Erosional corridor evolution in south-central British Columbia: insights from ground-penetrating radar surveys

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ABSTRACT
On the southern Fraser Plateau a 40 km long and ~0.25-1 km wide erosional corridor, with long upslope sections, contains both hummocks and an esker. Ground-penetrating radar (GPR) data reveal that the erosional corridor is eroded into till and filled with two distinct broad deposits that are overlain by the esker. Slump blocks from the esker flank cap the fill. These data suggest five phases of erosional corridor development: 1) subglacial erosion of the partly ice-walled erosional corridor; 2) subglacial deposition of gravel sheets; 3) subglacial deposition of a corridor-wide gravel dune field; 4) esker deposition either on the waning stage of the event that infilled the erosional corridor as the broader flows focused into conduits, or during subsequent and unrelated events; and 5) ice retreat, removal of ice support and subsequent esker reworking. These data suggest the erosional corridor was filled by high magnitude, bankfull flow and hence operated as a tunnel channel.

RÉSUMÉ
La portion sud du Plateau Fraser contient un corridor d'érosion d'une largeur de ~0.25 -1 km et s'étendant sur une longueur de 40 km. Ce corridor contient des sections où la pente grimpe vers l'aval, ainsi que des zones de formes bosselées et un esker. Des relevés au géoradar révèlent que ce corridor est érodé dans le till et rempli de deux dépôts distincts qui sont recouverts par l'esker et des sédiments provenant de blocs faillés le long des flancs de cet esker. Ces données suggèrent cinq phases de développement des corridors d'érosion: 1) l'érosion sous-glaciaire du corridor dans le till et partiellement dans la glace avoisinante; 2) la déposition sous-glaciaire de nappes de gravier; 3) la déposition sous-glaciaire d'un champ de dunes de gravier sur toute la largeur du corridor; 4) la déposition de l'esker soit pendant la phase de déclin de l'écoulement qui a remblayé le corridor d'érosion lorsque l'écoulement s'est progressivement canalisé, ou lors d'événements ultérieurs sans relations aux écoulements qui ont formé le corridor; et 5) retrait de la glace, perte du soutien latéral de la glace et le remaniement ultérieur de l'esker. Ces données suggèrent que le couloir d'érosion a été formé par des écoulements de grande ampleur remplissant complètement le conduit. Ce corridor est donc une vallée tunnel.

1 INTRODUCTION

Tunnel valleys are the erosive expression of channelized subglacial meltwater flows that are common on the beds of former ice sheets in North America and Europe (e.g., Wright, 1973; Piotrowski, 1994; Brennand and Shaw, 1994; Bradwell et al., 2008). Tunnel valleys are elongate troughs cut into sediment or bedrock that may be up to 100 km long and 4 km wide (Ó Cofaigh, 1996; Russell et al., 2007), occurring individually or as anabranched or dendritic networks (Ó Cofaigh, 1996). They often truncate subglacial bedforms and/or till, have long upslope paths, and contain landforms such as eskers and gravel dunes (e.g., Shaw and Gorrell, 1991; Brennand and Shaw, 1994; Ó Cofaigh, 1996; Fisher et al., 2005). Tunnel valley walls may be discontinuous and so partly ice-walled (e.g., Shaw 1983).

Although it is largely accepted that tunnel valleys are the expression of channelized subglacial meltwater erosion, there is much controversy surrounding the magnitude and frequency of formative flows (Ó Cofaigh, 1996). Broadly speaking, there are four main hypotheses of tunnel valley formation:

1) Headward tunnel valley growth by steady-state subglacial sediment deformation into a laterally migrating R-channel that was initiated by piping (e.g., Boulton and Hindmarsh, 1987). Squeezing of the substrate into the R-channel, where it was removed by glaciofluvial erosion, allowed the formation of tunnel valleys over a prolonged period of time.
2) Synchronous occupation and incision of the entire tunnel valley system by catastrophic subglacial outburst floods (e.g., Wright, 1973; Shaw and Gorrell, 1991; Brennand and Shaw, 1994; Sjogren et al., 2002; Russell et al., 2007).
3) Time-transgressive tunnel valley development via catastrophic drainage of subglacial lakes close to a retreating ice margin (Wingfield, 1990). Channel segments were eroded in a headward fashion by plunge pool incision and tunnel valleys formed time-

