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Melt water flow through a rapidly de-glaciating glacier and foreland catchment system, Virkísjökull SE Iceland

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Abstract

The aim of this study is to characterise the glacial and pro-glacial hydrology of a rapidly de-glaciating system at Virkísjökull in SE Iceland, and to determine the water velocities through the glacier and pro-glacial area. This was achieved using dye tracer tests, river discharge measurements and studies of conduits within the foreland and lower glacial ablation zone using Ground Penetrating Radar (GPR). Tracer testing through the glacier via a moulin demonstrated rapid flow of 0.58 m s^{-1} which is comparable to the flow velocities within the pro-glacial river. A subsequent test at the end of the winter season demonstrated slower but still rapid flow of 0.02 m s^{-1}. A tracer test through the proglacial foreland shows that the large proglacial lake does not substantially attenuate flow, with velocities of 0.03 m s^{-1}. GPR profiles suggest the presence of a buried conduit system enabling the rapid transit of water through this area. The pro-glacial foreland contains buried ice which represents the remains of the retreating glacier; therefore this conduit system may be the remains of an en- and sub-glacial conduit flow-path. Buried conduits may be common in other de-glaciating ice cored forelands, and this study reveals that these may by-pass large proglacial lakes, which has implications for understanding hydrological response times in glacial catchments. The pro-

glacial river is highly responsive to melt as a result of the fully developed conduits in both the sub-glacial and pro-glacial areas. Flow in the river is perennial, suggesting that the conduit systems in the glacier and buried ice remain open and active all year, and that glacial melting occurs in winter as well as in summer, enhancing the rapid deglaciation.

Introduction

Changes in glacier mass-balance as a result of temperature increases are the cause of widespread glacial-retreat in Iceland (Jóhannesson *et al.*, 2006; Ađalgiersdóttir *et al.*, 2011; Haugen & Iversen, 2005; Fenger, 2007; Halldorsdóttír *et al.*, 2006); and throughout the world including in the Andes (Vergara, 2007), Alaska (Arendt *et al.*, 2002), Greenland (Howat *et al.*, 2005) and the Alps (Fyffe *et al.*, 2014). Deglaciation is resulting in the rapid evolution of englacial conduits (Nienow *et al.*, 1998), increased development of crevasses and moulins (Catania & Neumann, 2010), and an increased risk of Jökulhlaup flood events (Watanabe *et al.*, 1994). The IPCC (2014) reports that increased melting is resulting in changes to proglacial river systems with consequences for the management of water resources and hazards (Cisneros *et al.*, 2014). In some catchments an overall increase in river discharge is observed due to the increase in meltwater volume (Nolin *et al.*, 2010). In some rivers there is a particular increase in winter flows due to increased winter temperatures (Fountain & Tangborn, 1985), and in some cases there is predicted to be a transition from ephemeral to perennial river flows (Jóhannesson *et al.*, 2007). Deglaciation causes rapid changes in proglacial forelands with some glaciers in Iceland retreating at rates of 14 m a^{-1} during the period 1990 – 2004 (Bradwell *et al.*, 2013). Proglacial lakes may be rapidly formed (Bennet & Evans, 2012; Ageta *et al.*, 2000; Kirkbride, 1993) or may disappear (Bjornsson & Pálsson, 2008). This study uses tracer tests, river discharge measurements and Ground Penetrating Radar (GPR) to characterise the glacial and pro-glacial hydrology of a rapidly de-glaciating system, and to determine the water velocity through the glacier and pro-glacial area.

The study was conducted at Virkísjökull glacier in southeast Iceland which is an example of a very rapidly deglaciating maritime glacier. Sequential field photographs show that there has been substantial mass loss over the last 20 years (Bradwell *et al*., 2013). The glacier margin has retreated nearly 500m since 1996, and there has been a decrease in the glacier surface elevation of 8 m a^{-1} in the lowest reaches since 2012. The rate of retreat is accelerating, with an increase from 14 m a^{-1} of retreat between 1990 and 2004 to 33 m a^{-1} of retreat between 2005 and 2011 (Bradwell *et al.*, 2013). Annual net balance (1991 – 2006) calculated using a robust stratigraphic method is found to range from – 10 m at the terminus to + 5 m above 1800m (Björnsson *et al.*, 1998; Björnsson & Pálsson, 2008). This average mass balance gradient is the strongest in Iceland, and Virkísjökull is one of the highest turnover glaciers in Europe.

The proglacial foreland is evolving in response to the rapid deglaciation, and a large pro-glacial melt water lake has formed within the last 10 years. Based on a digital elevation model and field photographs this ice-cored lake has an area of approximately 1 km^2 (Fig. 1) and is characterised by regions of collapse features and braided melt water channels.

In this study all three components of the system are investigated: the glacier, the proglacial foreland (including the lake), and the pro-glacial river (Fig. 1).

Methods

Overview

Sub-glacial tracer tests were undertaken from glacial moulins to the outlet at the glacier snout to investigate meltwater velocities at the start and end of the main meltwater season in this rapidly deglaciating system. Ground Penetrating Radar (GPR) was used to investigate the thickness of the glacier ice and the location of the conduits within it.

Tracer testing was also undertaken in the proglacial foreland to investigate whether glacial meltwater sinking into the buried ice near the glacier terminus discharges into the lake via diffuse or point inputs, and to investigate whether water flow from the subglacial conduit outlet is attenuated during transit through the large lake area. Ground Penetrating Radar (GPR) surveys were undertaken prior to this study at Virkísjökull in the proglacial area to investigate the presence and nature of buried ice below the glacial outwash (Philips *et al.,* 2014). They are used here to determine whether conduits are present within the buried ice.

River discharge was measured in the pro-glacial river over three years, to determine temporal changes in meltwater discharge, and establish whether melting occurs on a perennial basis. Tracer dilution gauging was attempted to evaluate the technique as an alternative to current meter gauging, given the difficulties of current meter use in high flows, and to estimate the flow velocity over an extended distance within the pro-glacial river. The tracer tests therefore enable the comparison of velocities through the glacier, the pro-glacial lake, and in the pro-glacial river.

Tracer tests

Glacial tracer tests were carried out at the end of the melt season in September 2013 and August 2014, and at the start of the melt season in May 2014. The glacier comprises two arms split by a rocky ridge (Fig. 1). Tracer tests were carried out in September 2013 and May 2014 from a moulin on the east arm, 1.5 km from the terminus (Point 1 on Fig. 1). The flow into the moulin was substantially lower during the May test (Fig. 2). Monitoring was carried out in the stream where it exits at the glacier terminus. Two tracer tests were carried out from large moulins on the lower ablation zone of the western arm of the glacier in May and August 2014 (Point 2 on Fig. 1). These were 0.839 km and 0.668 km from the terminus respectively. In all four tests a 40% sodium fluorescein dye solution was injected, and monitoring was carried out in the river just downstream from where it discharges through the glacier snout (Point 3 on Fig. 1).

The pro-glacial lake area represents a geologically constrained basin with only one outlet for meltwater into the pro-glacial river. The lake sits on top of a sediment covered ice core which is the remains of the former glacier. Water discharging at the glacier snout flows across the debris surface for approximately 50 m before it sinks back into this buried ice. There is no visible outlet for this water. In order to investigate the hydrological functioning of the pro-glacial lake and foreland, tracer was injected into the river where it emerged at the glacier terminus, 20 m upstream of where it sank into the proglacial foreland buried ice (Point 8 on Fig.1). The test was carried out in September 2013 and a 40 % rhodamine WT solution was injected. This dye was used because it is not susceptible to photochemical decay if exposed to light during transit through the lake. The east and west side of the lake outlet channel, and the downstream river gauging station were monitored for dye tracer return (Points 4, 5 & 6 on Fig.1).

GGUN-FL in-situ field fluorometers designed at Neuchatel University were used at all monitoring sites to measure fluorescence and turbidity at 2 minute intervals. The fluorometers were calibrated to standard concentrations before field work commenced.

Monitoring was carried out for as long as possible before and after the tests to ensure background fluctuations in fluorescence were well characterised, and to capture the tail of the breakthrough curves. On occasion monitoring periods were shortened because high flows put equipment at risk. Monitoring was continued for a minimum of 2 days after the dye injection where possible.

Fig. 1 shows the location of the injection and detection points for all the tracer tests and further details are provided in Table 1.

Tracer recoveries were not calculated at the lake outlet and the glacier terminus because it was not safe to measure the river discharge. They were not calculated at the downstream monitoring point in the proglacial river because although discharge data are available, the dilution gauging showed that there is incomplete tracer mixing within the channel and therefore it is not possible to estimate an accurate recovery.

