Borehole-based characterization of deep crevasses at a Greenlandic outlet glacier

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Key Points:

- Borehole logging by optical televiewer reveals the presence of numerous deep, high-angle crevasse traces at Store Glacier, Greenland.
- Crevasse traces are observed to a depth of 265 m and crevasses are inferred to a depth of 400 m.
- Crevasse traces present planes of weakness and show evidence of multiple reactivation phases, in some cases annual, during passage through the glacier.

Abstract Despite the inferred importance of deep surface crevasses to ice-mass deformation, mass redistribution and latent energy transfer, no observations of such crevasses have yet been reported. Optical televiewer (OPTV) logging of a 325 m-deep borehole drilled within a crevassed region of fast-moving Store Glacier, Greenland, revealed the presence of 35 high-angle crevasse traces that cut across the background primary stratification. Twenty-eight of these traces were formed of a bubble-free layer of refrozen ice that hosted one or more mm-thick laminae of bubble-rich 'last frozen' ice. Several such last-frozen laminae were observed in four traces, indicating multiple episodes of crevasse reactivation. In one notable case, nine such laminae were observed, indicating four, likely annual, reactivation events. Inspection of the full OPTV log reveals the frequency of crevasse traces generally decreased with depth, and that the deepest detectable trace was 265 m below the surface. This is consistent with the extent of the warmer-than-modelled englacial ice layer in the area, which extends to ~400 m below the surface. We conclude that deep crevassing is pervasive across Store Glacier, and therefore also at dynamically similar Greenlandic (and likely Antarctic) outlet glaciers. Once healed, the traces of these crevasses represent planes of weakness that may be reactivated during their subsequent passage through the glacier. Given their depth and continuing formation, it is highly likely that such traces survive ablation to reach the glacier terminus, presenting foci for rifting and iceberg calving.

Plain language summary Crevasses in the surface of ice masses allow ice-mass motion and the transfer of water and the heat it holds into the ice mass' interior. Crevasses extending deeper than some tens of metres have been inferred from indirect evidence such as radar scattering but have not been observed directly. Here, we evaluate the presence and properties of such deep crevasses by using an optical televiewer (OPTV) to record a continuous high-resolution image of the complete wall of a borehole drilled to a depth of 325 m in Store Glacier, a heavily-crevassed fast-moving outlet glacier of the Greenland Ice Sheet. The OPTV image intersects multiple now-closed crevasses to a depth of 265 m, many of which show evidence of multiple phases of opening and closing as they have been carried through the glacier to the location of our borehole. We infer that, at least in fast-moving glaciers such as Store, deep crevasses are pervasive, may be regenerated several times, and are able to transfer water and heat from the surface to depths of at least 400 m. The crevasses we image are sufficiently deep to survive surface melting to the glacier terminus, where they act as planes of weakness, enabling ice fracture and iceberg calving.

1 Introduction

Surface crevasses facilitate bulk ice motion and serve as pathways to transfer water and its thermal energy from the surface of an ice mass to its subsurface. Theoretically, creep closure limits the depth of dry crevasses to some tens of metres (see Mottram & Benn, 2009; Nye, 1955), but this can be increased substantially by hydrofracturing of water-filled crevasses (van der Veen, 1998; Weertman, 1973). Once initiated, hydrofracture extends the tip and, if enough water is available to increase its depth accordingly, propagation can continue to the glacier bed. This mechanism has been used to explain the rapid transfer of meltwater to the glacier bed during discrete surface lake drainage events (e.g., Chudley, Christoffersen, Doyle, Bougamont, et al., 2019; Das et al., 2008; Doyle et al., 2013), common through ice up to ~1 km thick around the margins of the Greenland Ice Sheet (GrIS) and at a smaller scale on, for example, Ellesmere Island, Canada (Boon & Sharp, 2003).

In the absence of lake-drainage driving crevasses to the glacier bed, at least three general lines of evidence suggest the presence of deep, englacially-terminating crevasses. First, such crevasses have been considered responsible for englacial warming as a result of energy released by meltwater refreezing within them (e.g., Colgan et al., 2011; Poinar et al., 2017), so-called 'cryo-hydrologic warming' (Phillips, Rajaram, & Steffen, 2010). For example, Lűthi et al. (2015) favoured refreezing of meltwater in 300 – 400 m deep crevasses to explain englacial ice temperatures measured in boreholes near the western margin of the Greenland Ice Sheet. These ice temperatures were recorded to be 10 – 15 °C warmer than modelled by heat diffusion alone. More recently, Seguinot et al. (2020) invoked the same process to explain measured englacial warming at a rate of 0.39 °C a^{-1} in englacial ice advancing towards the terminus of tidewater Bowdoin Glacier, Greenland. Here, it was considered that the crevasses, which possibly initiated in association with band ogives several km upglacier, may have warmed the glacier's full thickness of ~250 – 300 m. Such inferences are not limited to the margins of the GrIS. For example, Gilbert et al. (2020) suggested that meltwater refreezing in crevasses may be responsible for deep warming, inferred from surface ground-penetrating radar, at Rikha Samba Glacier, Nepal, a process that may also contribute to anomalously warm englacial ice measured in boreholes across the debris-covered tongue of Khumbu Glacier, Nepal (Miles et al., 2018). Second, deep englacially terminating crevasses have also been invoked to explain englacial radar scattering. For example, Catania et al. (2008) interpreted single englacial diffractors from 1 MHz surface radar transects in Greenland as reflections from the apex of deep crevasses, in this case from depths of ~100 – 200 m (their Figure 2). Third, inferred crevasse traces have been observed in ice near the glacier bed during exploration of ice-marginal subglacial cavities and tunnels (e.g., Hubbard, Cook, & Coulson, 2009; Lovell et al., 2015), implying initial propagation to depths of greater than some tens of metres. Hambrey (1976) calculated this depth (from cumulative overburden mass loss) to be at least 80 m from observations of surface crevasse traces in the ablation area of Charles Rabots Bre, Norway. However, in these cases there is some uncertainty involved in determining confidently whether such features were originally deep surface crevasses or local basal crevasses.

