water contents, we have used this technique to compare our results with the published literature.

Basal reflection power (BRP)

Numerous researchers (Copland & Sharp 2001; Pattyn *et al.* 2005) have calculated BRP as follows:

$$P \equiv \frac{1}{2(t_2 - t_1 + 1)} \sum_{i=t_1}^{t_2} A_i^2 \tag{7}$$

where P is returned power, A is the sum of the squared amplitudes and $t_1 - t_2$ is the time window. Because increasing ice thickness will affect reflection strength, this has to be compensated for by calculating residual BRP (BRP_r) as follows:

$$BRP_r = \frac{BRP(measured)}{BRP(predicted)} - 1. \tag{8}$$

The theoretical reflectivity R is calculated as a ratio between BRP_{rmax} and BRP_r.

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