Turbidity increases fluorescence, and fluctuates in response to discharge (Wilson *et al.*, 1986). It is therefore highly variable in glacial environments where discharge varies considerably due to both diurnal fluctuations in melting, and in response to rainfall (Schnegg, 2002). Turbidity measurements were collected concurrently with fluorescence data, enabling identification of fluorescence changes that were likely to be due to turbidity fluctuations rather than changes in tracer concentration.

Proglacial river discharge measurements

River discharge was monitored 2.92 km downstream of the lake outlet (Point 6 on Fig. 1). An automatic river gauging station was set up in order to generate continuous river flow data. Water level ('stage') was monitored continuously using submersible level transmitters attached to the adjacent road bridge. An Ott Kalesto V surface velocity sensor was also deployed to provide a cross-check, indicating when changes in channel morphology were likely to have affected the stage-discharge rating. Flows were measured with an Ott C-31 current meter by wading and with a 3m rod from the bridge. The resultant 56 flow gaugings in 2011-2014 allowed the stage data to be converted to flow at 15-minute intervals for the period 16/09/11 until 15/11/14. Because ice affects the stage-discharge relationship, photographs were taken three times daily, at 9:00, 12:00 and 15:00, to identify ice development in the channel during winter periods. Periods when ice was present in the channel or around the banks were removed from the discharge record. Temperature was measured at hourly intervals at the gauging station using a temperature probe. Discharge data from the period of the tracer testing are presented in Fig. 3.

GPR

Ground Penetrating Radar (GPR) surveys were performed at Virkisjökull in April 2012 and 2013 as part of an ongoing study into the structural glaciology (Philips *et al.*, 2013; 2014). The results of these surveys have been used in this paper as an indication of possible water flow pathways within the glacier. A GPR survey was also conducted in the proglacial area in September 2012. A PulseEKKO Pro system with 50 MHz and 100 MHz antennae was used. Antennae were aligned perpendicular to travel direction and towed manually across the surface, with the radar being triggered at 0.25 m and 0.5 m spacing by an odometer wheel. Where the ice surface was fractured, the antennae were moved stepwise and the radar was triggered manually. Positional data were stored alongside GPR trace data using a standalone Novatel SMART-V1 GPS antenna. Raw GPR data were processed in EKKO View Deluxe (Sensors and Software, 2003). The processing consisted of applying a dewow filter, 2-D migration (for clean ice surveys), SEC (Spreading and Exponential Compensation) gain, and topographic correction. For the clean glacier ice, a radar wave velocity of 0.156 m ns^{-1}, previously calculated for Virkisjökull (Murray *et al.*, 2000) was used.

Dilution gauging

Dilution gauging was attempted downstream of the river gauging station as a means of evaluating the method as an alternative to bridge gauging with a current meter, and to measure the water velocity over an extended reach in the proglacial river. Rhodamine WT dye was injected directly into the centre of the pro-glacial river channel downstream of the lake outlet (Point 7 in Fig.1). Monitoring was carried out 2.92 km downstream of the injection point on the east and west side of the river to check for tracer mixing (Fig. 1). River discharge is calculated using the method outlined by Leibundgut (1998):

$$Q = \frac{M}{\int_0^{\infty} C_t \times \Delta t} \tag{1}$$

Where Q is the discharge in flowing water; M is the mass of injected tracer; C_t is the tracer concentration at time *t* (the integral is the area under the breakthrough curve); and Δt is the length of the constant time interval.

The velocity in the pro-glacial river over the 2.9 km section was calculated based on the measured time to first arrival of tracer and time to peak tracer concentrations.

Results

Sub-glacial tracer tests

Tracer breakthrough at the glacier snout was very rapid in September 2013, occurring 50 minutes after dye injection into the moulin on the eastern arm of the glacier (Fig. 4a). Peak concentrations were about 58 minutes after injection. Monitoring stopped 11 hours and 43 minutes after injection when the water level dropped below the level of the fluorometer because the water was diverted naturally into a different channel. Tracer concentrations decline to below background levels 8 hours and 33 minutes after the dye injection. However turbidity was lower and decreasing at this time suggesting that background fluorescence was lower than at the start of the test, and therefore it is difficult to determine exactly when tracer ceased to be discharged. At the end of a full melt season the melt-water transmission to the glacier margin (based on the time to peak of the tracer test) was rapid at 0.58 m s^{-1}.

In May 2014, at the beginning of the ablation season, the tracer breakthrough along this flowpath was less rapid, occurring 5 hours and 18 minutes after the tracer injection (Fig. 4a). Peak tracer concentrations were also later (5 hours and 36 minutes after injection). Tracer concentrations had not returned to background when monitoring stopped 744 minutes after injection due to high flows which put the fluorometer at risk. At the beginning of the melt season the melt-water transmission to the glacier margin (based on the time to peak of the tracer test) was 0.07 m s^{-1}.

The tracer test from a moulin on the western arm of the glacier in May 2014 resulted in no observable break-through curve at the glacier snout (Fig.4b). The changes in florescence have a similar pattern to the changes in turbidity, and apparent increases in tracer concentration around 491 minutes after injection coincided with turbidity increases suggesting that they are natural background fluctuations rather than tracer being discharged. Monitoring stopped 717 minutes after injection because of movement of the outlet channel putting equipment at risk. Whilst it is possible that the tracer breakthrough could have occurred after monitoring stopped, this is unlikely given that the monitoring continues for substantially longer that the time taken for tracer breakthrough from the moulin on the eastern arm which was carried out during the same period under similar discharge conditions suggesting that there should have been enough flow in the injection moulin to flush the tracer through the system. In addition, the moulin on the western arm is much closer to the glacier snout monitoring point than the moulin on the eastern arm. It is therefore quite likely that sufficient tracer was injected and that tracer breakthrough at the glacier snout would have occurred within the monitoring period if there was a connection.

In August 2014 a tracer test was carried out from another moulin on the western arm of the glacier but there was also no observable break-through curve (Fig. 4c), despite two days of monitoring following injection and higher flow rates during the test (Fig. 3). This suggests that drainage from the western arm of the glacier may not be connected to the outlet at the glacier snout.

Pro-glacial tracer test

A breakthrough curve was obtained at the lake outlet west bank monitoring site (Fig. 5). Tracer breakthrough was at 19:00 on 17/09/13, 7.5 hours after injection (Fig. 5a). Peak concentration occurred at 20:30, 9 hours after injection.

Two smaller peaks 630 and 1475 minutes after injection are within measured background levels and coincide with increasing turbidity suggesting that they are variations in background fluorescence rather than tracer discharge. There was no tracer breakthrough at the lake outlet east monitoring point (Fig. 5b).

Tracer was detected at the bridge monitoring site in the river (approximately 1.5 km downstream of the lake outlet monitoring point). Although there is only a very small rise in dye concentration the turbidity is declining at this point suggesting that it is tracer being discharged (Fig. 5c). Tracer arrival is around 20:00 giving a travel time of 1 hour from the lake outlet to the road bridge within the main river channel which is similar to the travel time observed in the river tracer test (see below). Peak concentrations occurred at 04:30 on 18/09/13. Tracer concentrations remained above background for the longest time at this detection point reflecting dispersion along the river channel. They did not decrease to below background levels until 19.09.2013 at 13:00, 72 hours after the injection. Table 1 summarises the main findings for direct comparison between the three systems.

River discharge

Measured river discharge fluctuations at Virkisá are typical of sub-arctic meltwater rivers which are dominated by seasonal and diurnal temperature fluctuations (Shaw *et al.*, 2011).

There is a strong seasonal variation in flow which increases between May and September and decreases between September and November (Fig. 3). Data for the winter are limited due to icing, but the photographs and direct measurements in all seasons indicate that the river flows throughout the year, even when ice is present. This suggests that ice melt continues in the winter period and conditions are not cold enough to cause conduits to freeze and close. This is despite daytime temperatures in winter falling below 0 °C for up to 5 consecutive days. Over the 3 year monitoring period the mean flow was 5.6 $m^3 s^{-1}$, and the maximum flow was 39 $m^3 s^{-1}$ on the 6th of December 2013.