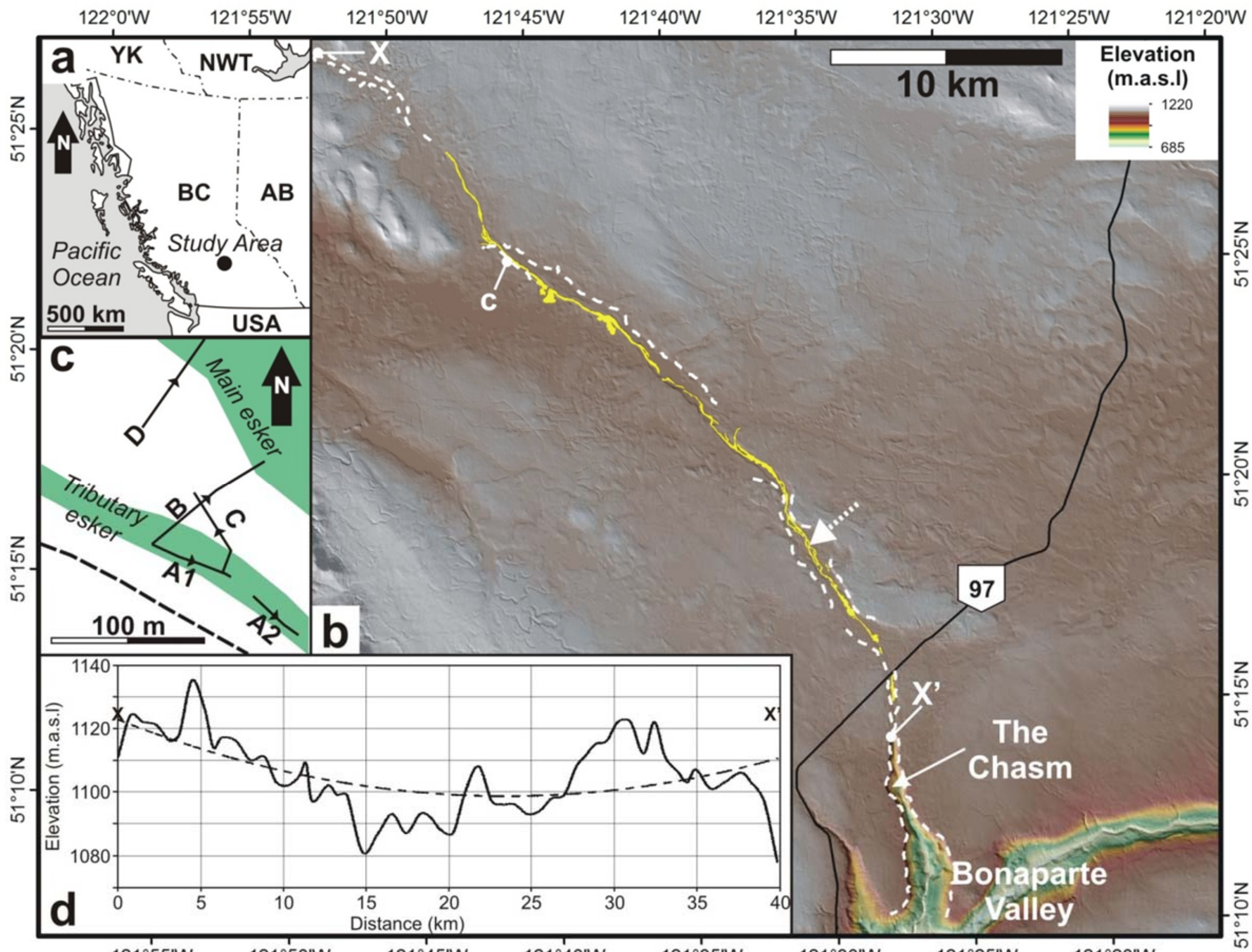

Figure 1. a) The location of the study area (Fraser Plateau) is indicated by the labelled black dot in British Columbia. b) A hillshaded elevation model of the southern Fraser Plateau (Geobase®). The dashed white line highlights the location of the erosional corridor; the yellow line is the main esker. The arrow labelled "c" indicates the location of the GPR grid shown in c) and dashed arrow shows the location of hummocks similar to those at c). X and X' correspond to up- and down-flow ends of the elevation profile presented in d). c) Map of the GPR grid (arrowed lines). The green areas highlight the relative positions of the main and tributary eskers. The dashed line is the wall of the erosional corridor. d) Topographic variation along the erosional corridor axis (solid line). The stippled line is a first order polynomial trend.

transgressively with ice retreat. Each segment was eroded by a separate outburst flood.

4) Seasonal formation of tunnel valley segments close to the ice margin from water supplied by supraglacial melting (Mooers, 1989). Ice retreat and continued segment erosion formed tunnel valleys time-transgressively.

Such varied hypotheses have led to strict terminological usage. Valleys that are inferred to have formed by bankfull flow are referred to as tunnel channels (e.g., Brennand and Shaw, 1994; Fisher et al., 2005), whilst theories that invoke less than bankfull flow refer to them as tunnel valleys (e.g., Wright, 1973; Mooers, 1989). More recently, glaciofluvial (or erosional) corridors that do not have continuous valley walls have been identified (e.g., Rampton, 2000; Utting et al., 2009).

As interpretations of tunnel valley formation aid reconstruction of Late Wisconsinan ice sheet hydrology and dynamics, it is important to fully understand the mechanisms responsible for their development (Brennand et al., 2006). Significant insight into the meltwater regime that flowed through a tunnel valley can be gained from analysis of valley fill (e.g., Shaw and Gorrell, 1991; Piotrowski, 1994; Russell et al., 2003; Fisher et al., 2005; Jørgensen and Sandersen, 2008). The sedimentary structure and architecture of tunnel valley fills have been described on the basis of their geomorphology and from exposures (e.g., Shaw and Gorrell, 1991) and geophysics (e.g., Pugin et al., 1999; Fisher et al., 2005). Tunnel valleys are often filled, partially or completely, with sand and gravel that is typically arranged into bedforms or hummocks (e.g., Shaw and Gorrell, 1991; Piotrowski, 1994; Russell et al., 2003; Fisher et al., 2005; Utting et al., 2009) and tabular sheets (Pugin et al., 1999; Russell et al., 2003) that record high magnitude flow. However, although tunnel valleys/channels in Canada are often inferred to have developed during single events, evidence from some European tunnel valleys suggest palimpsest development over multiple glaciations (e.g., Jørgensen and Sandersen, 2008).

