Refining the pattern and style of deglaciation on the southern Fraser Plateau and environs

by

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Abstract

The Cordilleran Ice Sheet (CIS) represents an ideal paleo-analogue for studying the pattern and style of deglaciation across an area of moderate relief. Hypotheses tested on the CIS landscape may be applied to our understanding of the processes involved in modern deglaciation. This thesis addresses current shortcomings in our understanding of the pattern and style of CIS deglaciation over the southern Fraser Plateau in south-central British Columbia (BC). During the last glacial buildup, ice advancing from mountain ranges in the east and west met somewhere over the southern Fraser Plateau (Tipper 1971a), forming an area of significant ice accumulation, according to models of glacioisostatic rebound (Fulton and Walcott 1975, Johnsen and Brennand 2004), and potentially increasing in thickness to form an ice divide during glacial maximum (Wilson et al. 1958). Deglaciation through ice stagnation and downwasting has been hypothesized for the plateau (Fulton 1967, 1976, 1991). This thesis tests the hypothesis that the lateglacial landform and sediment record supports the large-scale regional stagnation of the CIS over south-central BC.

Through the integration of aerial photographs, digital elevation models, field observations of landform morphology, stratigraphy and sedimentology, and shallow geophysics (electrical resistivity tomography, ground-penetrating radar) a map of lateglacial landforms and lakes associated with deglaciation of the last CIS is produced. Ice-marginal indicators, including ice-marginal lake systems and moraines are used to reconstruct ice-marginal positions, demonstrating retreat from southeast to northwest across the southern Fraser Plateau. Fields of pristine glaciotectonic moraines indicate active retreat rather than passive melting of plateau ice. Analysis of morphosedimentary relationships of esker systems and associated glacier hydrology demonstrates ice retreat was characterized by low-sloping thin ice where ice-dammed lakes were important for meltwater storage. Crevasse-fill ridges and eskers formed in ice-walled canyons record areas of localized ice stagnation.

Keywords: Cordilleran Ice Sheet; Esker; Moraine; Ice-marginal lake; Electrical resistivity tomography; Ground-penetrating radar.
To my loving wife, whose undying faithfulness, patience, and encouragement will be a model for me for the rest of my days

“If they keep quiet, the stones will cry out”
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1. Introduction
Introduction

The waxing and waning of ice sheets in recent geologic time has demonstrated their vast influence as geomorphic agents within the shifting limits of the cryosphere and as global climatologic agents. Changes in modern ice sheets have the potential to contribute vast amounts of freshwater to the ocean, impacting sea level and ocean circulation (e.g., Alley et al. 2005, Condron and Winsor 2012). Ideally, present-day ice sheets would be monitored directly to yield insight into the rate and scale of ice sheet change; however, several factors limit these investigations. 1) Access to the internal structure and ice/bed interface of modern ice sheets is limited, preventing ready observation of ice/bed interactions, and meltwater production, storage and routing. 2) Observations on modern ice sheets provide a relatively short-term snapshot of long-term ice-sheet evolution, limiting our ability to infer process over long time periods. Although recent work in remote sensing of ice sheets has progressed dramatically, one of the best ways to circumvent the limitations of ice sheet monitoring is to study the beds of past ice sheets. These landscapes are currently laid bare for investigation, have commonly undergone relatively minor geomorphic change since the last glacial period, and may retain a palimpsest of landforms and depositional sequences recording the last glacial period and prior. Through studying these landscapes and verifying our understanding against modern glacier behaviour, the tangled web of landforms and depositional sequences may be unravelled, leading to new insight into ice sheet patterns and dynamics. In turn these conclusions initiate further understanding of the pattern and dynamics of possible future change of present-day ice sheets. This thesis aims to explore the relict landscape near the centre of one such past ice-sheet, the last Cordilleran Ice Sheet (CIS), so that its development and disappearance can be better understood, and future changes to modern ice sheets may be better predicted.

Study area

Southern Fraser Plateau

The southern Fraser Plateau is located in the intermontane region of the southern Canadian Cordillera (Holland 1976), an area of broad plateaus and highlands surrounded
by ranges similar to the moderately rugged topography underlying parts of the Greenland and Antarctic ice sheets (Livingstone et al. 2012). It is a relatively low-relief surface generally ranging 1200-1500 m asl, but is bordered by peaks reaching elevations over 2200 m asl (Marble Range). The plateau is bordered on the west by the Fraser River and the Camelsfoot Range of the Coast Mountains, and on the east by the Thompson River and the Shuswap Highland (Holland 1976). Its southern limit adjoins the Thompson Plateau in the Arrowstone Hills area just north of Ashcroft and it extends north up to the area around 100 Mile House, overlapping parts of the subcategorized Bonaparte and Green Timber plateaus (cf. Huntley 1996). Bordered on the west by the Fraser River, this study area (Fig. 1.1) covers seven complete (92P 2, 3, 4, 5, 6, 7, and 12) and three partial (92O 1, 8, and 9,) 1:50 000 NTS mapsheets, incorporating an area of 7 842 km². This study area is consistent in chapters 2-5, and in Appendix A.

Gently sloping sub-horizontal Miocene and Pliocene olivine basalts of the Chilcotin Group (Campbell and Tipper 1971, Dohaney et al. 2010), underlie most of the southern Fraser Plateau, the major exception being the limestone of the Marble Range. These basalts overlie a basement of plutonic rock, with minor surface exposure of batholiths (Campbell and Tipper 1971). The present configuration of drainage networks on the plateau is likely the combined result of Tertiary drainage enhancing the gentle upwarping of the basalts during the late Miocene (Mathews 1989), and Quaternary glacial and meltwater erosion. The low-relief surface provides few opportunities for bedrock exposure (Andrews and Russell 2008), as the Plateau has a near-ubiquitous cover of till and meltwater deposits typically ≤ 20m in thickness (though ranging up to 50 m in thickness in places, Andrews et al. 2011).

The southern Fraser Plateau was covered by the south-central portion of the last CIS. It is located just south of a putative ice divide (Wilson et al. 1958, Tipper 1971a), an idea supported by reconstructions of lateglacial glacioisostatic rebound (Fulton and Walcott 1976, Johnsen and Brennand 2004). It is just east of a recently hypothesized area of large-scale, lateglacial ice flow reorganization (Margold et al. 2013). A suite of subglacial, ice-marginal and proglacial landforms on the Plateau have been previously documented at the reconnaissance scale (Tipper 1971c), and partly remapped by more recent larger-scale efforts (Bednarski 2009, Huscroft 2009, Plouffe 2009a, b), though the genetic connotations of many of these landforms have not been explored in detail.
Figure 1.1. A) General location of study area (grey polygon) within British Columbia, Canada. FP – Fraser Plateau, TP – Thompson Plateau. B) General outline of study area (white dashed line) overlain on hillshaded digital elevation model (Geobase 2007). Solid black line delimits the Fraser Plateau (Holland 1976). C) NTS mapsheets covered by study area. Green outline indicates mapsheets (1:50 000) recently mapped by the GSC. Mapping completed by Tipper (1971c) (1:250 000) covers the entire area.
CIS pattern and dynamics

Interglacial periods of the Pleistocene epoch were witness to glacier coverage similar to modern extent (Clague and James 2002), with most glaciation in British Columbia limited to alpine environments. Emergence from interglacial periods through glacier advance is thought to have occurred in four stages (Fulton 1991, Clague and James 2002) starting with the expansion of alpine glaciers, as climate deteriorated (cooler temperatures, greater precipitation). Subsequent buildup of ice in mountain ranges resulted in glacier coalescence into high elevation ice fields, and extension of alpine valley glaciers onto lowlands. As valley glaciers flowed into lowlands they merged, eventually forming a largely topographically-controlled sheet of ice over most of British Columbia. Finally, thickening of the ice sheet resulted in ice flow independent of topography in some areas. Configuration of this final stage was likely complex (e.g., Fig. 1.2), resembling a number of local ice domes centred over topographic highs joined by saddles over topographic lows rather than radial flow from a single central ice dome.

Over the south-central interior of British Columbia during Marine Isotope Stage-2 (MIS-2), initial ice advance took the form of coalesced piedmont glaciers advancing both east from the Coast Mountains, and west from Cariboo Range of the Columbia Mountains (Tipper 1971, Plouffe et al. 2011). Merging near the Fraser River, over the southern Fraser Plateau (Heginbottom 1972, Huntley 1997), flow direction appears to have deflected both north and south (based on the orientation of streamlined forms and striae). In some places the last CIS is proposed to have been several km’s thick at local Last Glacial Maximum (LGM) (Clague and James 2002), but over the high elevation southern Fraser Plateau, based on trim-line elevations (e.g. Huntley 1997) and periglacial landforms (e.g., Lian 1997), it may not have exceed much more than 600 m in thickness. However, recently published research suggests ice overtopped the Marble Range during the last glacial maximum (Margold et al. 2013, 2014) indicating it may have reached at least 1000 m in thickness over the Plateau.

Deglaciation of the southern interior plateau has been characterized as a top-down process wherein a rapid rise in equilibrium line altitude is necessary to allow surface melting of the topographically-confined interior ice mass (Fulton 1991). Evidence cited in support of this idea includes hummocky stagnation topography in the south-
Figure 1.2. A) CIS limits and ice flow patterns near local LGM from Clague and James (2002). Study area outline shown by grey polygon. B) Detail of ice flow patterns over southern Fraser Plateau from Plouffe et al. (2011).
central interior of BC, a lack of large recessional moraines whose presence would indicate an orderly ice retreat (Fulton 1967, Tipper 1971, Plouffe et al. 2011), and the presence of large ribbon-shaped, lateglacial ice-dammed lakes in the fjord-like basins bounding the southern interior plateaus (Nasmith 1962, Fulton 1965, Johnsen and Brennand 2004). However, it has been noted that the scale of past analyses may have been insufficient to allow the identification of recessional moraines in the region (Tipper 1971), and that ice-dammed lakes in the Thompson and Okanagan basins were coeval with melting plateau ice (Johnsen and Brennand 2006; Lesemann 2012). Furthermore, glacioisostatic tilt modelled from glacial lake shorelines in the Thompson and Nicola basins indicates maximum rebound to the northwest, suggesting thickest and/or longest lasting ice on the southern Fraser Plateau (Fulton and Walcott 1976, Johnsen and Brennand 2004) or Coast Mountains (Margold et al. 2013).

**Landforms used in CIS reconstruction**

As ice retreated during the last deglaciation it revealed a palimpsest landscape replete with glaciofluvial, glaciolacustrine, and glaciotectonic landforms and sediments, overprinted on landforms and sediments from previous glacial and interglacial periods (e.g., Tipper, 1971b; Huntley 1995; Lian et al. 1999; Bednarski, 2009; Plouffe, 2009a, b; Huscroft, 2009; Margold et al., 2011). The identification, relative ordering and genetic interpretation of landforms and sediments that provide first-order records of ice-margin positions or processes are the focus of this thesis. These landforms and sediments included meltwater channels (proglacial, subglacial, ice-marginal), eskers, glacial lake fans (subaqueous), deltas, lake-bottom sediments and spillways, and moraines. Landforms were identified from stereographic aerial photograph pairs (≤ 1:40 000, Appendix B) and digital elevation models (all elevation data retrieved from Canadian Digital Elevation Dataset available from Geobase®, provided as 100 m horizontal grid cells subsampled to 25 m grid cells, 90% of vertical measurements accurate to within 5 m of true elevation (refer to CDED 2007 for more information)), and digitally mapped on 0.5 m resolution orthophotograph mosaics (GeoBC, 2010) using ArcGIS 10.1-10.2 (ESRI®). Wherever possible (limited by access) the mapped landform classification was confirmed by field observations (geomorphology, sedimentology, stratigraphy) and/or shallow geophysical surveys.
Thesis scope and objectives

This thesis seeks to test the hypothesis that the lateglacial landform and sediment record supports the large-scale regional stagnation of the CIS over south-central BC. This is accomplished by refining our understanding of deglacial pattern and style of the last CIS across the southern Fraser Plateau through:

1. Detailed mapping of lateglacial landforms (chapters 2-5, Appendix A) in order to establish foundational data for evaluating deglacial pattern and style;

2. Regional reconstruction of ice-marginal lake system evolution and the implied pattern and style of deglaciation (chapters 2 and 3);

3. Furthering our understanding of moraine (chapter 4) and esker (chapters 2 and 5) genesis on the Plateau, in order to infer glaciotectonic and hydrologic conditions associated with deglaciation of the plateau.

The context and rationale for each objective are elaborated below.

Detailed mapping of lateglacial landforms

The southern Fraser Plateau is covered by a wide variety of lateglacial landforms including meltwater channels, eskers, moraines, and glaciolacustrine deltas. These landforms were first mapped and described at a scale of 1:250 000 by Tipper (1971a, b, and c), an effort supported largely by small-scale aerial photograph interpretation with little opportunity for field observation. More recently, lateglacial landforms on the eastern part of the southern Fraser Plateau have been mapped at larger scale (1:50 000) but the purpose of this mapping effort did not generally include the genetic interpretation of landforms sufficient to infer deglacial pattern and style (Bednarski 2009, Huscroft 2009, Plouffe 2009a, b). The lack of genetic context has led to difficulties in reconstructing lateglacial CIS pattern and style in the region. For example, support for regional stagnation has been drawn from absence of evidence arguments (e.g., the absence of large recessional moraines, Tipper (1971)) though this perceived absence may be more related to the scale of analysis in a heavily forested landscape, than a true absence (Tipper 1971). Therefore, the first stage in reconstructing the lateglacial story of the southern Fraser Plateau is the detailed mapping of lateglacial landforms. Chapters 2 and 3 document lateglacial landforms integral to reconstructing lateglacial ice-marginal lake systems,
including glaciolacustrine deltas, fans, grounding-line moraines, ice-marginal channels, and spillways. Chapters 4 and 5 look specifically at the distribution and size of eskers and moraines on the Plateau. All lateglacial landforms and lateglacial lakes were mapped at 1:40 000 scale and are available as spatial data files in the attached digital repository (Appendix J). They are also presented as a poster-sized 1:175 000 scale map (Appendix A).

**Ice-marginal lake reconstruction and implications for deglacial pattern and style**

Ice-marginal lake systems provide an opportunity to reconstruct the ice-margin positions based on the elevation and position of each lake surface. Most work on lateglacial lake systems in south-central British Columbia has focused on the long, narrow, fjord-like valley bottom lakes that occupied the Thompson (Johnsen and Brennand 2004, 2006), Nicola (Fulton 1965, Fulton and Walcott 1975), and Okanagan (Nasmith 1962, Lesemann 2012) basins. But observations of pockets of lacustrine sediment on the Plateau surface indicate that significant lateglacial lakes also developed atop the plateau as deglaciation progressed. The narrow and deep configuration of valley-bottom lakes translates into a relatively low amount of ice-contact at dam locations, whereas the extensive and shallow configuration of plateau lakes suggests they may record ice margin positions over broader areas and thus offer a better understanding of regional deglacial pattern. Therefore, this thesis seeks to: 1) reconstruct and employ ice-marginal lakes as indicators of ice-margin positions in a region where traditional ice-marginal indicators (moraines) were thought to be absent (chapters 2 and 3); and 2) utilize the paleogeography of successive intra-basin ice-marginal lakes to track the position of the ice margin over time and thus infer deglacial pattern and style (chapter 3).

**Regional reconstructions of moraine and esker genesis: implications for deglacial pattern and style**

Moraine and esker systems reflect significant characteristics of the glacier or ice-sheet that formed them, including glacier hydrology, ice thickness, ice-surface slope, pattern of ice-retreat, and ice/bed interactions (Brennand 2000, Bennett 2001). However, many of these relationships and interpretations have been tested on individual or small
groups of moraines (Bennett et al. 2004) and eskers (Shreve 1985, Boulton 2007) and have not been verified against multiple or larger populations. As a result, confident interpretations of moraine and esker genesis still require individual interpretation of ridge morphology in the context of internal sedimentary architecture; however, significant advantages exist in the potential to interpret moraine and esker genesis from morphology alone, especially in areas where field access is limited or study areas are large in size (e.g., Storrar et al. 2014). This thesis seeks to explore the utility of linking limited, detailed field investigation with broader, remotely sensed observation to determine the range of genetic forms on the landscape prior to large-scale morphogenetic classification. This is accomplished by: 1) integrating moraine and esker distribution and form with internal architecture in order to infer the range of modes of landform genesis for the study area (chapters 4 and 5); and 2) inferring CIS deglacial pattern and style from these landform patterns and morphogenetic relationships (chapters 4 and 5).

**Thesis format**

This thesis is written in paper format with four substantive chapters (and Appendix A) intended to stand-alone as publishable journal articles. Consequently, some limited repetition is necessary between chapters. Content, however, progresses between successive chapters with each chapter building on information available from the previous chapter(s). In Chapter 2 the formation of a lateglacial meltwater landform (previously interpreted as an esker (Tipper 1971a) or ice-contact, poorly-sorted, stratified gravel deposit (Plouffe 2009)) is explored within the larger context of basin evolution, and of several deglacial lakes, in order to infer associated deglacial pattern and style. It establishes the need for detailed morpho-sedimentary investigations of landform genesis in paleoglacial reconstructions, an approach also applied in chapters 4 and 5. Chapter 3 continues the examination of plateau basins, reconstructing the evolution of three ice-marginal lake systems and exploring their implications for understanding the pattern and style of ice-margin retreat. In Chapter 4 the distribution and the mode of formation of moraines (ice flow transverse ridges) are mapped and identified under or at the margins of the CIS, refining understanding of the pattern and style of last CIS advance, intermediate and retreat stages. In Chapter 5 the distribution and the mode of formation for eskers is mapped and identified in order to elucidate meltwater routing, meltwater
supply and controls on esker formation during CIS decay. The map (and associated documentation) on which the detailed investigations of chapters 2-5 are based is included here as appendix A because of its significance to each individual chapter as a reference for the spatial extent of landforms and interactions between landform groups, and because research from each chapter improved its accuracy and completeness.

**Published thesis chapters**


I was responsible for the majority of the data collection, analysis and synthesis and for the first manuscript draft and figure production for these chapters. My co-authors contributed to project conceptualization, data collection, analysis and manuscript writing through thoughtful discussion of concepts, comments on article structuring, and editing.
References


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2. Genesis of an esker-like ridge over the southern Fraser Plateau, British Columbia: implications for paleo-ice sheet reconstruction based on geomorphic inversion
Abstract

Robust interpretations of meltwater systems operating during ice sheet decay are integral to reconstructing deglacial patterns and style. Yet over reliance on meltwater landform morphology with limited attention to morpho-sedimentary relationships, and basin-scale geomorphic and stratigraphic context can lead to unreliable geomorphic inversion-based paleo-ice sheet reconstructions. This problem is illustrated by the evolution of Young Lake esker-like ridge (YLER) formed in the Young Lake basin (YLB) on BC’s southern Fraser Plateau during decay of the last Cordilleran Ice Sheet (CIS). We integrate data from digital elevation models, aerial photographs, sedimentary outcrops, water wells and shallow geophysics (ground-penetrating radar, electrical resistivity tomography). Previous interpretations of YLER as both an esker and an ice-contact, poorly-sorted, stratified deposit emplaced by westerly flowing meltwater, imply an eastward retreating ice margin. Geophysical data from a flat-topped component of YLER reveal slipface and planar-bedded sand and gravel overlying lacustrine sediments, characteristic of a Hjulstrom delta. Eastward-dipping foresets in a Gilbert delta exist at the eastern terminus. Contextually our observations suggest, despite esker-like morphology, YLER was not deposited within a subglacial ice tunnel. Instead, it formed through deposition of subaerial outwash between and/or on dead ice in front of a regionally backwasting ice margin. The complex deglacial evolution of YLB, including a drainage reversal and formation of two glacial lakes, supports northwestward backwasting of the CIS and dead ice within YLB. We conclude that accurate geomorphic inversion of meltwater landforms for deglacial paleo-ice sheet reconstruction requires knowledge of landform-scale morpho-sedimentary relationships and basin-scale geomorphic and stratigraphic context.
Introduction

Paleo-ice sheet reconstructions from geomorphic inversion (the use of the palimpsest glacial landforms such as till lineations, eskers, and meltwater channels to establish time-slice reconstructions of ice sheet properties and patterns) typically make assumptions about glacial landform genesis (e.g., Kleman and Borgstrom, 1996; Greenwood and Clark, 2009; Kleman et al., 2010). Accurate reconstructions of ice sheet growth and decay at the regional to local scale, therefore, rely heavily on the correct identification of landforms and their associated paleoglacial implications. Eskers form synchronously or time-transgressively in subglacial or englacial ice tunnels, or in supraglacial or ice-marginal ice-walled channels (Price, 1966; Banerjee and McDonald, 1975; Shreve, 1985; Hebrand and Åmark, 1989) and are defined here sensu Banerjee and McDonald (1975) as stratified alluvial sediment depositionally constrained by ice on both sides. Although these different formative environments reflect different deglacial styles and patterns (Shilts et al., 1987; Warren and Ashley, 1994; Brennand, 2000), subclassification of eskers along these lines (e.g., Warren and Ashley, 1994) has been rarely attempted. Rather, glacial geomorphic inversions aimed at deglacial paleo-ice sheet reconstruction have typically mapped esker ridges and assumed that all eskers formed time-transgressively near the ice margin during ice sheet decay despite some sedimentologically- and geophysically-based studies to the contrary (e.g., Brennand, 1994, Burke et al., 2012). It is our assertion that the accuracy of paleo-ice sheet reconstructions is often compromised when such reconstructions are derived from geomorphic inversions that rely solely on rudimentary landform identification, and assume landform genesis with little corroborating evidence.

The deglacial record of the last Cordilleran Ice Sheet (CIS) is replete with meltwater landforms (Tipper, 1971a,b,c; Kleman et al., 2010) and sediments (Ryder et al., 1991; Clague, 2000). However, their classification and interpretation is typically too general to allow accurate reconstruction of the pattern and style of paleo-ice sheet decay. This is, in part, because many landforms are heavily forested or relatively inaccessible by road, and consequently there is scant knowledge of their composition and morpho-sedimentary architecture. In addition, basin-scale geomorphic and stratigraphic records of deglacial events contained in basins set within the plateau surface have rarely been
exploited (e.g., Plouffe et al., 2011). Here we explore the morphology, composition and sedimentary architecture of one meltwater landform, an esker-like ridge within Young Lake basin, BC (Fig. 2.1B), placing it within the context of regional stratigraphy and basin evolution. We demonstrate that the integration of landform- and basin-scale data from digital elevation models (DEMs), aerial photographs, sedimentary outcrops, water wells and shallow geophysics (ground-penetrating radar, GPR; electrical resistivity tomography, ERT) allow more confident interpretations of landform genesis and thence deglacial style and pattern. We argue that such data integration should be integral to geomorphic inversion aimed at accurate paleo-ice sheet reconstructions, at least based on our present understanding of the general relationships between esker morpho-sedimentary character and formative process (and hence paleoglaciological implications).

**Study area and previous work**

Young Lake basin is located at the northeastern end of the Bonaparte valley on the southern Fraser Plateau of BC’s south-central interior (Holland, 1976; Fig. 2.1A and B). The basin incorporates the Young Lake esker-like ridge (YLER), modern Young Lake, the upper Bonaparte and Rayfield Rivers as well as the unnamed creek draining north from Deadman valley and is contiguous with the lower Bonaparte River to the SW (Fig. 2.1A). The southern Fraser Plateau surface ranges from 1,200 to 1,800 m asl and is bordered on the west by the Coast Mountains and on the east by the Shuswap Highlands and Columbia Ranges. The relatively flat surface of the Fraser Plateau is a reflection of its underlying geology: successive basalt flows, dating to the late Eocene to Miocene (Bevier, 1983; Andrews and Russell, 2008). The plateau surface is generally mantled by a thin unit of till that is streamlined in places (Fig. 2.1B, streamline orientations roughly align NW-SE; Tipper, 1971a,b,c; Plouffe et al., 2011). Where basalt bedrock protrudes through the thin till cover, it displays one to two striae sets (Fig. 2.1A). Striae sets, streamlined bedforms, and till geochemistry have been used to reconstruct the evolution of ice flow direction for the last CIS (Plouffe et al., 2011). Initial ice advance onto the Fraser Plateau was from accumulation centres in the east (Tipper, 1971a,b). As westward-flowing ice accumulated on the plateau, it encountered eastward-flowing ice from the Coast Mountains, somewhere in the vicinity of the Fraser River valley.
Figure 2.1.  (A) Location of the study area (box, Fig. 2.1B) on the Fraser Plateau (hash-marked polygon), and south of 100 Mile House (100MH).  (B) Shaded DEM of Young Lake basin and surrounding plateau area (Geobase®). Young Lake basin is centred over modern Young Lake, incorporates Deadman Valley, upper Bonaparte River, Rayfield River, Young Lake esker-like ridge (YLER), and is contiguous with the lower Bonaparte River. The locations of ground-penetrating radar grid 1 (GPR1) and 2 (GPR2), the electrical resistivity tomography (ERT) line, well logs A (WLA) and B (WLB), and the Bonaparte section (BS) in Young Lake basin are shown. White dashed box indicates extent of Fig. 2.2B. Striae data are from Bednarski (2009) and Plouffe (2009).

(Heginbottom, 1972). This convergence resulted in ice flow diversion to the north and south (Tipper, 1971a,b). Late stage ice flow in the vicinity of YLER was to the SE (Plouffe et al., 2011; Fig. 2.1A).

Decay of the southern interior sector of the last CIS is thought to have occurred through a complex process dominated by downwasting, accompanied by limited backwasting of the ice margin from south to north (Tipper, 1971a; Fulton, 1991). On the Fraser Plateau the deglacial landform record includes meltwater channels, eskers,
outwash sediments, hummocky deposits, minor moraine ridges, and glaciolacustrine sediments, including deltas and lake-bottom sediments (Tipper, 1971c; Bednarski, 2009; Huscroft, 2009; Plouffe, 2009). Ice-marginal positions are poorly known (indeed a clear pattern of deglacial retreat has been confounded by the lack of large recessional end moraines, Fulton 1991, Plouffe et al., 2011) yet may be defined by meltwater landforms through geomorphic inversion, if the landforms are accurately classified. A genetic classification of eskers in the region is presently lacking, in part because poor access, forest cover and shallow sediment exposures (typically < 1 m high) has limited the development of form-process models.

**Young Lake esker-like ridge (YLER)**

The YLER was initially classified as an esker or esker complex by Tipper (1971c) with formative paleoflow direction from east to west. Owen (1997) suggests that positive identification of a true esker be left to situations where geomorphology, sedimentology, and regional setting are all well understood. Perhaps with this in mind, and because few sedimentary exposures exist to confirm its sedimentary architecture, Plouffe (2009) conservatively interpreted the landform as an ice-contact, poorly-sorted, stratified deposit. Based on regional meltwater interpretations and the broad surface slope of glaciofluvial sediments within the Young Lake basin, paleoflow was inferred from east to west.

**Methods**

The evolution of Young Lake basin and the genesis of the YLER were investigated by combining detailed geomorphic mapping and shallow geophysics on local ridge segments within the context of regional stratigraphy. Stereographic aerial photographs (1:40, 000), regional digital elevation models (DEM) (25 m horizontal grid cells 1 m vertical resolution (90% of measurements within 5 m of true elevation), Geobase®), and local DEMs generated from real-time kinematic (RTK) differential global positioning system (dGPS) measurements (Leica System 500, decimetre accuracy) were used for geomorphic mapping. An average systematic elevation offset of 13 m between elevation data collected directly from dGPS measurements and Geobase® DEM elevations in the field area resulted in the adjustment of all measurements in this study to local dGPS
elevations. The landform elements of the YLER are characterized based on their surface morphology and relative relief. Geomorphic interpretations are supplemented by field observation of available sedimentary exposures and surface materials.