The times of the tracer tests are presented on Fig. 3 to show the hydrological conditions during the tests. Three tests were carried out in September 2013, towards the end of the meltwater season. The east arm glacier tracer test was conducted during moderately high flows (5.5 $m^3 s^{-1}$). The flow was decreasing during the proglacial foreland tracer test, and was lower (4 -3 $m^3 s^{-1}$). The proglacial river tracer test was conducted when the flow was about 3 $m^3 s^{-1}$. The second east arm glacier tracer test was carried out at the start of the following meltwater season in May 2014 when flow was 2 $m^3 s^{-1}$.

GPR

Sub-horizontal to gently up-ice dipping reflective surfaces within the glacier are apparent in profiles for the lower parts of the glacier (Fig 6A)). These reflectors are, in places, very high amplitude. In several areas their polarity is reversed, indicating a higher dielectric permittivity and lower wave velocity in the material below interface, suggesting the presence of water, or wet sediment. These sub-horizontal reflectors are longitudinally continuous for distances in excess of 100 m; they have been interpreted as thrust planes (Phillips et al. 2013), where the fractured ice potentially provides a zone for water flow and conduit development. Field observations (Fig 6B) confirmed the presence of wet, graded (waterlain) sediment, and conduits in one of these thrust planes. A prominent down-ice dipping reflector extends from the glacier surface, where it occurs in association with three moulins, down to the glacier bed approximately 50 m below the surface (Fig 6C). The reflector joins a zone of lateral fractures at the ice surface, and is interpreted as part of a down-ice dipping fault system where part of the glacier is collapsing (Phillips et al 2013). The focusing of moulins at the fault zone indicates that the system acts at an effective water flow conduit. Collectively, the results from the glacier suggest that there is a pattern of conduit formation in the lower glacier which is associated with ice structures (thrusts, faults), ice-surface topography and the position of moulins which occur predominately on the eastern side of the glacier where there may be a high meltwater input to the fault and thrust plane network (Fig. 6D).

GPR profiles in the proglacial area (shown in Fig. 7A) are characterised by an upper unit of horizontal and gently undulating reflectors overlying a generally less reflective unit (Fig 7B). Field observations confirm that the upper unit is stratified outwash sand and gravels which are 1-2 m in thickness, and that the lower less reflective unit is buried ice (Figs 7B,C). The top of the buried ice is characterised by a number of hyperbolae, which may represent water-filled conduits or cavities close to the ice surface. The base of the ice is marked by a transition back to higher amplitude reflectors (Fig 7B). Clear, reversed polarity hyperbolae occur in several places in the ice (for example, Fig. 7D). These are interpreted as water-filled conduits and, in places, are associated with a thickened zone of chaotic reflectors in the sands and gravel above, representing collapsing ground (Fig. 7D). A marked zone of muted or absent reflections was observed in a number of profiles where they crossed a distinct linear zone that had been particularly affected by collapse holes. The exact reason for poor reflection in this zone is not known, but it may be related to a turbulent subterranean river that was observed sinking underground in this zone at the time of survey (Fig 7A). Field observations indicate that kettle holes and collapse features in the pro-glacial zone intercept a freely draining system as melt water from the terminus is regularly redirected into one of these features. Collectively the radar data and observations from the proglacial area demonstrate the presence of an extensive mass of ice buried, with numerous conduits and voids, beneath the outwash sands and gravels.

River channel dilution gauging

Although the dilution gauging was carried out over a distance of 2.9 km the tracer was not fully mixed across the channel at the monitoring point, illustrating the difficulty of achieving dye mixing in complex braided mountain river channels and within acceptable distances. The peak concentration and tracer recovery were higher on the east bank monitoring point than on the west bank, and tracer arrival was faster on the east side (Fig. 8a). Data from the west side give an overestimate of the flow obtained from the gauging station, whereas the data from the east side underestimate flow. Because full mixing was not achieved it is not possible to obtain an accurate flow estimate from the dilution. However, the flow estimates from the dilution gauging are broadly similar to the continuous flow estimates during the test (Fig. 8b).

The river channel dilution gauging demonstrates rapid velocities along the river. Velocity based on time to first arrival of tracer is 0.6 m s^{-1}, whilst the velocity based on the time to peak is 0.4 m s^{-1}.

Discussion: Melt water velocities

Meltwater velocities through the glacier from the eastern arm are extremely rapid at the end of the main melt season (0.58 m s^{-1}). This is almost identical to the velocity of 0.6 m s^{-1} observed in the proglacial river channel, suggesting that meltwater transfer is as efficient within the glacial conduit system as it is within the braided proglacial river channel. It is also substantially higher than the velocities observed during tracer tests in karst conduits and caves. Worthington *et al* (2009) compiled velocities from 3015 karst tracer test which had a median velocity of 0.02 m s^{-1}. The Haut Glacier D' Arolla in Switzerland was found to have velocities in glacier conduits at between 0.37 – 0.72 m s^{-1}, so although the glacier water flow is rapid, it is within observed velocities for other glacial tests (Table 2).

A repeat test undertaken at the start of the following ablation season demonstrates a lower velocity of 0.07 m s^{-1}. Flow in the pro-glacial river is substantially lower during this tracer test (~2 m^3 s^{-1} compared to 5.4 m^3 s^{-1}). Flow in the injection moulin is also substantially lower (Fig. 2).

Changes in flow velocity and dispersivity have been interpreted as a change from a distributed system to a channelized one (Seaberg *et al.*, 1988; Willis *et al.,* 1990; Fountain, 1993; Nienow *et al.,* 1998; Hock & Hooke, 1993). However this assumption has not been validated in glaciers where the drainage system configuration is independently known (Gulley *et al.,* 2012a). Velocities lower than 0.4 m s^{-1} have been interpreted as flow in a distributed system in previous studies (Nienow *et al.*, 1998; Mair *et al.,* 2002; Hubbard & Glasser, 2005). These velocities are an order of magnitude higher than the median velocity of 0.02 m s^{-1} for karst aquifers reported by Worthington *et al* (2009) and it is therefore likely that the even these lower velocities observed in subglacial tracer tests are indicative of well-developed ice conduit systems. It is well known from karst systems that velocities along the same flow-path can vary greatly in response to discharge, with faster velocities under higher flow conditions (e.g. Stanton & Smart, 1981; Göppert & Goldscheider, 2007). Whilst this can be due to activation of a slightly different karst conduit flow-path under different flow conditions, it can also be due to changes in dispersivity and pooling along the same flow-path. By analogy, it is therefore difficult to draw conclusions about morphological changes in glacial conduit systems using tracer tests.

At Virkísjökull, repeat visits suggest that the conduit persists throughout the winter and its morphology appeared not to vary between repeat field visits from 2012 – 2014 when moulins and the outlet were observed to remain in the same location. According to the GPR survey the ice at the terminus overlying the conduit is not thick (<50m), which results in negligible creep closure rates. Nienow *et al* (1996) showed that velocities in glacial conduits were lower during tests in which the flow in the injection moulin was lower, whilst velocities did not seem to relate to changes in flow in the outlet channel.

It seems likely that in this study, reduced input flow (melt) into the injection moulin is the main reason for the lower velocities (0.07 m s^{-1}) observed at the start of the melt season, rather than a dispersed drainage system. Tracer may move less rapidly in the channel at low flows as it is slowed down in pools and around boulders due to the increased tortuosity (Hauns *et al.*, 2001; Benn & Evans, 2014). At high flow boulders are completely submerged thereby reducing the amount of back-eddy current and temporary storage (Gulley *et al.*, 2012a).

Meltwater velocities in glacial systems vary greatly because of discharge variations, and have been shown to do so in response to diurnal changes (Schuler & Fischer, 2009). The two glacier tracer tests that produced breakthrough curves were carried out under very different conditions: the end of the main melt season when flow in the glacial outlet channel was above average (Fig. 3.), and flow in the injection moulin was high (Fig. 2); and the start of the melt season when flows were substantially lower. In the study by Schuler *et al.* (2004) diurnal variations in velocities were from 0.34 to 0.75 m s^{-1} in tests during August 2000, and from 0.15 – 0.58 m s^{-1} in tests during September. These variations are relatively small compared to the order of magnitude variation observed in this study. The second tracer test was conducted when flow in the injection moulin was very low and it is unlikely that a tracer test could be carried out under much lower velocities. Therefore the velocity of 0.07 m s^{-1} is likely to be fairly close to the minimum velocity in this system, although velocities may be lower from other moulins. It is likely that at times flow into the injection moulin is higher than it was during the high flow test in September, and it is therefore possible that velocities may be higher at times. Slightly higher velocities have been observed in other glacial tracer tests (Table 2). However the results from these subglacial tracer tests are broadly comparable to these other published studies.