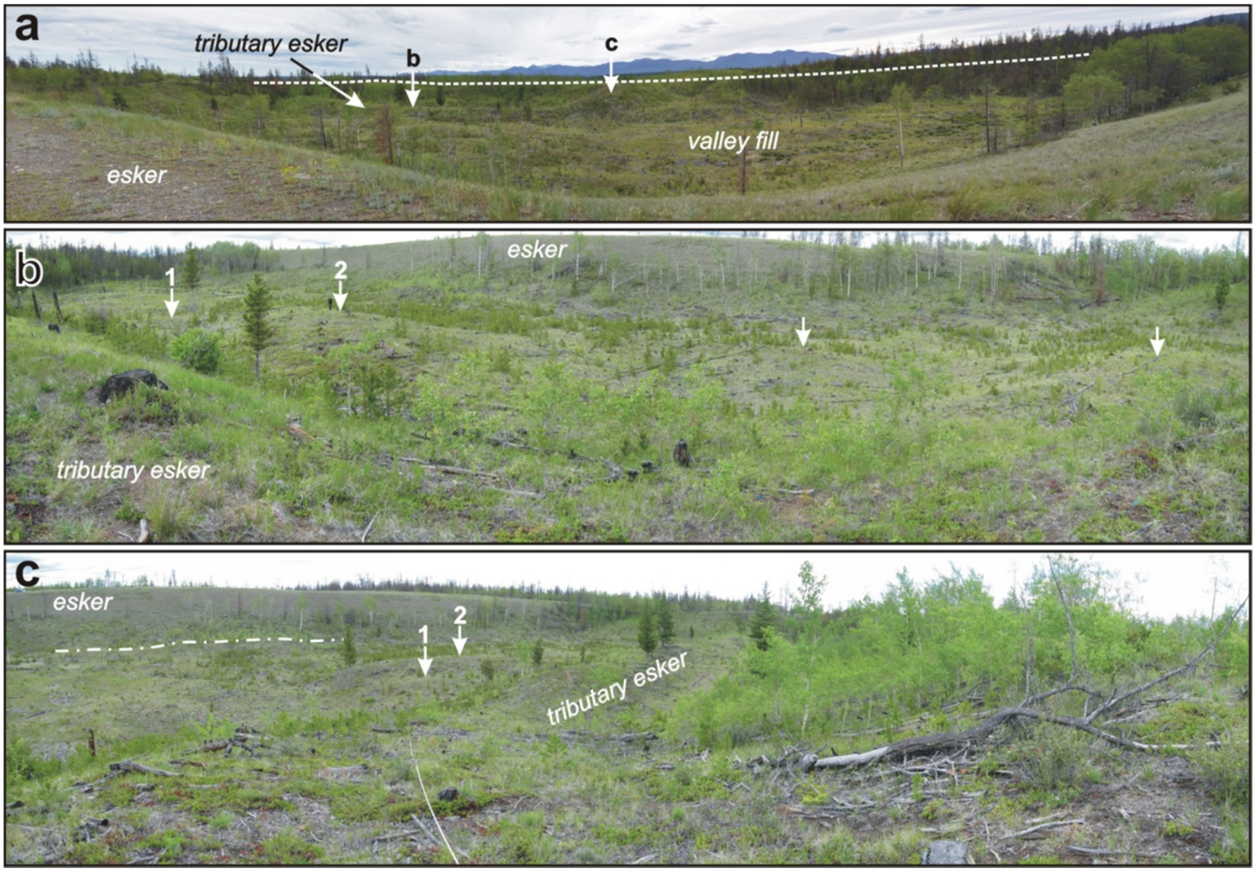

Figure 2. Landforms within the erosional corridor at the study site: a) A large main esker (esker) dominates the corridor fill. A small tributary esker appears to drape hummocks on the corridor floor. The erosional corridor wall is highlighted by the dashed white line. Photograph taken from the main esker crest looking southwest. The arrows labelled "b" and "c" indicate the positions from which the photographs in b) and c) were taken, respectively; b) view of the corridor fill taken from the tributary esker and looking northeast. Some of the hummocks are arrowed, with those labelled 1 and 2 corresponding to the same hummocks labelled in c); c) view of the eskers and corridor floor from the erosional corridor wall, looking southeast. The labelled arrows indicate stacked hummocks; the dot-dashed line highlights a linear ridge at the base of the main esker.

To-date most Canadian work has focused on Laurentide Ice Sheet tunnel valleys (e.g., Sjogren et al., 2002; Russell et al., 2003). We have limited knowledge of tunnel valleys from the Cordilleran Ice Sheet (e.g., Booth and Hallet, 1993; Booth, 1994; Lesemann and Brennand, 2009). In this paper we utilize ground-penetrating radar (GPR) to investigate landform relationships within an erosional corridor (tunnel valley) on the Fraser Plateau. We use the term erosional corridor as the landform does not have continuous valley walls. This work gives a significant insight into the nature of erosional corridor evolution in south-central British Columbia, improving our understanding of the processes operating beneath the Cordilleran Ice Sheet.

2 DEPOSITIONAL SETTING

Glaciation of the Fraser Plateau began around 30.5 cal ka BP (Booth et al., 2003) and reached its maximum at around 14-13 cal ka BP when the southern plateau was covered by 600-1100 m of ice (Huntley and Broster, 1994). Rapid ice sheet decay was complete by 10.7 cal ka BP (Clague and James, 2002), with Fulton (1991) ascribing the lack of large recessional moraines (c.f. Tipper, 1971) to stagnation and downwasting.

On the Fraser Plateau (Figure 1), whilst large moraines are absent, there are a number of meltwater landforms that can give insight into the dynamics and hydrology of the decaying Cordilleran Ice Sheet. A ~40 km long, ~0.25-1 km wide erosional corridor that begins abruptly at a ~0.5 km wide bedrock basin, feeds a large bedrock cataract that has been eroded headward from the Bonaparte valley (The Chasm, Figure 1b). As this erosional corridor is only slightly sinuous (sinuosity value of 1.05, as defined by the ratio between actual channel length and straight line valley length), has an undulating long profile with long up-slope sections (Figure 1d), is cut through till and bedrock, and contains an esker (Figure 1b), we interpret it to have formed subglacially. The esker within the erosional corridor is ~30 km long and has a varied morphology that ranges from flat topped, to round crested and anabranched. At ~15 km from the head of the erosional corridor, a recent clear cut on the valley floor reveals a number of hummocks that have crestlines strongly oblique (though variably orientated) to the corridor axis. These hummocks appear to be draped by the esker (Figure 2). These hummocks are located within a widening of the southwestern wall of the erosional