Shallow geophysical techniques were employed to delineate the sedimentary architecture and composition of the YLER. Two pseudo three-dimensional (3D) grids of 100 MHz GPR data (totalling ~ 2.3 km in linear distance) were collected. Grid 1 was collected on a relatively undisturbed, flat-topped segment of the YLER (Fig. 2.1B), whereas grid 2 was collected in a gravel pit (Fig. 2.1B) where 3-4 m of surface material had been excavated (except for the very edges of the pit). Data collection occurred with the antennas co-polarized, perpendicular broadside to the survey lines using a Sensors and Software Inc. pulseEKKO Pro system. During common offset (CO) surveys, antennas were kept at a constant separation of 1 m and data were collected in step mode (0.25 m). Two common mid-point (CMP) surveys provide an estimated average subsurface velocity of 0.107 ± 0.004 m/ns, which was used for data processing and to convert two-way travel time (TWTT) into depth. Processing of GPR data was completed in REFLEXW v5.6 and included time zero correction, low pass and bandpass filtering, migration, background removal, the application of a gain function, and topographic correction. Offsets in GPR reflections are interpreted as faults associated with removal (melting) of ice support (cf. Fiore et al., 2002). However, the full extent of faults (relative to landform thickness) may not be imaged owing to variations in fault orientation (relative to GPR survey direction) and GPR resolution. Elevation data for topographic correction was collected with an RTK dGPS. Data were collected in grids allowing for confident interpretation of bounding surfaces that could be traced throughout the grid; however, only those GPR lines sufficient to illustrate representative architectural forms are presented in this paper.

Two-dimensional (2D) ERT data were collected in a west-east orientation along the landform at GPR grid 1 (Fig. 2.1B) using an Advanced Geosciences, Inc. Supersting R8 112 passive electrode system arranged in a dipole–dipole configuration (max \( n = 6 \), max dipole = 12). Electrode spacing of 1-m provided high grid resolution to match the shallow, high resolution GPR data. Contact resistance for electrode takeouts was kept near or below 3 kΩ. Data were processed using Earthimager 2D (Advanced Geosciences, Inc.) with a smooth model inversion and a finite element forward model/pseudosection starting model. Negative resistivity measurements, data spikes, and misfit data points
were removed (to a maximum of 10% total data points) based on histogram analysis of root mean square (RMS) error between the starting model and the inverted pseudosection. Topographic information, collected with an RTK dGPS, was added prior to inverting the profiles.

Sediment exposures were logged at the decimetre scale and included observations of unit thickness, lateral extent, nature of the lower contact, texture, clast characteristics, structure, and paleoflow measurements. Two water well records (well tag numbers 45126 and 45278 (BCME 2011), renamed well logs (WL) A and B, respectively, in this paper; Fig. 2.1A) drilled in fans within the Young Lake basin were used in conjunction with sediment exposures and inferences from geophysical data to reconstruct regional stratigraphic context.

**Results**

**Geomorphology and surface character of the YLER**

The YLER and adjoining terraces (sub-classified as segments 1-4) extend for over 6 km in a west-east orientation, west of modern Young Lake, in the Young Lake basin (Figs. 2.1B and 2.2). Segment 1 consists of two landform elements (Fig. 2.2B) that extend from the western end of the ridge proper toward Rayfield River over a ridge-parallel distance of ~ 2.5 km. The northern landform element is in contact with bedrock on its NE flank and forms a relatively flat-topped (local relief of ~ 2 m), 400-m-wide terrace that is gently sloped toward segment 2. The southern landform element is an ~ 2-km-wide, hummocky terrace (mounds and hollows with local relief of > 8 m) that is ~ 10 m lower in elevation than the flatter, northern landform element (947 m asl compared to 957 m asl). A bedrock knob breaks the surface of the hummocky terrace near its centre. The ridge proper is composed of three segments of varying character (segments 2-4, Fig. 2.2). Segment 2, the westernmost section of the ridge, is ~ 2 km long, 85-260 m wide, and relatively flat-topped with a maximum surface elevation of 952 m asl (segment 2, Fig. 2.2). It is bounded to the north by enclosed, dry, sediment-floored depressions and to the south by the Bonaparte River (Fig. 2.2B). Slump depressions and scarps are evident along its surface and flanks. Surface material on the southern side of segment 2 is sandy gravel...
characterized by subrounded to rounded pebbles, cobbles, and rare boulders (Fig. 2.2G). However, the north side of segment 2 is capped with low amplitude (0.5 m) sand dunes (Figs. 2.2D-F). At the eastern terminus of segment 2, the ridge surface abruptly drops in elevation to 930 m asl (adjoining segment 3). Segment 3 is an 800-m-long, round-crested ridge that narrows to < 40 m wide in places and has an undulating long profile (segment 3, Fig. 2.2A-B). The ridge is bounded to the north by an enclosed pond-filled depression (Fig. 2.2B). This pond appears to be hydraulically connected to Young Lake because it shares an equivalent water surface elevation (923 m asl) and has no apparent surface outlet. The south side of segment 3 is rimmed by depressions (now incised and modified by the Bonaparte River) and scarps formed by post-depositional slumping. Segment 4 is 1.3 km long, up to 1 km wide, and relatively flat-topped (segment 4, Fig. 2.2A-B). The uniform flat ridge surface is broken by a large (~ 500-m-long) pond-filled enclosed depression that appears to be hydraulically connected to Young Lake (they have the same water surface elevation), and a gap in the ridge surface is filled by Young Lake (Fig. 2.2B). Several similar depressions (now modified by the Bonaparte River) form the slump rimmed southern side of the ridge. This segment gently slopes from 948 to 939 m asl toward the east, terminating 16 m above modern Young Lake (923 m asl). The first-order trend surface of all four segments of the YLER gently slopes (~ 3 m/km) toward modern Young Lake in the east (Fig. 2.2B).
Figure 2.2. (A) Topographic long profile of Young Lake esker-like ridge (8x vertical exaggeration, Geobase®, refer to (B) for profile path). Bars below profile correspond to legend in (B), and numbers refer to landform segments discussed in the text. Linear extent of geophysical survey grids 1 and 2 are indicated by solid black lines above profile. (B) Geomorphic map of YLER and associated landforms superimposed on an orthophotograph (Province of British Columbia, 2010; refer to Fig. 2.1B for location) displaying geophysical survey grid locations. (C) Topography around grid 1 interpolated from RTK dGPS elevation measurements (shown as points with decimetre accuracy) along GPR lines (see elevation
scale in (C)) superimposed on an orthophotograph (Province of British Columbia, 2010). Location of photos in (E)-(G) shown. (D) Perspective view of the DEM shown in (C). Refer to text for explanation of dunes and slump. (E) Surface relief looking east along YLER at grid 1. Arrows highlight two aeolian dunes on the north side of the ridge. Truck for scale (circled). (F) Small dune exposure on the north side of segment 2. (G) Boulders, cobbles, pebbles, and sand exposed on the south side of segment 2. Camera lens cap is 6 cm in diameter.

YLER radar elements

High amplitude radar reflections (labelled R1-R8 in Fig. 2.3 and Appendix E.2, E.3) traceable over multiple GPR lines were used to reconstruct the bounding surfaces of radar elements that are labelled A-L in Figs. 2.3 and 2.4 and referred to as RE-A to RE-L in the text. All radar elements are observed in GPR grid 1 except RE-J and RE-K, which are found only in grid 2 (Fig. 2.4). Radar elements and bounding surfaces are described in detail below.
Figure 2.3. Processed GPR profiles of (A) line X3, (B) line Y12, (C) line Y6, and (D) parts of line X7 from grid 1 (Figs. 2.1B, 2.2B, and C). Radar bounding surfaces (bold lines labelled R1-R9) delineate nine radar elements labeled A-F, H, I, and L (G is recorded elsewhere in line X7 (see supplementary data); RE- J and K are recorded in grid 2, Fig. 2.4). The locations where GPR lines intersect are shown by labelled double-headed arrows. Dashed lines mark offset reflections. GPR profiles are shown with no vertical exaggeration.
Figure 2.4.  (A) Fence diagram of representative processed GPR lines for grid 2 (refer to Figs. 2.1B and 2.2B for grid location) measured from the base of a partially excavated gravel pit. Relative line locations are shown by the solid lines on the grid map (bottom right). (B) Line X7 was measured on undisturbed ground 5 m above the pit floor. Dashed lines demark offset reflections. Broad diffractions at depth are caused by offline trees. GPR profiles are shown with no vertical exaggeration.

**RE-A**

RE-A is the lowest radar element imaged and can be identified throughout the grid by rapid signal attenuation below its upper bounding surface (R1). Few ‘real’ reflections are visible in this element and the lower bounding surface is not discernible. The rapid signal attenuation below R1 (Fig. 2.3) suggests a boundary between a highly transmissive medium (above) and a low transmissive medium (below, RE-A; Neal, 2004). R1 (940 m asl) is above modern Young Lake (923 m asl), and the Bonaparte River; therefore R1
cannot represent the water table. Till and bedrock are not observed at a depth that corresponds to R1 within WLA and WLB (Fig. 2.5), but a unit of silt (WLA-2) is recorded at this stratigraphic position in local water wells (Fig. 2.5). Therefore, RE-A is inferred to record relatively fine-grained sediments such as silt.

**Figure 2.5.** Stratigraphic logs from Young Lake basin; refer to Fig. 2.1B for log names and locations. The electrical resistivity tomography (ERT) log is interpreted from resistivity values (RU, resistivity units) in the ERT line at grid 1 (Fig. 2.6). The logs are arranged in spatial order from west (ERT log) to east (BS log). For each log, the X-axis records grain size and the Y-axis indicates depth below land surface (m). Dashed lines join units interpreted to have been deposited coevally and do not necessarily represent laterally or vertically continuous deposition; erosional unconformities are not displayed. Each log records a repetitive sequence of fine (lacustrine) and coarse (outwash) sediments; diamicton is absent (except for surficial sediments of WLB). The position of Deadman Valley, the elevations of inferred ice and sediment dams (Figs. 2.8 and 2.9), the inferred delta topset/foreset boundary at grid 2 (Fig. 2.4) and modern Young Lake are also shown.

**RE-B and RE-E to RE-I**

RE-B (~ 4-m-thick) is a tabular, moderately continuous element with an irregular lower bounding surface (R1, Fig. 2.3) that truncates reflections within RE-A (e.g., at ~ 20 m on line X7, Fig. 2.3D). Internal reflections are planar, subparallel and onlap R1. Lines
connect offset reflections crossing R2 into RE-B (dashed lines, Fig. 2.3) trending east and west on X3 (Fig. 2.3B, dip of 40 to 50°) and north on Y6 (Fig. 2.3C, average dip of 45°).

RE-E to RE-I are tabular in flow-parallel lines where they are at least 2-3x longer than RE-C and RE-D (Fig. 2.3B), resulting in a length to depth ratio that is about double that of RE-C and RE-D. They are trough-shaped in flow-perpendicular lines. RE-E to RE-I are composed of planar, subparallel reflections that onlap lower bounding surfaces (R4 to R8), although some of these planar reflections in RE-I dip slightly east. The sinuous and trough-shaped lower bounding surfaces of RE-E to RE-I are low angle (except for the lower bounding surface of RE-F (R5) that dips 22° to the west at its eastern extent in line X3, Fig. 2.3A), continuously traceable across all lines, and truncate deeper reflections. The lower bounding surface of RE-I (R8) truncates RE-E (Fig. 2.3). Aligned offset reflections within RE-E and RE-F continue through R4 and R5, trending both east and west (Fig. 2.3, dips of 25 to 60°). Aligned offset reflections fully contained within RE-E to RE-G are steeper (close to 90° in some cases) and trend east and west as well as north and south in Y lines (Fig. 2.3).

Strong radar signal return from RE-B and RE-E through RE-I (Fig. 2.3) and surface observations (Fig. 2.2G) suggests these radar elements are composed of sand and gravel. The continuity (in both X and Y lines) and the subparallel, planar nature of reflections within RE-B and RE-E to RE-I suggest that these are plane beds (Burke et al., 2008; Rice et al., 2009) and were deposited in gravel sheets (Rice et al., 2009). Bounding surfaces R1 and R4-8 truncate lower reflections and are reactivation surfaces indicating distinct erosional events between the deposition of gravel sheets. These events may record adjustments in accommodation space in the gently sloping YLER (Wooldridge and Hickin, 2005).

**RE-C and RE-D**

RE-C and RE-D are trough-shaped elements up to 3 m thick, 25 m long (Fig. 2.3A), and 20 m wide (Fig. 2.3C). Internal reflections are downlapped onto lower bounding surfaces and exhibit an average apparent dip of 26-30° to the east in X3 and X5, and 2-3° to the south in Y6 (Figs. 2.3A and C). Lower bounding surfaces (R2 and R3) are continuous and truncate deeper reflections (Fig. 2.3). Aligned offset reflections (Fig. 2.3)
trend east and north in X and Y lines (dip of 75° and 45° respectively), crossing from RE-C into RE-B.

The steeply dipping reflections within RE-C and RE-D (Fig. 2.3) are interpreted as avalanche surfaces, recording progradational slip-face deposition (Wooldridge and Hickin, 2005) at the edge of a scour into RE-B (recorded by R2 and R3, Fig. 2.3). Dip direction of slip-face deposits in RE-C suggest progradation, and hence formative flows, to the east, consistent with the general slope of the YLER (segments 1-4, Fig. 2.2A).

**RE-J to RE-K**

RE-J is one of two radar elements imaged in grid 2 (Fig. 2.2B) and is characterized by sets of variably dipping parallel reflections (23-30° dip toward 92° east in X5, Fig. 2.4A) that are undulatory to planar in cross section (Y lines, Fig. 2.4A). The lower boundary of RE-J is obscured by signal noise, and the upper boundary contacts the ground wave in the floor of the gravel pit (Fig. 2.4A). RE-K is the second radar element imaged in grid 2 (Fig. 2.4B) and is only visible in X7, located on an undisturbed, heavily forested area adjacent to the gravel pit and 5 m above the gravel pit floor. RE-K is composed of planar, subparallel reflections with offset reflections that align with shallow (~ 2-m-deep) topographic depressions on the surface. Signal noise from trees close to the survey line prevented imaging of the contact between RE-K and RE-J. However, the spatial relationship between the lines suggests that these radar elements are stratigraphically stacked.

The steeply dipping reflections of RE-J are interpreted as eastward-prograding foreset beds within a prograding, Gilbert-style delta (Postma, 1990; Jol and Smith, 1991). Foreset beds with shifting dip angles suggest slight changes in direction of lobate delta progradation (Kostic et al., 2005). The planar, subparallel reflections of X7 are interpreted as topsets (only imaged on the unmodified landform surface (X7) because 3-4 m of surface material has been removed from the gravel pit where radar grid 2 is centred). Offset reflections within X7 associated with small topographic depressions probably relate to later collapse of the delta surface, either resulting from contact and deposition overtop of melting ice blocks or settling after dewatering as lake level in the Young Lake basin fell.
**RE-L**

RE-L is thin (< 2-m-thick on average), wedge-shaped and characterized by irregular reflections. These reflections are typically horizontal, but in some places dip (~20°) eastward (Fig. 2.3D) and downlap the lower bounding surface that is conformable with deeper reflections (Fig. 2.3D).

The slightly dipping reflections in this radar element suggest some vertical accretion and eastward progradation. Gullies cut through this unit expose massive, silty fine sand (Fig. 2.2F). Based on surficial geomorphic expression of undulatory ridges, sandy composition and planar horizontal to slightly dipping sedimentary architecture, RE-L is interpreted as recording aeolian dunes and sand sheets comparable to those observed nearby on the Fraser Plateau by Lian and Huntley (1999) and Plouffe (2009).

**YLER resistivity units**

The ERT line had a maximum depth of penetration of ~ 48 m below the surface of the YLER. Five resistivity units are identified and hereafter labelled RU-1 to RU-5 (Figs. 2.5 and 2.6). For the descriptions below, we group RU-1, 3, and 5, and RU-2 and 4 because they have similar resistivity characteristics (Figs. 2.5 and 2.6). Undulating boundaries at unit contacts are probably the result of signal noise within the data set (Kilner et al., 2005).

**RU-1, RU-3, and RU-5**

RU-1 (≥ 10-m-thick, no lower contact observed) and RU-3 (~ 8-m-thick) average ~850 Ωm (Fig. 2.6). However, RU-1 displays a gradient from higher resistivity values close to RU-2 toward lower resistivity values at depth. RU-5 is an irregular unit (~ 1-m-thick) that contacts the surface of the YLER in places and has a relatively low average resistivity (~800 Ωm).

RU-1, 3, and 5 display resistivity values within the range of fine-grained sediments such as sandy-silt and fine-grained diamicton (Palacky, 1987; Reynolds, 1997). The stratigraphic position of RU-3 directly corresponds to RE-A, interpreted as silt in GPR grid 1 (Fig. 2.3). The rapid GPR signal attenuation within RE-A (Fig. 2.3) is consistent with the
low resistivity values measured in the ERT lines. Furthermore, at the landform surface and in gully walls along the north side of YLER segment 2 (Fig. 2.2F), an ~ 1-m-thick unit of silty fine sand corresponds to the low resistivity RU-5. Consequently, RU-1, 3, and 5 probably record fine material such as silty fine sand or silt. The gradation to lower resistivity values with depth in RU-1 (~ 922 m asl) is attributed to low resistance approaching the water table (Bonaparte River level is 920 m asl, and lakes in enclosed depressions to the north are at 923 m asl).
Figure 2.6 Interpreted electrical resistivity tomography (ERT) line (dipole–dipole array, 2-m electrode spacing) overlain by radar elements interpreted from GPR line X5 (collected at the same location in grid 1). Solid lines demark boundaries between resistivity units, labelled RU-1 to RU-5. Dotted lines (R1-R5 and R8) define the radar elements labelled A-F and I. RE-A corresponds to the relatively low resistivity values in RU-3, whereas RE-B to RE-I correspond to the relatively high resistivity values of RU-4. The relatively low resistivity values in RU-5 reflect reworked silty-fine sand at the surface. Refer to Fig. 2.5 for a stratigraphic log derived from interpretation of this ERT profile, and to the text for further discussion of these resistivity units.
**RU-2 and RU-4**

RU-2 (~ 14-m-thick) averages ~ 3000 Ωm (can be as high as 6000 Ωm), whereas RU-4 (~ 6-m-thick) averages ~ 5000 Ωm (can be as high as 8100 Ωm). These relatively high resistivity values suggest that RU-2 and RU-4 record gravel and/or sand (Palacky, 1987). This is consistent with the presence of cobbles, pebbles, and sand on the surface of the YLER, where RU-4 extends to the landform surface (Figs. 2.2G and 2.6), and with the good electromagnetic signal penetration through RE-B to RE-I in GPR grid 1 (correlative with RU-4; Fig. 2.6). RU-2 has similar resistivity values, and nearby water well records also record sand and gravel in similar stratigraphic positions to RU-2 and RU-4 (Fig. 2.5). We therefore interpret these resistivity units as sand and gravel.

**The Bonaparte section**

The Bonaparte section (BS) is a 41-m-thick, valley-side exposure consisting of five stratigraphic units (BS-1 to 5, Figs. 2.5 and 2.7A-B) exposed in a river cut 10 km upflow from the YLER, along the upper Bonaparte River (BS, Fig. 2.1B). The lowest unit (BS-1, Fig. 2.7A) is at least 6.4-m-thick. Beds of 0.5-m-thick moderately consolidated, well-sorted, planar, cross-bedded medium sand alternate with 1-m-thick cosets of type B cross lamination (15 mm crest height, 35° angle of climb) and minor amounts of type A cross lamination that record paleoflow toward the SW. Based on texture and structure, BS-1 was deposited in a subaqueous environment, likely lacustrine, with SW flowing turbidity currents (Lowe, 1982).

BS-2 is a 17.2-m-thick, upward-fining unit of subrounded to rounded, clast-supported cobble to boulder gravel in a sparse matrix of coarse sand to granule (Fig. 2.7B) with a sharp lower contact. The unit includes 0.5-m-thick lenses of planar-bedded
Figure 2.7. Sediments and geomorphology of the Bonaparte section (BS, Fig. 2.1B). (A) Bonaparte section exposure showing unit (BS-1 to BS-5) boundaries. Refer to text for unit descriptions. (B) Close-up of putative deglacial units within the Bonaparte section, including finely laminated sediments of BS-4 and coarse gravel of BS-5.

medium sand. The sharp lower contact indicates partial erosion of BS-1 prior to deposition of BS-2. Given clast size, roundness, and framework support, the unit was likely deposited in a high energy fluvial environment.

BS-3 is a 0.5-m-thick unit of massive, coarse sand containing isolated angular to subrounded pebbles concentrated within the lower half of the unit (Fig. 2.7B) and has a sharp lower contact. The large proportion of massive sand combined with isolated pebbles suggests deposition in a hyperconcentrated flow (Postma et al., 1988), possibly into a lake (see BS-4 below). The large, rounded clasts of BS-2—contrasted with the small, angular clasts in BS-3—suggest the hyperconcentrated flow was not powerful enough to erode or incorporate much of BS-2. Yet the sharp lower contact of BS-3 is an unconformity, probably reflecting erosion from another (earlier) agent of which BS-3 was the waning stage or which BS-3 directly followed.
BS-4 is a 2.5-m-thick unit of horizontally-laminated fine sand and silt with three beds of massive, medium sand up to 5 cm in thickness (Fig. 2.7B). These beds dip 10° toward the Bonaparte River, and the unit has a sharp lower contact. The horizontally-laminated fine sand and silt is characteristic of suspension settling in a slackwater environment such as a lake (Ashley, 1975; minimum elevation of lake surface 984 m asl based on the maximum elevation of BS-4). The beds of massive medium sand likely record episodic turbidity flows (Lowe, 1982) across the lake bottom.

BS-5 spans the upper 5 m of the exposure and contains moderately sorted, weakly upward-fining, mainly clast-supported cobble gravel with a coarse sand to granule matrix (Fig. 2.7B). Clasts are rounded to well-rounded, and some display worn glacigenic wear features (Krüger, 1984). The surface of the unit is gently sloped toward the Bonaparte River and is pocked by circular depressions up to 3 m deep. The well-rounded clasts and moderate sorting of BS-5 suggest deposition in a moderately energetic fluvial environment. The upward-fining trend suggests an overall decrease in energy within the fluvial system, perhaps as a result of channel migration away from deposition at the exposure location. The worn glacigenic wear features suggest the clasts are reworked glacigenic material. The circular depressions on the surface are interpreted as kettle holes related to the melt-out of buried ice (cf. Maizels, 1992; Fay, 2002).

**Geomorphic evolution and drainage reversal of Deadman Valley**

The upper Deadman Valley has been recently classified as a large meltwater channel (Huscroft, 2009), but no paleoflow direction has been attributed to it. Today, an unnamed creek drains upper Deadman Valley northward to Young Lake. However, a raised channel and erosional remnants elevated 35 m above the modern floor of Deadman Valley (Fig. 2.8) suggest that this drainage may have been reversed in the late glacial period, linking it to the upper Bonaparte River valley and the Bonaparte section. Therefore, when it was an active meltwater channel, the Deadman Valley may have supported southward drainage contiguous from the upper Bonaparte River valley. A 2.3-m/km southward-dipping slope connects the top of BS-5 (991 m asl; Figs. 2.5 and 2.8C) with the upper surface of the erosional remnants and the floor of the raised channel (~984 m asl; Fig. 2.8C). This slope is consistent with proglacial outwash streams.
Figure 2.8. (A) Terrain hillshade (Geobase®) showing Young Lake basin and location of a topographic profile starting in upper Bonaparte River valley (Y) and terminating in Deadman Valley (Y'). Dashed box indicates location of (B). (B) White arrows highlight movement direction of coalescent slump blocks that impounded Deadman Valley during deglaciation (deglacial event stage 3a, Figs. 2.5 and 2.9D) and now form a drainage divide along modern Deadman Valley. (C) Topographic profile from Y to Y’. Vertical exaggeration is 30x. The gray bar shows an inferred regional gradient of 2.3 m/km for the paleo upper Bonaparte–Deadman Valley drainage system (deglacial event stage 2A and B, Figs. 2.5 and 2.9A and B) connecting (from Y to Y’) the top of unit BS-5 (Fig. 2.5), the upper surface of the erosional remnants and the base of the raised channel. It intersects with the landslide dam (Y’) responsible for reversing flow in Deadman Valley. Bar width accounts for vertical error in the DEM.

elsewhere (Maizels, 1979; Chew and Ashmore, 2001), and therefore BS-5 likely records progacial floodplain deposition by the late glacial/early Holocene Bonaparte River (paleo-Bonaparte River terrace) prior to its incision to modern level (950 m asl). The two erosional remnants (ER, ~ 985 m asl, Fig. 2.8C) adjacent to Young Lake are composed of ice-contact sediments and hummocky till (cf. Bednarski, 2009) and are just slightly higher than the floor of the raised channel (Fig. 2.8C). These erosional remnants were probably part of the paleo-Bonaparte drainage through Deadman Valley and were partially eroded as the river incised through Deadman Valley fill.
The bathymetry of Young Lake displays a series of closed depressions below the water surface, from 20 to 60 m deep (Balkwill, 1980). These depressions likely record the positions of ice blocks within the Young Lake basin that blocked the westward flow of water, forcing drainage of the paleo-Bonaparte River south through Deadman Valley. Furthermore, many of the fans deposited by streams draining into the Young Lake basin are pocked by circular depressions resembling kettle holes. This suggests that the position of modern Young Lake was probably occupied by stagnating, inactive ice following active ice retreat from the Young Lake basin. A long, low drainage divide in the valley 3.4 km south of Young Lake is formed from several coalescent slump blocks from the walls of Deadman Valley (Huscroft, 2009). This earthen dam likely terminated southerly drainage of the paleo-Bonaparte River through Deadman Valley.