The velocity in the buried ice channel through the foreland area is 0.03 m s^{-1} which is similar to the velocity in the subglacial conduit system in May. This suggests a well-developed conduit system which discharges through a discrete point into the lake and then continues as channelized flow without any substantial ponding or dispersion before reaching the lake outlet channel. In this deglaciating system meltwater is efficiently drained through the entire system – the glacier, the proglacial area and the river.

The glacial drainage system

The results of this study allow conclusions to be drawn about the connection between the location of moulins and the main drainage system as a result of the glacier structure. GPR suggests that the lower part of the glacier is characterised by a well-developed, structurally influenced, interconnected drainage system. In 2012 this system extended to a depth of 30 m below the surface at the clean ice margin; this ice is now slowly being covered by outwash as it becomes part of the proglacial area. These results indicate that the conduit system is connected to a structural system within the glacier that can be easily accessed by large moulins. Indeed moulins on Virkísjökull are observed in association with down ice dipping faults on the eastern arm of the glacier (Philips *et al.*, 2013; 2014). These features are concentrated along a north-south line on the eastern side of the glacier and the tracer testing data suggest that these feed directly into a north-south flowing main conduit drain beneath the glacier. Gulley *et al* (2012b) suggest that moulin locations, determined by the location of supraglacial streams and crevasses, control the location of subglacial recharge. It is not possible to estimate the number of conduits in this small survey area, but GPR shows that the drainage is tied to the glacier structural collapse features which suggests that melt water flow is likely to be very efficient and interconnected. How far this system propagates up-glacier is unknown, but large moulins are present higher up the glacier, suggesting that the main drain is likely to extend to the altitude of the ice fall at 400 m asl.

Tracer injected into moulins on the western side of the glacier was not detected in the glacier terminus outlet channel. The amount of water observed in this terminus outlet stream appears to be substantially less than the amount of water in the outlet channel from the lake, although it was not safe to measure the discharges in these channels. This suggests that the drainage from the western arm of the glacier may be connected directly to the buried ice in the proglacial area through active conduits within the dead ice, which resurface at discrete points in the lake area. Therefore in addition to the efficient conduit draining the eastern arm, there may be an additional major conduit system draining the western side which has not yet been located.

The proglacial foreland drainage system

The pro-glacial foreland at Virkísjökull is a formerly glaciated, ice cored region that currently comprises a series of down-wasting glacial features, braided melt water channels and an extensive lake area. This region has very recently been exposed, with the lake area only forming within the last 10 years. It is therefore a rapidly evolving and dynamic region.

Despite this, the pro-glacial buried ice appears to have retained its main melt water conduit enabling rapid transport of glacial melt water through the pro-glacial area.

The GPR data from the proglacial area demonstrate the presence of an extensive mass of ice buried beneath 1-2 m of outwash sands and gravels. Cavities and conduits are evident within the buried ice, and a more significant drainage route is also present indicated on the surface by the presence of collapse features.

The tracer test demonstrates that drainage is via a channelized system. Water emerging from the Virkísjökull glacier snout rapidly sinks into the buried ice in the foreland via a large kettle hole and flows through a conduit system in the buried ice to re-emerge within the large proglacial lake. The tracer test indicates a point input into the lake rather than a diffuse input. The rapid velocities indicate that once meltwater has emerged from the conduit system into the lake, the flow remains channelized, and dispersion into more stagnant areas of the lake is fairly minimal. The main melt-water flow exits the foreland through the western side of the lake outlet channel suggesting that the sub-glacial conduit also discharges on the west side of the lake. A flowing “channel” is visible moving through the west side of the lake, which is presumably fed by the main drainage conduit. However, the conduit outflow is not apparent, and may originate in an inaccessible area of the lake.

The GPR profiles of the proglacial area demonstrate that this region is associated with collapse features and void spaces as well as conduits. The lake extends over a large area, and many parts of it have no apparent flow but provide substantial off-line storage of water from the glacial catchment. The flow of the Virkisá (the pro-glacial river) is therefore predominantly influenced by the channelization of flow within the proglacial lake area.

The east side of the lake outlet channel has a strong flow and yet dye was not detected here during the proglacial test. This could be because the water on the eastern side of the lake originates from another conduit system discharging from the western side of the glacier, as discussed above.

The GPR and dye tracing results from the pro-glacial lake suggest that there is an efficient conduit within buried ice in the proglacial area. This conduit may be the remains of the original sub-glacial conduit that has been buried in the foreland after terminus retreat. Rapidly retreating glaciers can produce a transitional environment (Fig. 9). This transitional environment may form due to the slowing of glacier ice movement during deglaciation. Glacier stagnation can be linked to the drainage system (Gulley *et al.*, 2012a). Conduits with an efficient hydraulic capacity can drain surface meltwater more efficiently and thereby decrease subglacial water pressures and glacier sliding (Mair *et al.*, 2002; Anderson *et al.*, 2004). This results in the collapse of the distributed conduit system (which have a low hydraulic capacity) and leads to the collapse of flow into an integrated network of conduits (Kamb, 1987). Ice can then be buried by the accumulation of debris transported from higher reaches where ice is still flowing as a result of the steep gradient of the ice fall (Fig. 9(1)). The remains of active channels that exploit planes of weakness in the ice begin to collapse back due to being closer to the surface and covered by only a thin layer of ice and sediment. This exposes the water moving through the area (Fig. 9(2)) and a lake begins to form where ponding of water and the formation of surface pools occur (Fig. 9(3)). The current proglacial region has a surface river that originates sub-glacially and sinks back below buried ice into a conduit that was formerly connected to the active glacier system. This conduit connects to the lake that formed as a result of the collapse and decay of the ice on the far side of the proglacial area (Fig. 9 (4)). This conceptual model is based on the connection proven by tracer testing and GPR in the proglacial area, and illustrates the principle, but may be a simplification of a more complex network of conduit pathways which formed in a similar manner within the ice cored foreland.

Buried ice is observed in other places in Iceland through geo-electrical resistivity surveys carried out at recently deglaciated sites (Everest & Bradwell, 2003), and has also long been observed in other parts of the world (French & Harry, 1990 ; Evans & England, 1992). At Skeiđarárjökull, Kötlujökull and Hrútarjökull in Iceland, areas of buried ice are thought to have persisted in a stable condition for 50 - 200 years as photographic and lichenometric evidence suggest that the overlying debris has remained relatively stable indicating very slow melting (Everest & Bradwell, 2003; Kjaer & Krüger, 2001). Yet until now the movement of water though these areas has not been considered and how this influences the connectivity of the catchment discharge processes. Buried ice is characterised by many surface collapse features. GPR in this study has indicated voids and conduits within the buried ice at shallow depths. These present a significant hazard to persons and infrastructure in rapidly deglaciating catchments particularly following high flows that can undermine the structural integrity of buried ice conduits.

Conclusion

Tracer testing in the glacial and pro-glacial areas of a rapidly deglaciating system demonstrate very rapid transit of melt water at the end of the melt season with velocities of 0.58 m s^{-1} and 0.03 m s^{-1} respectively. This shows that at Virkísjökull the pro-glacial river is highly responsive to melt as a result of well-developed conduits in both the sub-glacial and pro-glacial areas. Melt water velocities through the glacier are similar to those in the river which demonstrates that the conduit system is a highly efficient means of transporting water.

Rapidly de-glaciating catchments such as Virkísjökull create a pro-glacial setting which is transitional between the glacial and longer established pro-glacial areas resulting in an extensive system of buried ice containing the relic conduits of the former ablation zone through which melt water can be transferred rapidly to the pro-glacial river. These findings have improved our understanding of the formation and role of lakes in the hydrology of de-glaciating landscapes, as well as highlighting the potential hazards in proglacial areas where buried ice may contain shallow and unstable conduits.

Buried ice in proglacial forelands is likely to become more common as a result of deglaciation, and understanding the hydrology of these areas is important to enable appropriate catchment modelling and hazard mitigation. This study shows that ice-stagnation terrain does not appear to impede the hydrological connectivity between the glacier and foreland during deglaciation in a humid temperature environment.

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Table 1: Summary of all tracer tests performed at Virkísjökull and their main findings. Velocity is calculated using the distance and time to peak.