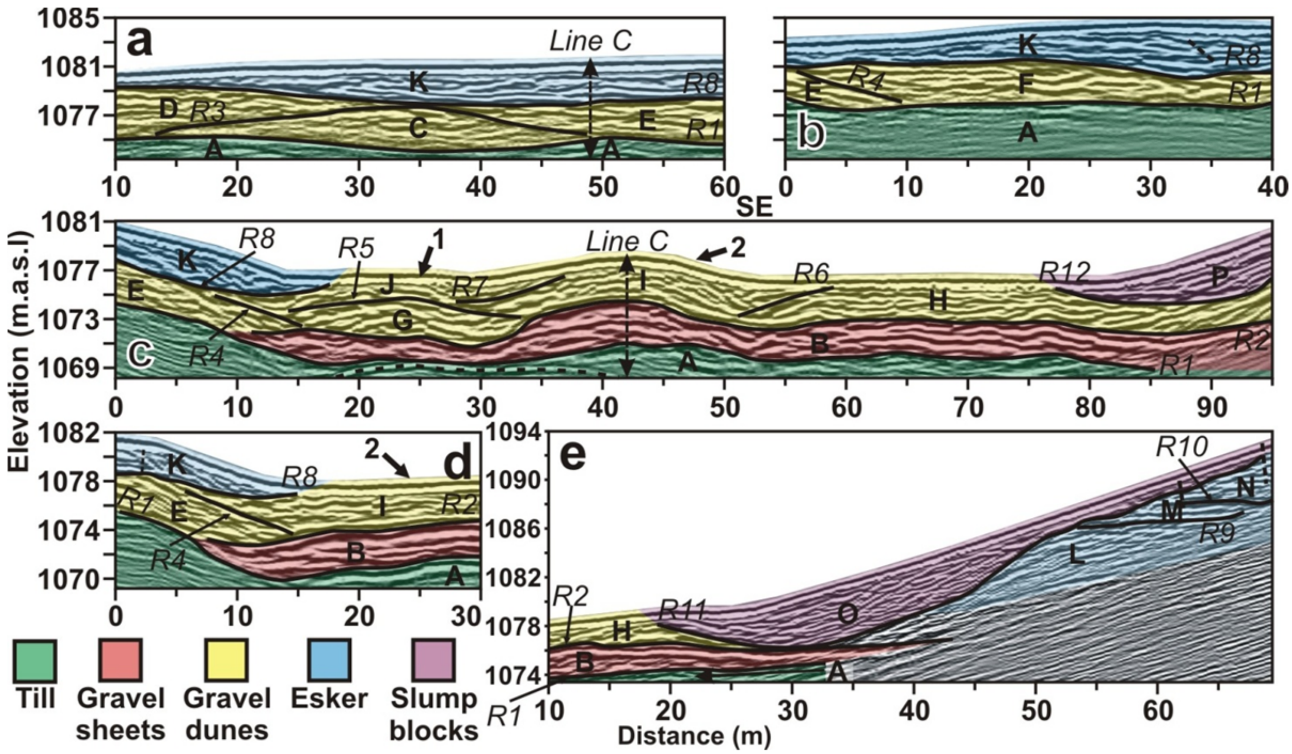

Figure 3. Processed GPR profiles: a) line A1; b) line A2; c) line B; d) line C; e) line D. See Figure 1 for locations. Radar bounding surfaces are highlighted by the bold lines and are labelled R1–R12. These bounding surfaces define twelve radar elements labelled A–P. The locations where line C intersects line A1 and line B are shown by the labelled double arrows. The stippled line in c) highlights a diffraction (from a surface reflector) that has been distorted by topographic correction during data processing. Dashed lines mark offset reflections. The arrows labelled 1 and 2 highlight the radar elements that correspond to the same labelled hummocks in Figure 2.

corridor. Similar hummocks can be identified on aerial photographs within a further erosional corridor widening ~18 km down-flow (Figure 1). Here we utilize GPR to investigate landform and sediment relationships within the erosional corridor.

3 METHODOLOGY

A total of 330 m of 100 MHz common offset (CO) GPR lines (Figure 1c) were collected as an irregular grid within a clear cut on the floor of the erosional corridor using a Sensors & Software Inc. pulseEKKO PRO system. The presence of tree stumps and slash within the clear cut prevented collection of regularly spaced lines. During CO data collection, the GPR antennas were co-polarised and perpendicular-broadside to the survey line, to reduce reflections from offline sources (Arcone et al., 1995). Antennas were kept at a constant separation of 1 m and data were collected in step mode (0.25 m) along the lines to improve ground coupling and trace stacking (32 traces). Two common mid-point (CMP) profiles, collected on the main esker, provide an estimated subsurface average velocity of 0.121 ± 0.008 m/ns, which was used to convert two-way travel time (TWT) into depth and for data processing. All lines were surveyed for topographic variation using a real-time kinematic differential global positioning system (dGPS). GPR data processing was carried out in REFLEXW v5.6 (Sandmeier, 2010) and included time-zero correction, 'dewow' filtering, bandpass filtering, migration, background removal filtering, application of a gain function, and topographic correction.

4 RESULTS

4.1 Description of Radar Elements

Radar bounding surfaces (R1-R12, Figure 3) are high amplitude, continuous reflections that can be traced across multiple GPR profiles and demarcate the termination of reflections above and below. These bounding surfaces define sixteen radar elements (labelled A–P in Figure 3 and referred to as RE-A to RE-P in the text).

4.1.1 RE-A

RE-A can be identified throughout most of the grid, though no lower bounding surface is imaged due to rapid signal attenuation (Figure 3). RE-A is largely reflection free, though some weak, discontinuous, chaotic reflections can be identified in places (Figure 3c–e). The upper bounding surface of RE-A, R1, can be traced throughout the grid and has an irregular geometry and often truncates reflections within RE-A (e.g., at ~47 m and ~79 m on line B, Figure 3c).

4.1.2 RE-B and RE-C to RE-J

R1 is primarily draped by moderately continuous, subhorizontal reflections within the sheet-like RE-B. However, at the south-western end of the grid (lines B and C) there is a step (or terrace) in the geometry of R1 (at 10-17 m on line B, and at ~7-15 m on line C, Figure 3).

Here, reflections within RE-B onlap R1. The upper bounding surface of RE-B, R2, has an irregular geometry that is subparallel to the ground surface and can be identified throughout the grid. R2 truncates reflections within RE-B (Figure 3c–e).

Eight lenticular, convex-up, cross-cutting radar elements (RE-C to RE-J) are located above RE-B. These are draped onto R2 throughout most of the grid, except at the south-western side (e.g., Figure 3a–d) where RE-B is not identified. Here the elements are draped onto R1. Individually, these cross-cutting radar elements form hummocks that are up to 4 m high and 38 m long (as imaged by the GPR: e.g., RE-C, Figure 3a) and in some cases (RE-I and RE-J, Figure 3c) conform to hummocks on the ground surface (1 and 2 in Figure 2b–c). Reflections within the cross-cutting hummocks are moderately continuous, but have an irregular geometry. However, on line B (Figure 3c) RE-G, RE-I and RE-J show some similarity. Reflections are slightly arched normal to the lower bounding surface, but on the eastern side of the hummocks, reflections dip northeast at *c.* 15˚ from horizontal. In a profile that crosses perpendicularly to line B (line C, Figure 3d) the dipping reflections within RE-I are subparallel to the lower bounding surface.