Strong inferential evidence therefore exists for the occurrence of crevasses extending >10s m below the ice surface. However, gaining direct access to such crevasses is effectively impossible and no first-hand observations of their presence or depth have yet been reported. One means of acquiring such evidence lies in imaging the wall of a borehole drilled to intersect crevasse traces that have deformed such that they are no longer vertical. In this paper we report on the application of borehole optical televiewer (OPTV) logging to ascertain the presence and properties of deep crevasses within Store Glacier, a fast-moving outlet glacier of the GrIS.

2 Field site and methods

Store Glacier (Greenlandic name *Sermeq Kujalleq*) is a fast-moving ice stream draining the central-west sector of the GrIS into Davis Strait via Ikerasak Fjord in the Uummannaq Fiord system (Figure 1).

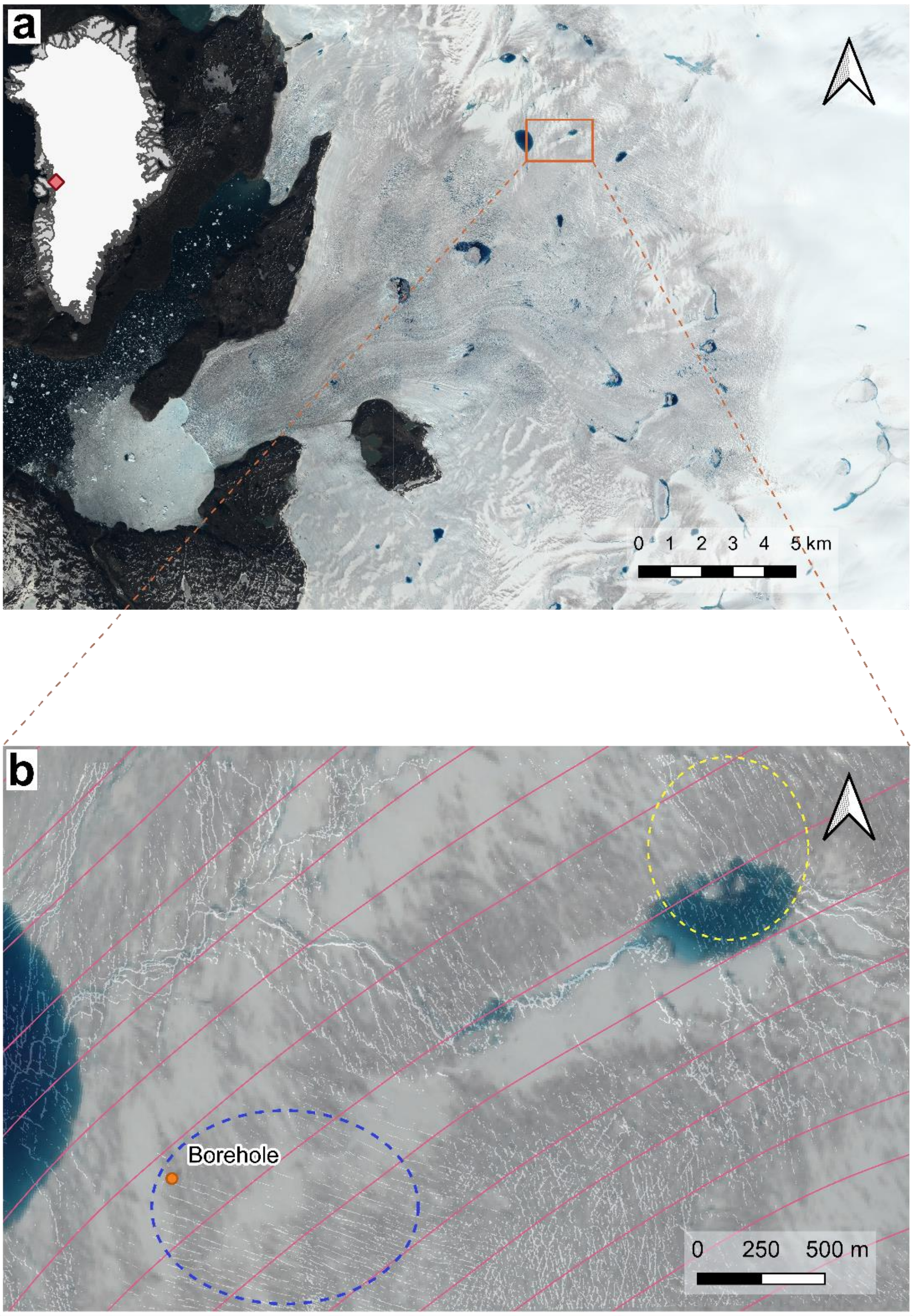

Figure 1. (a) Basemap image of Store Glacier, Greenland, and (b) expansion of (a) showing superimposed crevasse patterns from a UAV-generated digital elevation model (Chudley, Christoffersen, Doyle, Abellan, & Snooke, 2019). The smooth red lines indicate ice surface flow lines (NE to SW) derived from the MEaSUREs velocity dataset (Joughin, Smith, Howat, Scambos, & Moon, 2010), and the areas enclosed by the dashed lines identify dominant (planform) crevasse orientations of 290°/110° around the borehole (blue) and of 350°/170° ~3 - 4 km upflow of the borehole (yellow).