Location	Injection time & date	Dye	Dye (g)	Distance (km)	Time to peak (min)	Velocity (m/s)	Post-injection monitoring (hr)
East	12:04, 12.09.13	Fluorescence	500	1.5	43	0.58	11.5
East	14:16, 03.05.14	Fluorescence	500	1.5	318	0.07	12.5
West	16:02, 04.05.14	Rhodamine WT	400	0.668	NA	NA	12
West	11:18, 04.08.14	Fluorescence	580	0.839	NA	NA	48
Foreland	16:48, 17.09.13	Rhodamine WT	2199	1	450	0.03	72
River	11:45, 15.09.13	Rhodamine WT	128	2.92	90 - 115	0.6	4

Table 2: Summary of the findings from selected dye tracer studies through glacier conduits.

Velocity ($m\ s^{-1}$)	Time to peak (mins)	Distance (km)	Number of successful tests	Glacier	Region	Publication
0.2 - 1.5	No number	No number	57	Pasterzengletscher	Austria	Burkimsher (1983)
0.008 - 0.228	No number	No number	15	Midtalsbreen	S. Norway	Willis *et al* (1990)
0.085 – 0.157	75 - 255	various	16	Brewster glacier	New Zealand	Willis *et al* (2009)
0.047 - 0.32	35-485	0.485-3.335	No number	South Cascade	USA	Fountain (1993)
0.07 - 0.72	20.35 – 39.60	3.3	415	Haut Glacier D'Arolla	Switzerland	Neinow *et al* (1998)
0.04-1.49	No number	1.5-14	43	Leverett Glacier	Greenland	Cowton *et al* (2013)
0.6 – 1.7	23.7 - 189	2.1-4.3	12	Gangori glacier	Himalaya	Pottakkal *et al* (2014)
0.07 – 0.88	45 -240	0.6	9	Rieperbreen	Svalbard	Gulley *et al* (2012a)
0.07 - 0.58	43-318	1.5	2	Virkísjökull	SE Iceland	(this study)

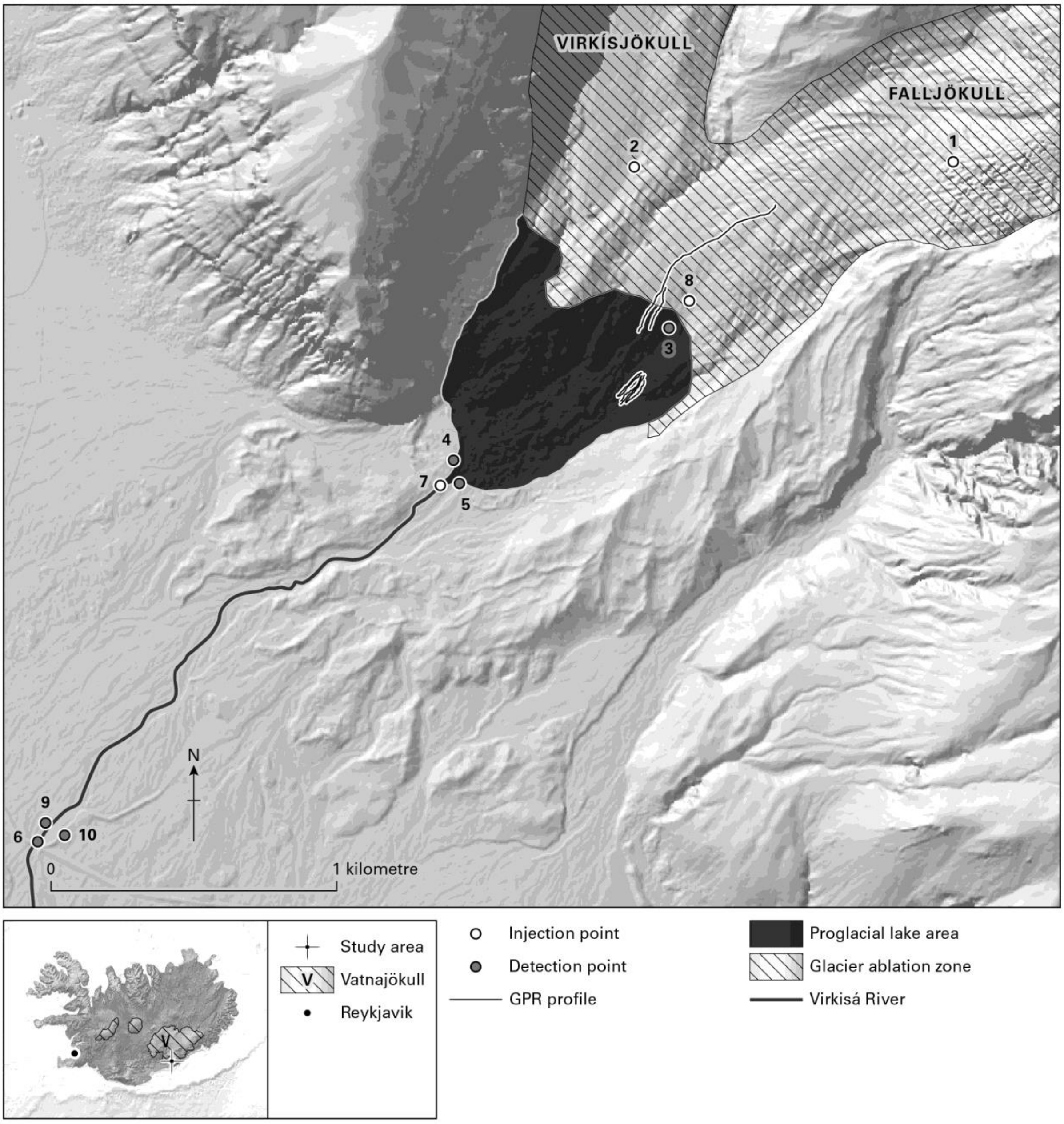

Figure 1: The lower part of the Virkísjökull catchment including the lower ablation zone and the pro-glacial lake area and outlet river. Points labelled on the map are (1) East arm injection moulin (2) west arm injection moulin May 2014 & August 2014 (3) Glacier snout outflow monitoring point (4) lake outlet west monitoring point (5) lake outlet east monitoring point (6) Proglacial river monitoring point (for proglacial foreland tracer test) point (7) Dilution gauging river injection point (8) proglacial foreland river sink injection (9) West bank dilution gauging monitoring point (10) East bank dilution gauging monitoring point.

Figure 2 The eastern arm glacier injection moulin in September 2013 (A) and April 2014 (B).