4.1.3 RE-K to RE-N

Six further radar elements (RE-K to RE-N) are imaged above the cross-cutting hummocks and below R11. R8 is draped by moderately continuous and subparallel reflections within RE-K, the extent of which forms a small tributary esker (Figure 2). Some offset reflections are imaged (short dashed lines, Figures 3b and d). RE-L, RE-M and RE-N are lenticular radar elements within the main esker (Figure 3e), though signal attenuation prevents observation of the esker-hummock contact. RE-L to RE-N are composed of continuous, sub-parallel reflections, though reflection offsets are imaged in places (short dashed lines, Figure 3e).

4.1.4 RE-O and RE-P

R11 truncates reflections within RE-L to RE-N and is the lower bounding surface of RE-O. R12 is the lower bounding surface of RE-P that caps RE-H (a hummock). RE-O forms an asymmetrical trough that increases in thickness from offset reflections at the esker crest to the base of the esker, where it conforms to a small linear ridge that is parallel to the esker flank (Figure 2c). RE-O is composed of chaotic reflections and diffractions (Figure 3c). RE-P is similar in geometry and composition to RE-O, though the GPR profile is too short to fully image the RE.

4.2 Interpretation of Radar Elements

4.2.1 Till (RE-A)

The deepest bounding surface imaged is R1, which because it is a significant attenuator, may be interpreted as an interface between a resistive aggregate and bedrock, or a resistive aggregate and till. The expected contrast in dielectric properties between a resistive aggregate and bedrock would be less than between a resistive aggregate and till (see table 2, Neal, 2004). Thus, R1 is more consistent with a resistive aggregate-till interface because it is a strong reflection, the signal rapidly attenuates below it, and RE-A is largely reflection free (e.g., Beres and Haeni, 1991; Ékes and Hickin, 2001).

4.2.2 Corridor Fill Elements (RE-B to RE-J)

The good signal penetration above R1 in the corridor fill suggests the presence of low conductivity material. There were no sediment exposures in the corridor floor, but a small section through the main esker reveals it is composed of sand and gravel. Because there is no distinct change in signal return between the eskers and corridor floor, and small cobbles are present at surface on the corridor floor, the up to 9 m thick material above R1 is likely composed of sand and gravel. The corridor fill elements can be grouped into two distinct element associations: 1) gravel sheets; 2) gravel dunes.

RE-B is draped onto R1 and is the initial deposit within the erosional corridor. Reflections within RE-B are for the most part subparallel to R1 and interpreted as plane beds similar to those identified in other GPR surveys (e.g., Burke et al., 2008, 2010; Heinz and Aigner, 2003; Rice et al., 2009). As these plane beds are extensive, they are likely broad gravel sheets indicative of high velocity flow and wide, continuous deposition (Heinz and Aigner, 2003; Rice et al., 2009). Similar gravel sheets have been identified in tunnel valleys beneath the Oak Ridges Moraine in Ontario (e.g., Russell et al., 2003). However, RE-B cannot be identified throughout the grid, suggesting that gravel sheet deposition may not have occurred across the full width of the corridor, similar to observations from tunnel valleys in southern Ontario (cf. Russell et al., 2003).

As R2 truncates some reflections within RE-B it is interpreted as an erosion surface. Subsequent corridor fill is recorded by a number of stacked hummocks in the GPR profiles, some of which conform to morphological hummocks in the ground surface (Figure 2a–b). Internally these hummocks have irregular reflection geometries, but where GPR profiles are close to perpendicular to the hummock crestline (e.g., RE-G, RE-I and RE-J, Figure 3c) stacked, dipping reflections are identified. These dipping reflections are interpreted as foreset beds that indicate progradation of asymmetric dunes, similar to those identified in tunnel valleys elsewhere (e.g., Shaw and Gorrell, 1991; Piotrowski, 1994; Pugin et al., 1999; Russell et al., 2003; Fisher et al., 2005). The morphology of (Figure 2a–b) and variation in reflection geometry within (Figure 3) these dunes indicate they are complex 3-D forms. Where GPR profiles are subparallel to local flow direction, foreset beds are clearly imaged. However, if the GPR profile is oblique to flow direction forests become increasingly unclear as line deviation from flow direction increases. The inferred local flow direction appears to be oblique to the erosional corridor wall, which is likely due to the development of flow vortices immediately down-flow of the corridor widening.

Based on the extent of RE-I, which can be directly linked to a surface expression of the bedform (2, Figure 2b-c), the gravel dunes are up to ~4 m high, have stoss and lee side lengths of ~15 m and ~10 m respectively, are ~60 m wide, and have a wavelength of ~25 m. These hummocks are of a similar scale to those identified in