The borehole investigated herein (BH18c) is located within a crevassed area ~30 km from the terminus of Store Glacier and within some km of a supraglacial lake that drains rapidly via hydrofracture during most summers (Chudley, Christoffersen, Doyle, Bougamont, et al., 2019). Although the area is generally crevassed, BH18c was drilled ~5 m away from the nearest open crevasse, both for reasons of operator safety and to minimise the chances of the vertical borehole intersecting the (assumed vertical) crevasse. Locally, ice surface velocity is ~700 m a^{-1}, achieved largely through basal motion focussed at a temperate interface (Doyle et al., 2018) between ice and an underlying deformable subglacial sediment layer (Hofstede et al., 2018). Although drilling (by hot pressurized water) was halted temporarily to allow logging, the borehole was eventually drilled to the bed at a depth of ~949 m. The uppermost 325 m of this borehole was logged by optical televiewer (OPTV), providing a geometrically accurate RGB image of the complete borehole wall at a resolution of ~1 mm both vertically and laterally. Such OPTV images have the capacity to inform on annual accumulation (Philippe et al., 2016), the presence and scale of refrozen layers within firn (Hubbard et al., 2016), marine ice accretion (Hubbard et al., 2012) and visible structural features intersecting the borehole wall (Roberson & Hubbard, 2010). For example, the last of these studies identified eight separate ice structures, including high-angle crevasse traces to a depth of some tens of metres, in the ablation area of Midtre Lovénbreen, Svalbard. Since OPTV images extend around the entire (cylindrical) borehole wall, they are typically unrolled for 2D presentation, extending left to right around the compass (N-E-S-W-N or 0°-90°-180°-270°-0°). Planar layers appear as sinusoids on such 2D images, with the amplitude of the sinusoid representing the layer's dip and the phase of the sinusoid its azimuth or strike (Hubbard, Roberson, Samyn, & Merton-Lyn, 2008). Structural analysis and plotting were carried out using software packages BiFAT (Malone, Hubbard, Merton-Lyn, Worthington, & Zwiggelaar, 2013), WellCAD, and Stereonet v.11 (Cardozo & Allmendinger, 2013).

Borehole temperatures were recorded by sensor strings installed into two boreholes, BH 18b and BH18d, both located <10 m from BH18c. Englacial temperature was measured using thermistors (see Doyle et al., 2018) and DS18B20 digital temperature sensors (Table S1; Figure S1). Undisturbed ice temperatures were estimated following established methods (Doyle et al., 2018; Humphrey & Echelmeyer, 1990). A theoretical borehole temperature profile was also created by inverting the 2016 annually observed MEaSUREs surface ice velocity using the higher-order Community Ice Sheet Model 2.0, which conserves momentum, mass and thermal energy in three dimensions. The approach follows that used by Price et al. (2011) for Greenland and Bougamont et al. (2019) for Antarctica. The observed motion was simulated by first prescribing a no-slip basal boundary condition, and then subtracting the derived internal ice deformation from the observed surface velocities. We then iterated basal traction and sliding rates to an equilibrium in which ice temperature, effective ice viscosity, and ice velocity fields converged with the target velocities. To constrain the inversions, we used BedMachine v.3 ice thickness and bed topography (Morlighem et al., 2017) resampled to the 1 km fixed grid of our model. Surface air temperature was specified at the same 1 km resolution using output from the RACMO regional climate model (Noël et al., 2018).

3 Results

3.1 OPTV log

The complete 325 m long OPTV log of BH18c (Figure 2a) shows two general characteristics.