Discussion

Stratigraphic correlations within Young Lake basin

The sediments located in the Bonaparte section, well logs A and B, and inferred from geophysical investigations along the YLER can be correlated across Young Lake basin (Fig. 2.5). The stratigraphically lowest and therefore oldest sediments exposed in Young Lake basin are recorded by BS-1 and BS-2, RU-1 and WLA-1, 2, and 3 (Fig. 2.5). No correlation is attempted for these units as they likely represent deposition prior to the deglacial period. Rather, we focus on time-stratigraphic correlation based on stratigraphic position, elevation, composition, and environmental interpretation for the upper units in the basin fill that are likely deglacial in age. Units are grouped into four deglacial event stages (DES, Fig. 2.9). The fine-grained units at YLER (RU-1, RU-3, and RE-A, Figs. 2.3, 2.5, and 2.6) and in the well logs in Young Lake basin (WLA-3, WLA-5, and WLB-2, Fig. 2.5) are too low in elevation to correlate with the lake sediments of BS-3 and BS-4; and no lake sediments or landforms are found in the rest of Young Lake basin at elevations similar to those in the BS. Therefore, ice must have been present everywhere in Young Lake basin, except the upper Bonaparte River valley, in order to prevent expansion of this lake (herein named glacial Lake Spectacle) and more widespread deposition. This represents the first stage in the deglacial event sequence within Young Lake basin (DES-1, Figs. 2.5 and 2.9A).
Figure 2.9. Geomorphic evolution of the Young Lake basin during CIS retreat. 
(A) Deglacial event stage (DES) 1: The ice margin is downwasting and backwasting to the NW, opening the upper Bonaparte River valley, forming glacial Lake Spectacle (gLS), and resulting in the deposition of glacial lake sediments (BS-3 and BS-4, Figs. 2.5 and 2.7B). (B) DES-2a: The ice margin retreats further, allowing drainage from the paleo-upper Bonaparte River south through Deadman...
Valley, resulting in deposition of river terrace gravel (BS-5, Figs. 2.5 and 2.7B-C) and the sediments within the erosional remnants (Fig. 2.8A), and incision of the Deadman Valley raised channel (RC, Fig. 2.8A). (C) DES-2b: Ice retreat opens up the Rayfield River valley, but downwasting ice blocks the lower Bonaparte River valley resulting in drainage east into Young Lake basin overtop of stagnating ice (recorded by RU-2, Figs. 2.5 and 2.6; WLA-4, WLB-1, Fig. 2.5) and south down Deadman Valley (raised channel from DES-2a is abandoned after further glaciofluvial erosion of Deadman Valley). This is the first stage in the formation of YLER. (D) DES-3a: Coalescent slumps (Fig. 2.8) into Deadman Valley dam this drainage route forming glacial Lake Young (gLY) and resulting in the deposition of glaciolacustrine sediments (recorded by RE-A, Figs. 2.3 and 2.5; RU-3, Figs. 2.5 and 2.6; WLA-5 and WLB-2, Fig. 2.5). DES-3b: Evolution of a Hjulstrom-style delta occurs as sediment progrades eastward over lake bottom sediments and amongst wasting ice blocks in the lower Bonaparte River valley at GPR grid 1 (recorded by RE-B to RE-I, Fig. 2.3; and RU-4, Figs. 2.5 and 2.6). A Gilbert-style delta progrades eastward over lake bottom sediments at GPR grid 2 (recorded by RE-J and K, Fig. 2.4), as inflows encounter deeper lake water. At this time (or perhaps slightly later) meltwater drainage from the plateau surface to the north deposits outwash sediments in valley marginal fans within the Young Lake basin (recorded by WLA-6 and WLB-3, Fig. 2.5); (E) DES-3c: Lake level rises, overtops the ice dam in the lower Bonaparte Valley (the lowest possible outlet at the west end of Young Lake), and rapidly drains, establishing the path for modern Bonaparte River drainage. (F) DES-4: All dead ice has melted out and the modern lower Bonaparte River drains west from modern Young Lake. Rayfield River no longer drains into Young Lake basin and south through Deadman Valley, but has cut through the ice contact sediments at the west end of the YLER and joined the westward-flowing lower Bonaparte River. Ice margin locations are schematic in (C)-(E).

Coarse-grained unit BS-5 is inferred to correlate with units RU-2, WLA-4, and WLB-1 (Fig. 2.5) based on grain size and stratigraphic position. The sediments, which are consistent with deposition in a proglacial outwash system, are in line with confluent southward flow of the paleo-upper Bonaparte River and paleo-Rayfield River out of Young Lake basin through Deadman Valley. RU-2, WLA-4, and WLB-1 connect flow out of the paleo-Rayfield River with Deadman Valley. These units likely record later stage floodplain deposition from paleo-Rayfield River atop of and amongst dead ice where modern Young Lake and the YLER now exist, after the active ice front had retreated from Young Lake basin (Figs. 2.5 and 2.9C, DES-2a-b).
In the western arm of Young Lake basin, units RU-3, WLA-5, and WLB-2 all record fine-grained sedimentation into a high stage lake up to at least 986 m asl (herein named glacial Lake Young, Figs. 2.5 and 2.9D) in Young Lake basin (Figs. 2.5 and 2.9D, DES-3a). This lake formed because the southern outlet of the Rayfield–Bonaparte drainage system through Deadman Valley was blocked by slumped material (Fig. 2.8), likely caused by debuttressing as a result of deglaciation and slope undercutting by earlier meltwater flow through Deadman Valley. The modern route of Bonaparte River through the lower Bonaparte River valley to the west is inferred to have been blocked by ice (Fig. 2.9D).

Coarse-grained, deltaic units within YLER (RU-4 including RE-B to RE-K) cap the glaciolacustrine sequence locally (Figs. 2.3, 2.5, and 2.6). WLA-6 and WLB-3 are interpreted to be deposited coevally with RU-4 because of their similar coarse-grained character and stratigraphic position directly atop finer-grained units below but above the elevation of modern Young Lake (Fig. 2.5). The coarse-grained character of sediments in WLA-6 and WLB-3 is consistent with deposition in fluvially-dominated alluvial fans (e.g., Ryder, 1971; Ballantyne, 2002) or low energy subaqueous fans (Lowe, 1982; Winsemann et al., 2007), suggesting that WLA and WLB were drilled into valley-side, kettled fan sediments. The fans are connected north to the plateau surface by long, deeply incised channels with small modern catchments. Consequently, the fans were likely built by sediment deposited by proglacial outwash streams (DES-2b to 3b, Figs. 2.9C and D). Kettles holes suggest that fan sedimentation occurred atop stagnating ice in the Young Lake basin, an inference consistent with that made for RU-4 of the YLER. As ice retreat on the plateau progressed, these meltwater channels were ultimately abandoned.

RU-5 (including RE-L, Figs. 2.3, 2.5 and 2.6) and WLB-4 cap their respective stratigraphic sequences and, although disparate in sedimentology, record local postglacial processes (Figs. 2.5 and 2.9F, DES-4). The fine-grained sediments of RU-5 (Fig. 2.6) record postglacial aeolian activity; whereas the thin, coarser-grained diamicton of WLB-4 records colluvial activity.

**Genesis of the Young Lake esker-like ridge**

Two hypotheses have been proposed for the genesis of the YLER: (i) it is an esker formed in an ice-walled channel with meltwater flows from east to west (Tipper, 1971c);
and (ii) it is an ice-contact, poorly-sorted, stratified deposit aggraded by meltwater flowing from east to west (Plouffe, 2009). New mapping shows that the landform has a general slope to the east with varied morphology (Fig. 2.2). It is composed of coarse outwash material (RU-2, Figs. 2.5 and 2.6) overlain by sandy silt or silty lacustrine sediments (RE-A and RU-3, Figs. 2.3, 2.5, and 2.6; cf. Pellicer and Gibson, 2011) typical of the distal end of a deltaic environment (Smith and Ashley, 1985), topped by gravelly deltaic sediments (RE-B to RE-K, RU-4, Figs. 2.3 to 2.6) recording delta progradation and flow from west to east atop and amongst dead ice.

Tabular gravel sheets and scour fills are characteristic of subaerial braided stream floodplains (Heinz and Aigner, 2003; Sambrook Smith et al., 2006) and Hjulstrom deltas fed by proglacial rivers (Hjulstrom deltas are characterized by rapid sedimentation in high energy, spatially restricted, shallow-water depositional environments, cf. Church and Gilbert, 1975; Jol and Smith, 1991; Postma, 1990) rather than subglacial esker fills. Subglacial esker fills deposited by flow under hydrostatic pressure are typically characterized by deposition in large convex-up bedforms and macroforms (e.g., Brennand, 1994; Burke et al., 2008). Therefore, the YLER is not a classic subglacial esker (cf. Brennand, 2000). The low relief, gently-sloping character of the terrace that forms the northern landform element of segment 1 is attributed to deposition by an outwash stream directly on land. The flat-topped character of segment 2 (Fig. 2.2A and B), along with its largely undisturbed primary bedding and minor amounts of faulting with consistent dips toward the landform flanks (corresponding to slump scarps at the landform surface), suggest that deposition in segment 2 was also mainly on land rather than ice. Conversely, if the landform had been deposited on top of ice, we would expect highly disturbed primary bedding and more extensive, irregularly oriented faults. The presence of distinct fault scarps and kettle holes on the north side of the ridge at segment 2 suggests that it was deposited atop of and in contact with dead ice and that its flanks partially collapsed after melting of this ice support (Maizels, 1992; Fay, 2002). By extension, we attribute the hummocky landform element of segment 1, undulatory character of segment 3, and the pond-filled depressions north of segment 3 and within segment 4 (Fig. 2.2B) to sedimentation atop of and in contact with dead ice. The surface of YLER is capped by postglacial aeolian sand sheets and dunes (RE-L and RU-5, Figs. 2.3 and 2.6), recording Holocene reworking and sedimentation.
In summary, the YLER formed as ice-marginal glaciolacustrine sediments and braided outwash/deltaic sediments were deposited on top of decaying ice and/or within an ice-walled canyon in decaying ice. Consequently, the YLER is an ice-contact deposit (Plouffe, 2009) that is best classified as a composite landform composed of a kame terrace (northern landform element of segment 1), deltas and kames that include elements of deposition atop ice (segment 4 and southern landform element of segment 1), and an esker deposited on top of and/or containing buried ice within a subaerial ice-walled canyon (segments 2 and 3). The Hjulstrom and Gilbert deltas (RE-B to RE-K, Figs. 2.3 and 5; RU-4, Fig. 2.6) were deposited by a stream flowing east from Rayfield River valley (Figs. 2.2B and 2.9C-E).

**Deglacial and early Holocene evolution of Young Lake basin**

Reconstruction of the evolutionary history of Young Lake basin must accommodate regional geomorphology and stratigraphy of the basin and eastward paleoflow directions recorded by the YLER. The valleys that are confluent on Young Lake basin (Deadman Valley, Rayfield River valley and upper Bonaparte River valley) are antecedent pre-glacial valleys (Andrews et al., 2011; Plouffe et al., 2011) that were deepened during the last glacial period by the erosion of meltwater channelled from the Fraser Plateau surface and the decaying CIS margin (DES-1 to 4, Fig. 2.9).

As ice began to decay over Young Lake basin, the first event (DES-1) recorded in the basin fill was the formation of the small, ice-dammed glacial Lake Spectacle within the upper Bonaparte River valley, marked by the deposition of BS-3 and BS-4 (Figs. 2.5 and 2.7B, DES-1, Fig. 2.9B).

Broad regional ice retreat NW toward the ice divide over the Coast Mountains (Tipper, 1971a) would have caused ice retreat away from the entrance to Deadman Valley opening an outlet (south through Deadman Valley) through which glacial Lake Spectacle was able to drain. Ice in the west arm of Young Lake basin blocked the modern lower Bonaparte River valley forcing the paleo-upper Bonaparte River to also flow south through Deadman Valley (DES-2a, Fig. 2.9B). During this period, the raised channel in Deadman Valley (RC, Fig. 2.8A) was incised and a river floodplain aggraded. The latter is recorded in the sediments within the erosional remnants along Deadman Valley (ER, Fig. 2.8A) and
the gravel in BS-5 (Figs. 2.5 and 2.7B). With further ice retreat, downwasting continued within Young Lake basin, resulting in detachment of dead ice blocks from the main backwasting ice margin. Formation of the YLER was first characterized by deposition of gravel and sand in an outwash stream flowing east overtop and between this stagnating ice along the lower Bonaparte River valley (RU-2, Figs. 2.5 and 2.6; WLA-4 and WLB-1, Fig. 2.5) through the area occupied by modern Young Lake before joining the paleo-upper Bonaparte River as it flowed south through Deadman Valley (DES-2b, Fig. 2.9C).

Opposing slump block failures in Deadman Valley (Fig. 2.8B) formed an earthen dam at what is now a drainage divide along Deadman Valley, ponding glacial Lake Young in Young Lake basin to an elevation of ~949 m asl (maximum elevation of Gilbert-style delta foreset/topset boundary; DES-3a and b, Fig. 2.9D), and allowing the accumulation of fine-grained lake bottom sediments (RE-A, Fig. 2.3; RU-3, Figs. 2.5 and 2.6; WLA-5 and WLB-2, Fig. 2.5). During this phase, an outwash stream was still delivering sediment from the Rayfield River valley through an unroofed, ice-walled channel (within the dead ice) into glacial Lake Young. This resulted in the deposition of a shallow water, Hjulstrom delta (RE-B to RE-I, Figs. 2.3) within the ice walls where lake depth was low and, following progradation, a Gilbert-style delta (Fig. 2.4; YLER delta, Fig. 2.5) where lake depth increased near the dead ice margin in Young Lake basin (DES-3a and b, Fig. 2.9D). Also at this time, meltwater channels were still delivering sediments from the northern plateau surface to valley marginal fans overtop of stagnating ice in the Young Lake basin (WLA-6, WLB-3, Fig. 2.5).

With the rising water level in glacial Lake Young and decaying ice in the Bonaparte Valley, a natural spillway developed to the west over the lowest point in the downwasting ice. This spillway was proximal to the southern boundary of the YLER and generally followed the course of the modern lower Bonaparte River (DES-3c, Fig. 2.9E). Water flow through the lake spillway enhanced outlet incision through the ice dam (via thermomechanical melting of the underlying ice) and facilitated rapid lake drainage, releasing up to 0.145 km³ of water down the Bonaparte Valley.

Following lake drainage, deltaic and lake-bottom sedimentation ceased and the modern outlet for Young Lake through the lower Bonaparte River valley was established (DES-4, Fig. 2.9F). The Rayfield River also responded to the new base level by incising
through segment 1 of YLER (Fig. 2.2B) to establish its modern confluence with the Bonaparte River west of Young Lake. Subsequent loss of lateral and buried ice support through melting led to exposure of the YLER, localized slope failure, and kettle hole development. Immediately following deglaciation, and prior to stabilization by vegetation, aeolian (Figs. 2.2D-F; RE-L, Fig. 2.3; RU-5, Fig. 2.6) and colluvial sedimentation (WLB-4, Fig. 2.5) were active on the surface of the YLER and within the broader Young Lake basin.

In summary, geomorphic inversion of YLER based on detailed morpho-sedimentary knowledge at the landform scale and basin-scale geomorphic and stratigraphic insight suggests northwestward ice-marginal retreat. This is in contrast to existing interpretations relying on reconnaissance level geomorphology which favour eastward retreat (Tipper, 1971c; Plouffe et al., 2011), but is consistent with the regional pattern of CIS retreat from SE to NW across the southern interior plateaus inferred from the glacioisostatic tilt of proglacial lake water planes to the south (Fulton and Walcott, 1975; Johnsen and Brennand, 2004).

Implications for robust geomorphic inversion

Through detailed morpho-sedimentary and basin analysis we have shown that YLER is an ice-contact meltwater deposit (Plouffe, 2009) that is best classified as a composite landform composed of a kame terrace, kames, delta, and esker deposited by a westward flowing outwash stream through and around dead ice. This insight supports northwestward ice-margin retreat across Young Lake basin, in contrast to the eastward retreat implied by existing interpretations based largely on small-scale mapping (Tipper 1971b, Bednarski, 2009; Plouffe et al. 2011).

Our data confirm that eskers do not always form in subglacial R-channels (cf. Shreve, 1985; Boulton et al., 2009), and so the presence of an esker alone cannot be used to make generic inferences concerning deglacial patterns and subglacial hydrology (Owen, 1997). Esker formation in Young Lake basin took place within an unroofed, ice-walled channel and included supraglacial, subaerial, and subaqueous deposition. Such esker formation must have been supplied by water produced from sources in addition to basal melting that is often assumed to be the primary water supply in numerical models.
used to explain esker patterns and infer ice dynamics (Boulton et al., 2007; Hooke and Fastook, 2007). Only if the morpho-sedimentary relationships in eskers support deposition in subglacial ice tunnels should eskers be interpreted as such in paleoglaciological reconstructions based on geomorphic inversion.

This study demonstrates that the accuracy of paleo-ice sheet reconstructions is often compromised when such reconstructions are derived from geomorphic inversions that rely primarily on rudimentary landform identification and geomorphic context, and assumptions about landform genesis (e.g., Kleman and Borgstrom, 1996). Rather, we argue that accurate paleo-ice sheet reconstructions require integrated knowledge of basin-scale geomorphic and stratigraphic context, and of landform-scale morpho-sedimentary relationships. In field areas with relatively poor sedimentary exposures the application of multiple shallow geophysical tools may be pivotal to developing an understanding of subsurface sedimentary architecture, allowing for robust paleo-ice sheet reconstructions.
References


3. Refining the pattern and relative timing of Cordilleran Ice Sheet retreat: paleogeography, evolution and implications of lateglacial ice-marginal lake systems on the southern Fraser Plateau, British Columbia
Abstract

Decay of the last Cordilleran Ice Sheet (CIS) near its geographic centre has been conceptualized as being dominated by passive downwasting (stagnation), in part because of the lack of large recessional moraines. Yet, multiple lines of evidence, including reconstructions of glacioisostatic rebound from paleo-glacial lake shoreline deformation suggest a sloping ice surface and a more systematic pattern of ice-margin retreat. Here we reconstruct ice-marginal lake evolution across the subdued topography of the southern Fraser Plateau in order to elucidate the pattern and style of lateglacial CIS decay. Lake stage extent is reconstructed using primary and secondary paleo-water-plane indicators: deltas, spillways, ice-marginal channels, subaqueous fans and lake-bottom sediments identified from aerial photograph and digital elevation model interpretation combined with field observations of geomorphology and sedimentology, and ground-penetrating radar surveys. Ice-contact indicators, such as ice-marginal channels, and grounding-line moraines are used to refine and constrain ice-margin positions. Results show that ice-dammed lakes were extensive (average 27 km$^2$; max. 116 km$^2$) and relatively shallow (average 18 m). Within basins successive lake stages appear to have evolved by expansion, decanting or drainage (GLOF (glacial lake outburst flood), outburst flood or lake maintenance) from southeast to northwest, implicating a systematic northwestward retreating ice margin (rather than chaotic stagnation) back toward the Coast Mountains, similar in style and pattern to that proposed for the Fennoscandian Ice Sheet. This pattern is confirmed by cross-cutting drainage networks between lake basins and is in agreement with numerical models of North American ice sheet retreat and recent hypotheses on lateglacial CIS reorganization during decay. Reconstructed lake systems are dynamic and transitory and likely had significant effects on dynamics of ice-marginal retreat, the importance of which is currently being recognized in the modern context of the Greenland Ice Sheet, where >35% of meltwater streams from land-terminating portions of the ice sheet end in ice-contact lakes.
Introduction

Recent inventories of ice-marginal lakes at the terrestrial margins of the Greenland Ice Sheet demonstrate the effectiveness of such systems in retaining glacier meltwater and in so doing buffering the passage of climate-induced melt to the ocean (Carrivick and Quincey 2014). Ice-marginal lakes may also affect ice stability and dynamics, weather and climate (e.g. Stokes and Clark 2004; Carrivick and Tweed 2013), and their drainage often results in distinctive geomorphic and hydrologic change (e.g., Baker 2002). These linkages, and the fact that the current pattern of deglaciation is globally increasing the size and number of ice-marginal lakes (Carrivick and Tweed 2013), reinforce the importance of understanding the records of paleo-ice-marginal lakes and their implications for past climate, hydrologic and geomorphic change, and ice dynamics and decay. Reconstructions of glacioisostatic rebound (e.g. Teller and Thorleifson 1983; Lewis et al. 2005; Konfal et al. 2013) and patterns of paleoglacial retreat have been significantly improved by elucidating paleo-ice-marginal lake geography, evolution and drainage (Jansson 2003; Etienne et al. 2006; Carrivick and Tweed 2013).

The Cordilleran Ice Sheet (CIS) last occupied the southern Fraser Plateau of interior British Columbia (BC) between 24 and 11.5 cal. ka BP (Clague and James 2002; Dyke et al. 2003). Limited radiocarbon ages suggest that ice sheet decay over interior BC was relatively rapid, commencing ~15 cal. ka BP and almost entirely complete by 11.5 cal. ka BP (Clague and James 2002). However, the pattern of deglaciation through the Interior remains poorly defined, despite recent contributions on large-scale patterns of glacial retreat (e.g. Margold et al. 2013). The apparent lack of large recessional moraines (Fulton 1967; Tipper 1971; Plouffe et al. 2011) and presence of landforms reflecting downwasting ice (e.g. nested ice-marginal channels, kettle and kame topography) have led to speculation that late stages of deglaciation in the Interior were characterized by stagnation (passive decay) of ice masses stranded as ice thinned over moderate topography (Fulton 1967, 1969), leaving a complex mosaic of dead and dying ice tongues in valleys (Clague and James 2002). Yet, multiple lines of evidence, including reconstructions of glacioisostatic rebound from paleo-glacial lake shoreline deformation in the Thompson and Nicola valleys to the south (Fulton and Walcott 1975; Johnson and Brennand 2004), and lateglacial drainage evolution in the Young Lake basin to the east (Perkins et al. 2013)
suggest a sloping ice surface and a more systematic pattern of ice-margin retreat. Accompanying deglaciation, glacial lakes formed both within deep inter-plateau valleys (Nasmith 1962; Fulton 1965, 1969; Fulton and Walcott 1975; Johnsen and Brennand 2004, 2006) and on the more subdued topography of the Plateau (Valentine and Schori 1980; Plouffe et al. 2011). Though the former lake systems have been investigated in detail, the latter have not been investigated in sufficient detail to contribute to reconstructions of deglacial retreat patterns.

Here, we reconstruct ice-marginal lake positions and evolution on the southern Fraser Plateau in order to determine if the last CIS retreated in a coherent pattern or decayed more chaotically near its geographic centre. In so doing we also investigate patterns of glacioisostatic adjustment. We explore the broader implications of our results for ice dynamics during decay, and the utility of the approach for reconstructing retreat patterns in regions lacking more obvious ice margin indicators (e.g., moraines).

**Study area and previous work**

We reconstruct ice-marginal lakes over an area of about 8 000 km² centred just north of the village of Clinton, on BC’s southern Fraser Plateau (Fig. 3.1B), and located in the intermontane region of the southern Canadian Cordillera (Holland 1976), near the geographic centre of the CIS at its maximum extent. The area is relatively low in relief, generally ranging 1200-1500 m a.s.l., but is bordered by peaks reaching elevations over 2200 m a.s.l. (Marble Range, Fig. 3.1B). Most of the area is underlain by subhorizontal Miocene and Pliocene olivine basalt flows of the Chilcotin Group (Campbell and Tipper 1971; Dohaney et al. 2010) that have undergone gentle tectonic warping sometime after the late Miocene (Fig. 3.1C; Mathews 1989). Bedrock is commonly mantled by up to ~10 m of till, glaciofluvial or glaciolacustrine sediments (Plouffe et al. 2011) and surface outcrops of bedrock are infrequent (Andrews and Russell 2008; Dohaney et al. 2010). Glacial landforms include subglacial, ice-marginal and proglacial channels, kame-deltas, glaciofluvial deltas, subaqueous fans, eskers, moraines (hummocky, minor ridges, over-ridden ridges) and streamlined landforms (Tipper 1971; Plouffe et al. 2011; Burke et al. 2012a, b; Perkins et al. 2013).
Figure 3.1. A. The study area (box B) located on the southern Fraser Plateau, BC. B. Hillshaded digital elevation model of the study area showing the general locations (black boxes) of the deglacial lake systems described in this paper, and the location of Fig. 3.9A (white dot-dashed box). Maps of the deglacial landforms and sediments associated with these lake systems are presented in Fig. 3.3. Dashed line labelled X-X’ denotes position of topographic profile in ‘C’. C. Topographic profile X-X’ showing relief of tectonically-warped basin and upland system across southern Fraser Plateau (cf. Mathews 1989). Abbreviations in B and C refer to locations mentioned in the text: BBB = Big Bar basin; BBV = Big Bar valley; BCB = Brigade Creek basin; BCV = Brigham Creek valley; BRV = Bonaparte River valley; CCV = Canoe Creek valley; DCV = Dog Creek valley; DCB = Dog Creek basin; DRV = Deadman River valley; FRV = Fraser River valley; LLV = Loon Lake valley; MLB = Meadow Lake basin; PCB = Pigeon Creek basin; WLB = White Lake basin; YLB = Young Lake basin.
At Local Last Glacial Maximum (LLGM; ~17 cal. ka BP, Clague and James 2002) ice was >1200 m thick over the Plateau (Margold et al. 2014). As deglaciation progressed several large, well-documented lateglacial ice-dammed lakes deposited thick sediment successions in the narrow, deep valleys that transect the Plateau (e.g. Nasmith 1962; Clague 1988; Johnsen and Brennand 2004), including several, as yet unnamed, lakes formed within the Fraser River valley and its tributaries adjacent to the Marble Range (Huntley 1996). Soil (Valentine and Schori 1980) and surficial geology (Bednarski 2009; Huscroft 2009; Plouffe 2009) surveys reveal that pockets of glaciolacustrine sediment also occupy the subdued topographic basins of the tectonically-warped plateau (tectonic warping was posposed by Mathews (1989) based on his identification of a subtle topographic dome through the centre of the study area). The shallow and highly discontinuous extent of these sediments suggests that the lakes in which they were deposited were not as deep or as long-lived as their valley counterparts. But shallow lake depth combined with a relatively low relief plateau surface would have produced lakes with expansive areal coverage resulting in broader spatial exposure to potential topographic spillways. Broad spatial exposure to potential spillways increases the probability of multiple lake stages, helpful in reconstructing the relative position of the ice margin over time. Furthermore, expansive lakes require greater contact with the ice margin, when contrasted with deep, narrow lakes within valleys, and therefore offer a more extensive record of ice-margin position.

**Methods**

**Reconstructing lateglacial lake extent**

*Identification and classification of paleo-water-plane and ice-contact indicators*

Paleo-water-plane (primary and secondary) and ice-contact indicators used to reconstruct lateglacial lake extent and associated ice-margin positions, and the basis for the classification of each in this study, are summarized in Tables 3.1 and 3.2 and are shown in Figs 3.2 and 3.3. Indicators were identified and provisionally classified by combining published research (e.g. Tipper 1971; Plouffe et al. 2011), aerial photograph
(≤1:40 000) analysis and digital elevation models (DEM, Geobase® 2007 (25 m horizontal grid cells with 1 m vertical resolution (90% of measurements within 5 m of true elevation), Geobase®) in a geographic information system (GIS) (cf. Jansson 2003; Johnsen and Brennand 2004; Rosentau et al. 2007). Where accessible, landform and sediment classifications were later confirmed by field observations at exposures and by hand augering (sediment texture) or ground-penetrating radar (GPR) surveys (sedimentary architecture; Fig. 3.2E) where exposures were not available. GPR data were collected in pseudo-3D grids with 100 MHz antennas co-polarized, perpendicular broadside to survey lines with a Sensors and Software Inc. PulseEKKO IV system. Common offset surveys were completed with antennas separated by 1 m and data were collected in step mode (0.25 m steps). Common mid-point surveys provided an estimated average subsurface velocity (0.102 m ns⁻¹ ±0.005) used for data processing, and conversion of two-way travel time into depth. Processing of GPR data was completed using REFLEXW v5.6 and included time zero correction, low pass and bandpass filtering, migration, background removal, the application of a gain function and topographic correction (based on floating RTK dGPS surveys completed with a Leica system 500). Exposure logging (decimetre scale) included observations on unit thickness, lower contacts, grain size, shape and roundness (where applicable), sorting, and structure (Fig. 3.4). The apparent linear distribution of water-plane indicators (Fig. 3.3) is largely a reflection of sampling accessibility along road networks.

Table 3.1. Paleo-water-plane indicators, southern Fraser Plateau.