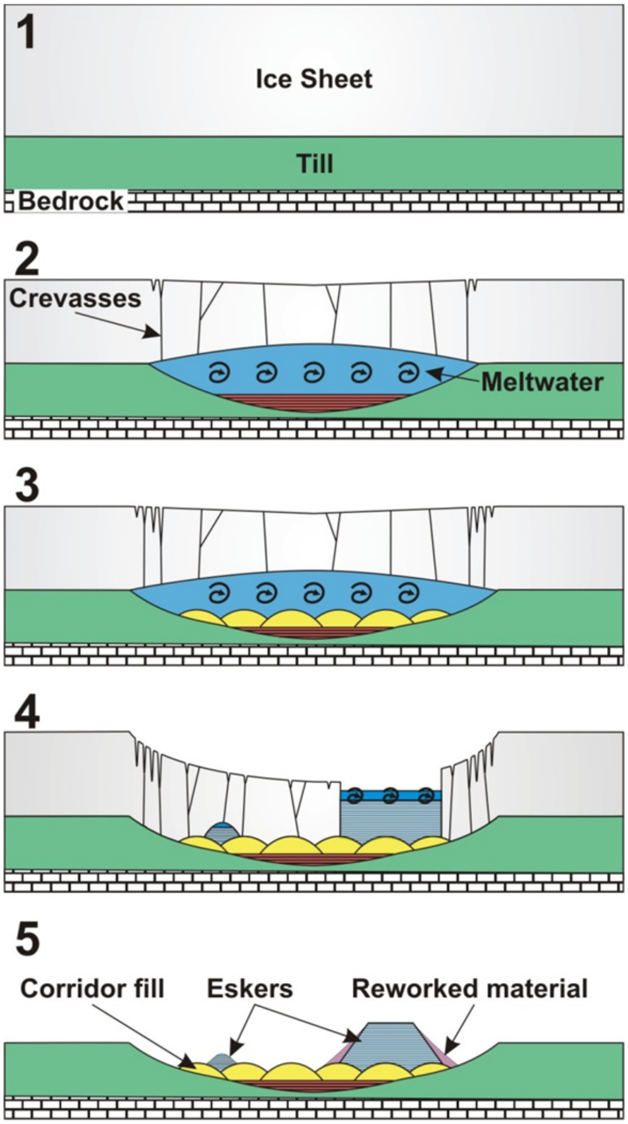

Figure 4. Cartoon depicting erosional corridor evolution on the southern Fraser Plateau. Following erosion of the channel (1), gravel sheets (2), dunes (3), and eskers (4) were deposited. Ice sheet decay removed lateral support from esker sediment and resulted in esker flank collapse (5). Curled arrows denote turbulent flow. See text for full description.

erosional corridors elsewhere (e.g., Utting et al., 2009). If subaerial bedform wavelength and height to flow depth scaling relationships can be applied to those identified in the erosional corridor (wavelength to depth ratio of 5:1 and dune height to flow depth ratio of 1:6; cf., Allen, 1984), a flow depth of ~5-20 m can be estimated. Consequently, flow depth would be greater than the erosional corridor wall height along most of its length, suggesting flow was contained by an ice roof.

As dune morphology appears intact (Figure 2) there must have been little, if any, waning stage reworking of the corridor fill. This, and the lack of a fine carapace, point to a rapid flow waning stage following dune deposition. In addition, the undisturbed internal architecture of the dunes suggest the sediment deformation expected from ice bed recoupling did not take place (e.g., Brennand, 1994).

4.2.3 Esker Elements (RE-K to RE-N)

Draped onto the dunes is RE-K, which corresponds to the tributary esker. Whether dunes extend below the main esker (RE-L to RE-N) is unclear. The limited extent of post-depositional reworking within the eskers suggests they were not deposited englacially or supraglacially (cf. Brennand, 1994, 2000). However, the main esker has a flat-topped morphology adjacent to the study site, consistent with deposition within a subaerial, unroofed, ice-walled canyon (e.g., Russell et al., 2001; Burke et al., 2008). In contrast, the smaller tributary esker has a more rounded crest that would suggest R-channel deposition.

4.2.4 Postglacial Elements (RE-O and RE-P)

The presence of offset reflections within the eskers indicates faulting due to removal of ice support (cf. Fiore et al. 2002; Woodward et al. 2008; Burke et al. 2008, 2010). The stratigraphically uppermost radar elements within the corridor fill are asymmetrical troughs on the esker flank (RE-O and RE-P). We interpret these as slumped blocks of esker material due to removal of ice support because: 1) the head of RE-O corresponds directly to faults within the esker; 2) RE-O and RE-P are composed of chaotic reflections indicating poorly sorted or disturbed material; and 3) RE-O and RE-P form parallel linear ridges at the base of the esker flank consistent with slump blocks (Figure 2b).

5 DISCUSSION

The large bedrock cataract at the down-flow end of the erosional corridor is eroded into bedrock and so probably generated over multiple glaciations (cf. Brennand and Shaw 1994). This topographic low would have acted as a low pressure sink at the ice sheet margin or subglacially and so focused regional meltwater flow. However, the up-flow erosional corridor is eroded into till deposited during the last glaciation (cf. Plouffe et al., 2010) (stage 1, Figure 4). Consequently, where the current erosional corridor is directly eroded into till, corridor genesis must have taken place during the last glaciation as new meltwater routes towards the Chasm were established. However, the slightly sinuous, single-thread erosional corridor planform, coupled with an abrupt channel head, indicates waters likely originated from a point source (most likely a supraglacial lake) rather than a distributed network.

Although erosional corridor incision into till may have taken place via a single or multiple events, the character of the corridor fill suggest that it was at least filled by high magnitude flow, similar to that ascribed to tunnel channels elsewhere (e.g., Shaw and Gorrell, 1991; Brennand and Shaw, 1994; Sjogren et al., 2002; Russell et al., 2003; Fisher et al., 2005). As this fill is stratigraphically between the eskers and till, and within an erosional corridor that has long upslope sections, we interpret the full sequence to have been deposited subglacially, consistent with other tunnel channels and erosional corridors (Shaw and Gorrell, 1991; Fisher et al., 2005; Utting et al., 2009).

The deepest deposits within the corridor fill, gravel sheets, suggest high velocity flow across much of the corridor width (stage 2, Figure 4). This high magnitude flow would have resulted in ice surface collapse and crevassing along the flow path (Björnsson, 2002). Later, most likely during the same event, the corridor was widened slightly at the margins and the gravel sheets were partially eroded before the large gravel dunes were deposited, the scale of which suggest bankfull flow and that the erosional corridor operated as a tunnel channel (stage 3, Figure 4). The transition from gravel sheets to dunes suggests a reduction in discharge, stream power, and/or an increase in flow depth.

The lack of dune reworking and fine-grained deposits capping dunes, point to a rapid flow waning stage after dune deposition. Following the corridor-wide flow that deposited the dunes, discharge was rapidly focused into an R-channel and a subaerial, unroofed, ice-walled canyon preventing corridor fill reworking or burial in finer sediment (stage 4, Figure 4). Indeed, such collapse of broad flows into esker forming conduits has been documented from contemporary outburst floods (cf. Burke et al., 2008, 2010) and suggested for tunnel valley-esker relationships elsewhere (e.g., Shaw, 1983). However, the flat-topped main esker morphology at this site suggests esker deposition within a subaerial, ice-walled canyon, indicating the ice was thin enough to allow at least partial conduit unroofing. Similar local collapse of channel roofs has been identified during Icelandic jökulhlaups (Björnsson, 2002).