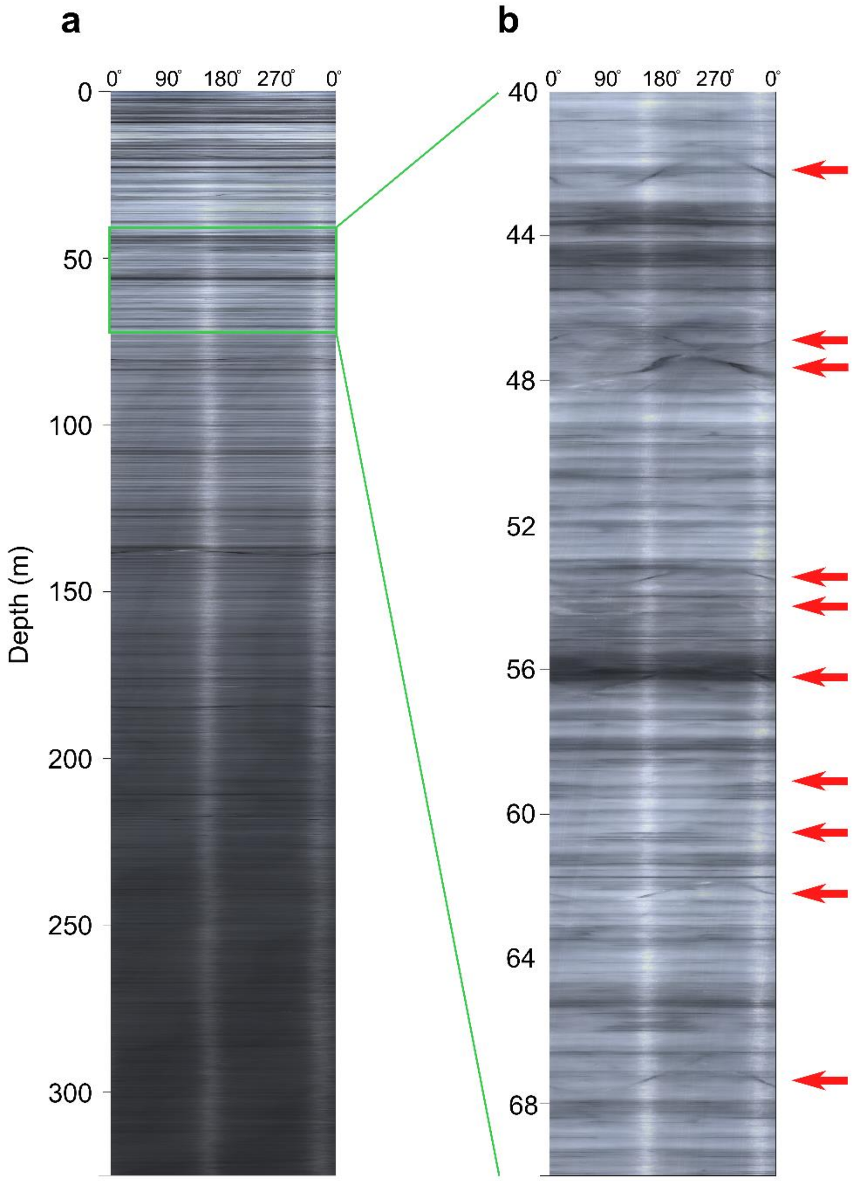

Figure 2. OPTV log of BH18c. (a) The complete 325 m-long log (note ~150 x vertical compression), and (b) expansion (~15 x vertical compression) of depth range 40 - 70 m showing 10 high angle sinusoids (planes in 3D space), each marked by a red arrow, cutting across low-angle background layers. With a borehole diameter of ~12 cm, the width of these OPTV images is the borehole circumference, ~0.38 m. Note, the reflective (brighter) sub-vertical streaks at ~160° and ~340° are drilling artefacts, probably caused by contact between the hose and borehole wall.

First, the log shows an overall decrease in luminosity with depth. This is common to many OPTV logs recovered from glacier boreholes (Hubbard et al., 2013), with light scattering decreasing with depth as the material forming the borehole wall progresses from highly reflective snow and firn through bubble-rich ice of intermediate reflectivity to almost transparent bubble-poor ice at depth. Since BH18c is in the ablation area of Store Glacier, it intersects no near-surface snow or firn, and the darkening in this case represents a general decrease in bubble content with depth. Second, numerous alternating decimetre-scale light and dark bands signify alternating bubble-rich 'white' ice and bubble-poor transparent ice (Figure 2b). This banding extends throughout the full depth of the OPTV log and, although appearing uniform and horizontal at the (vertically compressed) scale illustrated in Figure 2a, an expanded section (Figure 2b) reveals that the banding dips shallowly and is of highly variable thickness around the borehole.

Closer inspection of the OPTV log of BH18c also reveals the presence of 35 distinct high-angle layers (examples of which are indicated by red arrows in Figure 2b) that cut across the less distinct, lower-angle banding noted above. Enlargements of the log (Figure 3) reveal that these layers form regular sinusoids, indicating correspondingly uniform and planar layers in 3D space. Each layer is formed of one or more thin (~mm scale) laminae that is bright (reflective) and hence bubble-rich. In 28 cases these laminae are enveloped by cm- to tens of cm-thick layers of ice that appear dark and are hence transparent and devoid of reflective bubbles. Of these 28 layers, 24 host a single central lamina (e.g., Figure 3a and b), one hosts two such laminae, two host three such laminae, and one (a ~0.25 m-thick layer at a depth of ~184 m) hosts nine such laminae (Figure 3c).