<table>
<thead>
<tr>
<th>Indicator</th>
<th>Description</th>
<th>Explanation</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Primary: true indicator of lake water plane</td>
<td>Gilbert-type delta</td>
<td>Large, lobate landform with gently sloping top and steep foreslope; above modern lakes. Bedded sand and gravel and laminated silt and clay organized into topsets/foresets and bottomsets, respectively.</td>
<td>Where known, contact between foresets and topsets is used as an indicator of water plane. Delta plain slightly over-estimates lake level.</td>
</tr>
<tr>
<td></td>
<td>Paleo-spillway: lake maintenance</td>
<td>Channel in bedrock or till.</td>
<td>Paleo-spillways in bedrock or bouldery till typically maintained a stable or gradually decreasing lake level (maintenance) at the spillway height.</td>
</tr>
<tr>
<td>Indicator</td>
<td>Description</td>
<td>Explanation</td>
<td>Ref.</td>
</tr>
<tr>
<td>-----------------------------------</td>
<td>-----------------------------------------------------------------------------</td>
<td>-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------</td>
<td>------</td>
</tr>
<tr>
<td>Secondary: under-estimation of lake water plane</td>
<td>Paleo-spillway: GLOF or outburst flood through breached earthen dam</td>
<td>Channel incised through remnant earthen dam.</td>
<td>7–8</td>
</tr>
<tr>
<td>Ice-marginal channel</td>
<td>Limited geomorphic signature at inferred ice-dam position. Presence of hummocky moraine where spillway is likely to have existed. Absence of lake sediments where ice is proposed.</td>
<td>GLOF or outburst flood drainage through breached ridges (e.g. moraines and eskers) or fans occurs when lake drainage incises through the dam, rapidly creating a new spillway. The spillway floor represents an under-estimation of lake level.</td>
<td>8</td>
</tr>
<tr>
<td>Lake-bottom sediment</td>
<td>Nested, sinuous channel sets subparallel to topographical contours.</td>
<td>GLOF drainage atop a breached ice-dam. The modern topography where the spillway is inferred to have existed represents an under-estimation of lake level.</td>
<td>9–10</td>
</tr>
<tr>
<td>Subaqueous fan</td>
<td>Low relief surface containing normally graded or horizontal-, planar- or cross-laminated sand, silt and/or clay.</td>
<td>Channels formed in contact with ice and land that grade to lake level. The elevation of the channel mouth floor under-estimates lake level.</td>
<td>11</td>
</tr>
<tr>
<td>Secondary: over-estimation of lake water plane</td>
<td>Earthen dam</td>
<td>Sedimentary landforms (e.g. eskers, moraines) that prevent lake drainage by blocking potential outlets.</td>
<td>7–8</td>
</tr>
</tbody>
</table>

## Table 3.2. Ice-contact indicators, southern Fraser Plateau.

<table>
<thead>
<tr>
<th>Indicator</th>
<th>Description</th>
<th>Utility for locating ice margin</th>
<th>Explanation</th>
<th>Ref.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ice-marginal channel</td>
<td>Nested, sinuous channel sets subparallel to topographical contours.</td>
<td>Estimates ice-marginal position against a topographical slope.</td>
<td>Form where most efficient topographical drainage pathway is blocked by ice.</td>
<td>1–2</td>
</tr>
<tr>
<td></td>
<td>Ridge-like to fan-shaped deposit with a steep stoss slope and gentle lee slope. Composed mainly of inclined beds of gravel and sand, sometimes with diamicton interbeds. Found in association with lake-bottom sediments and may be bisected by GLOF or outburst flood spillway.</td>
<td></td>
<td>Follows most efficient drainage pathway between ice and topographical slope.</td>
<td></td>
</tr>
<tr>
<td>Grounding-line moraine</td>
<td>Extensive, undulating surface deposits composed of diamicton and commonly mantled by large (&gt;1 m b-axis), angular basalt (local) boulders.</td>
<td>Marks ice-contact position.</td>
<td>Grounding-line moraine may form where sediment-laden meltwater exits the ice margin and flows into a lake.</td>
<td>3–4</td>
</tr>
<tr>
<td>Ice-rafted debris</td>
<td>Outsize clast or groups of clasts occurring randomly within lake-bottom sediments.</td>
<td>Occurrence suggests deposition in an ice-contact lake but does not locate the ice margin.</td>
<td>Where lake-bottom sediments are absent, but located nearby, it is likely that buried ice raised incipient hummocky terrain sediments above the surrounding water plane.</td>
<td>5–6</td>
</tr>
<tr>
<td>Lack of alternative damming mechanism</td>
<td>Open-ended valleys where lake-bottom sediments and associated landforms cease to exist, but no topographical or pre-existing sediment dam is apparent.</td>
<td>Ice-contact location is somewhere beyond the last occurrence of waterplane indicators (Table 3.1).</td>
<td>Absence of topographical or pre-existing sediment dam suggests a glacier or ice block must have impounded the lake.</td>
<td>7</td>
</tr>
</tbody>
</table>

Figure 3.2. Delta morphology and sedimentology in Dog Creek basin (lake system 3) (refer to Fig. 3.1B for location of Dog Creek basin and Fig. 3.3C for location of deltas and other deglacial landforms). A. Oblique orthographic projection of a hillshaded digital elevation model (Geobase®) showing five deltas built into the Dog Creek basin. Meltwater channels (MC), now abandoned, once supplied water and sediment to these deltas. GPR grid shown in E was collected on
delta 3. Black arrowed lines show location and direction of topographic profiles in B. Gridlines overlain on DEM show equal areas of 400 m x 400 m. B. Topographic profiles of upper and lower deltas recording sedimentation into high and low stage lakes, respectively. C. Surface boulders at the apex of delta 4 in A. D. Clasts in delta topset material (delta 4 shown here) are typically framework-supported, subrounded basalt pebbles and cobbles (grid has 5 cm subdivisions). E. Selected lines from a pseudo 3-D ground-penetrating radar grid collected from the floor of a gravel pit in delta 3 (refer to appendix E.5 for full grid and text for details of interpretation). Solid black letters refer to interpreted radar elements and thick black lines represent radar bounding surfaces between elements. Dipping, subparallel reflections in lines X3 and X6 are interpreted as foresets (strike/dip of 152°/21°) prograding to the SW. GPR line X9, collected on the undisturbed land surface above the gravel pit, shows sub-horizontal, undulatory reflections characteristic of topset beds.
Figure 3.3. Deglacial landforms and sediments within (A) the Brigade Creek basin (BCB; lake system 1), (B) the Big Bar Lake (BBB), Meadow Lake (MLB), and White Lake (WLB) basins (lake system 2), and (C) the Dog Creek (DCB) and Pigeon Creek (PCB) basins (lake system 3). Dog Creek basin deltas are numbered 1-5. Exposure locations (Ex 1-3) refer to the lithostratigraphic log and images shown in Fig. 3.4. Refer to Fig. 3.1 for regional location of A, B, and C. Note: some spillways may have operated as proglacial rivers in addition to operating as spillways during lake drainage.
Figure 3.4. Exposures of lake sediments and corresponding lithofacies associations (S1-S3, F1-F2). Refer to Fig. 3.3 for exposure locations and Table 3.3 for lithofacies descriptions and interpretations. A. Exposure 1: high energy lake-bottom sediments including type A ripple cross-laminated medium sands grading upwards into Type B and S lamination. B. Exposure 2: low energy lake-bottom sediments including graded fine sandy silt and clay. C. Exposure 3: lithostratigraphic log of subaqueous fan sediment proximal to inflow.

Paleo-water plane definition

Reconstruction of lake surface elevation was completed using both primary (approximation of water plane) and secondary (under or over-estimation of water plane) water plane indicators (Table 3.1). Primary indicators in the study area are limited to foreset/topset contact in Gilbert-type deltas (where visible in section or from GPR survey interpretation) and lake maintenance (non-outburst flood) paleo-spillways. Where the
topset/foreset contact of Gilbert-type deltas was not measurable, the delta brink-point is used as an estimate of the maximum water-plane elevation. Lake spillways in the study area were identified through aerial photograph interpretation of abandoned drainages and the iterative process of flooding DEM surfaces to projected lake levels whilst observing where lake planes intersect potential topographic drainage pathways (cf. LaRocque et al. 2003). Glacial lake outburst flood (GLOF) or outburst flood (lake drainage) spillways are created by sudden, high-magnitude meltwater flows associated with rapid glacial lake or lake drainage (Tweed and Russell 1999; Carrivick and Tweed 2013). Where formed in contact with ice, GLOF spillways can be further subdivided based on their drainage position: supraglacial, ice-marginal, and subglacial (Tweed and Russell 1999). Lake maintenance spillways form where sustained meltwater flow maintained paleolake stillstands. Spillways were classified based on their substrate type (e.g., ice, sediment or bedrock; a proxy for channel stability and likelihood of lake-stage maintenance or lake drainage) and landform associations. Generally lake maintenance spillways are those formed over resistant substrates such as bedrock or bouldery till (such as that common on the Plateau, Huntley 1996). Till-floored spillway stability is facilitated by boulder lags and the typically gentle slopes of the Plateau. GLOF and outburst flood spillways are associated with earthen dam breaches (i.e. eskers, grounding-line moraines, alluvial fans) and/or ice-contact drainage (supraglacial or ice-marginal). Secondary paleo-water-plane indicators used in this study include lake-bottom sediments and subaqueous fan surfaces (both under-estimations of water plane, Table 3.1) as well as the crest elevation of breached earthen dams (overestimations of water plane, Table 3.1). Where deltas or lake maintenance spillways are absent, the highest lake-bottom sediments and subaqueous fans define the minimum lake level elevation. In order to maintain consistency between landforms and lake reconstructions elevation data were extracted from the DEM (Geobase®).

This iterative process of assessing lake extent required the emplacement of dams at points where it was known no lake sediments existed or where the modern DEM failed to correctly model paleolake boundaries because of Holocene erosion subsequent to lake existence. The lack of a topographic dam today at some locations means these lateglacial dams were either formed of sediment that has since been removed, either during lake drainage or Holocene erosion, or of ice (an ice margin or ice block) that has since melted.
Where geomorphic evidence of a sediment dam (e.g. breached moraine ridge) was not observed, and ice-marginal landforms (e.g. ice-marginal channels, moraines, ice-contact deltas) or hummocky terrain were present nearby an ice dam was inferred (Table 3.2).

We use the term lake system to refer to all lakes that evolve within each of three tectonically-warped basins in the study area on the southern Fraser Plateau. Each system includes one or more lakes (named), each within its own topographic sub-basin. Each lake exhibits one or more stages (numbered stillstands) controlled by different paleo-spillways or dams. Lakes and their stages evolved over time as the ice margin position and/or lake basin geometry changed.

The influence of glacioisostatic rebound on paleo-water-plane tilt was modeled by performing first order trend analysis on primary water-plane indicators within the same lake stage. It was tested by comparing results to nearby glacioisostatic tilt estimates (Fulton and Walcott 1975; Johnsen and Brennand 2004). No reliable glacioisostatic tilt could be calculated from the available data and the data did not fit nearby glacioisostatic tilt estimates (Fig. 3.5; elaboration in Appendix G). Therefore, all stillstands are reconstructed from water plane (Table 3.1) and ice contact (Table 3.2) indicators and modelled using a horizontal plane.

**Reconstructing lake size and evolution**

Lake stage extent, depth and therefore volume are calculated by subtracting the modern DEM surface from reconstructed water-plane elevations and are therefore dependent on how truthfully the modern DEM models paleolake bathymetry. We acknowledge that Holocene geomorphic activity, although limited by low plateau slopes, has resulted in erosion of (and some deposition on) the land surface and this is inherited in the DEM. Therefore calculations of lake extent, depth and volume are likely over-estimates of actual values.
Figure 3.5. A. Extent of glacial Lake Dog (stage 2, cf. Fig. 3.8) modelled with zero glacioisostatic tilt and with a regional glacioisostatic tilt of 1.7 m km\(^{-1}\) up toward 321° (cf. Johnsen and Brennand 2004). The tilt plane is anchored at the southeastern extent of the stage 2 water plane (as determined by the extent of glacial lake sediments and reconstructed spillways, see Fig. 3.8). B. Glacioisostatic rebound derived from water plane reconstructions in the Thompson basin.
(dotted line; Johnsen and Brennand 2004) graphed against water-plane indicators in glacial Lake Dog. DS-2 = glacial Lake Dog spillway stage 2. The Thompson basin glacioisostatic tilt does not fit delta elevations. Rather, deltas form two groups (upper and lower) that define two non-tilted lake stages (1 and 2, respectively) maintained by spillways (DS-1a and DS-1b) at stage 1 (spillways located to the east of this figure, Fig. 3.8B), and dammed by Canoe Creek esker at stage 2 (Fig. 3.8C). Stage 2 drained through DS-2 on breaching Canoe Creek esker. However, deltas within these groups are likely non-synchronous (1-2 and 3-5 occur at progressively lower elevations) and built into a lake with a gradually or episodically lowering water surface elevation as the ice margin receded to the northwest (i.e. changes in lake surface elevation are mainly related to changes in lake-basin geometry rather than glacioisostatic rebound).

Three principles guide reconstruction of the relative timing and evolution of lateglacial lake stages (stillstands). i) Within lake systems, stillstands occur when water-plane elevation reaches the level of one or more lake maintenance spillway(s) able to accommodate all lake outflows. This water-plane elevation is maintained until a new, lower spillway is opened by dam (ice or sediment) failure forming a GLOF or outburst flood spillway, or as retreat of the ice margin changes basin geometry or reveals a new lake maintenance spillway. Therefore, successive stillstands exhibit a progressive reduction in water-plane elevation when dammed between a downwasting or retreating ice margin and a positive topographic slope (cf. Fulton 1967; Huntley 1996; Clague and James 2002; Carrivick and Tweed 2013). ii) Cross-cutting geomorphic relationships (e.g. cross-cutting spillways) allow the relative timing of different lake stages within the same system and between lake systems to be determined. iii) Geographically separate lakes (stages) within a lake system are considered essentially coeval when they are dammed by a contiguously reconstructed ice margin within the same basin, or are connected by short spillways that display limited incision (demonstrating the presence of a raised local base level in the form of another lake). Where possible, GLOF or outburst flood volumes were calculated by differencing lake-basin volumes before and after drainage.
Results

Reconstruction of lateglacial lake extent

Identification and classification of paleo-water plane and ice-contact indicators

Paleo-water plane and ice contact indicators were identified in three lake systems within the tectonically-warped topographic basins on the southern Fraser Plateau (Fig. 3.1B). Lake system 1 (LS-1) formed in the eastern part of the study area and at the highest elevation of all three lake systems, within the modern Brigade Creek basin. The Big Bar Lake, Meadow Lake, and White Lake basins just north of the Marble Range comprise the extent of lake system 2 (LS-2). Lake system 3 (LS-3) exists within the Dog Creek and Pigeon Creek basins in the northwest part of the study area, directly north of LS-2. The landforms used to delineate the extent of these lake systems are described below.

Gilbert-type deltas

The most reliable indicators of paleo-water-plane in the study area are Gilbert-type deltas; however, these are only identified in LS-3 (Fig. 3.3), probably because the other two lake systems lacked major channelized meltwater inflows delivering significant sediment supply, were too shallow, or too short-lived to allow formation. Deltas are identified by their fan-like shape, gently-sloping (~1°) plains (topsets), and steep (8.4°-22.15°) terminal slopes (foresets) (Fig. 3.2A, B) (cf. Smith and Jol 1997; Winsemann et al. 2007). They are located at the end of ice-marginal channels (Figs 3.2A, 3C), classified as such because they exhibit reaches that do not follow topographic slope and thus suggest ice-marginal control. Several deltas exhibit numerous enclosed depressions (10-12 m deep) on the delta plain interpreted as kettle holes (cf. Kostic et al. 2005). Only topsets are visible in available delta exposures (>3 m thick) and no macroscale structures were observed, but stacking of imbricate clasts is visible confirming flow directions towards the Dog Creek basin in LS-3 (Fig. 3.3). Sedimentary exposures within delta topsets reveal poorly- to moderately-sorted, framework-supported, subangular to subrounded pebble- to cobble-sized clasts in a granule to coarse sand matrix (Fig. 3.2D, E; G2 in Table 3.3) (cf. topsets of Kostic et al. 2005).
<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Lithofacies description</th>
<th>Unit lower contact</th>
<th>Unit thickness (m)</th>
<th>Lithofacies associations</th>
<th>Lithofacies interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>G1: inclined gravel beds</td>
<td>Inclined gravel beds, only imaged in GPR data (RE-A to C, Figs 3.2, appendix E.5).</td>
<td>No lower contact observed</td>
<td>≥5.0</td>
<td>G2</td>
<td>Foreset deposition via avalanching over delta brinkpoint (Church and Gilbert 1975; Gustavson et al. 1975).</td>
</tr>
<tr>
<td>G2: planar-beded gravel</td>
<td>Poorly to moderately sorted, framework-supported, subrounded pebble to cobble gravel in a coarse sand to granule matrix, rare boulders (RE-D, Fig. 3.2, appendix E.5).</td>
<td>Sharp</td>
<td>≥4.0</td>
<td>G1</td>
<td>Topset deposition from bedload transport in proglacial rivers onto delta surface (Church and Gilbert 1975; Gustavson et al. 1975; Kostic et al. 2005).</td>
</tr>
<tr>
<td>S1: planar-beded sand</td>
<td>Planar-beds (≥1 cm thick) of medium sand, fining upward into fine sand. Very rarely contains clusters of angular pebbles.</td>
<td>Sharp</td>
<td>0.01–0.08</td>
<td>F1, S2</td>
<td>Deposition from traction transport in low-density turbidity flows (Lowe 1982).</td>
</tr>
<tr>
<td>S2: cross-laminated sand</td>
<td>Planar or trough cross-laminated fine or medium sand, typically fining and exhibiting an upward transition from type A to B to S3 upward in section.</td>
<td>Sharp</td>
<td>0.3–0.8</td>
<td>F1, S1</td>
<td>Deposition from traction transport on the waning stage of low-density turbidity flows (Ashley et al. 1982; Winsemann et al. 2007).</td>
</tr>
<tr>
<td>S3: massive silty-fine sand</td>
<td>Structureless silty-fine sand, typically observed at the top of sections. Commonly bioturbated.</td>
<td>Draped or conformable</td>
<td>0.1–0.9</td>
<td></td>
<td>Deposition from wind transport. Aeolian sediment is typical at the land surface in south-central BC (Lian and Huntley 1999).</td>
</tr>
<tr>
<td>F1: normally graded silt with minor fine sand</td>
<td>Normally graded beds of silt with minor amounts of fine sand. Sometimes include flame and dish structures.</td>
<td>Sharp or conformable</td>
<td>0.1–0.9</td>
<td>S1, S2</td>
<td>Deposition by suspension settling in distal lacustrine environments (Smith and Ashley 1985), or during the waning stage of low-density turbidity flows (Lowe 1982). Sometimes deformed by subsequent subaqueous slumping.</td>
</tr>
<tr>
<td>F2: Horizontally laminated clay</td>
<td>Structureless to finely laminated clay.</td>
<td>Sharp</td>
<td>&lt;0.01</td>
<td>F1, S1</td>
<td>Deposition by suspension settling during the waning stages of turbidity flows (Johnsen and Brennand 2006).</td>
</tr>
</tbody>
</table>
Clast lithology is dominantly basalt reflecting local bedrock (Campbell and Tipper 1971), but also includes metamorphic and granitic clasts glacially transported from bedrock source areas to the east and west (cf. Huntley 1996; Plouffe et al. 2011). Rounded basalt boulders, with b-axis diameters of up to 1 m occur where ice-marginal channels merge with delta plains. Four radar elements RE-A to RE-D are interpreted from a pseudo-3D GPR grid collected in a gravel quarry in one of the deltas (Appendix E.4, E.5). RE-A, -B and -C are wedge-shaped elements containing dipping reflections (true dip 21°, striking 152°) that downlap their lower bounding surfaces (the lower bounding surface of RE-A is below the depth of maximum signal penetration) (Fig. 3.2E, Appendix E.5). They are interpreted as recording foresets within separate lobes prograding to the SW, the result of streamflow switching and deposition from avalanching grainflows down a gilbert-style delta foreslope (Jol and Smith 1991) (Fig. 3.2E, Appendix E.5; G1 in Table 3.3). Line X9, taken from an undisturbed section of the delta ~4 m above the pit floor, contains only RE-D, a 3-4 m thick set of mainly discontinuous, sub-horizontal, undulatory reflections with some trough-shaped and downlapping reflections characteristic of gravel sheets with minor scour and fill (Kostic et al. 2005). This radar element is interpreted as topsets accreted in a braided stream environment. The foreset/topset contact is not visible in line X9 because of signal attenuation, but is presumed to occur near 1098 m a.s.l. based on maximum foreset and minimum topset reflection elevations (Fig. 3.2E, Appendix E.5).

**Paleo spillways**

GLOF and outburst flood spillways contain coarse gravel deposits (Carling 2013), and landforms associated with outburst flood discharges, including scabbed topography (Kehew 1982), expansion bars (Maizels 1997) and deeply incised canyons (Lamb and Fonstad 2010). Maintenance spillways are cut in bedrock or bouldery till (Table 3.4) where stability is facilitated by high clast concentrations and the low slopes of the Plateau. Spillways also include: (i) ice-marginal channels - single-walled channels that do not follow topographic slope (cf. Syverson and Mickelson 2009); (ii) proglacial channels - double-walled channels following topographic slope (cf. Kehew 1982); and, (iii) supraglacial channels - where geomorphic evidence for a suitable drainage route is absent but drainage was still necessary between lake stages and evidence for the existence of ice (e.g. hummocky terrain) is present.
### Table 3.4. Physical characteristics and interpretations of lateglacial lake spillways on the southern Fraser Plateau.

<table>
<thead>
<tr>
<th>Spillway name</th>
<th>Lake system</th>
<th>Intrabasin timing</th>
<th>Lake stage name</th>
<th>Channel type</th>
<th>Lake proximal channel substrate</th>
<th>Flow style(^1) (drainage volume (km(^3)))(^2)</th>
<th>Spillway floor elevation (m a.s.l.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>BS-1</td>
<td>LS-1</td>
<td>2</td>
<td>Glacial Lake Brigade</td>
<td>Spillway</td>
<td>Till</td>
<td>Lake maintenance</td>
<td>1162</td>
</tr>
<tr>
<td>BS-2</td>
<td></td>
<td>3</td>
<td>Paleolake Hudson</td>
<td>Spillway (esker breach)</td>
<td>Gravel</td>
<td>Outburst flood (0.002)</td>
<td>1100</td>
</tr>
<tr>
<td>BBS-1</td>
<td>LS-2</td>
<td>1</td>
<td>Glacial Lake Big Bar stage 1</td>
<td>Spillway</td>
<td>Till</td>
<td>Lake maintenance</td>
<td>1122</td>
</tr>
<tr>
<td>BBS-2</td>
<td></td>
<td>2</td>
<td>Glacial Lake Big Bar stage 2</td>
<td>Spillway</td>
<td>Till</td>
<td>Lake maintenance</td>
<td>1108</td>
</tr>
<tr>
<td>MS-1</td>
<td></td>
<td></td>
<td>Glacial Lake Meadow stage 1(^3)</td>
<td>Spillway</td>
<td>Bedrock</td>
<td>Lake maintenance</td>
<td>1100</td>
</tr>
<tr>
<td>BBS-3</td>
<td></td>
<td>3</td>
<td>Glacial Lake Big Bar stage 3</td>
<td>Ice-marginal spillway</td>
<td>Ice then bedrock</td>
<td>GLOF (0.37)</td>
<td>1037</td>
</tr>
<tr>
<td>MS-2</td>
<td></td>
<td></td>
<td>Glacial Lake Meadow stage 2</td>
<td>Spillway</td>
<td>Bedrock</td>
<td>Lake maintenance</td>
<td>1077</td>
</tr>
<tr>
<td>WS-1</td>
<td></td>
<td></td>
<td>Glacial Lake White stage 1</td>
<td>Supraglacial evolving to till-floored spillway</td>
<td>Ice then till</td>
<td>GLOF evolving to lake maintenance</td>
<td>1072</td>
</tr>
<tr>
<td>WS-2a</td>
<td></td>
<td>4</td>
<td>Glacial Lake White stage 2</td>
<td>Ice-marginal spillway</td>
<td>Ice and till</td>
<td>Lake maintenance</td>
<td>1066</td>
</tr>
<tr>
<td>WS-2a</td>
<td></td>
<td>5</td>
<td>Paleolake White</td>
<td>Till-floored spillway</td>
<td>Till</td>
<td>Lake maintenance</td>
<td>1066</td>
</tr>
<tr>
<td>WS-2b</td>
<td></td>
<td>6</td>
<td>Paleolake White</td>
<td>Spillway (grounding-line moraine breach)</td>
<td>Gravel</td>
<td>Outburst flood (0.09)</td>
<td>1042</td>
</tr>
<tr>
<td>LS-1</td>
<td></td>
<td></td>
<td>Paleolake Long</td>
<td>Spillway (grounding-line moraine breach)</td>
<td>Gravel</td>
<td>Outburst flood (0.18)(^4)</td>
<td>991</td>
</tr>
<tr>
<td>DS-1a, b</td>
<td>LS-3</td>
<td>1</td>
<td>Glacial Lake Dog stage 1</td>
<td>Spillway</td>
<td>Till</td>
<td>Lake maintenance</td>
<td>1120</td>
</tr>
<tr>
<td>Spillway name</td>
<td>Lake system</td>
<td>Intrabasin timing</td>
<td>Lake stage name</td>
<td>Channel type</td>
<td>Lake proximal channel substrate</td>
<td>Flow style¹ (drainage volume (km³))^2</td>
<td>Spillway floor elevation (m a.s.l.)</td>
</tr>
<tr>
<td>---------------</td>
<td>-------------</td>
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<td>-----------------</td>
<td>--------------</td>
<td>---------------------------------</td>
<td>--------------------------------------</td>
<td>---------------------------------</td>
</tr>
<tr>
<td>DS-2</td>
<td>2</td>
<td>Glacial Lake Dog stage 2</td>
<td>Spillway (esker breach)</td>
<td>Gravel</td>
<td>GLOF (1.33)</td>
<td>1072</td>
<td></td>
</tr>
<tr>
<td>DS-3</td>
<td>3</td>
<td>Glacial Lake Dog South stage 3b</td>
<td>Spillway</td>
<td>Till</td>
<td>Lake maintenance</td>
<td>1087</td>
<td></td>
</tr>
<tr>
<td>DS-4</td>
<td>4</td>
<td>Glacial Lake Dog stage 4</td>
<td>Supraglacial evolving to till-floored spillway</td>
<td>Ice then till</td>
<td>GLOF</td>
<td>&lt;1004</td>
<td></td>
</tr>
<tr>
<td>DS-5</td>
<td>5</td>
<td>Paleolake Dog</td>
<td>Spillway (alluvial fan breach)</td>
<td>Gravel</td>
<td>Outburst flood (0.03)</td>
<td>910</td>
<td></td>
</tr>
</tbody>
</table>

¹Flow style is defined by substrate type. Erosion resistant substrates (bedrock, till sheet) support lake maintenance or gradual lowering of the water plane over time; less stable substrates (ice, gravel and till ridges i.e., eskers and moraines) support catastrophic drainage when incised, resulting in rapid lake drainage and GLOFs or outburst floods.

²The approximate discharge volume of GLOFs or outburst floods is calculated based on elevation measurements of the pre-drainage water plane elevation and spillway floor elevation from the DEM. This was not possible where the spillway elevation was uncertain.

³Earlier stages of glacial Lake Meadow were likely synchronous with glacial Lake Big Bar stage 1, but no sediments or landforms were found to confirm the co-existence of these still-stands.

⁴This discharge volume includes 0.09 km³ inferred to have discharged from paleolake White into paleolake Long.

**Ice-marginal channels**

Ice-marginal channels represent locations high on valley slopes where water flow along topographic gradient was blocked by ice. These channels occur in each lake system (Fig. 3.3) and their nested relationship allows reconstruction of ice-margin position through time. Channels are only mapped as ice-marginal where they are nested and/or do not follow topographic slope. Throughout the study area ice-marginal channels range in elevation from 1620 to 939 m a.s.l., the largest range of which is visible along the NE side of the Marble Range (Fig. 3.1B). These channels are an average of 1500 m in length, 2-32 m deep, and are horizontally spaced 100-300 m apart and vertically separated by ~30 m. The average bed-slope of ice-marginal channels is 7% (this figure may include subglacial chute reaches which are difficult to separate out from the overall channel). Many abruptly terminate in deltas (Fig. 3.2A, B) or subaqueous fans at the margins of paleolake basins suggesting that water in these channels followed the ice margin until
emptying into an extant ice-dammed lake, which acted as local base-level. When terminating in a delta, the elevation of the topset/foreset contact is used as an estimate of water plane elevation. When terminating in a subaqueous fan, the elevation of the channel mouth floor is taken as an underestimate of water-plane elevation. Ice-marginal channels are also used to locate successive ice-margin positions within and between lake systems.