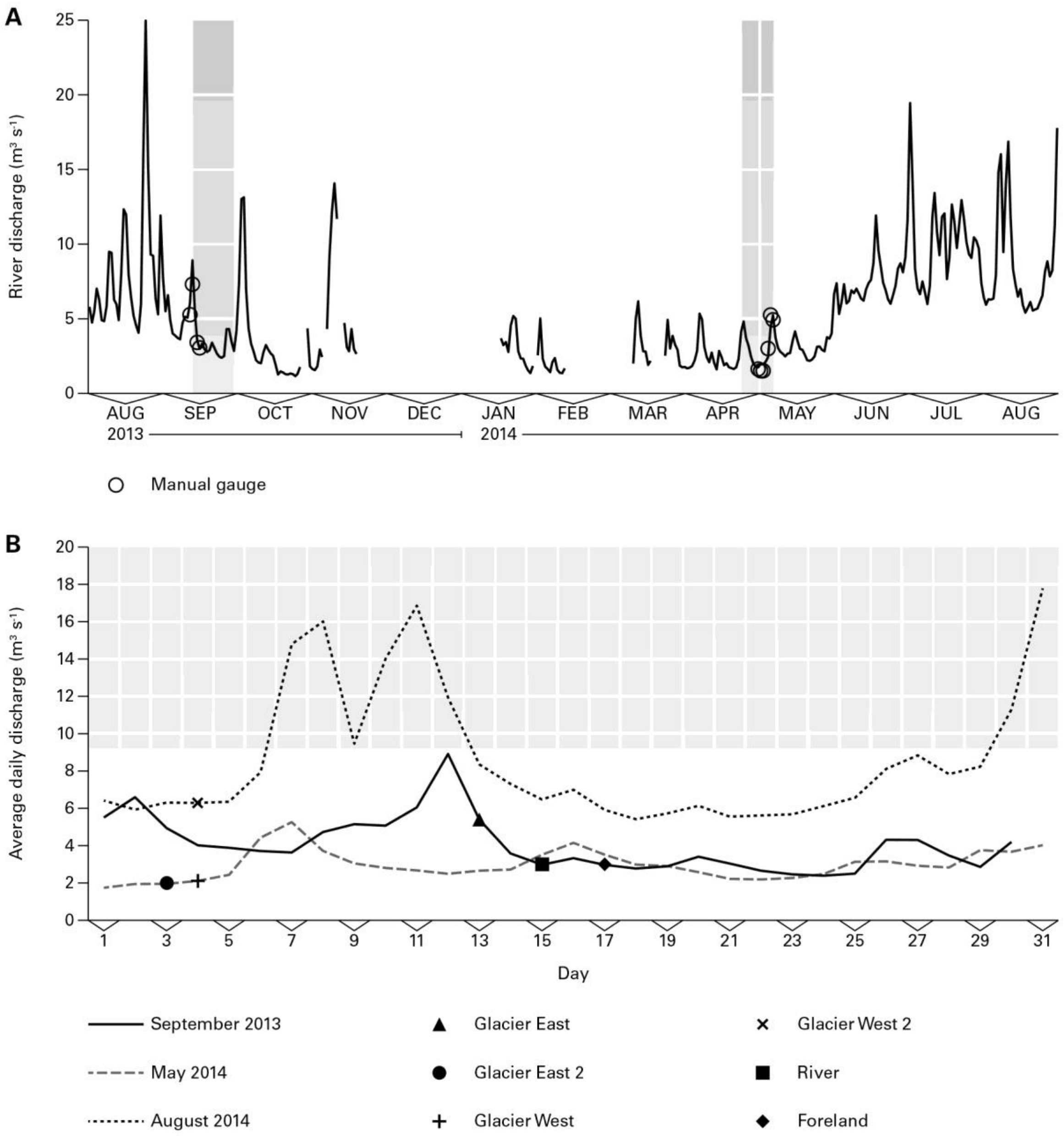

Figure 3: (a) River discharge record from August 2013 to August 2014 showing areas in grey when tracer tests were carried out and values for manual river gaugings. Gaps in the record are dates removed due to channel ice. (b) Detail of the river discharge measurements during the tracer tests.

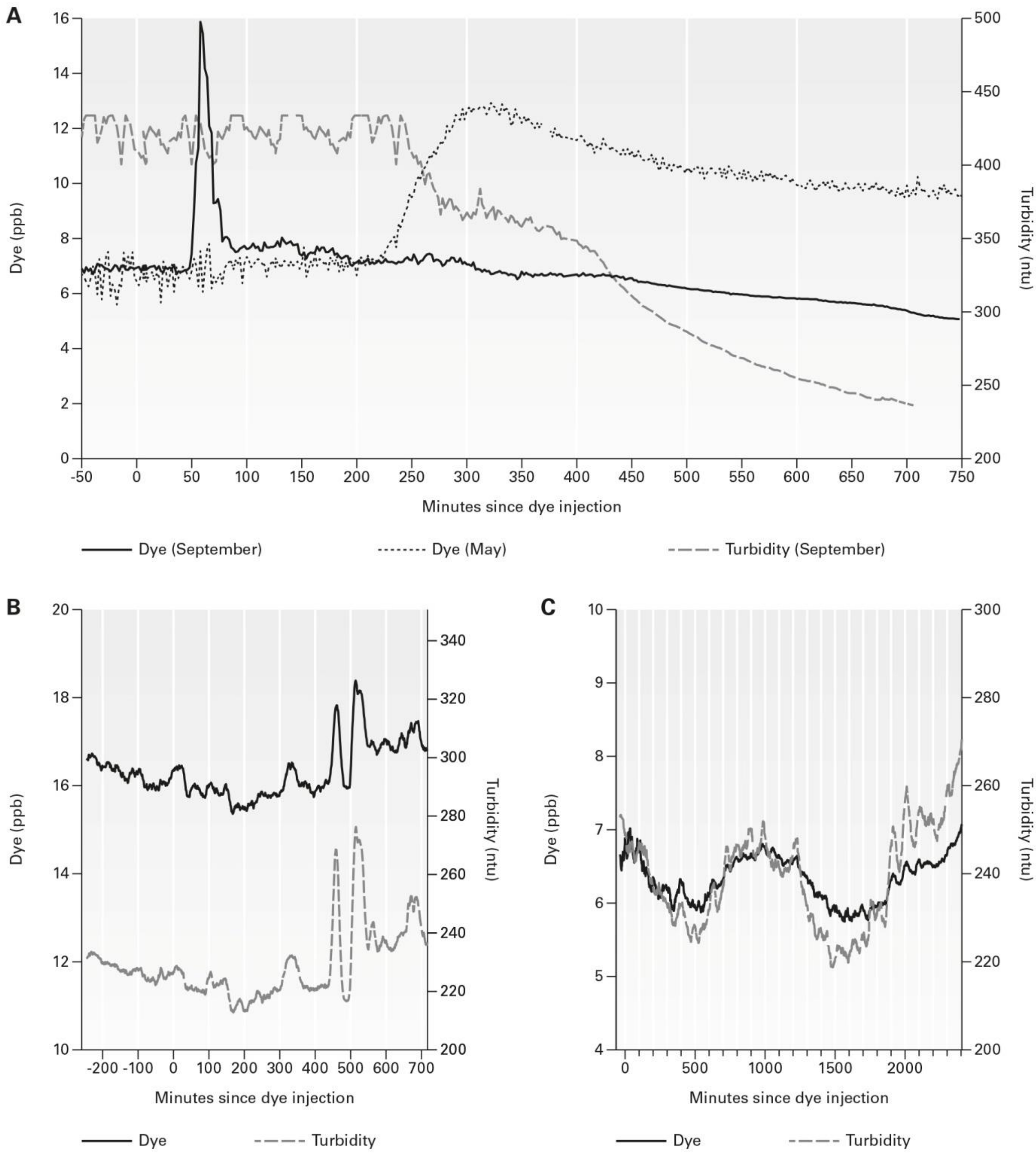

Figure 4: A. Tracer breakthrough curves during the glacier tracer tests on the eastern glacier arm in September 2013 and May 2014. B. results from the test undertaken from the moulin on the western glacier arm in May 2014 showing no breakthrough (fluctuations in florescence are a result of turbidity, shown in grey). C. results from the test undertaken from the moulin on the western glacier arm in August 2014 showing no breakthrough (fluctuations in florescence are the result of turbidity, shown in grey).