The lack of corridor fill deformation, which is thought to accompany ice bed recoupling following collapse of broad flows (e.g., Brennand, 1994), further supports the presence of thin ice. However, although this reasonable hypothesis suggests a single event for at least dune and esker deposition, we acknowledge that the channel fill and esker deposition may have taken place during unrelated events. Following removal of ice support the esker flanks collapsed through both normal faulting and slumping (stage 5, Figure 4).

6 CONCLUSIONS

A relatively low sinuosity erosional corridor that dissects the Fraser Plateau in south-central British Columbia is shown to have formed subglacially as a tunnel channel. Although the erosional corridor is directly related to down-flow erosion of a large bedrock cataract, presumably over multiple glaciations, GPR data reveal that the erosional corridor upflow is eroded into till and so was likely generated during the last glaciation. This contrasts with development of buried tunnel valleys over multiple glaciations in Europe (e.g., Piotrowski, 1994; Jørgensen and Sandersen, 2008). Erosional corridor fill indicates a relatively simple event sequence, consistent with high magnitude flow and bankfull conditions, as suggested for tunnel channels elsewhere in Canada (e.g., Shaw and Gorrell, 1991; Brennand and Shaw, 1994; Sjogren et al., 2002). Initial deposition of gravel sheets represents the highest magnitude flow, whilst the stratigraphically higher gravel dunes indicate a reduction in flow magnitude and/or an increase in flow depth. The lack of waning stage corridor fill reworking or burial by fines suggests abrupt termination of corridor-wide flow. Instead, waning flow may have been focused into esker-forming conduits. Whilst current findings point to a single event sequence, detailed investigation of esker morphology and sedimentary architecture may better elucidate the nature of esker genesis and the relationship to erosional corridor formation.

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REFERENCES

Allen, J.R.L. 1984. *Sedimentary Structures Their Character and Physical Basis*, Elsevier, New York, NY, USA.

Arcone, S.A., Lawson, D.E. and Delaney, A.J. 1995. Short-pulse radar wavelet recovery and resolution of dielectric contrasts within englacial and basal ice of Matanuska Glacier, Alaska, U.S.A., *Journal of Glaciology*, 41(137): 68-86.

Beres, M. and Haeni, F.P. 1991. Application of ground-penetrating radar methods in hydrogeologic studies, *Ground Water*, 29(3): 375-386.

Björnsson, H. 2002. Subglacial lakes and jökulhaups in Iceland, *Global and Planetary Change*, 35: 255-271.

Booth, D.B. 1994. Glaciofluvial infilling and scour of the Puget Lowland, Washington, during ice sheet glaciations, *Geology*, 22: 695-698.

Booth, D.B. and Hallet, B. 1993. Channel networks carved by subglacial water: observations and reconstruction in the Puget lowland of Washington, *Geological Society of America Bulletin*, 105: 671-683.

Booth, D.B., Troost, K.G. Clague, J.J. and Waitt, R.B. 2003. The Cordilleran Ice Sheet, *Developments in Quaternary Science*, 1: 17-43.

Boulton, G.S. and Hindmarsh, R.C.A. 1987. Sediment deformation beneath glaciers: rheology and sedimentological consequences, *Journal of Geophysical Research*, 92: 9059-9082

Bradwell, T., Stoker, M.S., Golledge, N.R., Wilson, C.K., Merritt, J.W., Long, D., Everest, J.D., Hestvik, O.L., Stevenson, A.G., Hubbard, A.L., Finlayson, A.G. and Mathers, H.E. 2008. The northern sector of the last British Ice Sheet: maximum extent and demise, *Earth-Science Reviews*, 88: 207-226.

Brennand, T.A. 1994. Macroforms, large bedforms and rhythmic sedimentary sequences in subglacial eskers, south-central Ontario: implications for esker genesis and meltwater regime, *Sedimentary Geology*, 91: 9-55.

Brennand, T.A. 2000. Deglacial meltwater drainage and glaciodynamics: inferences from Laurentide eskers, Canada, *Geomorphology*, 32: 263-293.

Brennand, T.A. and Shaw, J. 1994. Tunnel channels and associated landforms, south-central Ontario: their implications for ice-sheet hydrology, *Canadian Journal of Earth Sciences*, 31: 505-22.

Brennand, T.A., Russell, H.A.J. and Sharpe, D.R. 2006. Tunnel channel character and evolution in central southern Ontario. In: Knight, P.G. (ed.), *Glacier Science and Environmental Change*, Blackwell Publishing Ltd, Oxford: 37-39.

Burke, M.J., Woodward, J., Russell, A.J., Fleisher, P.J. and Bailey, P.K. 2008. Controls on the sedimentary architecture of a single event englacial esker:

Skeiðarárjökull, Iceland, *Quaternary Science Reviews*, 27: 1829-1847.
Burke, M.J., Woodward, J., Russell, A.J., Fleisher, P.J. and Bailey, P.K. 2010. The sedimentary architecture of outburst flood eskers: a comparison of ground-penetrating radar data from Bering Glacier, Alaska and Skeiðarárjökull, Iceland, *Geological Society of America Bulletin*, 122: 1637-1645.
Clague, J.J. and James, T.S. 2002. History and isostatic effects of the last ice sheet in southern British Columbia, *Quaternary Science Reviews*, 21: 71-87.
Ékes, C. And Hickin, E. 2001. Ground penetrating radar facies of the paraglacial Cheekye fan, southwestern British Columbia, Canada. *Sedimentary Geology*, 143: 199-217.
Fiore, J., Pugin, A. and Beres, M. 2002. Sedimentological and GPR studies of subglacial deposits in the Joux Valley (Vaud, Switzerland): backset accretion in an esker followed by an erosive jökulhlaup, *Géographie physique et Quaternaire*, 56(1): 19-32.
Fisher, T.G., Jol. H.M. and Boudreau, A.M. 2005. Saginaw lobe tunnel channels (Laurentide Ice Sheet) and their significance in south-central Michigan, USA, *Quaternary Science Reviews*, 24: 2375-2391.
Fulton, R.J. 1991. A conceptual model for the growth and decay of the Cordilleran Ice Sheet, *Géographie Physique et Quaternaire*, 45(3): 281-286.
Heinz, J. and Aigner, T. 2003. Three-dimensional GPR analysis of various Quaternary gravel-bed braided river deposits (southwestern Germany), In: Bristow, C.D. and Jol, H.M. (Eds.), *Ground Penetrating Radar in Sediments*, Geological Society, London, Special Publications, 211: 99-110.
Huntley, D.H. and Broster, B.E. 1994. Glacial Lake Camelsfoot: a Late Wisconsinan advance stage proglacial lake in the Fraser River Valley, Gang Ranch area, British Columbia, *Canadian Journal of Earth Sciences*, 31: 798-807.
Jørgensen, F. and Sandersen, P.B.E. 2008. Mapping of buried tunnel valleys in Denmark: new perspectives for the interpretation of the Quaternary succession, *Geological Survey of Denmark and Greenland Bulletin*, 15: 33-36.
Lesemann, J-E. and Brennand, T.A. 2009. Regional reconstruction of subglacial hydrology and glaciodynamic behaviour along the southern margin of the Cordilleran Ice Sheet in British Columbia, Canada and northern Washington State, USA, *Quaternary Science Reviews*, 28: 2420-2444.
Mooers, H.D. 1989. On the formation of the tunnel valleys of the Superior Lobe, central Minnesota, *Quaternary Research*, 32: 24-35.
Neal, A. 2004. Ground-penetrating radar and its use in sedimentology: principles, problems and progress, *Earth-Science Reviews*, 66: 261-330.
Ó Cofaigh, C. 1996. Tunnel valley genesis, *Progress in Physical Geography*, 20(1): 1-19.
Plouffe, A., Bednarski, J.M., Huscroft, C.A., Anderson, R.G.McCuaig, S.J. 2010. Geochemistry of glacial sediments of the Bonaparte Lake map area (NTS 92P) south central British Columbia, *Geological Survey of Canada*, Open File, 6440, CD-ROM.
Piotrowski, J.A. 1994. Tunnel-valley formation in northwest Germany – geology, mechanisms of formation and subglacial bed conditions for the Bornhöved tunnel valley, *Sedimentary Geology*, 89: 107-141.
Pugin, A., Pullan, S.E. and Sharpe, D.R. 1999. Seismic facies and regional architecture of the Oak Ridges Moraine area, southern Ontario, *Canadian Journal of Earth Sciences*, 36: 409-432.
Rampton, V.N. 2000. Large-scale effects of subglacial meltwater flow in the southern Slave Province, Northwest Territories, Canada, *Canadian Journal of Earth Sciences*, 37: 81-93.
Rice, S.P., Church, M., Wooldridge, C.L. and Hickin, E.J. 2009. Morphology and evolution of bars in a wandering gravel-bed river; lower Fraser river, British Columbia, Canada, *Sedimentology*, 56: 709-736.
Russell, H.A.J., Arnott, R.W.C. and Sharpe, D.R. 2003. Evidence for rapid sedimentation in a tunnel channel, Oak Ridges Moraine, southern Ontario, Canada, *Sedimentary Geology*, 160: 33-55.
Russell, A.J., Gregory, A.R., Large, A.R.G., Fleisher, P.J. and Harris, T.D. 2007. Tunnel channel formation during the November 1996 jökulhlaup, Skeiðarárjökull, Iceland, *Annals of Glaciology*, 45: 95-103.
Russell, A.J., Knudson, Ó., Fay, H., Marren, P.M., Heinz, J. and Tronicke, J. 2001. Morphology and sedimentology of a giant supraglacial, ice-walled, jökulhlaup channel Skeiðarárjökull, Iceland: implications for esker genesis, *Global and Planetary Change*, 28: 193-216.
Sandmeier, K.J. 2010. REFLEXW (v5.6) Processing program for seismic, acoustic and electromagnetic reflection, refraction and transmission data. K.J. Sandmeier, Karlsruhe, Germany.
Shaw, J. 1983. Drumlin formation related to inverted melt-water erosional marks, *Journal of Glaciology*, 29(103): 461-479.
Shaw, J. and Gorrell, G.A. 1991. Subglacially formed dunes with bimodal and graded gravel in the Trenton drumlin field, Ontario, Canada, *Géographie Physique et Quaternaire*, 45: 21-34.
Sjogren, D.B., Fisher, T.G., Taylor, L.D., Jol, H.M. and Munro-Stasiuk, M.J. 2002. Incipient tunnel channels, *Quaternary International*, 90: 41-56.
Tipper, H.W. 1971a. Multiple glaciation in central British Columbia, *Canadian Journal of Earth Sciences* 8: 743-752.
Utting, D.J., Ward, B.C. and Little, E.C. 2009. Genesis of hummocks in glaciofluvial corridors near the Keewatin Ice Divide, Canada, *Boreas*, 38: 471-481.
Wingfield, R. 1990, The origin of major incisions within the Pleistocene deposits of the North Sea, *Marine Geology*, 91: 31-52.
Woodward, J., Burke, M.J., Tinsley, R., and Russell, A.J., 2008. Investigating reworking of proglacial sediments using GPR: Skeiðarárjökull, Iceland, In: Rodgers, C.D.F. and Chignell, R.J. (eds.), Proceedings of the 12th International Conference on Ground Penetrating Radar: Birmingham, University of Birmingham, ISBN 978-0-9561121-0-1, 1099 p.
Wright, H.E. 1973. Tunnel valleys, glacial surges and subglacial hydrology of the Superior Lobe, Minnesota, In: Black, R.F., Goldthwaite, R.P. and Willman, H.B. (eds.), *The Wisconsinan stage*, Memoir 136, Boulder, Co. Geological Society of America, 136: 251-76.