**Lake-bottom sediments**

Lake-bottom sediments, up to 3 m thick atop bedrock or diamicton, are the most common indicator of the presence or absence of paleolakes in the study area (Fig. 3.3). Two lithofacies associations were identified in all lake systems. Lithofacies association 1 is most frequent and is characterized by cross-laminated medium sand (S2) with silt drapes, thinning upwards into graded fine sandy silt (F1) (Fig. 3.4A; Table 3.3). Lithofacies association 2 is composed of graded fine sandy silt (F1) alternating with clay (F2) (Fig. 3.4B; Table 3.3).

Within a lake-bottom environment ripple cross-laminated medium sand is indicative of deposition from traction transport within turbidity currents (Donnelly and Harris 1989; Winsemann 2007). Graded fine sand and silt beds topped with laminated clay may represent classic glaciolacustrine varves deposited through a combination of seasonal underflow and suspension settling (Shaw 1977; Smith and Ashley 1985), but annual periodicity cannot be confirmed. The rarity of laminated clay within lake-bottom sediment exposures suggests that if couplets represent annual cycles of deposition from suspension settling and turbidity flows (Lowe 1982; Smith and Ashley 1985), the lakes were relatively short lived, on the order of tens of years. This is consistent with a relatively rapid rate of ice-margin retreat, which would have opened up new lake spillways relatively quickly from year to year. Lithofacies association 2 indicates relatively low energy deposition (suspension settling) in an ice-marginal lake, characteristic of an increasingly distal or off-axis position with respect to turbid inflows (Smith and Ashley 1985).

Rare, exotic (igneous intrusive on the southern Fraser Plateau) lonestones that deform or penetrate underlying laminae and are draped by overlying laminae are interpreted as dropstones (Thomas and Connell 1985). Dropstones are present in lake-bottom sediments of LS-1 and LS-3 and are consistent with calving into ice-contact lakes or lake-ice delivery (Bennett et al. 1996).
Subaqueous fans

Subaqueous fans are most common in LS-2 and LS-3 (Fig. 3.3B, C). They are fan-shaped, 0.5-1.9 km² in area, with surface slopes of 1-5°. They occur downflow of abandoned drainage channels. Exposed sediments commonly transition upward from planar-bedded sand to cross-laminated (types A-B-S) sand to normally-graded sandy silt (S1-S2-F1 in Fig. 3.4C; Table 3.3) near the fan apex and grade downslope into laminated finer sediments (e.g. F2 in Table 3.3). Subaqueous fans identified within LS-2 are deposited adjacent to the Chasm esker, a ~40 km long esker that records a subglacial channelized flood from the Pigeon Creek basin (Fig. 3.1B; Burke et al. 2012a). Fans are near orthogonal to esker ridge crestline, stratigraphically above esker sediments, contain sediment sizes inconsistent with esker sediments (the esker is composed of pebble to cobble gravel and the fans are composed of silt to coarse sand (Burke et al. 2012b)), and are connected to abandoned drainage channels incised orthogonally through the esker. Therefore we interpret these fans to have been deposited in an expansive subaerial lake sometime after esker formation (cf. Burke et al. 2012b). The lithofacies associations in these fans (Fig. 3.4C; Table 3.3) are diagnostic of subaqueous formation. The upward transition from planar-beds to cross-laminae (types A-B-S) into normally-graded sandy silt (S1-S2-F1 in Fig. 3.4C; Table 3.3) records deposition from traction and suspension within turbidity flows (Lowe 1982; Winsemann et al. 2007). The downslope grading into finer-grained laminated sediments (e.g. F2 in Table 3.3) indicates the transition to deposition from suspension in lake-bottom deposits (Ashley 1975).

Moraines

Moraines used for lake reconstruction were all found within LS-2 and include a linear diamicton ridge, two stratified sediment ridges, and two areas of hummocky terrain (cf. Huntley 1996) (Fig. 3.3B). Each ridge acted as a dam during lake formation. Hummocky terrain likely represents the meltout of sediment-rich ice (Table 3.2).

The linear diamicton ridge is round-crested, 2500 m long, 400 m wide, 26 m high, and composed of sandy-silt, clast-rich diamicton. It is aligned orthogonal to the axis of the Meadow Lake basin (LS-2, Fig. 3.3B) and has a slightly steeper ice-proximal slope (based on assumed direction of ice retreat from adjacent ice-marginal channels). It is bounded to the northeast by hummocky, kettled topography and is ~10 m above the local bedrock.
surface. The ridge is interpreted as a moraine because it is composed of diamicton and
because it is located amongst hummocky, kettled topography suggesting it formed at, or
near, the ice margin during deglaciation.

Two stratified sediment ridges identified within LS-2 formed within the narrow
White Lake basin. They have well-defined ridge morphology, ranging from 5-15 m high,
25-90 m wide, and 140-630 m long. They exhibit asymmetrical cross profiles: ice-proximal
slopes have angles up to 15°, and ice-distal slopes up to 20°. These ridges contain closed
circular depressions, indicating post-depositional meltout of buried ice blocks. Ridges are
composed of a range of materials including moderately-sorted, sub-rounded pebble to
cobble gravel, laminated silt and fine sand, and diamicton containing angular clasts
exhibiting minor glacigenic wear features such as bullet noses and plucked ends (cf.
Krüger 1984). They are interpreted as grounding-line moraines because of their cross-
valley orientation, asymmetric topographic profiles, kettled surface, and composition
(Sharpe and Cowan 1990; Benn 1996). The two ridges within the White Lake basin exhibit
gaps along the valley axis where they have been cut by post-depositional water flow
inferred to be associated with lake drainage events (outburst flood spillways).

Hummocky terrain capped by large (> 1 m b-axis) boulders of local lithology is
present at the western margins of LS-2, and NE of the moraine in the Meadow Lake basin
(Fig. 3.3B). Limited exposures indicate an internal composition of heterogeneous
diamicton, likely till. Hummocky terrain is interpreted to record the decay of a zone of dirty
stagnant ice (Benn 1992; Huntley 1996). The presence of this landform adjacent to
exposures of lake-bottom sediments, suggests the ice margin, or detached ice blocks,
dammed a lake.

Lake size and evolution

Lake extents are constrained by the elevation of water plane indicators (Table 3.4),
the position at which the reconstructed water plane intersects topography, and the position
of paleo-dams (ice or sediment dams). Modern river drainage in all three paleolake basins
is to the west or northwest (Fig. 3.1), therefore dams were required in the northwest
sectors of each basin in order for paleolakes to have formed, and thus to account for the
geomorphic and sedimentary record. Paleolake extents, volumes and associated dams
and spillways are reconstructed for each lake system (Figs 3.6-3.8; Tables 3.4 and 3.5), and their evolution is described below.

**Paleolake System 1**

Paleolake system 1 contained two lakes (glacial Lake Brigade and paleolake Hudson), one of which is ice-contact; each has one reconstructed stillstand (Fig. 3.6). Meltwater channels eroded into bedrock and, in places, containing gravel deposits, follow topographic slopes and record initial proglacial drainage from the Brigade Creek basin into the Deadman River valley (Figs 3.3A, 3.6B). Ice-marginal channels on the southern slope of the Brigade Creek basin (Figs 3.3A, 3.6C) suggest that the ice margin retreated westward in this basin. The stillstand for glacial Lake Brigade is defined by a till-floored spillway at 1162 m a.s.l. (BS-1, Fig. 3.6C; Table 3.4). No deltas or subaqueous fans were identified in this lake, but lake-bottom sediments occur between 1140 and 1159 m a.s.l. and are spatially restricted to an area west of BS-1 (Fig. 3.6B). Type A and B ripples in lake-bottom rhythmites record eastward paleoflows (Exposure 1, Fig. 3.4), consistent with inflows from an ice margin to the west. Therefore, we infer that as the ice margin retreated west, it exposed a topographic high in the vicinity of BS-1, resulting in the ponding of glacial Lake Brigade between this sill and the ice margin (Fig. 3.6C). Following retreat of the ice margin, the lake gradually lowered below BS-1 and decanted into a lower lake (paleolake Hudson) to the west.

Paleolake Hudson was dammed by an esker ridge (Tipper 1971) at the western end of its basin. The extent of paleolake Hudson is inferred based on the maximum water-plane elevation (1110 m a.s.l.) that could have been retained by the dam (Fig. 3.6D). Poor access and a lack of exposures prevented identification of lake-bottom sediments associated with this lake stage. With no natural outlet, the lake filled from water decanting from glacial Lake Brigade and then local runoff until water level naturally overtopped the esker (or the esker decreased in elevation due to melting of buried ice), resulting in rapid incision (BS-2, Fig. 3.6D; Table 3.4) and release of a small outburst flood into the Loon Lake valley (0.002 km³, Table 3.4). Planar-bedded cobble gravel transitioning upwards into planar-bedded sand is exposed downstream of the esker breach and is likely associated with deposition during the waning stages of lake drainage.
**Figure 3.6.** A. Overview of lateglacial lake system 1 (LS-1). Maximum, non-synchronous lake extents are shown in dark to light blue (early to late evolution). Lake evolution followed westward ice-margin retreat and is detailed in B-D. Refer to text and Table 3.4 and 3.5 for detailed explanation of lake evolution and Fig. 3.1 caption for abbreviations. B. Time 1: Prior to the formation of glacial Lake Brigade proglacial rivers drained southeastward towards Deadman River valley. C. Time 2: Glacial Lake Brigade formed in the Brigade Creek basin and lake level was maintained by northeastward flow through a till-floored spillway (BS-1, 1162 m a.s.l.). Continued westward ice retreat opened up a lower western basin, allowing lake level to gradually fall below the elevation of BS-1 while draining into the basin to the west. D. Time 3: Once ice had evacuated the Brigade Creek basin, paleolake Hudson (1110 m a.s.l.) formed, impounded against an esker. The esker ridge likely eventually failed by piping or was overtopped and breached (BS-2), either by melt-out of ice within the esker ridge resulting in lowering of dam height, or by local runoff raising the level of paleolake Hudson until lake level exceeded the dam height. The breach likely resulted in an outburst flood (0.002 km³) into the Loon Lake valley.
Figure 3.7  A. Overview of lateglacial lake system 2 (LS-2). Maximum, non-synchronous lake extents are shown in dark to light blue (early to late evolution). Lake evolution (expansion, decanting and drainage) followed northwestward ice-margin retreat and is detailed in B-E. Refer to text and Table 3.4 and 3.5 for full explanation of lake system evolution and to Fig. 1.1 caption for abbreviations. B. Time 1: glacial Lake Big Bar stage 1 (S1) developed in the Big Bar basin and lake level was maintained by eastward flow through a till-floored spillway (BBS-1, 1122 m a.s.l.). C. Time 2: lake level in glacial Lake Big Bar stage 2 (S2) was maintained by southeastward flow through a till-floored spillway (BBS-2, 1108 m a.s.l., slightly higher than modern Big Bar Lake). Glacial Lake Meadow stage 1 formed in the Meadow Lake basin and lake level was maintained by southeastward flow through a bedrock spillway (MS-1, 1100 m a.s.l.). D. Time 3: lake level in glacial Lake Meadow stage 2 was maintained by westward flow through a bedrock spillway (MS-2, 1077 m a.s.l.), spilling into the adjacent White Lake basin, forming glacial Lake White stage 1. Glacial Lake White filled until it overtopping its ice dam to the south forming a supraglacial, evolving to ice-marginal, channel (WS-1) as flow incised through the ice dam. Flow through WS-1 (floor at 1072 m a.s.l.) maintained lake level in glacial Lake White stage 1 and entered glacial Lake Big Bar stage 3 (1072 m a.s.l.; bracketed between the lowest ice-marginal channels and the maximum elevation of lake-bottom sediment and spillway elevations), a lake dammed by a detached ice block to the south. As this ice dam downwasted it was either overtopped or floated, and glacial Lake Big Bar stage 3 drained (GLOF volume 0.37 km³) to the south (BBS-3) into the Fraser River basin. E. Time 4: Thermo-mechanical melting of the stagnant ice block adjacent to WS-1 opened a second, lower ice-marginal channel (WS-2a into the Big Bar basin and eventually into the Fraser River valley, 1066 m a.s.l.), leading to the development of glacial Lake White stage 2. Glacial Lake White stage 2 was impounded by ice and a grounding line moraine to the west. F. Time 5: Palaeolake White formed following westward ice retreated. It was dammed by a grounding line moraine and its stage (1066 m a.s.l.) was maintained by flow through WS-2a. Ice retreat stabilized at a point further west where it impounded glacial Lake Long (1060 m a.s.l.) and formed a second grounding line moraine across White Lake valley. Glacial lake Long was a closed-basin lake. G. Time 6: Following ice retreat from White Lake valley, palaeolake Long (1060 m a.s.l.) was impounded by the second grounding line moraine; it was a closed-basin lake. Kettle holes on the west side of the palaeolake White earthen dam (grounding line moraine) indicate meltout of ice in this ridge, which resulted in crestandline lowering, dam breaching (WS-2b) and drainage of glacial Lake White stage 2 (outburst flood volume 0.09 km³), likely into palaeolake Long. In domino-fashion, the rapid rise in water level in palaeolake Long likely resulted in its drainage due to overtopping and breaching (LS-1) of its earthen dam (grounding-line moraine) sending water (0.18 km³) west towards the Fraser River valley.
Figure 3.8. Overview of lateglacial lake system 3 (LS-3). Maximum, non-synchronous lake extents are shown in dark to light blue (early to late evolution). Lake evolution (expansion, decanting and drainage) followed northwestward ice-margin retreat and is detailed in B-F. Refer to text and Table 3.4 and 3.5 for an explanation of lake evolution, and to Fig. 3.1 caption for abbreviations. B. Time 1: glacial Lake Dog stage 1 formed in the Dog Creek basin and lake level was maintained by southeastward and southwestward flow through till-floored spillways (DS-1a and 1b, respectively at 1120 m a.s.l.). C. Time 2: glacial Lake Dog stage 2 expanded to the northwest, allowing the successive formation of deltas 3-5 (stage 2 reconstruction is based on delta 4, but lake stage lowered between deltas 3 through 5 as the lake expanded westward). Glacial Lake Dog stage 2 also decanted into the Pigeon Creek basin, where it was impounded behind Canoe Creek esker (CCE, crest elevation of 1098 m a.s.l.). It drained (GLOF volume 1.33 km3) down Canoe Creek valley (DS-2) once the esker was overtopped and breached, likely due to esker surface lowering from ice block melt-out. D. Time 3: glacial Lake Dog stage 3 shown after GLOF drainage from stage 2 was complete. Glacial Lake Dog is temporarily divided into north and south sub-basins by an interfluve. Glacial Lake Dog North (stage 3a, 1097 m a.s.l.) has no outlet. Glacial Lake Dog South (stage 3b, 1087 m a.s.l.) is maintained by a till-floored spillway (DS-3, 1087 m a.s.l.). E. Time 4: glacial Lake Dog stage 4 formed, impounded by remnant ice in the Dog Creek basin. It likely drained (discharge volume unconstrained) supraglacially (<1004 m a.s.l., from the channel floor elevation of inflowing ice-marginal channels) west through the Dog Creek valley into the Fraser River valley. F. Time 5: once ice had evacuated the Dog and Pigeon creek basins water supplied from the plateau surface to the north was channeled down Brigham Creek valley (BCV), forming an alluvial fan at its juncture with Dog Creek basin. This fan (932 m a.s.l.) impounded paleolake Dog. Overtopping of this sediment dam resulted in a dam breach, formation of DS-5 (910 m a.s.l.) and catastrophic drainage (outburst flood volume 0.13 km3) from paleolake Dog down Dog Creek valley and into the Fraser River basin. This flow deposited an expansion bar and produced a boulder lag downflow of the breach.

Paleolake System 2

Paleolake system 2 was comprised of four ice-contact lakes (glacial lakes Big Bar, Meadow, White and Long) and two non-glacial lakes (paleolakes White and Long) (Fig. 3.7A-E). Glacial Lake Big Bar formed north of the Marble Range below the lowest ice-marginal channels on these slopes and was dammed against regional topographic slope to the north and east (Figs 3.1, 3.7). The stage 1 extent of glacial Lake Big Bar (Table 3.5) is constrained by the highest lake sediments (1122 m a.s.l.) in this lake system, and its
lake level was maintained by a till-floored spillway (BBS-1, Fig. 3.7B; Table 3.4). Stage 1 is inferred to have been dammed by ice to the northwest because till is present at elevations that would otherwise have been inundated by lake water (above the elevation of stage 2 of this lake), and the spillway (BBS-1) would not have been formed without an ice-dam present in the basin. Furthermore, ice-marginal channels on the northeast side of the Marble Range (Figs 3.1B, 3.3B, 3.7B, D) record successive ice-margin positions from southeast to northwest in this region of the Plateau. From their southeasterly path paralleling the lower slopes of the Marble Range, these channels plunge downslope to the northeast, suggesting that the ice margin extended normal to the southeast trend of the Marble Range, and provided a dam against regional slope. A second, lower stillstand of glacial Lake Big Bar (glacial Lake Big Bar stage 2, Fig. 3.7C; Table 3.5) is reconstructed based on the presence of a lower, till-floored spillway (BBS-2, Table 3.4), the inferred pattern and direction of ice-retreat based on nested ice-marginal channels and the presence of a modern, westward draining lake in this basin (which implicates the necessity of a paleo-ice dam to impound a larger lake). Stage 2 of glacial Lake Big Bar was larger than stage 1 (Table 3.5). Lake-bottom sediments at 1072 m a.s.l. provide a minimum elevation for the lowest, and largest stillstand of this lake (glacial Lake Big Bar stage 3, Fig. 3.7D; Table 3.5). Remnant hummocky terrain (Huntley)

Table 3.5. Physical characteristics of lateglacial lakes on the southern Fraser Plateau.

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<th>Lake system</th>
<th>Intrabasin timing</th>
<th>Lake name</th>
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*aPaleolake depth and volume estimates are derived from modern DEM elevations and the horizontal water plane. Therefore, post-lake incision and basin fill contribute towards error in these estimates. However, relatively minor modern drainages occupy paleolake basins so these over- and under-estimates are likely within the error of the DEM (±10 m vertical, at 99% of data points, Geobase®).
1996) suggests a large zone of stagnant ice constrained the northwest extent of this lake. Stage 3 drained through BBS-3, a steep-walled bedrock canyon that initially acted as an ice-marginal spillway (Fig. 3.7D; Table 3.4); ice filled the broader, shallower valley to the east (Fig. 3.3B). Ice-marginal channels on the southwestern slopes of the Marble Range east of BBS-3 (Fig. 3.3B) are oriented northeast to southwest, confirming successive southeast to northwest ice margin positions towards glacial Lake Big Bar (stage 3), and suggesting that the ice occupying the valley adjacent to BBS-3 was probably a detached ice block. A nearby kettled terrace suggests the presence of inactive ice (Fig. 3.7D). Drainage of glacial Lake Big Bar (stage 3) past its southern ice dam through Big Bar Creek and into the Fraser River basin (Figs 3.3B, 3.7D) likely occurred as a GLOF (1.33 km$^3$, Table 3.4) based on potentially rapid ice-marginal incision through the existing ice dam and the deeply incised channel downflow of the bedrock spillway (visible on Fig. 3.1B; though some of this incision is likely from early Holocene river processes, the current discharge through the valley being low).

Glacial Lake Meadow stage 1 was the largest lake reconstructed in this paleolake system (63.6 km$^2$, 34 m max. depth, Fig. 3.7C; Table 3.5). Stage 1 is constrained by secondary water-plane indicators (subaqueous fans and lake-bottom sediments) at 1090-1100 m a.s.l., and was maintained by eastward flow through a bedrock-floored spillway (MS-1, Table 3.4) across a drainage divide (an earlier subglacial erosional corridor associated with Chasm esker, Burke et al. 2012a) (Fig. 3.7C). It was dammed by the ice margin to the west as inferred from the presence of a moraine ridge, and consistent with the inference of an in-line ice margin damming glacial Lake Big Bar stage 2 (Fig. 3.7C). Lake-bottom sediments within glacial Lake Meadow stage 1 abruptly terminate at the edge of large depressions that surround Chasm esker suggesting ice blocks marginal to the esker prevented deposition there during lake occupation (Fig. 3.7C, cf. Burke et al. 2012b). Subaqueous fans adjacent to Chasm esker (Fig. 3.7C) likely developed during the drawdown of lake level between lake stages 1 and 2 (Fig. 3.7C, D): water ponded on the northeast side of Chasm esker drained over low points in the ridge forming short channels dissecting the ridge and terminating in subaqueous fans on the southwest side of Chasm.
esker. Glacial Lake Meadow stage 2 is reconstructed based on the presence of subaqueous fans (elevations 1074-1075 m a.s.l.) and a bedrock-floored spillway (MS-2, Figs. 3.3B, 3.7D; Table 3.4), which allowed water to decant west into the White Lake basin. Co-existence of these lakes is supported by the limited length and depth of MS-2, suggesting it was connected to the local base-level provided by glacial Lake White stage 1.

Two stages of glacial Lake White are reconstructed based on the presence of spillways, grounding-line moraines and a subaqueous fan. Ice-marginal channels to the north of the White Lake basin (Fig. 3.3B) confirm successive east to west positions of the ice margin in this region. The extent of glacial Lake White stage 1 (Table 3.5) is constrained by the existence of a till-floored spillway (WS-1, Table 3.4) which supported initial flow south into glacial Lake Big Bar stage 3 (Fig. 3.7D). Co-existence of these lakes is supported by the limited length and depth of WS-1, indicating flows encountered a local base-level (glacial Lake Big Bar), limiting incision (Fig. 3.7D). Spillways WS-1 and WS-2a are nested and sub-parallel to one another but are at significantly different elevations and their path does not follow regional topography. These characteristics are best explained by invoking ice-marginal drainage against a detached ice mass to the southeast (Figs 3.3B, 3.7D). Prior to the occupation of WS-1 and the glacial Lake White stage 1 stillstand, the route where WS-1 now sits was likely blocked by ice. Glacial Lake White stage 1 filled until it overtopped ice to the south (the lowest available drainage route, Table 3.5). Initial supraglacial drainage likely resulted in a small southward GLOF (volume unconstrained) into glacial Lake Big Bar stage 3, until the spillway stabilized as an ice-marginal, till-floored channel (WS-1, Table 3.4). Thermo-mechanical erosion from ice-marginal drainage through WS-1 led to the development of the lower ice-marginal, till-floored spillway WS-2a (Table 3.4). Glacial Lake White stage 2 (Table 3.5) was dammed by ice to the west, and its water level was maintained by this lower spillway (WS-2a, Fig. 3.7E). WS-2a can be traced to the intersection of modern drainage within Big Bar basin towards the Fraser River valley, suggesting that glacial Lake Big Bar stage 3 had likely drained around the time or shortly after this spillway was established. A grounding-line moraine was deposited at this ice margin. As ice retreated further west, paleolake White was impounded by this grounding-line moraine (crestline at 1066 m a.s.l.) and likely maintained its outlet at WS-2a until water levels fell below this outlet (Fig. 3.7F). Further west, glacial Lake Long was
dammed between the ice margin and the earthen dam for paleolake White; it had no apparent outlet. A second grounding-line moraine was deposited at this new western ice margin (Fig. 3.7F). Following further ice retreat out of the basin, paleolake Long was impounded behind this second grounding-line moraine (crestline 1060 m a.s.l., Fig. 3.7G). The grounding line moraine that impounded paleolake White contains kettle holes in its west side and an undulatory ridge surface, suggesting meltout of buried ice. Associated dam lowering may have triggered westward drainage of paleolake White (WS-2b, Fig. 3.7G; Table 3.4) resulting in a small outburst flood (0.09 km³, Table 3.4). The grounding-line moraine impounding paleolake Long does not exhibit kettle holes and therefore likely failed due to piping or overtopping (Clague and Evans 2000). The latter scenario is best explained by the co-existence of the two paleolakes, with the drainage of paleolake White sending an influx of water into paleolake Long forcing the overtopping and breaching of its moraine dam and resulting in an outburst flood (0.18 km³, Table 3.4) down the Canoe Creek valley towards the Fraser River valley (Fig. 3.7G).

**Paleolake System 3**

Paleolake system 3 includes one large ice-contact lake (glacial Lake Dog) and one non-glacial paleolake (paleolake Dog), all topographically-confined to the Dog Creek and Pigeon Creek basins (Figs 3.1B, 3.3, 3.8) with an inferred ice-dam to the northwest. Stage 1 (Table 3.5) of glacial Lake Dog is ~70 km² in area and is constrained by two deltas (brinkpoint elevations for deltas 1 and 2 are 1122 and 1121 m a.s.l., respectively) (Fig. 3.2B; labelled 1 and 2 in Figs 3.2A, 3.8B), lake-bottom sediments (1120 m a.s.l.), and two till-floored spillways in the southeast (DS-1a) and southwest (DS-1b) (Fig. 3.8B; Table 3.4). In order for this lake to have formed, an ice-dam, demarcated by the ice-marginal channel connecting to delta 2 (Figs 3.3C, 3.8B), must have been present to the northwest. Because lake-bottom sediments are not found at similar elevations within the Pigeon Creek basin to the south, this suggests that this lake stage was largely restricted to the northern Dog Creek basin. Glacial Lake Dog, stage 2 is the largest of all lakes reconstructed in this study (115.9 km², Fig. 3.8C; Table 3.5). Its water plane elevation is constrained by deltas (3-5 in Figs 3.2A, 3.3C, 3.8C; foreset/topset contact of delta 3 is between 1096 and 1098 m a.s.l., Figs 3.2E, 3.5B) and lake-bottom sediments (1080-1098 m a.s.l., Fig. 3.8C). Delta brinkpoint elevation decreases to the northwest (1101-1096 m a.s.l., Figs 3.2B, 3.5B), and we infer from this that deltas were non-synchronously formed.
Rather, they were built in series into an episodically or gradually lowering glacial Lake Dog as the ice margin receded to the northwest and the geometry of the lake basin changed. Stage 2 is reconstructed at the brinkpoint elevation of delta 4 (1098 m a.s.l., Table 3.5) for convenience, but acknowledge the evolving nature of the lake during this time. This stage is dammed by ice to the west and Canoe Creek esker (Fig. 3.8C) to the south. The ice-margin position is demarcated by the ice-marginal channel connecting to delta 5 (Figs 3.2A, 3.3C, 3.8C). Glacial Lake Dog stage 2 was too low to access spillway DS-1a but the Canoe Creek esker is breached by a gravel-floored spillway (DS-2, Fig. 3.8C; Table 3.4). Water likely accumulated in glacial Lake Dog stage 2 until the crestline of the Canoe Creek esker was lowered due to the melting of buried ice (a more likely scenario than lake filling to the point of overtopping the esker, based on dynamic lowering of the overall lake system as evidenced by successively lower delta elevations to the NW), resulting in a GLOF (1.33 km³, Table 3.4) down Canoe Creek valley (Fig. 3.8C).