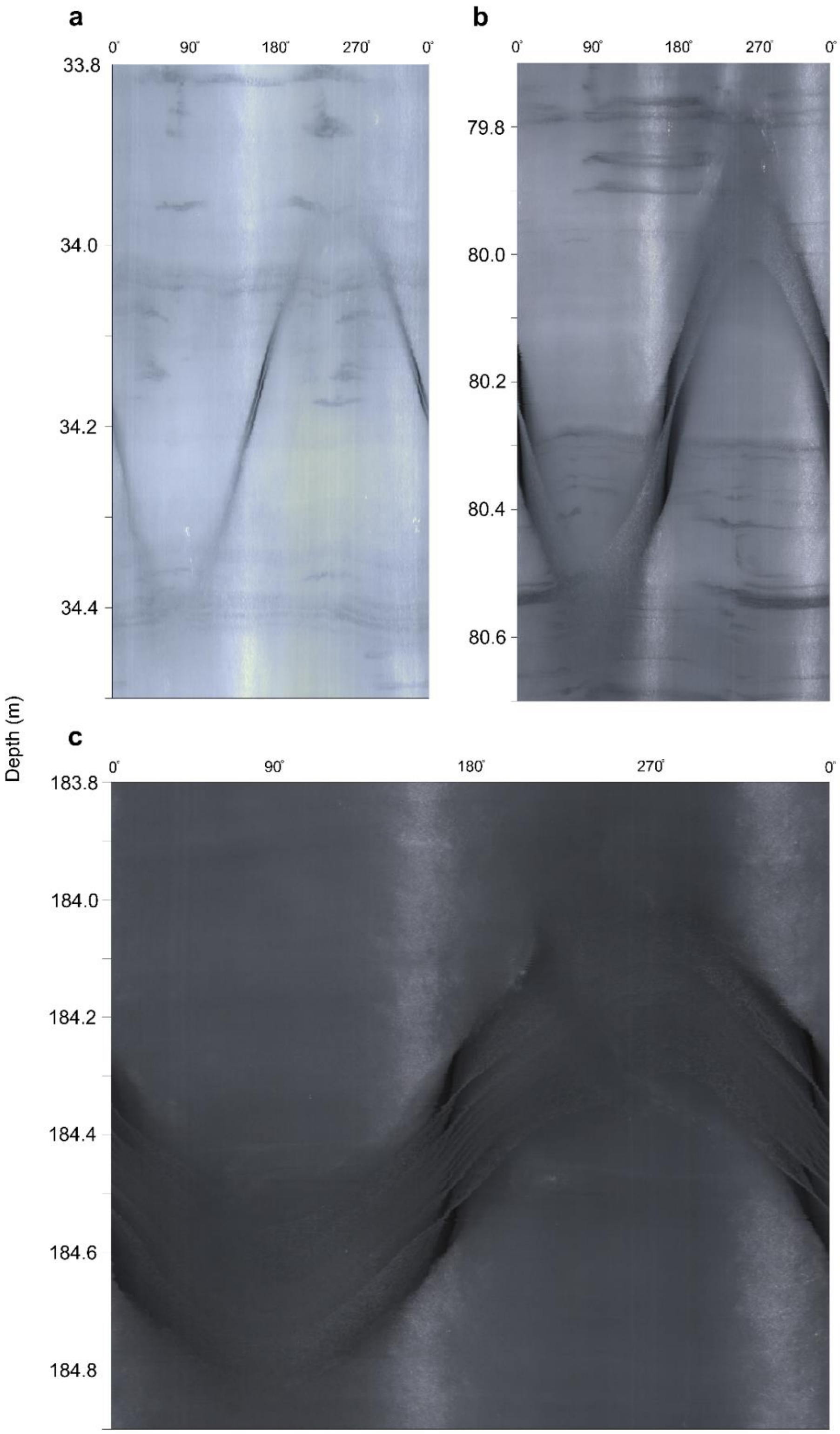

Figure 3. Selected expanded sections of the OPTV image of BH18c (shown in Figure 2), illustrating high-angle planar layers, apparent as a sinusoid in these 2D representations: (a) 33.8 – 34.5 m depth, (b) 79.7 – 80.7 m depth, and (c)

183.8 – 184.9 m depth. Note the general decrease in luminosity with depth and the central bubble-rich (bright) planar layers bisecting the high-angle layers imaged in (a) and (b) and the presence of multiple such layers in (c).

Analysis of the depth distribution of the 35 high-angle layers intersected by BH18c (Figure 4a) shows a general decline in frequency with depth, with the deepest observed plane being 265 m below the glacier surface. It is quite possible that the borehole intersected similar planes below this depth but that they were indistinguishable from the background ice. This is because (visibly bright) included bubbles are progressively expelled from all ice as pressure increases at depth, homogenizing the darkening OPTV image.