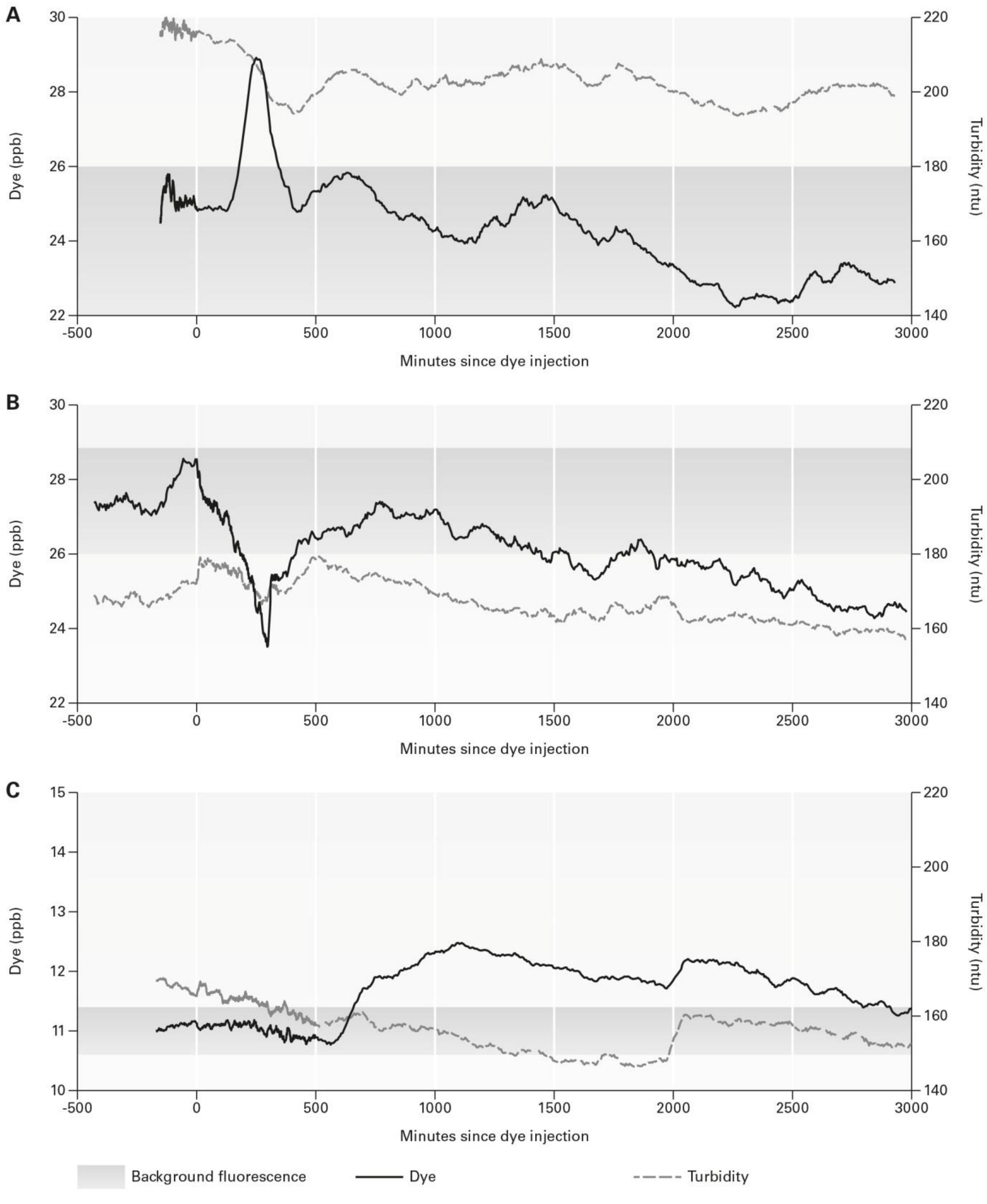

Figure 5: Rhodamine WT breakthrough curves during the proglacial tracer test. (a) at the Lake Outlet channel west (b) at the Lake Outlet channel east (c) at the downstream river gauging station. Background fluorescence is the signal naturally detected as a result of particles in the water

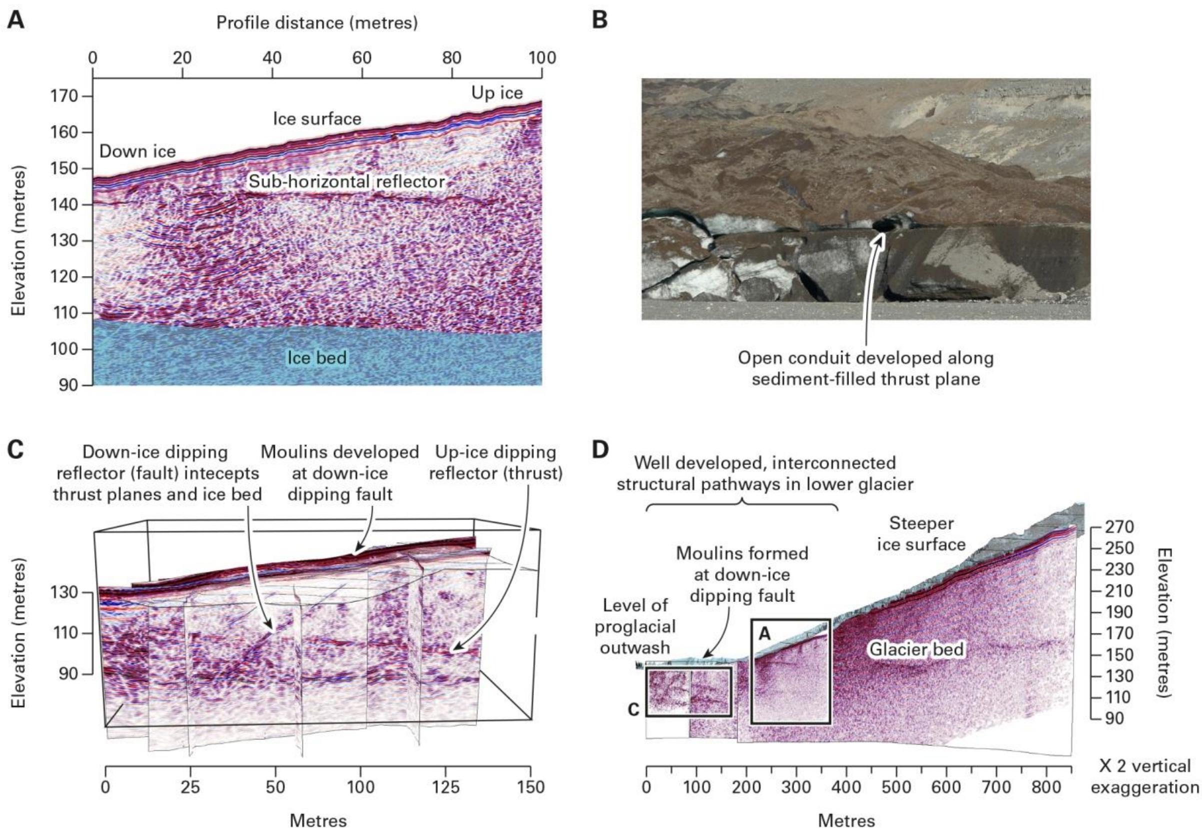

Figure 6. A. Continuous sub-horizontal reflector interpreted as thrust plane. B. Field photograph at debris covered ice margin showing a conduit developed within a sediment filled thrust plane. C. Enlarged composite image showing up-ice and down-ice dipping reflectors in the lower glacier, interpreted as thrust and fault planes. D. Composite of glacier GPR surveys showing a zone in the lower glacier with clear reflectors representing a well-developed englacial structural network

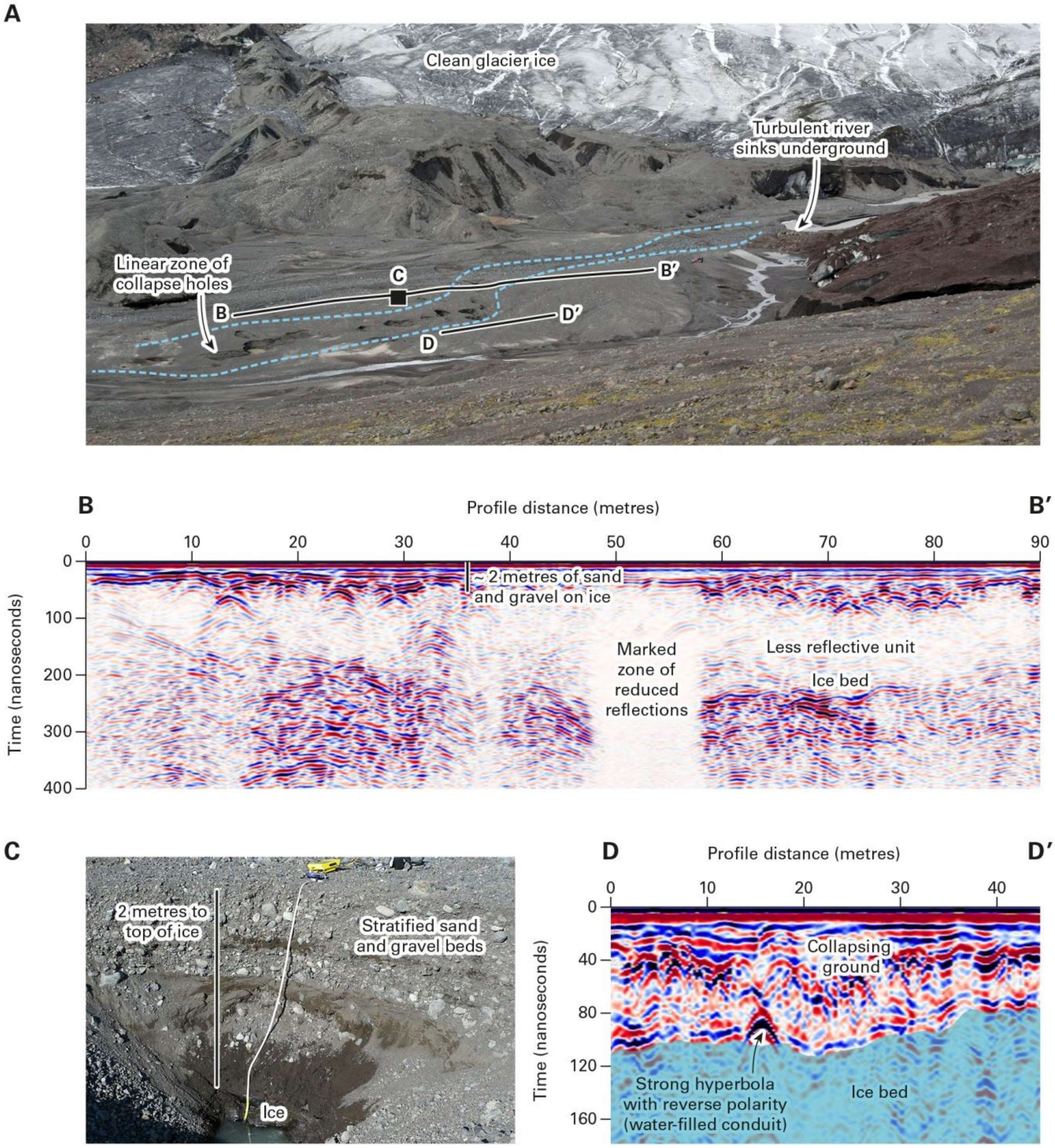

Figure 7.GPR profiles in the pro-glacial area. A. Photograph showing the location of GPR profiles B-B’ and D-D’, and observation point C. B. Un-migrated GPR profile across the proglacial area. The marked zone of reduced reflections coincides with the linear track of collapse features in the photograph in A. C. Field photograph showing ~2 m of stratified sand and gravel overlying buried ice. D. Un-migrated GPR profile showing a collapse structure and fill material (C.F) overlying a strong hyperbola interpreted as a water-filled conduit.