The drainage of glacial Lake Dog stage 2 and the corresponding lowering of the water plane cut-off the northern arm of the lake at a drainage divide (1097 m a.s.l.), temporarily leaving it with no outlet (glacial Lake Dog North stage 3a, Fig. 3.8D; Table 3.5). Drainage from stage 2 continued with water in the Pigeon Creek basin draining south through the breached Canoe Creek esker until water level lowered to the elevation of a topographic sill (DS-3, Table 3.4) forming glacial Lake Dog South stage 3b (Table 3.5). The remaining water to the south of the sill evacuated through the Canoe Creek esker breach and resulted in incision of DS-2 to 1072 m a.s.l. (DS-2, Table 3.4). The disconnection between the northern and southern arms of glacial Lake Dog was only temporary; the basins were reconnected as ice retreated to the northwest (Fig. 3.8E). The water plane of glacial Lake Dog stage 4 (Fig. 3.8E; Table 3.5) is constrained by the elevation of the mouth of an ice-marginal channel at its outlet into the Dog Creek basin (1004 m a.s.l.). The mouth of this ice-marginal channel is in line with the apex of a subaqueous fan (950 m a.s.l.) deposited in the lake. Assuming that the ice-marginal channel was graded to stage 4 lake level, the channel mouth elevation underestimates lake level (Table 3.1) and thus the mapped extent of stage 4 is a minimum (Fig. 3.8E). Ice-marginal channels demarcate the ice margin (dam) to the west of the lake and lake-bottom sediments (observed at elevations ranging from 901-950 m a.s.l.) in the bottom of the basin (Fig. 3.8E) are present further east than the extent of glacial Lake Dog stage 3.
(Fig. 3.8D) and above the level of paleolake Dog (see next paragraph), indicating that stage 4 represents an independent stillstand. No significant geomorphic evidence for a spillway associated with this lake stage was found in the Dog Creek valley walls between the point where lake sediments terminate and the junction with the Fraser River. Because regional gradient necessitates drainage out of this outlet, and no geomorphic evidence is diagnostic of a spillway for time 4 (Fig. 3.8E) we presume DS–4 must have operated as a supraglacial (<1004 m a.s.l., based on reconstructed lake extent) spillway (discharge volume unconstrained).

During lateglacial drainage a cross-valley alluvial fan at the outlet of Brigham Creek valley (Fig. 3.8F) formed in the Dog Creek basin, damming a lake to the east (paleolake Dog). Observations of lake-bottom sediments confirm its presence (Fig. 3.8F). This lake could have attained a water plane elevation of 932 m a.s.l. and an area of 4.7 km² (Table 3.5) before overtopping the fan. The fan has been incised on its southern edge to an elevation of 910 m a.s.l. This incision event likely records fan overtopping and breaching associated with drainage of paleolake Dog (DS–5, Fig. 3.8F; Table 3.4). Downflow of the spillway (DS–5) an expansion bar and boulder lag exist (Fig. 3.8F). These suggest relatively rapid incision of the sediment dam, releasing water down the lower Dog Creek valley as an outburst flood (0.13 km³, Table 3.4).

Discussion

Regional pattern of CIS retreat

Our data support a lateglacial regional northwestward retreat of the CIS across the southern Fraser Plateau rather than regional stagnation (cf. Fulton 1967); ice stagnated locally in areas of kettled topography. The emergent pattern of intrabasin lake development suggests successive ice-marginal lakes expanded progressively west and northwestward across the southern Fraser Plateau. The evolving position of the ice margin, as it formed an effective dam for these lakes, places the ice margin consistently on the west-northwest side of evolving lakes. This suggests the presence of a contiguous ice mass receding to the west-northwest during deglaciation. This general pattern of retreat is in agreement with that inferred for nearby Young Lake basin (Perkins et al. 2013)
(Fig. 3.1B) and inferences of glacioisostatic tilt from shorelines in the Thompson (Johnsen and Brennand 2004) and Nicola (Fulton and Walcott 1975) basins to the south.

Interbasin (between lake systems) northwestward ice retreat and consequently sequential development of lake systems 1-3 is supported by relative elevation and cross-cutting relations. LS-1 formed at the highest elevation and is also the furthest east. Downwasting of the ice sheet would have deglaciated higher elevation areas before lower elevations were ice free (Fulton 1991; Margold et al. 2014). Therefore we infer that LS-1 was the first of the lake systems to develop.

Both paleolake Long (LS-2, Fig. 3.7E) and glacial Lake Dog stage 2 (LS-3, Fig. 3.8C) drained into the Fraser River valley via the southern portion of Canoe Creek valley, allowing the use of cross-cutting geomorphic relationships to determine their relative timing. The head of Canoe Creek valley is a straight, 8.5 km long, 250-600 m wide, recessional cataract averaging ~60 m deep (Figs 3.1B, 3.9). The near-vertical walls of this valley are composed of columnar basalt (Campbell and Tipper 1971) overlain by till (Fig. 3.9D). North of the cataract knickpoint a wide, flat-bottomed and scabbed system of braided channels emanates from the Pigeon Creek basin (Fig. 3.9B). These channels are largely incised into glaciolacustrine silts from glacial Lake Dog but also cut through Canoe Creek esker (Fig. 3.9B). Immediately down-flow (south) of the knickpoint is a semi-lunate-shaped bar form with a planar surface that dips gently to the southwest and is covered with metre-scale boulders (Fig. 3.9D) comparable to the boulder-covered expansion bars commonly associated with outburst floods (cf. Maizels 1997). The Canoe Creek fan is located at the confluence of Canoe Creek and Indian Meadows Creek valleys (Fig. 3.9A). Significantly, the fan has been incised by drainage from glacial Lake Dog and not from paleolake Long, suggesting it was deposited after the drainage of paleolake Long. On the basis of this we infer that all lakes in LS-2 had drained at least by the time glacial Lake Dog stage 2 in LS-3 existed. This confirms northwestward ice margin recession between lake systems 2 and 3.

Lake system 3 contained the only lake (glacial Lake Dog) in which paleo-deltas were observed (Fig. 3.8). Interestingly the deltas all formed on the north side of the lake basin. This is consistent with the regional picture of northwesterly ice retreat as the largest meltwater and sediment flux would likely be delivered from the north/northwest of the lake
basin where the largest amounts of remnant ice existed. Regionally, this parallels the pattern observed in the Thompson valley to the south where the bulk of delta sedimentation into glacial Lake Thompson occurred from tributaries to the north of the valley (Johnsen and Brennand 2006). Also, paleo-deltas in LS-3 decrease in elevation to the northwest indicating a gradual lowering of lake level as glacial Lake Dog grew in that direction, following the retreating and lowering ice margin.

Figure 3.9. Geomorphic relationship between drainageways of lake systems 2 and 3. Refer to Fig. 3.1 for location of A. A. Perspective view of cross-cutting drainageways from paleolake Long (LS-2, time 6, Fig. 3.7G) and glacial Lake Dog, stage 2 (LS-3, time 2, Fig. 3.8C). Both drainages emptied into the Fraser River valley to the west. Canoe Creek valley is 8.5 km long from DS-2 to Canoe Creek fan apex. Water from paleolake Long drained through Indian Meadows valley prior to the formation of the Canoe Creek fan. The GLOF from glacial
Lake Dog stage 2 down Canoe Creek (northeast to southwest) removed the western side of the Canoe Creek fan, consequently paleolake Long drained prior to glacial Lake Dog stage 2. B. Scabbed zone upflow from drainage knickpoint and esker. C. View of the edge of the spillway wall just downflow of expansion bar from the top of the Canoe Creek esker (dam for glacial Lake Dog stage 2). D and E. Bouldery expansion bar and spillway wall composed of columnar basalt topped by till (boundary denoted by black dashed line).

**Implications of regional glacioisostatic tilt patterns**

The glacial lakes defined and described in this paper formed in areas occupied by erosional corridors and esker systems previously interpreted to have formed under relatively thin ice (Burke *et al.* 2012a). We were unable to define a glacioisostatic tilt from primary water-plane indicators on the southern Fraser Plateau (Fig. 3.5, and Appendix G). A zero tilt was also inferred on glacial lake water planes in the nearby Fraser River valley (Huntley and Broster 1994). Thin ice and zero tilt on glacial lake water planes on the southern Fraser Plateau may suggest that glacioisostatic rebound was complete prior to lake development, confirming rebound must have occurred rapidly as suggested for most of southern British Columbia (cf. Clague and James 2002). Should zero tilt be real and not an artefact of the poor spatial (and temporal) resolution of the data, the fact that glacial lakes in the Thompson and Nicola valleys to the south record significant rebound (Fulton and Walcott 1975; Johnsen and Brennand 2004) suggests that the timing of those lakes was just early enough to be affected by glacioisostatic rebound. Sufficient time must have elapsed between the formation of these southern valley lake systems and the development of the plateau lake systems in this study for the majority of glacioisostatic rebound to be complete. That is, the glacial lakes on the southern Fraser Plateau were younger than the glacial lakes in the Thompson and Nicola basins. This reasoning reinforces the interpretation of regional northwestward ice retreat across the Nicola and Thompson basins and the southern Fraser Plateau, rather than regional downwasting into valleys (cf. Fulton 1967, 1975, 1991).

A lateglacial regional northwestward retreat of the CIS across the southern interior implies a southeastward ice surface slope. This slope aspect is consistent with the pattern of eskers (Plouffe *et al.* 2011). It is also consistent with numerical and thermo-mechanical
models of the CIS (e.g. Roberts 1991, Tarasov and Peltier 2006) and supports the recent hypothesis of CIS decay by a combination of downwasting and frontal retreat to the Coast Mountains (Margold et al. 2013) in a pattern and style similar to the Fennoscandian Ice Sheet (Boulton et al. 2001). The formation of ice-marginal lakes at the margins of the receding CIS is also predicted by numerical models of North American ice sheet hydrology (Tarasov and Peltier 2006). Inferences on ice margin positions, ice-surface slope and isostatic adjustment derived from plateau-lake reconstructions should be included in future numerical or thermo-mechanical models of the CIS.

**Broader implications for ice sheet decay and dynamics**

Currently >35% of meltwater streams from land-terminating portions of the Greenland Ice Sheet (GrIS) end in ice-contact lakes (Lewis and Smith 2009), a number currently on the rise (Carrivick and Quincey 2014). These ice-contact proglacial lake systems appear to maintain many of the same dynamics as their marine-terminating counterparts (Warren 1991; Trüssel et al. 2013). Although flotation of the ice margin (and therefore efficient calving), was not likely for most of the ice-marginal lakes of the southern Fraser Plateau given the shallow water depths involved (Table 3.5), other similarities in behaviour were likely. For example, decreased effective pressures and a locally raised water table likely contributed to enhanced sliding rates in lake-terminating sectors of the ice sheet (Benn 1996; Tsutaki et al. 2011). Furthermore, the lower albedo and low thermal conductivity of proglacial lakes (in contrast to bare sediment or ice) would have moderated local air temperatures and contributed to enhanced thermo-mechanical melting of the in-contact ice margin (e.g., through notching just below the water level, Carrivick and Tweed 2013) increasing melt rates for lake-terminating ice on the Plateau, and accounting for the complex and transitory evolution of the lake systems. The implication of these effects is to enhance glacier retreat, at least partially decoupling it from broader climate patterns (Carrivick and Tweed 2013; Margold et al. 2014). Paleogeographic reconstruction of lake volumes and depths is therefore critical for understanding not just ice-margin position, but also retreat dynamics associated with lake-terminating glacial systems.
Conclusions

The decaying CIS left a series of ice-marginal (ice-dammed) lakes both in the deep valleys and on the Plateau surfaces of B.C.’s southern interior (Fulton 1965; Johnsen and Brennand 2004; Perkins et al. 2013). Ice-dammed lakes on the southern Fraser Plateau developed as generally shallow (average 17 m deep), but extensive (up to 115 km²) water bodies due to the gently sloping nature of the Plateau surface. These lake systems evolved in a southeast to northwest pattern as the ice margin retreated to the northwest. The identification of regional systematic retreat of the CIS margin across the southern Fraser Plateau is consistent with patterns of glacioisostatic rebound reconstructed in the nearby Thompson valley (Johnsen and Brennand 2004) and recently suggested models of lateglacial CIS reorganization (Margold et al. 2013) but contrasts with previous suggestions of regional stagnation with localized ice margin retreat (cf. Fulton 1967). It suggests that although the ice may have been thin at the time of lake formation (c.f. Huntley 1996; Burke et al. 2012a), the ice slope was sufficient to maintain systematic ice margin retreat. The fact that the lakes contain no record of glacioisostatic tilt confirms that the ice was thin when the lakes were extant and suggests that most glacioisostatic rebound had occurred prior to lake formation on the southern Fraser Plateau. Consequently, glacial lakes in the deep valleys to the south (Thompson and Nicola valleys, Fulton 1969; Fulton and Walcott 1975; Johnsen and Brennand 2004) that do record shoreline deformation, likely occurred earlier than those on the southern Fraser Plateau.

The success of this study in defining the retreating ice margin in an area where large recessional moraines are absent suggests that regional patterns of ice-dammed lake development can be integral in reconstructing the pattern and relative timing of deglaciation in areas where traditional ice-marginal indicators are absent or data resolution issues prevent large scale mapping of minor moraine systems. The process of using cross-cutting drainages between lake systems supported by geomorphic evidence is a useful way to internally corroborate intra- and inter-lake system patterns of ice margin retreat. This is especially true where repetitive drainage events (and routes), typical of GLOFs and outburst floods, provide increased opportunity to develop these cross-cutting relationships. Future work should explore the chronology of paleolake evolution.
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4. Active ice-marginal retreat at the interior of the last Cordilleran Ice Sheet: Glaciodynamic inferences from moraine systems on the southern Fraser Plateau, British Columbia, Canada
Abstract

Reconstructing the pattern and style of paleo-ice sheet retreat relies heavily on the identification and genetic interpretation of moraine systems. Where moraine systems are absent from areas that have undergone deglaciation, accurate reconstruction is considerably more difficult and specific glaciodynamic conditions (e.g., regional stagnation) are often inferred in order to explain their absence. The southern Fraser Plateau region within the footprint of the last Cordilleran Ice Sheet (CIS) is previously reported to have very few moraines. Extant moraine systems are systematically mapped from recently improved DEM and aerial photograph datasets, and classified based on morphology, and their internal composition is analyzed using shallow geophysics (electrical resistivity tomography, ground-penetrating radar), sedimentary exposures, and well logs. Based on morphology, sedimentary architecture and sedimentology, glaciotectonic moraines, grounding-line moraines, and esker-like crevasse fills are identified on the southern Fraser Plateau. Overridden moraines yield information about ice advance or subglacial dynamics, while pristine moraines indicate conditions during ice retreat. Genetic interpretation of these moraines suggests an active ice margin both during advance and retreat of the last CIS over the southern Fraser Plateau; stagnation was localized during deglaciation. Significantly, active moraine forming processes inferred to be functioning during deglaciation suggest a northwest to southeast sloping ice surface and a systematic northwestward retreat of the ice margin toward the Coast Mountains.
Introduction

Moraine ridges are commonly used to constrain the position of the ice margin during the retreat of major ice sheets (e.g., Bennett 2001, Bradwell et al. 2008, Kleman et al. 2010, Ó Cofaigh et al. 2012). In the absence of these moraine systems our ability to reliably reconstruct the ice margin at various time intervals is confounded. Regional stagnation (e.g., Flint 1929, Fulton 1967, 1975) or stagnation zone retreat (e.g., Flint and Gebert 1976, Koteff and Pessl 1981, McCabe and Ó Cofaigh 1996), or decay in association with the presence of a hydrologic system capable of efficient sediment transport away from the ice margin (e.g., Swift et al. 2002) are all common explanations for glaciated landscapes where moraines are absent. Recent work has emphasized the critical importance of moraines to inferring ice-margin glaciodynamics, especially where the moraines exhibit glaciotectonic deformation (Laberg et al. 2007, Golledge and Phillips 2008); affect the distribution and production of proglacial sediment (Knight et al. 2007); and are part of larger glacial flowsets (Evans et al. 2008). Furthermore, some moraine (ridge)-like forms have been attributed to subglacial processes. For example, ribbed moraine (sensu Dunlop and Clark 2006), has been linked with extending flow at basal thermal transitions (e.g., Hättestrand and Kleman 1999, Laberg et al. 2007), and compressive flow at topographic obstructions (e.g., Bouchard 1989) or resulting from bed dewatering (Stokes et al. 2008).

The existence and genesis of moraine systems is re-evaluated for the southern Fraser Plateau in south-central British Columbia (BC), near the geographic centre of the last Cordilleran Ice Sheet (CIS), where past research has highlighted a lack of large moraine ridges and the localized presence of ribbed terrain (e.g., Tipper 1971a, b, Plouffe et al. 2011, Kleman et al. 2010). Our research suggests (foreshadowed by Tipper (1971a)) that the perceived absence of moraines more reflects the difficulty in identifying relatively low relief moraine ridges in heavily forested terrain during small scale, regional mapping campaigns, rather than a true absence. We find that medium-to-large scale mapping (e.g., Plouffe 2009) and investigations into internal architecture and stratigraphy of moraine ridges provide critical advances in understanding ice dynamics and retreat patterns for this sector of the CIS.

In order to infer glaciodynamic conditions from moraine ridges, their morpho-sedimentary character must be understood (Bennett 2001, Menzies 2001, Lindén et al. 2007). Where sedimentary exposure is limited, characterization of sedimentary architecture and composition has successfully been accomplished with the aid of shallow geophysics such as ground-penetrating radar (GPR; e.g., Lonne and Lauritsen 1996, Bakker and van der Meer 2003, Bennett
et al. 2004, Parkes et al. 2009) and electrical resistivity tomography (ERT; e.g., Kristensen et al. 2009, Langston et al. 2011, Moore et al. 2011), although many of the latter investigations are primarily concerned with characterizing groundwater flow through moraine systems. We apply shallow geophysics and classic sedimentology to previously mapped and newly discovered moraine systems on the southern Fraser Plateau, in order to elucidate their composition and sedimentary architecture. The resulting observations lead to new inferences on glaciodynamic conditions associated with the advancing and retreating margin of the CIS over south-central BC.

**Study Area**

The study area covers ~8 000 km² of the southern Fraser Plateau in south-central BC, centred just north of the village of Clinton (Fig. 4.1b). The Plateau is a high elevation (~1 200 m asl), low relief feature between the Coast and Cariboo Mountains. The low relief is largely due to the surface expression of sub-horizontal Miocene and Pliocene basalt flows (Campbell and Tipper 1971, Bevier 1983, Dohaney et al. 2010). In places the bedrock may be covered by up to 20 m of till (Andrews et al. 2011), although bedrock/till contacts are not readily apparent in shallow exposures and thickness data relies on spatially restricted well-log records. Reconstructions from glacial lineations, striae, and pebble and geochemical dispersal data have suggested that ice flow originally inundated the southern Fraser Plateau from the east, meeting ice travelling west from the Cariboo Mountains in the vicinity of the Fraser River, before flow was diverted south and then southeast during the last glacial maximum (Huntley and Broster 1997, Lian and Hicock 2000; Plouffe et al. 2011). Recent work, however, suggests early deglaciation from the southern Fraser Plateau may have occurred under an ice mass which initially sloped eastward from the Coast Mountains (Margold et al. 2013) and only after the elevation of ice dropped below the level of, and was blocked by, local mountain ranges (e.g., the Marble Range) did flow evolve to a southeasterly direction (with northwesterly retreat) for the western portion of the study area (cf. Margold et al. 2013, chapter 3). Ice in the eastern portion of the study area is believed to have undergone *in situ* stagnation and complex retreat towards the Cariboo Mountains (Plouffe et al. 2011).
Figure 4.1  
(a) Location of the study area (dark grey polygon) on the southern Fraser Plateau (hachure polygon) (Holland 1976).  (b) Elevation (Geobase®) and putative ice flow directions (from Plouffe et al. 2011) on the southern Fraser Plateau.  (c) Moraine ridges and inferred ice flow directions on the southern Fraser Plateau.  Arrows indicate ice flow directions reconstructed from moraine positions (black = advance stage; pink = local LGM or intermediate stage; brown = retreat stage).  White outlined boxes highlight locations of individual moraine investigations and subsequent Figures reported in this paper.  Modern lakes are blue polygons.  Study area is outline by dashed white line.
Although significant moraine ridges were not documented by early mapping of the region (e.g., Tipper 1971a), recent surficial geology mapping in the study area has highlighted the existence of local moraine ridges (Plouffe 2009a, b, Bednarski 2009, Huscroft 2009, Appendix A). In the surrounding region minor moraine assemblages have been used to reconstruct an ice margin with a history of complex frontal retreat (e.g., annual moraines of Aylsworth 1975, Ryder 1976, and crevasse fills of Heginbottom 1972, Huntley 1996). The widespread ribbed terrain ("drift ridges of uncertain origin", Tipper 1971a, b) present in the study area has been compared with De Geer moraine (Tipper 1971a) formed subaqueously at the ice margin (c.f. Lindén and Möller 2005), and ribbed moraine formed by inward-transgressive subglacial thawing (Kleman et al. 2010), though little investigation of the internal architecture or stratigraphy of these ridges has been completed to ground these interpretations.

Methods

A Geographic Information Systems (GIS) database, including existing surficial and bedrock geology maps, glacial landform maps, orthophotograph mosaics (1 m resolution orthophotographs, BC Ministry of Environment), and a digital elevation model (DEM, 25 m horizontal grid cells, 1 m vertical resolution (90% of measurements within 5 m of true elevation), Geobase®), was assembled and combined with medium scale aerial photographs (≤1:40 000 stereopairs) in order to inventory and map the ridge-crest for moraines in the study area. Moraines were identified primarily based on the presence of a ridge and were differentiated from esker ridges by their asymmetric cross-profiles and geomorphic context (e.g., perpendicular to meltwater channels). All accessible moraines were visited in the field to investigate whether the ridge structure could be more simply explained by other formational mechanisms. Field-checking confirmed that most moraines were composed of diamicton or gravel, though, occasionally, laminated sediments were also found.

Moraine ridges of different shape, scale and context were selected for detailed field investigation based on field accessibility, availability of exposures and opportunity for application of geophysics. Detailed investigation of moraine ridges involved the analysis of available exposures for sedimentary composition and architecture. The maximum dip of clast \( ab \)-planes and the associated trend was recorded along with the orientation of the \( a \)-axis relative to the trend of the maximum dip (transverse, parallel, oblique) when measuring \( ab \)-plane fabrics (\( n \geq 30 \)) in gravel units. Trend and plunge of clast \( a \)-axes were measured for \( a \)-axis fabrics (\( n \geq 30 \)) in
diamicton units. Both types of fabrics were plotted on equal-area, lower hemisphere schmidt nets. Fabric orientations were quantified using the eigenvector method (Mark 1974), and visually inspected for modality. All principal eigenvalues are statistically significant at the 99% confidence level (Woodcock and Naylor 1983). Associated structural data including shear and fracture planes are plotted as poles to planes on fabric diagrams. Contour diagrams are calculated based on the 1% area method (i.e. 2, 4, 6 etc. measurements per 1% area of the plot). Rose diagrams were treated following Krumbein (1939) to avoid inflating visual dispersion. The orientation of bedding planes, direction of shear for augen structures and surface boulder orientations were also recorded. Where sedimentary exposure was not available ERT and GPR surveys were conducted, and water well records (BCME, 2013) facilitated interpretations.

Just over 1.4 km of ERT surveys (Appendix D.1) were conducted using a passive, 112 electrode, eight-channel system (SuperSting R8 Advanced Geosciences Inc.). A closely-spaced (1-1.5 m takeout spacing) dipole-dipole array (max n = 6, max dipole = 12; Appendix D.1) provided an appropriate compromise between signal depth penetration, and horizontal and vertical resolution (Samouëlian et al. 2005). Different takeout spacings were based on site-specific restrictions (e.g., necessary depth of investigation, overall line distance, difficult site conditions). Contact resistance was maintained below 3 kΩ whenever possible. Inversion of most ERT data was accomplished using EarthImager 2D v. 2.40. A finite element method was used as the forward model and a smooth-model inversion routine was applied. Negative resistivity values and data spikes were removed. A real-time kinematic differential global positioning system (RTK dGPS, Leica system 500) providing decimetre positional accuracy was used for topographic correction of ERT data. RTK dGPS data was post-processed (Leica Geo Office v. 7) where field conditions prevented constant communication between base-station and rover. Floating differential Global Positioning System (dGPS) measurements taken from landform surveys were corrected using CVGD 27 to align with the DEM. Based on goodness-of-fit following inversion, up to 10% of total data points were removed, and inversion repeated, allowing reduction in overall root mean square (RMS) error between the predicted resistivity model and the observed data. Order of magnitude changes in resistivity are interpreted as resistivity units, except where the water table was inferred to intersect the profile (based on the elevation of nearby waterbodies). Resistivity unit composition was refined based on sedimentary units observed in exposures or water well logs wherever this data was available. In one case a 3D ERT survey was completed. These data were processed using EarthImager 3D v. 1.5 but are graphically displayed as 2D planes extracted from the 3D grid.
One line (75 m) of GPR data was recorded using a pulseEKKO IV system (Sensors and Software Inc.). These data were collected as a common offset (CO) survey with antennas copolarized and arranged perpendicular broadside to the survey line (Arcone et al. 1995). 100 MHz antennas with a 0.25 m step size and 1 m antenna separation were used. GPR data were processed using REFLEXW v5.6. Processing included application of static correction, ‘dewow’ filter, ‘bandpass’ filter, migration, ‘background removal’ filter, gain function and topographic correction. High amplitude, continuous reflections within processed GPR profiles were selected as radar element bounding surfaces. Radar element interpretations were linked to sedimentary architecture visible in sediment exposures. Offset reflections are interpreted as faults and may be more continuous than imaged in GPR profiles given alignment and resolution issues (cf. Fiore et al. 2002, Woodward et al., 2008).

Results

A total of 432 individual moraines were mapped in the study area with a cumulative ridge-crest length of ~500 km (493.6 km). Ridge lengths ranged from 58 m to 8 200 m, and averaged 1 140 m. Analysis of aerial photographs and DEM’s allowed for the geomorphic identification of three moraine ridge types which were later interpreted genetically based on field investigations: 1) broad or narrow-crested, occurring discretely or in fields as linear, arcuate ridges oriented parallel to the ice margin – glaciotectonic ridges (Bennett 2001; Dunlop and Clark 2006); 2) single-crested, linear arcuate ridges with gentle ice-proximal, and steeper ice-distal slopes, often associated with glacial lake basins – grounding line moraines (Benn 1996, Chapter 3); and 3) multiple, low-elevation ridges, often reticulate in pattern and aligned normal or oblique to the ice margin – crevasse fill ridges (Sharp 1985) (Fig. 4.1c). Ridge dimensions and sub-types are summarized in Table 4.1.

Type 1a (Discrete glaciotectonic moraine, overridden) – 70 mile moraine

The 70 mile moraine is a relatively straight, 3 500 m long x 425 m wide, flat-topped ridge oriented WNW-ESE (perpendicular to the inferred MIS-2 ice advance margin, Plouffe et al. 2011) (yellow arrow, Fig. 4.1b) and rising 10-15 m above the surrounding plateau surface (Figs 4.2, 4.3). It is one of a series of ridges with similar scale and orientation, located on the rising western slope.
of the Green Lake basin (Fig. 4.1). Its western and northern flanks have been eroded by channelized meltwater (Plouffe 2009) and its southern flank exhibits slumping and subsidence. Lake levels during the field season were measured at an elevation of 1090 m asl (south of ridge) and 1082 m asl (north of ridge).