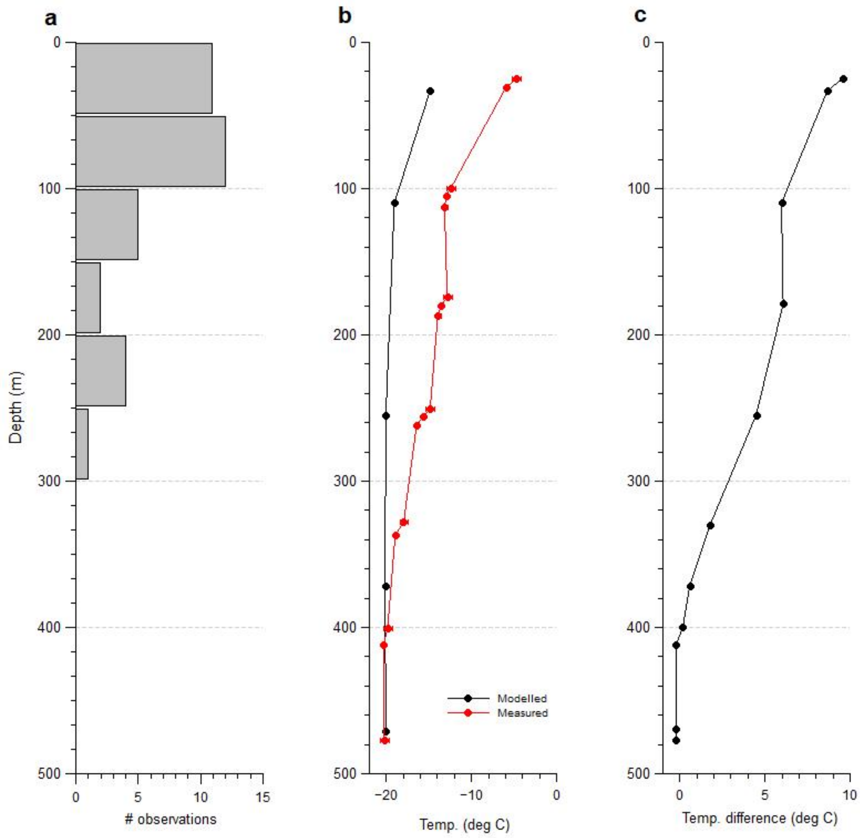

Figure 4. Key properties of the uppermost 325 m of BH18c at Store Glacier: (a) histogram of the number of high-angle planes identified in the OPTV log, (b) modelled (black dots) and measured (red dots) englacial temperatures for the overlapping sections of the two records, and (c) difference ('residual temperature') between the measured and modelled temperatures presented in (b).

Geometrical analysis of the orientation of these planes reveals a strong single maximum dipping approximately west to east (mean azimuth of pole to plane = 262°) at a mean dip of 63° (Figure 5 and Table 1).

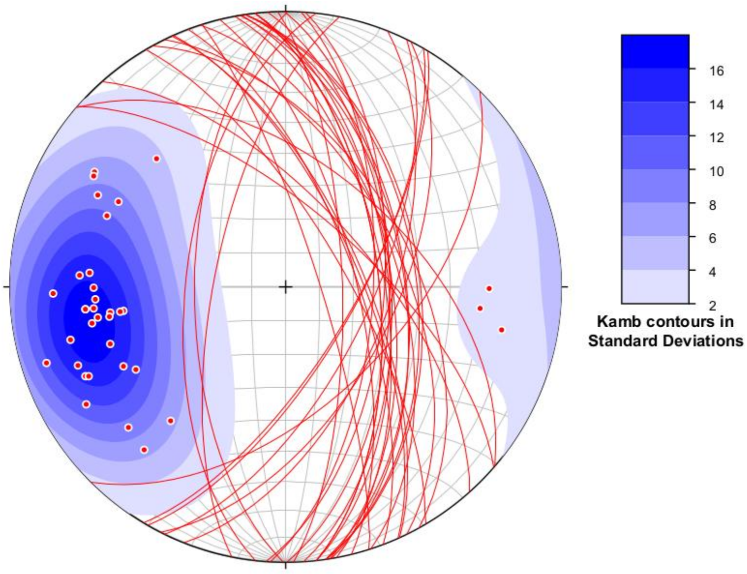

Figure 5. Lower hemisphere equal-area plot of all high-angle planes identified in the OPTV log of the uppermost 325 m of BH18c. Planes are plotted as red lines and poles to those planes as red dots, contoured by standard deviation following Kamb (1959). Eigen analysis data are given in Table 1.

Accordingly, eigen analysis of the poles to these planes (plotted as red dots on Figure 5) indicates a dominant principal eigenvalue of 0.83 at a principal vector of 262°/63° (Table 1).

Table 1. Eigen analysis data summarizing the 3D orientation of all 35 high-angle poles to planes, plotted as dots on Figure 5, identified in the OPTV log of the uppermost 325 m of BH18c.

Axis	**Eigenvalue**	**Eigenvector**	
		Azimuth (°)	**Dip (°)**
1	0.83	262	63
2	0.10	165	75
3	0.07	50	31

3.2 Englacial ice temperature

The borehole sensor strings recorded 32 discrete ice temperatures between the near-surface and the ice-bed interface (Figure S1). These temperatures decrease from ~-5 °C at a depth of ~25 m (the uppermost thermistor location in the borehole) to ~-22 °C at 600 – 700 m depth, to rise again sharply to temperate at the bed at a depth of ~949 m. Over most of the borehole length, the modelled temperatures correspond closely with the measured temperatures (Figure S1). However, modelled temperatures diverge from the observed record in the depth ranges 0 – ~400 m and 770 - 850 m. The former of these is relevant to the present study, and temperature data over the depth range 0 – 500 m are plotted in Figure 4b. Measured temperatures are notably higher than modelled temperatures, a difference (‘residual temperature’) that decreases with depth from ~10 °C near the surface to zero at ~400 m (Figure 4c).

4 Interpretation and discussion

The background (host) ice is formed of alternating sub-horizontal, but irregular, bands of bright bubble-rich ice and darker bubble-poor ice at the scale of cm to tens of cm. This is typical of the remnants of primary stratification in englacial ice, with the bands having originally been laid down as snowfall in the accumulation area of the GrIS. Such primary stratification has been identified elsewhere, both at the glacier surface (see review by Hambrey & Lawson, 2000) and within borehole OPTV logs (Roberson & Hubbard, 2010). Superimposed onto this primary structure, the high-angle planes have the physical properties of healed crevasses or crevasse traces. These properties include their secondary status (cutting across the primary stratification), high angle, and planar geometry. The clear nature of the transparent ice forming these layers is consistent with formation by the refreezing of a water-filled crack some cm to tens of cm across. Such ice is typical of refrozen ice from which gas has been expelled during the freezing process (e.g., Pohjola, 1994), as are their bubble-rich central laminae. Here, gas is rejected at the advancing ice front, increasing its concentration in the remaining reservoir of unfrozen water, eventually reaching saturation. At that point, bubbles form and are incorporated as a lamina into the last-frozen ice layer (Hubbard, 1991). The central position of the bubble-rich last-frozen lamina we observe in most of the OPTV image of BH18c (Figures 3a and b) is consistent with water freezing inwards at a similar rate from both edges a crevasse. Similar features, also interpreted as crevasse traces, have been identified cropping out at the glacier surface (e.g., Figure 6).