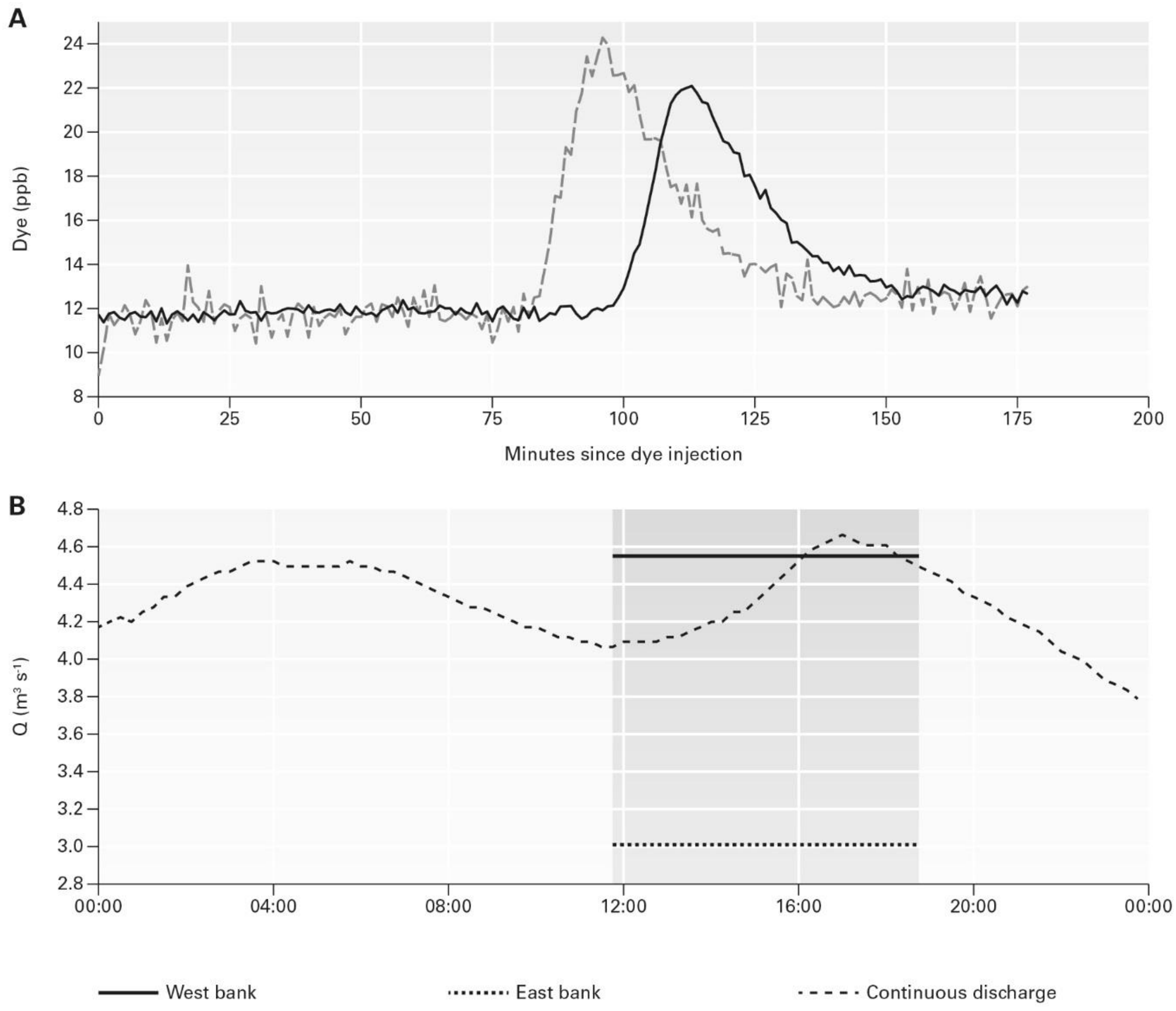

Figure 8: Proglacial river tracer test results (A) Tracer breakthrough curves (B) Comparison of flow measured at the gauging station with flow estimates from dilution gauging.

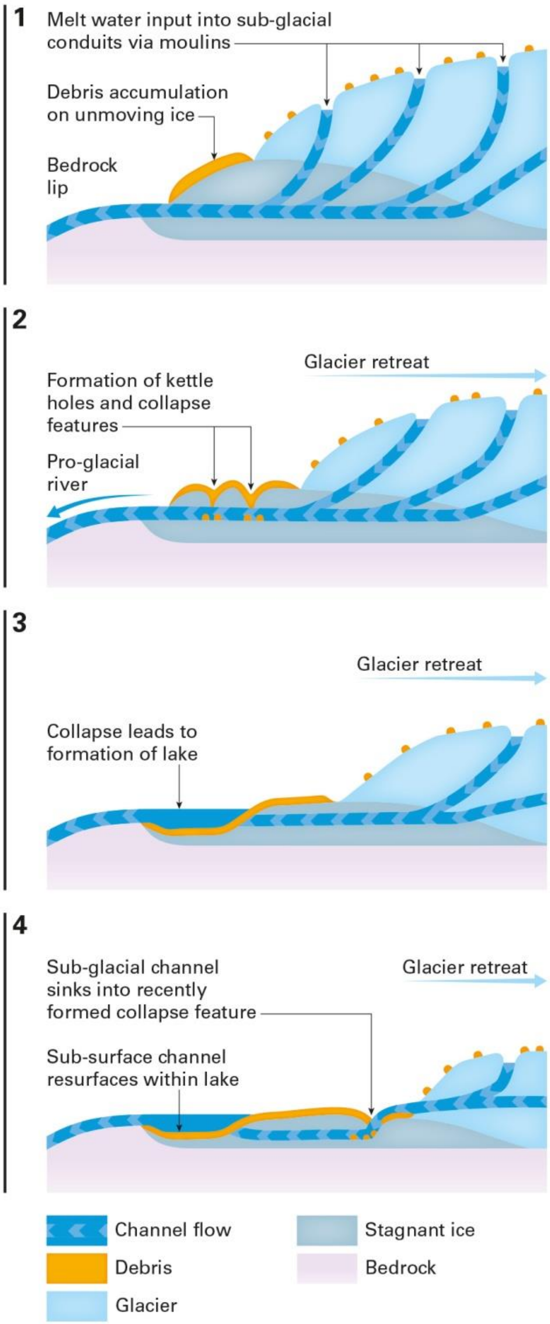

Figure 9: Diagrams showing conceptually how rapidly retreating glaciers produce a transitional environment and how they evolve and change. In the first stage of deglaciation the slowing of the glacier ice results in unmoving (stagnant) ice at the terminus of the glacier that is subsequently buried by the accumulation of debris transported by the active glacier margin. Meltwater is input into this system from moulins and conduits remain active in the stagnant ice (1). The remains of these active conduits within the buried ice will then begin to collapse to expose water moving through the proglacial buried-ice area (2). This process allows the formation of a pro-glacial lake that sits upon ice as the active glacier margin continues to retreat (3). The unmoving buried ice is insulated from rapid melting by the accumulation of debris. In the final stage (currently observed at Virkísjökull) the collapse of the active ice margin has exposed an en-glacial conduit. The meltwater, rather than flowing across the surface, exploits a collapse feature within the foreland to sink back into the conduit system within the buried ice to resurface within the newly formed pro-glacial lake system (4).