Three ERT surveys (two lines transverse (70M1 and 70M2) and one parallel (70M3) to ridge crestline) for a total of ~550 m were completed on this ridge (Figs. 4.2a, 4.2f, 4.3; Appendix D.4-D.6). Five resistivity units (RU1-RU5) are inferred from the inverted profiles based on resistivity values, elevations and the spatial proximity of profiles (Fig. 4.3). RU1 (70-300 ohm-m) is the lowermost unit and is at least 12 m thick (extends from ~1080 m asl to below the maximum depth of penetration) in profile 70M1. It is irregularly shaped, with a steep rise in elevation near the southwest end of the profile. A nearby water well within the moraine ridge (well log 16230, BC Well Log Database, Fig. 4.2a) encountered porous bedrock at 1079 m asl. Resistivity values for weathered mafic igneous rocks (consistent with those found in this region of the Plateau) are typically 5-90 ohm-m (Palacky 1987). Therefore, based on its resistivity values and elevation, we interpret RU1 as the irregular weathered bedrock surface below the moraine.

### Table 4.1 Moraine type classification

<table>
<thead>
<tr>
<th>Type number</th>
<th>Genetic nomenclature</th>
<th>n-size</th>
<th>Mean length (m)</th>
<th>Morphologic character</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a</td>
<td>Discrete glaciotectonic ridge (overridden)</td>
<td>12</td>
<td>1570</td>
<td>Broad-crested (occasionally streamlined), single arcuate ridges</td>
</tr>
<tr>
<td>1b</td>
<td>Discrete glaciotectonic ridge (recessional)</td>
<td>20</td>
<td>645</td>
<td>Sharp, narrow-crested, single arcuate ridges</td>
</tr>
<tr>
<td>1c</td>
<td>Glaciotectonic moraine field (subglacial, ribbed terrain)</td>
<td>21</td>
<td>3510</td>
<td>Broad-crested (occasionally streamlined) arcuate nested ridges, occurring in fields</td>
</tr>
<tr>
<td>1d</td>
<td>Glaciotectonic moraine field (recessional)</td>
<td>317</td>
<td>1155</td>
<td>Sharp, narrow-crested, sinuous, nested ridges occurring in fields</td>
</tr>
<tr>
<td>2</td>
<td>Grounding line moraine</td>
<td>11</td>
<td>280</td>
<td>Sharp, narrow-crested, single linear ridges with steep ice-proximal and gentle ice-distal slopes, located within paleo-lake basins</td>
</tr>
<tr>
<td>3</td>
<td>Crevasse-fill ridge</td>
<td>51</td>
<td>350</td>
<td>Low elevation ridges, reticulate (or subreticulate)</td>
</tr>
</tbody>
</table>
RU2 and RU4 (80-600 ohm-m, average 400 ohm-m) are tabular units that average 16 m and 8 m thick, respectively. RU2 is truncated by RU4 in 70M1 (Fig. 4.3), and RU4 is consistently visible at the ridge surface where it is not overlain by RU5. Both RU2 and RU4 are relatively

Figure 4.2 Geomorphology and sedimentary architecture of 70 mile moraine (see Figure 1a for location). (a) WNW-orientated 70 mile moraine (black outline; orthophotograph (Province of British Columbia 2012) overlain onto a
hillshaded DEM (Geobase®)). Locations of photos (b) and (c) are shown. The black cross marks the location of the clast a-axis fabric (Fab-70M1) in (d). The black dashed box marks the location of the boulder a-axis orientation measurements shown in (e). ERT profiles in (f) are represented as arrows extending from 70M1, 70M2, and 70M3 with the direction of survey indicated by the arrow. The white circle (WL) is the location of well log 16230 (BC well logs database). (b) Concentrations of angular boulders on the steep southern edge of the moraine ridge. (c) Diamict exposure within RU4 (Exp-70M2). Note high clast concentration and presence of boulders. Metre-stick for scale. (d) Three-dimensional stone a-axis fabric data (trend and plunge) taken from Fab-70M1 in RU2 near the eastern end of the moraine (Exp-70M1). V1 is the direction of the principal eigenvector (14°). Centre lines represent stone surface striae orientations and bold ticks around the stereogram perimeter represent the directions of plucked ends (Table 4.2 contains additional data). (e) Rose diagram representing surface boulder a-axis orientations (mean angle = 11°/191°) which exist on RU4. (f) Oblique view of moraine surface looking west and processed ERT profiles (70M1, 70M2, 70M3) in section (see Figure 3 for ERT interpretation). Double-ended black dashed arrow on moraine surface indicates median orientation of surface boulder a-axes (d).

high resistivity units consistent with diamicton reported elsewhere on the Plateau (Burke et al. 2012) and sedimentology at sparse exposures (Fig. 4.2). An a-axis pebble fabric taken from a consolidated sandy silt diamicton unit at a similar elevation (1089 m asl) to RU2 at the eastern end of the ridge (Fig. 4.2a, d) shows a high occurrence of glacigenic wear features (e.g., plucked ends) and striae (unlike those found in RU4, see below). Clasts are largely basalt but an elevated number of granitic clasts is comparable to counts of granitic pebbles found in till within the Green Lake basin (Plouffe et al. 2011). The fabric displays a spread bimodal distribution with a primary mode transverse to the ridge crestline (V1 = 14°) and a secondary mode parallel to the ridge crestline. Bimodal striae orientations correspond to these modes (Fig. 4.2d) and plucked ends exhibit no preferred direction. The occurrence of glacigenic wear features and the variety of lithologies in this consolidated diamicton and its high resistivity value suggest it is most likely part of the regional till sheet (Andrews et al. 2011). The presence of two orthogonal fabric modes in the clast fabric, perpendicular and parallel to the ridge, suggest the till may have undergone some compressive stress (c.f. Hicock et al 1996) as it encountered the bedrock ridge (RU1) underlying the deposit (Fig. 4.3). Alternatively, presuming a viscous till rheology (cf. Carr and Rose, 2003), it is possible extensional stresses dominated this system, resulting in both parallel and transverse fabric modes, though this seems less likely given the bedrock obstruction.

Where visible at the surface and in exposures, RU4 is composed of stone-rich sandy diamicton (~15% clast concentration; medium to fine sand matrix). Clasts range from pebble to
boulder sized angular basalt, some of which are deeply weathered. No striae or glacigenic wear features were observed on clasts, however where RU4 intersects the ridge surface there is a higher frequency of locally-derived basalt boulders than on the surrounding plateau surface (Fig. 4.2b). A few of the surface boulders exhibit faint striae (typically aligned with a-axis orientation) and some appear to have plucked ends and facets. Surface boulders have a mean a-axis orientation of 15/195° (Fig. 4.2e), roughly perpendicular to ridge crestline and parallel to putative early ice flow direction (southwest, Plouffe et al. 2011). The position of RU4 at the land surface, its stoney diamicton composition in combination with surface boulders with glacigenic wear features and aligned with early ice flow directions suggest RU4 may be a comminution till unit emplaced by southwesterly flowing ice. Angular basalt within and atop of RU4 may have been locally excavated as the glacier passed over top of bedrock with glacigenic wear occurring during overriding subsequent to RU4 deposition. The preservation of boulder alignment with early ice flow despite later ridge overriding from the northwest indicates these boulders may have been protected from reorientation during ridge overriding (i.e. sediment covered) and only exhumed during deglaciation.

RU3 (40-300 ohm-m) is a 5-10 m thick unit that is visible at the northeast end of 70M1 and is overlain or truncated by RU4 elsewhere (Fig. 4.3). The low resistivity values might suggest higher water content than the bounding units, but the continuous nature of the low resistivity values above the local water table as implied by lake levels to the north and south of the ridge suggest these low values likely record a high silt/clay content. The position of the unit between till units (RU2 and RU4) suggests that RU3 may record lake sediments deposited in a proglacial lake dammed by advance stage (early) southwest ice flow (Plouffe et al. 2011) and topographic obstructions within the Green Lake basin. Advancing ice excavated, sheared, transported, and stacked RU2, RU3 and RU4 forming the moraine ridge.
Figure 4.3  Processed and interpreted ERT profiles (70M1, 70M2, and 70M3) for 70 mile moraine (refer to Figure 2a for profile locations). Black lines denote interpreted boundaries between resistivity units (RU1-RU5). Refer to Appendix D for ERT survey details and processing results.
RU5 is a 0.1-3 m thick, lenticular to tabular, low resistivity (3-80 ohm-m, with minor higher resistivity spikes likely related to surface disturbances such as animal burrows) unit that overlies RU1, RU3 and RU4 (Fig. 4.3). It is composed of sandy silt where it is visible at the surface and in shallow exposures (e.g., above RU4). Its limited and discontinuous extent, stratigraphic position (Fig. 4.2a) and low resistivity values combined with surface observations of silty sand suggest it represents accumulations of windblown sediments deposited post-glacially (Lian and Huntley 1999).

Based on the above interpretations, the following formational sequence is suggested for the 70 mile moraine. RU2 likely records a till sheet deposited by the penultimate glaciation or an oscillation of the margin during the advance stage of the last glaciation, directly atop of bedrock (RU1), and preserved behind a bedrock obstruction (visible at the SW end of the 70M1 resistivity profile). The higher granitic content in RU2 matches an unexplained distribution of granitoids in till within the Green Lake basin, possibly attributable to recycling of material from a previous glaciation (cf. Plouffe et al. 2011). A proglacial water body was ponded in the Green Lake basin in front of southwestward advancing ice, resulting in the deposition of a local unit of fine-grained sediments. The diamicton of RU4 was emplaced as ice advanced from the Northeast, with associated boulder wear occurring during this phase of transport. As the material in RU2, RU3 and RU4 was overridden, the units were sheared and stacked into position behind the bedrock obstruction. Subsequent changes in ice-flow direction apparently did not significantly alter boulder alignment indicating boulders were only exhumed during deglaciation. Postglacial accumulations of aeolian material (RU5) followed deglaciation.

**Type 1b (Discrete glaciological moraine, recessional) – Loon Lake moraine**

Loon Lake moraine is a discontinuous, ~1 km long, 200-250 m wide, and 12 m high, round-crested ridge. It trends sinuously N-S across the Loon Lake valley (Fig. 4.4a). The valley slopes west towards Loon Lake, and glaciolacustrine sediments located low in the valley fill (~40 m above the 818 m asl level of modern Loon Lake) suggest this basin may have held a slightly larger lake during ice advance (Lian 1997). Previous work has
suggested the ridge formed as ice retreated to the northeast within Loon Lake valley (Plouffe et al. 2011), consistent with the apparent convex-southwest planform of the ridge (Fig. 4.4a).
Figure 4.4  Geomorphology and sedimentology of Loon Lake moraine (see Figure 1a for location). (a) Plan view of Loon Lake moraine (black
Four westward dipping stratigraphic units are distinguishable in an exposure within the Loon Lake Moraine (Fig. 4.4a and d). Unit 1 is a consolidated fine sandy silt diamicton containing subrounded to subangular, dominantly basaltic and granitic cobbles (~5% clast concentration) with rare metamorphic cobbles. Clasts exhibit glacigenic wear features: keels, plucked ends and striae with no preferred orientation or direction (Fig. 4.4e). An a-axis pebble fabric is spread unimodal with the principal eigenvector perpendicular to the ridge crestline (Fab-LL1, Fig. 4.4a, e, Table 4.2). The diamicton contains silty augen structures (up to 30 cm thick and 100 cm long), similar in composition to advance-stage glaciolacustrine sediments commonly found in the Loon Lake basin to the west of the ridge (Lian 1997). The augens exhibit tails extending from their lower surface to the southwest and material sheared from their upper surface extending to the northeast (Fig. 4.4g). Diamicton composition, the presence of glacigenic wear features lacking alignment on stones and the presence of augen structures suggest that this unit is till that underwent lodgement and deformation processes (Dreimanis 1988, Lian and Hicock 2000, Lian et al.)
2003). The transverse pebble fabric suggests that the till in unit 1 may have experienced compression (Hicock et al. 1996). The attenuation of augen structures and the eastward dip of fractures support shear to the northeast. Unit 2 is a 0.2-0.5 m thick unit of massive silt and fine sand with shears dipping 10-60° to the southwest (black dashed lines, Fig. 4.4d). The contact between unit 1 and 2 is a 0-0.2 m shear zone of highly deformed sandy silt dipping 12-40° to the southwest. Strings of silt, up to 2 cm thick and dipping 0-15° to the northeast, are injected from the shear zone into fractures within unit 1 (white dot-dashed lines, Fig. 4.4d). Unit 2 is likely derived from the same material as the augen structures in unit 1, and the presence of an extensive shear zone between unit 1 and 2 suggests unit 2 was sheared and stacked into place on top of unit 1. Unit 3 (0.8-1.2 m thick) is a massive, matrix-supported, medium sand diamicton (<5% clast concentration) containing laminated silt rip-ups and occasional angular clasts (pebbles to cobbles). Unit 4 is a silty sand diamicton with angular clasts, ~10-15% clast concentration, and has a sharp lower contact with unit 4. Clasts exhibit abundant glacigenic wear features (e.g., keels and plucked ends) but these lack a preferred orientation or direction and the alignment of clast a-axes is close to a random distribution (Fig. 4.4f). The position of the diamicton at the surface, its near randomly distributed fabric, and its localized spatial extent suggest it may be a glacigenic debris flow (Hicock et al. 1996) or highly deformed till (Hart et al. 2009).

A 3D ERT profile to the east of the ridge crestline (Fig. 4.5) reveals zones of high resistivity (200-2700 ohm-m) alternating with zones of lower resistivity (10-100 ohm-m) (Fig 5). The zones of high resistivity dip to the west (Fig. 4.5), similar to the attitude of the sedimentary units and shears within them in the exposure to the north (Fig. 4.4c, d).

<table>
<thead>
<tr>
<th>Table 4.2</th>
<th>Diamicton a-axis stone fabric data</th>
</tr>
</thead>
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<tr>
<td>Moraine</td>
<td>Exposure Fabric</td>
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<tr>
<td>70 mile</td>
<td>Exp-70M1 Fab-70M1</td>
</tr>
<tr>
<td>Loon Lake</td>
<td>Exp-LL1 Fab-LL1</td>
</tr>
<tr>
<td>Loon Lake</td>
<td>Exp-LL1 Fab-LL2</td>
</tr>
</tbody>
</table>
The Loon Lake moraine is interpreted as a discrete, glaciotectonic moraine generated by shearing and stacking of till and advance stage glaciolacustrine sediments, and possibly overlain by a glacigenic debris flow. The geometry of sedimentary and resistivity units, the attitudes of shears and fractures, the geometry of augen structures,
and the fabric of unit 1 suggest that the moraine was formed by ice advancing towards the east. The round-crested nature of the ridge and the lack of a continuous till cap indicates that the ridge has not been overridden by later ice advance and therefore it may be used to reconstruct the last major ice flow direction within the Loon Lake valley. This would place retreating ice on the west side of the moraine ridge, opposite to the ice position suggested by Plouffe et al. (2011), but consistent with an overall ice retreat pattern to the northwest inferred for this region of the Plateau based on ice-marginal lake reconstructions (Chapters 2 and 3). The gap in the middle of the moraine ridge may reflect an area of non-deposition or an area that was breached and removed by later glaciofluvial erosion (consistent with a potential lake drainage event within the Loon Lake basin as ice retreated; though evidence downvalley for such a drainage event has not been observed).

**Type 1c (Glaciotectonic ribbed terrain, subglacial) – Green Lake ribbed terrain**

A ~400 km² field of broad ridges (cf. “drift ridges of uncertain origin”, Tipper 1971c) with NE-SW oriented crestlines were mapped around Green Lake (Fig. 4.1c). The ridges are up to 1.1 km wide, 4-6 km in length with a relief of 4-25 m above the surrounding terrain. Many of the ridges are separated by basins containing lakes. They are overprinted by NW-SE oriented streamlined bedforms (Plouffe 2009b), and truncated by the Green Lake erosional corridor and esker system (Figs 4.1, 5.4; Burke et al. 2012).

An ERT survey (GL1) was completed across a 4-m high ridge within Green Lake ribbed terrain (1080 to 1084 m asl), perpendicular to Exp-GL1 (Fig. 4.6), and extending onto inter-ridge
Figure 4.6  Geomorphology and sedimentary architecture of Green Lake glaciotectonic ribbed terrain (subglacial). (a) Green Lake ribbed terrain (black outline) is oriented southwest to northeast at the southeast end of the Green Lake basin (hillshaded DEM (Geobase©)) (refer to Fig. 1c for location). White square labeled ‘Exp-GL1’ indicates location of exposure shown in (b). White arrow labeled ‘GL1’ shows location and direction of ERT survey GL1. White circle indicates location of water well log (WW, BC well tag number 41849). (b) Exp-GL1 in ridge showing three sedimentary units within RU3 (scale bar is 1 m long with decimetre increments). Refer to text for unit interpretations. (c) Interpreted ERT profile and nearby well log (WW; colours for well log interpreted at base of figure). Three resistivity units (RU1-RU3, separated by dotted lines) are present. Refer to Appendix D.8 for ERT survey details and processing results. ‘Exp-GL1’ indicates location of exposure shown in (b).

swales to the northwest and southeast (1076 and 1079 m asl, respectively). The resulting ERT profile contains three resistivity units (RU1 to RU3, Fig 6c). RU1 is a unit of highly variable resistivity (~29 to >47000 ohm-m) that is at least 10 m thick (extends upwards to 1071 to 1073 m asl), and is continuous across the ERT profile. It is characterized by
vertically-oriented zones of very high and very low resistivity. When elevations are matched to a nearby water-well log (BC well tag number 41849; WW, Fig. 4.6a and c), RU1 correlates to fractured bedrock. The vertical zones of high and low resistivity are characteristic of resistivity patterns in fractured bedrock (Reynolds 1997, where high and low resistivity may represent air and water-filled fractures respectively). RU2 (29 to 300 ohm-m) ranges from 4 to 8 m thick (1071-1079 m asl), extends to the surface in the southern inter-ridge swale, and appears to be continuous underneath the ridge, and under RU3 in the northern inter-ridge swale (Fig. 4.6c). This unit is characterized by low resistivity values (similar to RU1 in the 70 mile moraine resistivity profiles (Fig. 4.3); small differences in resistivity values are expected based on local conditions and differences in electrode spacing) and occurs at the same stratigraphic position as weathered bedrock documented in a nearby water well (Fig. 4.6a and c). Therefore this unit is interpreted as weathered bedrock or comminuted bedrock (aka comminution till, e.g., Elson 1988).

RU3 is a <6 m thick medium resistivity unit (750-1000 ohm-m; up to 12000 ohm-m under the compacted forestry road at Exp-GL1) that forms the moraine ridge proper, and occurs in the inter-ridge swale to the NW (Fig. 4.6c). A road cut exposure (Exp-GL1; ~1.5 m high) within this unit contains stratified diamicton (Fig. 4.6) – weakly stratified fine sand diamicton containing angular basaltic pebbles and cobbles (~10% clast concentration), with some plucked ends (unit 1) coarsens upward to diamicton (~5-7% clast concentration) containing more granitic clasts in a fine to medium sand matrix (unit 2). Some clasts are deeply weathered. Unit 3 is a heterogeneous, overturned layer of boulders and organic material associated with road building. The weakly stratified character of the diamicton and the presence of some plucked ends on clasts suggests that RU3 is likely a till, and its presence at the surface indicates it is likely part of the regional, last glacial till sheet (as mapped by Plouffe (2009) and Andrews et al. (2011)). The high concentration of local clasts and their angularity suggest a short transport distance from a locally excavated bedrock source. The inclusion of more granitic clasts in unit 2 is consistent with deposition of more far-traveled lithologies as ice advanced. The resistivity values in RU3 are slightly higher than those of other till units on the southern Fraser Plateau (cf. Burke et al. 2012), although a slight variance in resistivity values is expected based on textural variability and differences in electrode spacing. The geometry and attitude of RU3 is reminiscent of a hill-
hole pair (Aber et al. 1989) (Fig. 4.6c), suggesting that Green Lake ribbed terrain was formed by glaciotectonic stacking.

Overprinting by streamlined forms, incision by the Green Lake meltwater corridor and broad-crested moraine profiles suggest that the Green Lake ribbed terrain field was formed subglacially or at least overridden. The moraine field could be a remnant from the retreat of a previous glaciation or last glacial advance. However, it is more likely that the field formed subglacially because ridge orientation suggests a formative ice flow direction between that proposed for ice advance and local LGM (Plouffe et al. 2011) and because of their association with other subglacial landforms (streamlined bedforms (Tipper 1971a) and tunnel valleys (Burke et al. 2012)) also attributed to the last glacial. In any respects the ridges resemble ribbed terrain sensu Dunlop and Clark (2006).

Based on the observations and interpretations of Green Lake ribbed terrain, the following event sequence is suggested for its formation. Ice advanced (probably from the NW, Plouffe at al. 2011) overtop of fractured bedrock (RU1) and weathered bedrock or comminution till (RU2) initially depositing local then more exotic lithologies and forming a regional till sheet (RU3). Reorientation of ice flow to the southeast resulted in shearing, thrusting and stacking of the till into ridges along the weak décollement surface provided by the weathered bedrock, and some excavation of till (and likely some weathered bedrock) from inter-ridge swales (forming ridge equivalents of hill-hole pairs). Subglacial stacking may have been facilitated by compression against a topographic obstruction to the south (e.g., Mount Jim, Fig. 4.1c) (e.g., Bouchard 1989). Ice flow shifted slightly during local LGM, resulting in SE streamlining of the moraines. Alternatively, if the moraines did form during retreat of the penultimate glaciation streamlining may have occurred at any time during the last glacial period.

**Type 1d (Glaciotectonic moraine field, recessional) – Holden Lake moraine field**

The Holden Lake moraine field (~930 km²) exists on a topographic high surrounding the Dog Creek basin, and on an adverse slope (1.3%) with respect to the local lateglacial ice flow direction (Chapter 3) (Fig. 4.1). It is composed of numerous sinuous (mean sinuosity = 1.09), sub-parallel ridges that are 7-18 m high, 0.5-4 km long
and 80-200 m wide. The area has been previously mapped as containing pitted terrain and “drift ridges of uncertain origin” (Tipper 1971c). Within this field a 90 m long ERT survey (HL1) was completed across a 3.7 km long, 6 m high, 160 m wide ridge (Fig. 4.7a-c). Two nearby exposures, Exp-HL1, a 2.2 m high exposure containing 3 units and located in a ridge ~1 km west of the ERT line, (Fig. 4.7a-c) and Exp-HL2, a 1.7 m high exposure containing 2 units and located in a parallel ridge 250 m to the southwest of the ERT line, (Figs 4.7a, 4.8a) were logged at the decimetre scale. ERT profile HL1 reveals three resistivity units (RU1 to RU3, Fig. 4.7c) which are lithostratigraphically correlative to units observed in these nearby exposures. Units logged in the exposures are of similar thickness to those interpreted from the ERT profile, have compositions consistent with recorded resistivity values, and are arranged in similar stratigraphic order.
Figure 4.7 Geomorphology and sedimentology of Holden Lake moraine ridge (recessional). (a) Holden Lake ridge (black outline) is 14 m high, 160 m wide, and 3.7 km long and oriented northeast to southwest (orthophotograph (Province of British Columbia 2012) overlain onto hillshaded DEM (Geobase®)). Refer to Figure 1c for location. Location and direction of ERT survey (HL1) and topographic profile (T to T', (b)) shown by white arrow. Locations of exposures Exp-HL1 and Exp-HL2 shown by black circles. (b) Perspective view of processed ERT profile (HL1) in relation to ridge geomorphology. White arrows point to boulders on the ridge surface. (c) Interpreted ERT profile HL1. Resistivity units (RU1-RU3) are bounded by solid black lines (dots at surface represent electrode locations). Refer to Appendix D.9 for ERT survey details and processing results. High resistivity sediments (RU2), interpreted as gravels, compose the majority of the ridge. (d) Exp-HL1 is composed of massive diamicton (units 1 and 3) and openwork gravel with sand lenses (unit 2) (scale bar is 1 m long with decimetre increments). Dashed black lines
DENOTE SHEARS, SOLID BLACK LINES DENOTE FRACTURES, AND DOT-DASHED WHITE LINE INDICATES EXTENT OF A FINE SAND LENS IN THE SURROUNDING A BEDROCK CLAST IN UNIT 2. DASHED WHITE BOXES (FAB-HL1 AND FAB-HL2) LOCATE CLAST FABRICS AND SOLID WHITE BOXES LOCATE (E) AND (F). (E) SHEAR ZONE IN-FILLED WITH COARSE SAND (UNIT 2). SCALE HAS CM-SUBDIVISIONS. (F) FINE SAND LENS SURROUNDING BOULDER CLAST. (G) STRUCTURAL DIAGRAM FOR SHEAR (DASHED ARCS) AND FRAC TURE (SOLID LINES) PLANES IN UNIT 2. (H) AND (I) STONE A-AXIS FABRIC SCATTERGRAMS AND CONTOUR PLOTS FROM EXP-HL1 (FAB-HL1 AND FAB-HL2). SOLID BLACK LINE REPRESENTS THE ORIENTATION OF A BASAL KEEL (TABLE 4.2 CONTAINS ADDITIONAL DATA).

RU1 (Fig. 4.7c) is at least 18 m thick and is characterized by moderate resistivity (97 to 600 ohm-m). It is continuous under the ridge and in inter-ridge swales. Resistivity values are consistent with diamicton found elsewhere on the southern Fraser Plateau (e.g., Burke et al. 2012, Perkins et al. 2013) and unit 1 in nearby exposures (Figs 4.7a, d, 4.8a). At Exp-HL1, unit 1 (at least 1.8 m thick, Fig. 4.7d) is a massive diamicton, containing angular clasts with few glacigenic wear features (keels) and recording a-axis pebble fabrics with spread bimodal distributions – principal eigenvectors to the NNE (Fig. 4.7d, h and i, Table 4.2) and orthogonal secondary modes – and high dip angles (>25% of clasts sampled had dip angles ≥ 40°). Unit 1 (and RU1) is interpreted as part of the regional till sheet (Andrews et al. 2011) based on its architecture, distribution, diamictic composition, fabric shape, and presence of clast keels, though angular clasts combined with limited glacigenic wear features indicates a relatively short transport distance. An intermediate ice flow direction between initial ice advance and ice maximum (Plouffe et al. 2011) is inferred from the primary eigenvectors of fabrics (Fab-HL1 and Fab-HL2 of Exp-HL1, Fig. 4.7h and i and Table 4.2). The orthogonal (east-west) secondary fabric mode, roughly transverse to ridge orientation, and the high dip angles of clasts suggest that the till likely experienced compressive stresses during ridge construction (interpretation as compressive flow in a plastic till (cf. Hooke et al. 1997) is preferred over extensive flow in a viscous till (cf. Carr and Rose 2003) because of deposition against an adverse slope).
Figure 4.8  (a) Holden Lake moraine ridge (recessional), Exp-HL2 (refer to Fig. 5.7a for location) contains two units (unit numbers are correlative with those at Exp-HL-1): unit 2 is gravel and unit 3 is diamicton. Dashed boxes indicate locations of ab-plane gravel fabric (Fab-HL3) and a-axis diamicton fabric (Fab-HL4) in unit 2 and 3, respectively. White boxes (b, c) show location of (b) and (c). Scale bar is 1 m high with 0.1 m increments. (b) and (c) Cross-laminated sand drapes and occurs below boulders within unit 2. (c). (d) Pebble clast roundness for units 2 and 3. (e) Stone fabric scattergrams and contour plots from Exp-HL2. Ab-plane gravel (Fab-HL3) and a-axis diamicton (Fab-HL4) fabrics are plotted on lower hemisphere, equal-area Schmidt diagrams. Stones in the gravel fabric are classified according to their a-axis orientation with respect to the trend of maximum dip. Density contours are calculated based on the 1% area method (i.e. 2, 4, 6 etc. measurements per 1% area of the plot). V1 indicates direction of principal eigenvector (Table 4.2 contains additional data). ‘Ravg’ on Fab-HL3 indicates the mean vector of ripple measurements (n = 5) taken from sand lenses around Fab-HL3.