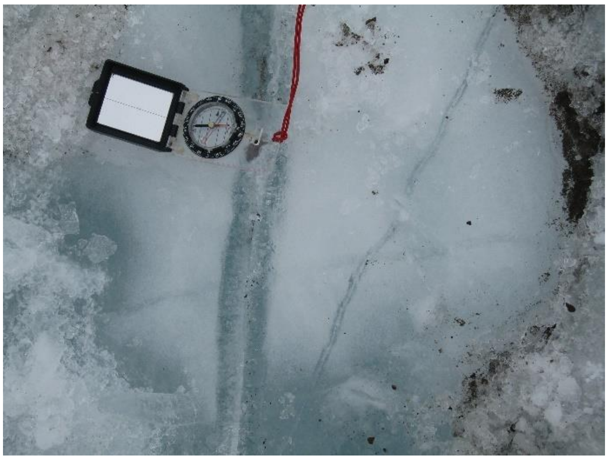

Figure 6. Photograph of a surface exposure of a set of sub-vertical crevasse traces at Trapridge Glacier, Canada (Hambrey & Clarke, 2019; their Figure 7d). Note the planar crevasse traces cut across the host ice and are generally clear with a bubble-rich central plane, like those in our OPTV images shown in Figure 3.

Where exposed at the glacier surface, such clear ice crevasse traces commonly erode preferentially, providing ready channels for supraglacial drainage (Hambrey & Müller, 1978) (also see Figure S2). We interpret multiple last-frozen laminae within a crevasse trace, such as that imaged at a depth of 184 m (Figure 3c), in terms of the effects of multiple separate refreezing events. Although multiple central laminae have not, to our knowledge, been reported elsewhere, reactivation of crevasse traces as thrust faults has been proposed (e.g., Goodsell, Hambrey, Glasser, Nienow, & Mair, 2005). Since the crevasse traces we image at Store Glacier overprint, but do not displace, the primary stratification (Figure 3), the crevasses appear to have formed by Mode I 'opening' with little or no accompanying shearing (Mode II 'sliding' or Mode III 'tearing'). Accordingly, we envisage reactivation of an existing crevasse trace to involve opening along its bubble-rich last-frozen lamina and splitting that layer into two thinner bubble-rich laminae. The open crevasse then refills with meltwater that eventually refreezes itself, forming a new (third) bubble-rich last-frozen lamina. After that, any subsequent reactivation can open any of the three pre-existing last-frozen laminae and the process is repeated, each time adding a pair of new laminae and therefore generally resulting in an odd total number of such laminae. This is consistent with our observations, and in particular with the feature at 184 m depth which has nine such laminae (Figure 3c), indicating the initial crevasse trace was reactivated four times. In contrast, we imaged one high-angle layer with an even number of central laminae (two). In this case, the crevasse could have reactivated along a new plane rather than along the pre-existing central laminae. If this were the case then, of the 28 filled crevasse traces imaged in the uppermost 325 m of BH18c, 24 showed no clear evidence of reactivation, three showed evidence of a single reactivation phase, and one showed evidence of four reactivation phases. Of these seven inferred reactivation phases, six occurred along a pre-existing bubble-rich last-frozen lamina. The seven bright laminae that were not visually bounded by transparent (refrozen) ice likely formed as dry crevasses, the traces of which consequently did not host a refrozen ice layer.

Geometrical analysis of the crevasse traces intersected in BH18c (Figure 5) indicates a strongly clustered distribution cantered on a dip of 63° in an approximately west-to-east direction. The poles to these planes are clustered at an azimuth of 262°, indicating that the strike of the plane is 352/172°. Reference to Figure 1b shows that this orientation is inconsistent with that of crevasses cropping out at the glacier surface locally (bounded by the dashed blue line in Figure 1b), which trend ~290/110°, but is almost identical to that of a crevasse field located ~3 - 4 km upflow of BH18c (bounded by the dashed yellow line in Figure 1b), which trends ~350/170°. All 35 crevasse traces imaged in BH18c therefore appear to have been formed some kilometres upflow and to have advected to the borehole. Indeed, it is this flow that has allowed them to deform sufficiently away from vertical to be intersected by the (vertical) borehole. This also explains why BH18c appears to have intersected no local crevasses (trending 270/90°), since these are presumably still undeformed and therefore close to vertical, reflecting their Mode 1 nature, while the borehole was drilled ~5 m equidistant from the nearest crevasses. Since local crevasse spacing is typically ~5 – 10 m (Figure S2), this absence of intersection indicates that these crevasses are shallow and/or do not deviate significantly from vertical over their entire depth range. Further, with a local surface ice velocity of ~700 m a^{-1}, it would take these newly formed crevasses ~5 a to advect to the borehole. The crevasse trace intersected at a depth of 184 m (Figure 3c), which shows four reactivation phases, could therefore have been reactivated annually. In this case the crevasse would be expected to open, providing a meltwater reservoir and possible drainage pathway, in the late spring or summer, when the ice speeds up and meltwater becomes available, and refreeze when that opening and filling are no longer able to resist inwards refreezing from the cold crevasse walls.

It takes ~40 years for ice to move from the location of our borehole to the front of Store Glacier, during which time ~100 m of surface ice would be lost to ablation (assuming a surface ablation rate of 2.5 m a^{-1}). Thus, crevasse traces deeper than this at the borehole site, as well as those formed deep enough farther down-glacier, would survive to the glacier's terminus, some to depths of 100s of m. Where this is the case, such traces will almost certainly continue to present sub-vertical planes of weakness, offering locations of preferential brittle failure for rifting and iceberg calving.

The frequency of intersected crevasses with depth is irregular but shows a general decrease (Figure 4a). Assuming this depth distribution represents that of the original crevasses, then only seven (or 20%) of the 35 intersected by the borehole were shallower than 40 m depth and 12 (34%) were deeper than 100 m. Although specific to our study site, the relationship between the crevasses present (%) per 30 m depth range ($C(30)_{\%}$) and depth (D, m), as illustrated in Figure 4a, can be approximated (R = 0.69) as a logarithmic function:

$$C(30)_{\%} = 40 - 6.2\ln(D) \qquad \text{Eq. 1}$$

This fit both suggests the presence of crevasses below the lowermost crevasse trace imaged by our OPTV log (at a depth of 265 m) and is consistent with the difference between measured and modelled englacial temperatures at our field site (Figure 4c), suggesting that crevasses warmed englacial ice to a depth of at least ~400 m. Fitting a logarithmic curve to this residual temperature (T_e, °C) against depth, as illustrated in Figure 4c, yields a close match (R = 0.95) described by:

$$T_e = 21.0 - 3.3\ln(D) \qquad \text{Eq. 2}$$

The similarity of the relationships between crevasse frequency and depth (Figure 4a; Eq. 1) and residual temperature and depth (Figure 4c; Eq. 2) lends support to our interpretation that crevasses propagate to at least 400 m below the surface of Store Glacier, warming ice to that depth by the presence and refreezing of crevasse-filling meltwater.

5 Summary and conclusions

Our OPTV log of a borehole drilled in a crevassed area of fast-moving Store Glacier, Greenland, has successfully imaging deep crevasses for the first time. Combining analysis of this log with thermo-mechanical modelling has revealed the following.

- Thirty-five traces of surface crevasses were imaged directly to a maximum depth of 265 m. It is possible that the borehole intersected crevasses below this, but that they could not be distinguished from the host ice. Approximately one third of these 35 traces were intersected below a depth of 100 m, and only 20% were intersected in the uppermost 40 m.
- The borehole intersected traces of crevasses that originally formed ~3 - 4 km upglacier. Despite drilling in an active crevasse field, and between two open crevasses spaced ~10 m apart, the borehole intersected no local crevasses. This indicates that the active local crevasses were shallow and/or did not deviate sufficiently from vertical to intersect the uppermost 325 m of the borehole.
- Of the 35 traces imaged, 28 showed evidence of having been filled with meltwater that subsequently refroze. This meltwater and its refreezing released heat, assumed to be responsible for warming the ice to a depth of ~400 m, suggesting crevasses extended to this depth - although not imaged below 265 m by our OPTV log.
- Refreezing of crevasse-fill water commonly creates a distinctive ice layer formed of a planar bubble-free ice layer some cm to tens of cm thick that envelopes a planar mm-thick central layer of bubble-rich last-frozen ice.
- Once formed, crevasse traces represent a plane of weakness that may be reactivated during advection through the glacier. Visual analysis of layering in the borehole OPTV image suggests that reopening occurs preferentially along the bubble-rich last-frozen layer(s), which then refill and refreeze, creating additional new last-frozen layers.
- Reactivation of crevasses may occur annually with crevasses opening and filling in the late spring and early summer. Once filled, refreezing would occur when the flow of meltwater is no longer able to resist inward freezing from the crevasse walls. The time separating opening and closing is not known but is likely to be within the range of hours to months. Although our OPTV log indicates that the refrozen ice layers are individually some cm to tens of cm thick, it is not known whether crevasses opened and filled to greater widths than this, subsequently closing through a combination of mechanical closure and refreezing.
- The crevasse traces imaged by our OPTV log, supplemented by new ones formed downglacier, extend deep enough to survive ablation and reach the glacier terminus. Here, being planes of weakness, these traces likely precondition the precise location of rifting, iceberg fracture and calving.

Acknowledgments

This research was funded by the European Research Council as part of the RESPONDER project under the European Union’s Horizon 2020 research and innovation program (Grant 683043). Hot water drilling and borehole logging equipment were also supported by an Aberystwyth University Capital Equipment Grant and NERC grants NE/J013544 and NE/K610026. TRC and RL were supported by a UK NERC Doctoral Training Partnership Studentship (Grant NE/L002507/1). We are grateful to Ann Andreasen and the Uummannaq Polar Institute for their kind hospitality, to Nicole Bienert and Sean Peters for assistance in the field, and to Katie Miles, Emma Docherty and Tom Chase for assistance in constructing sensor strings.

Data availability

The OPTV, temperature and modelling data presented herein are available from https://figshare*...

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