RU2 is a 2-6 m thick, high resistivity (600-2716 ohm-m) unit that intersects the land surface on the SE flank of the moraine and is absent in part of the SE inter-ridge swale. Its high resistivity values are consistent with sand and gravel found in other areas of the southern Fraser Plateau (e.g., Burke et al. 2012, Perkins et al. 2013) and unit 2 in nearby exposures (Figs 4.7a, d, 4.8a). In Exp-HL1, unit 2 (>1.8 m thick, lower contact obscured,
Fig. 4.7d) is composed of nearly openwork cobble gravel with a deficiency of coarse sand, imbricate clusters of larger clasts, and high mean dip angles on the imbricate plane of clasts similar to the heterogeneous unstratified gravel described by Brennand (1994). Shear planes and fractures are common (Fig. 4.7d and g) and generally dip northwest and southeast, respectively. Lenses of coarse to fine sand occur along shear zones (Fig. 4.7e), and surround large clasts (Fig. 4.7f). In Exp-HL2, unit 2 (>1.2 m thick, lower contact obscured, Fig. 4.8a) contains openwork, subrounded to rounded pebble gravel with imbricate clusters of pebbles and cobbles, and is interbedded (every 0.2-0.4 m) with lenses (at least 7 x 4 x 0.05 m in dimension) of planar-laminated or type A ripple-drift cross-laminated (eastward paleoflow, Fig 8e), well-sorted fine to medium sand. Sandy boulder scour fills exist under the largest clasts (Fig. 4.8b and c). An ab-plane pebble fabric (n = 30, Fab-HL3, Fig. 4.7a and e) records a SE paleoflow with the majority of clasts oriented with their a-axis transverse to the paleoflow direction. Unit 2 (and RU2) is interpreted as a meltwater deposit recording southeastward paleoflows and rolling traction transport of larger clast based on the presence of well sorted, planar-laminated and cross-laminated sand lenses and imbricate gravel clusters with a-axis orientations largely transverse to the dips of ab-planes. High mean dip angles of clasts suggest clast collisions were frequent during gravel deposition (cf. Rust 1972, Brennand 1994). The openwork nature of some lenses and the presence of coarse sand and granules (a grain size often absent in fluvial deposits (Shaw and Kellerhals 1982)) in the matrix of others indicates transport in powerful flows with high sedimentation rates (cf. Rust and Koster 1984, Brennand 1994). The interbedded sand sheets record episodes of moderate to lower energy flow, shifts in flow direction or leeside deposition behind boulders. Northwestward dipping shear planes (Fig. 4.7g) suggest glaciotectonic shearing and stacking of unit 2 by ice flowing to the southeast (consistent with till fabrics in unit 1) (Fig. 4.7g-i), resulting in the formation of the moraine ridge. The orientation of fracture planes is consistent with southeastward glaciotectonic shear and their presence records dewatering and brittle deformation following the main episode of ridge-building.

RU3 overlies RU2 on the NW moraine flank and in the NW inter-ridge swale (Fig. 4.7c). It is a sub-2 m thick unit of very low resistivity (97-300 ohm-m, Fig. 4.7c), consistent with resistivity values for diamicton found in other areas on the southern Fraser Plateau (e.g., Burke et al. 2012, Perkins et al. 2013), unit 3 in nearby exposures (Figs 4.7a, d,
4.8a), and in other shallow exposures south of ERT line HL1. Unit 3 in Exp-HL1 and Exp-HL2 (Figs 4.7d and 4.8a) is a 0.3-0.7 m thick, massive, poorly-consolidated, clast-supported (~50% clast concentration) coarse sand to granule diamicton, with an irregular, sharp lower contact. Clast size ranges from pebbles to boulders with an increased proportion of boulders near the top of the exposure (and at the land surface). Clasts are angular to subrounded (Fig. 4.8d). An a-axis pebble fabric from unit 3 (Exp-HL2) is spread unimodal with a principal eigenvector of 343° (Fab-HL4, Fig. 4.8a and e, Table 4.2). Unit 3 (and RU3) is interpreted as subglacial till based on its composition, location, distribution and fabric. Its diamictic composition and lateral extent suggest it is part of the regional till sheet (Andrews et al. 2011). Its presence largely on the upflow side of the ridge (Figs 4.7 and 4.8) suggests deposition during or following glaciotectonic ridge building. The absence of fractures indicate high mobility during active shear, and the presence of a relatively strong primary fabric mode perpendicular to ridge crestline (Fab-HL4) suggests that the till melted-out in place following ridge building; the larger clasts near the ground surface likely record englacial or supraglacial transport and meltout during final decay.

The Holden Lake moraine ridges are composed of gravel (unit 2) sandwiched between subglacial till units (units 1 and 3). Episodic (perhaps annual) readvances of the ice margin during retreat resulted in glaciotectonic shearing and stacking to form the moraine ridges (cf. Bennett 2001) against the adverse topography surrounding the Dog Creek basin (Fig. 4.1c). Consequently, unit 2 either records deposition in proglacial outwash formed between episodic readvances, or in broad subglacial meltwater channels or sheets; the presence of boulder scour fills perhaps favours subglacial formation (cf. englacial boulder scours and sand beds, Shaw (1982)). Given the presence of bedded sand in shear zones within unit 2 it is unlikely that the gravels were frozen during ridge building (cf. Kruger 1995). A non-frozen bed model for annual moraine formation similar to that of Eybergen (1986) for the push moraine at Turtmann glacier, Switzerland demonstrates analogous thrusting of proglacial outwash sediments on an adverse topographic slope.
Type 2 (Recessional, grounding-line moraines) – White Lake moraine

The remnants of several convex-west cross-valley ridges exist in the White Lake basin (Fig. 4.1c). The largest, best preserved example of these constructional ridges is the White Lake moraine at the west end of White Lake (Fig. 4.9a). This 780 m long and 290 m wide ridge reaches a maximum height of 31 m (1054 m asl) above the surrounding terrain, incorporates a large circular depression on its west flank (Fig. 4.9b) and is breached by a 40 m wide channel near its midpoint. Its eastern (ice-distal) slope is steeper (10°) than its western (ice-proximal) slope (7°). Surface exposures on the east side of the northern segment of the ridge reveal a >3 m thick unit of well sorted, rounded pebble to cobble gravel (unit 1) overlain by a 1-2 m thick unit of moderately sorted, mainly openwork, sub-angular pebble to cobble gravel in a sparse, medium sand matrix (unit 2) (Fig. 4.9c). Unit 1 contains interbeds of silt and sand exhibiting type A and B cross lamination recording paleoflows towards the NE, and brittle deformation structures including outcrop scale-thrust faults upthrust 1-2 cm to the east and NE (Fig. 4.9d). A single GPR profile along the eastern slope of the ridge (WLM, Fig. 4.9a, e) (the western side was heavily forested) exhibits significant signal attenuation but records six radar elements (RE-A to F) to a maximum depth of ~ 5 m below the ridge surface. RE-A is characterized by high signal attenuation, with little to no internal structure visible. RE-B is a planar, horizontal element. RE-C to RE-F are lenticular elements with continuous, concave, oblique internal reflections and curved lower bounding surfaces that appear to truncate lower units. The heavy signal attenuation in RE-A may be a response to a thin silt cap (observed at the surface of the southern segment of the ridge), or diamicton in the ridge core, though the presence of silty sand, similar to that in unit 1 (Fig. 4.9d) would likely produce similar attenuation at depth. Based on internal reflections and the elevation of corresponding sand and gravel in unit 1 on the northern segment of the ridge, RE-B is interpreted as horizontally-stratified gravel and sand. RE-C to RE-F are interpreted as slump deposits composed of reworked material from upslope, based on their curvilinear lower contacts and internal reflections that appear slightly rotated off their original depositional plane. The absence of offset reflections indicates each element likely moved as unit, resulting in little internal deformation. Overall the GPR profile suggests a larger amount of silty sand (or diamicton) at depth then is visible in the exposure in the northern segment of the ridge.
Figure 4.9 Geomorphology and sedimentary architecture of White Lake moraine. (a) White Lake ridge (black outline) is located at the western end of White Lake (orthophoto (Province of British Columbia 2012) overlain onto hillshaded DEM (Geobase®)) (refer to Fig. 1c for location). Lines X-X’ and U-U’ are locations of cross ridge topographic profiles in (b). ‘c and ‘d’ indicate position of photographs in (c) and (d). ‘WLM’ and black arrow indicates location and survey direction of GPR profile in ‘e’. ‘IMC’ and white arrows show the positions of ice-marginal channels. (b) Topographic profiles X-X’ and U-U’.’K’ notes the location of a closed depression. (c) The moraine is covered by a surficial unit of reworked material (from quarry activity), that demonstrates the general size of material exposed while digging lower down in the ridge (grid is 2 cm x 2 cm). (d) Below the surface colluvium, the moraine interior includes bedded sand and pebbles (knife handle is 3.5 cm long). (e) Interpreted GPR profile. Black lines denote bounding surfaces of radar elements (A-F).

The sharp nature of the ridge and lack of diamicton on the ridge surface indicate that this ridge was not overridden during the last glaciation. Channels northwest of the ridge that do not follow regional slope (interpreted here as ice-marginal channels) record the westerly downwasting and retreat of an ice mass within the White Lake basin (see also chapter 3). An ice-dammed lake (glacial Lake White) up to 1052 m asl was maintained
in this basin by the retreating ice (Chapter 3), allowing the ridge to have formed almost entirely subaqueously. Subaqueous fan deposition results in narrow ridges which sometimes form convexities on their ice-proximal edge (Sharpe and Cowan 1990), and commonly exhibit ice-proximal slopes that are steeper than ice-distal slopes (e.g., Benn 1996). The well sorted, rounded gravel and silty sand interbeds within the ridge support deposition by flowing water, and the presence of kettle holes and thrust faults in the silty sand interbeds suggest that ice was in active contact with the landform. Therefore this landform is interpreted to be a subaqueous grounding line moraine, formed at the margins of an actively retreating ice mass in contact with a proglacial lake in the White Lake basin. Following retreat of the ice margin away from the ridge, it is likely that the lake drained west, incising through the moraine (Chapter 3).

**Type 3 (Crevasse fill ridges) – Big Bar crevasse fills**

A series of low relief (1-2 m high) ridges form a rectilinear grid pattern (Fig. 4.10a) 12 km to the southwest of White Lake on the southern Fraser Plateau (Fig. 4.1b). The ridges are slightly curvilinear with N-S trending ridges convex to the east. These ridges are found on the northeast flank of Big Bar Mountain and are sometimes separated by enclosed, lake-filled depressions previously interpreted as kettle holes (Huntley 1996), or are interrupted by deposits of poorly-sorted gravels forming eskers or meltwater channel deposits (Heginbottom 1972). Ridges average 390 m long (but may be up to 1500 m), 2-3 m wide, and less than 2 m high, and the southern group of ridges (Fig. 4.10a, d) overlie a more prominent (5 m high, 2500 m long) flat-topped glaciotectonic moraine oriented oblique to the low relief ridges. At ridge intersections one ridge may be superposed on the other, or they may be confluent (cf. Gravenor and Kupsch 1959). Based on limited exposures, observations at ridge surfaces and shallow (1 m) boreholes, the ridges are composed of moderately sorted, rounded, pebble to cobble gravel (Fig. 4.10e) interbedded with lenses of well sorted, silty fine sand.
Figure 4.10  (a) Big Bar crevasse fill ridges are 150-1500 m long and 1-2 m high curvilinear ridges, arranged in a rectilinear grid (orthophotograph (Province of British Columbia 2012) overlain onto hillshaded DEM (Geobase®)) (refer for Fig. 1c for location). White boxes show locations of (b) and (d). (b) Orthophotograph showing northwestern crevasse fill ridges. White arrow indicates view direction of photograph in (c). (c) Ground view of crevasse-fill ridges (white arrows). (d) Orthophotograph showing southeastern crevasse-fill ridges superimposed on an overridden glaciotectonic moraine. (e) Crevasse-fill ridges are composed of moderately-sorted, rounded pebble to cobble gravel (trowel is 20 cm long).

Within the study area and vicinity some of these ridges have been interpreted as crevasse fills in stagnant ice formed by till squeezing (Tipper 1971a, Fulton 1976, Huntley and Broster 1997), as eskers (Huntley 1995), or as push moraines formed during minor winter readvance during deglaciation (Aylesworth 1975). A till squeeze mechanism would require the ridge core to be composed of diamicton (Fulton 1976, Sharp 1985). Rather, these ridges are composed entirely of waterlain stratified sediment and their distribution in a rectilinear pattern suggests channelized meltwater flow within crevasses at the retreating ice margin—essentially crevasse-controlled eskers. The continuity of the ridges supports formation in subglacial rather than supraglacial crevasse systems. Longitudinal
and transverse crevassing is attributed to ice margin oscillation and flexure in the large proglacial ice-dammed lake (glacial Lake Big Bar, Chapter 3) directly east of the ridges.

**Discussion**

Through reconstruction of moraine type and genesis on the southern Fraser Plateau a coherent picture of the dynamics of the last CIS emerges from the spatial pattern (including orientation) of moraine distribution (Fig. 4.1c).

**Ice advance**

Ice margin positions during last CIS advance are recorded by Type 1a moraine systems, including the 70 mile moraine (Fig. 4.1c). These moraines are orthogonal to inferred ice flow during advance (Plouffe et al. 2011), are composed of glaciotectonically thrust glacigenic sediments, and show evidence of being overridden. Generally, Type 1a moraines are offset from one another and are greater than 5 m high, suggesting they are more significant than seasonal push moraines (cf. Bennett 2001). It is possible that initial advance of the ice sheet and subsequent impounding of an ice-marginal lake against topographic obstructions to the southwest resulted in elevated pore-water pressure at the ice base and an increased mobility of sediments leading to construction of advanced stage push moraines in the 70 mile area. The sedimentary evidence of this lake comes in the form of fine material recorded in the ERT profile, and clay sediments recorded at depth in nearby well logs (lake likely existed ~1085 m asl). It is unlikely that this lake formed again during retreat of the ice sheet as retreat was to the northwest (Plouffe et al. 2011, Chapter 3), leaving an outlet for meltwater drainage through the Green Lake meltwater corridor. Furthermore lacustrine sediments are rarely exposed at the surface in this area (Plouffe 2009b).

**Ice advance to local LGM**

Detailed investigations into fields of “drift ridges of uncertain origin” (Tipper 1971a, b) and ribbed terrain (Kleman et al. 2010) on the southern Fraser Plateau have revealed two scales of glaciotectonic moraine fields (Type 1c and 1d, Fig. 4.1c). The broader, flat-
topped ribbed terrain (Type 1c, e.g., Green Lake ribbed terrain) appear to have been constructed as hill-hole pair-like ridges. Ridge alignment suggests an ice-flow direction between that of ice advance (as indicated by Plouffe et al. 2011 and the orientation of advance stage moraines in this study, e.g., 70 mile moraine) and ice retreat (supported by ice-marginal lake system evolution on the Plateau (Chapter 3), and the orientation of recessional moraines in this study, e.g., Holden Lake moraine), suggesting they likely formed subglacially due to compression as ice flowed SE against a topographic obstruction (Mount Jim environs, Fig. 4.1c). Their flat-topped surface and the presence of superimposed streamlined bedforms suggest the ridges continued to be overridden by ice after formation. A shift in thermal regime, perhaps due to ice thickening, is likely necessary to explain the transition from glaciotectonic ridge-building to streamlining and has been suggested previously for the southern Fraser Plateau in this time period (Lian and Hicock 2000).

**Ice retreat**

A field of narrow, sharp-crested ridges near Holden Lake (Type 1d moraines, Fig. 4.1c) are composed of heterogeneous unstratified gravel confined between two units of subglacial till. Ridges are transverse to putative ice flow at local LGM (Plouffe et al. 2011), indicating they may have formed any time between glacial maximum and deglaciation, but their sharp-crested form and the absence of overprinted streamlined bedforms implies that they were not significantly overridden after formation, and indeed that they formed as recessional moraines during ice retreat. Compressive fabrics within diamictons and shear planes within gravels indicate that glaciotectonic shear to the southeast, perpendicular to ridge crestline, was responsible for ridge building, and is consistent with northwesterly ice margin retreat inferred from ice-dammed lake evolution nearby (Fig. 4.1c) (Perkins et al. 2013, Chapter 3). The formation of a field of recessional glaciotectonic moraines requires an active ice margin and hints at the possibility of seasonal or annual formation (Bennett 2001). An active ice margin suggests that an ice surface slope (sloping from the northwest) existed during deglaciation, and this is in agreement with glacioisostatic tilts computed from lateglacial lakes in the Thompson and Nicola basins to the south (Fulton and Walcott 1975, Johnsen and Brennand 2004 see Fig. 1.1). These tilts suggest that the maximum crustal deformation and presumably greatest ice thickness existed somewhere to the
northwest, in the vicinity of the southern Fraser Plateau (cf. Wilson et al. 1958, Tipper 1971b) or Coast Mountains. The distribution of ice-marginal channels (Margold et al. 2013) and recent numerical modelling (Seguinot 2014) also suggest ice retreat toward the Coast Mountains.

Grounding line moraines (Type 2) in topographic basins confirm the existence of ice-contact proglacial lakes at the margin of the retreating ice sheet (cf. Chapter 3). Their presence indicates a significant amount of water and sediment were being delivered to the ice margin during glacier retreat.

Localized rectilinear grids of crevasse fill ridges (Type 3) formed at the bed of the retreating ice sheet and record a marginal zone of crevassed, thin ice abutting some proglacial lakes. Their presence confirms that localized stagnation was occurring on the southern Fraser Plateau during active retreat of the contiguous ice margin (cf. Perkins et al. 2013).

Conclusions

With improved datasets, we confirm the presence of a large number of glaciotectonic moraines formed during both advance and retreat of the last CIS across the southern Fraser Plateau. Ice advanced into the southern Fraser Plateau from the northeast, based on the identification of overridden glaciotectonic moraines. Subglacial ribbed terrain records the shift in direction of ice flow (from southwest to southeast) and the change in thermal regime as the ice sheet thickened. Identification of recessional glaciotectonic and grounding-line moraines suggests an ordered, active ice-marginal retreat northwestward toward the Coast Mountains, consistent with recent inferences based on the mapping of ice-marginal channels (Margold et al. 2013) and numerical modelling (Seguinot 2014). The presence of an active ice margin during retreat indicates that an ice surface slope towards the southwest must have been present. Based on retreat direction this places the maximum ice thickness to the northwest of the study area, affirming previous inferences of an ice divide near the southern Fraser Plateau based on landform interpretations (Wilson et al. 1958, Tipper 1971b) and reconstructions of glacioisostatic tilt from paleolake shorelines (Fulton and Walcott 1975, Johnsen and
Brennand 2004). Active ice-marginal lakes within depressions on the Plateau encouraged the formation of grounding-line moraines in topographic lows during ice retreat, and may have facilitated the formation of recessional moraines on the surrounding topographic highs due to the transference of basal pressure to these grounded areas. Pockets of localized stagnation, supported by the presence of crevasse-fill ridges, likely existed as large-scale retreat occurred across the Plateau surface.
References


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5. Towards a morphogenetic classification of eskers: implications for modelling ice sheet hydrology
Abstract

Validations of paleo-ice sheet hydrological models have used esker spacing as a proxy for ice tunnel density. Changes in crest type (cross-sectional shape) along esker ridges have typically been attributed to the effect of changing subglacial topography on hydro- and ice- dynamics and hence subglacial ice-tunnel shape. These claims assume that all eskers formed in subglacial ice tunnels and that all major subglacial ice tunnels produced a remnant esker. We identify differences in geomorphic context, sinuosity, cross-sectional shape, and sedimentary architecture by analyzing eskers formed at or near the margins of the last Cordilleran Ice Sheet on British Columbia’s southern Fraser Plateau, and propose a morphogenetic esker classification. Three morphogenetic types and 2 subtypes of eskers are classified based on differences in geomorphic context, ridge length, sinuosity, cross-sectional shape and sedimentary architecture; they largely record seasonal meltwater flows and GLOFs through sub-, en- and supraglacial meltwater channels and ice-walled canyons.

General principles extracted from these interpretations are: 1) esker ridge crest type and sinuosity strongly reflect meltwater channel type. Eskers formed in subglacial conduits are likely to be round-crested with low sinuosity (except where controlled by ice structure or modified by surging) and contain faults associated with flank collapse. Eskers formed near or at the ice surface are more likely to be sharp-crested, highly sinuous, and contain numerous faults both under ridge crest-lines and in areas of flank collapse. 2) Esker ridges containing numerous flat-crested reaches formed directly on the land-surface in ice-walled canyons (unroofed ice tunnels) or in ice tunnels at atmospheric pressure, and therefore likely record thin or dead ice. 3) Eskers containing macroforms exhibiting headward and downflow growth likely record flood-scale flows (possibly glacial lake outburst floods (GLOFs) where a lake can be inferred). These conclusions suggest that esker crest type, sinuosity and geomorphic context, when understood along with sedimentary architecture, largely reflect formational position with respect to the ice-surface. Reconstructions of ice sheet hydrology need to account for variation in esker morphology because basing hydrodynamic inferences on the presence or absence of an esker alone ignores encoded differences in water source, supply, flow magnitude and frequency, and conduit position.
Introduction

Models of ice dynamics fundamentally rely on an accurate representation of ice-sheet hydrology (e.g., Schoof, 2010; Bartholomaus et al., 2011; Colgan et al., 2011): efficient subglacial drainage systems (channelized meltwater) tend to moderate ice flow whereas distributed subglacial systems tend to enhance ice flow (e.g., Sole et al., 2011; Sundal et al., 2011). Eskers record the casts of ice-walled meltwater channels, and as such esker spacing has been invoked as a proxy for subglacial ice tunnel density in validations of models of paleo-glacier hydrology (e.g., Boulton et al., 2009). In addition, changes in cross-sectional shape along esker ridges within the footprints of paleo-ice sheets have typically been attributed to the effect of changing subglacial topography on hydro- and ice- dynamics and hence subglacial ice tunnel shape (Price 1966, Shreve 1985). However, both claims assume that eskers formed in subglacial ice tunnels and that all major subglacial ice tunnels produced a remnant esker. We use the term esker sensu Bannerjee and McDonald (1975) to reflect the deposition of stratified alluvial sediment within a channel bounded by glacial ice on both sides. Research linking esker ridge morphology, sedimentary architecture, and geomorphic context to esker genesis and hence meltwater channel type (i.e. sub-, en-, or supra-glacial, or ice-walled canyon) rarely accompanies such claims, though such inferences could be inaccurate if underlying assumptions are found to be wrong.

Field research has shown that eskers may form in subglacial ice tunnels, englacial conduits, and supraglacial channels, as well as in ice-walled canyons (e.g., Price, 1966; Bannerjee and McDonald, 1975; Syverson et al., 1994; Warren and Ashley, 1994). Past efforts to associate esker ridge morphology with esker genesis have generally focused on ridge cross-sectional shape (crest type) and the longitudinal topographic profile of the ridge crestline. Crest type has been associated with meltwater channel type: (1) sharp-crested ridges have been observed to form from the meltout and subsequent let-down of supraglacial channel sediments (e.g., Price, 1966; Syverson et al., 1994) but have also been associated with downslope water flow within subglacial ice tunnels (Shreve, 1985); (2) flat-crested ridges have been associated with water flow at atmospheric pressure or where non-pipe-full conditions maintain triple-point pressure (Hooke, 1984) and also with up-slope water flow (Shreve, 1985). Unfortunately, past work has provided little
quantification of cross-sectional shape (in many areas robust quantification awaits the availability of higher resolution DEMs), and has rarely provided clear definitions for cross-sectional shape classes (e.g., Burke et al., 2012b, 2015) and thus comparisons between studies are fraught with uncertainty. Ridge crestline longitudinal topographic profile has also been associated with meltwater channel type: upslope segments have been attributed to pipe-full flow in subglacial ice tunnels (Bannerjee and McDonald, 1975), and englacial conduit or supraglacial channel sediment may meltout (be lowered) onto adverse topographic slopes (Price, 1966; Syverson, 1994).

In this paper we pursue whether a combined consideration of esker morphometric variables and geomorphic context may be diagnostic of esker genesis and meltwater channel type by exploring the relationships between esker geomorphic context, morphology, sedimentary architecture and robust inference of genesis. To this end we map the distribution, and summarize the geomorphic and geologic context and morpho-sedimentary character (from ground penetrating radar (GPR) and electrical resistivity tomography (ERT) surveys, and limited exposure sedimentology) of eskers formed at or near the margins of the last Cordilleran Ice Sheet on British Columbia’s southern Fraser Plateau (Fig. 5.1). Differences in geomorphic context, sinuosity, cross-sectional shape, and sedimentary architecture suggest a morphogenetic esker classification, which aids in reconstruction of local patterns of ice sheet retreat, and is a step toward improving the use of eskers as verification for numerical ice sheet models that include channelized meltwater flow.

**Study Area and Previous Research**

Work was completed on a ~7800 km² area of the southern Fraser Plateau, confined by the Fraser River on the west, and the Bonaparte Lake region in the east (Fig. 5.1). The southern Fraser Plateau is located between the Coast and Columbia mountain ranges in south-central BC and averages around 1175 m asl (Fig. 5.1B, Holland, 1976; Geobase®). The relatively low relief plateau surface is primarily composed of Miocene basalt flows which have generally in-filled basement topography (Andrews et al., 2011); intrusive igneous and sedimentary rock outcrop or subcrop in a third of the study area. Regional mapping suggests bedrock is overlain by ≤ 20 m of till (Andrews et al., 2011), but study area outcrops consistently revealed typical thicknesses of 1-2 m. Glaciofluvial
sediments ( eskers, kames ) and glaciolacustrine sediments and landforms ( including deltas ) are also common ( Plouffe et al., 2011; chapters 2 and 3 ). The location of the study site was chosen based on its accessibility, preliminary work on glacial history ( Tipper 1971a, b, c; Plouffe et al., 2011 ) and the identification of a concentration of eskers, including one of the longest eskers in south-central BC ( the Chasm esker, Tipper, 1971b; Burke et al., 2012b ).
Figure 5.1. A) Location of the study area (dark grey polygon) on the southern Fraser Plateau (light grey polygon, Holland, 1976) in south-central British Columbia. B) Location and topographic setting (Geobase®) of esker types on the southern Fraser Plateau. Bold black line outlines the study area. Refer to the text for a description of esker type. White boxes (1-5) show the location of eskers explored in this paper: 1) Chasm esker; 2) Young Lake esker-like ridge; 3) Green Lake esker (Fig. 5.4); 4) Canoe Creek esker (Fig. 5.8); 5) Hooke Road esker (Fig. 5.11). Dot-dashed lines i-i’ and ii-ii’ indicate the location of transects used to estimate average distance between esker ridges on the plateau surface.
Regional ice flow history, reconstructed from the orientation of streamlined forms, meltwater channels, till geochemistry, pebble lithologies and supplemented by striae orientations (Tipper, 1971b; Plouffe et al., 2011), indicates that during Marine Isotope Stage 2 (MIS 2) ice flowed onto the plateau from source areas in the Coast Mountains to the west and the Cariboo Mountains (a sub-range of the Columbia Mountains) to the east, eventually coalescing somewhere west of the Fraser River and forcing flow north and south (Heginbottom, 1972; Tipper, 1971b, c; Plouffe et al., 2011). During deglaciation, the ice is thought to have regionally stagnated because of a rapid rise in equilibrium line altitude (Fulton, 1991), with minor marginal backwasting across the interior plateaus (Fulton, 1967; Clague and James, 2002). This concept is supported by the presence of hummocky topography interpreted as ice-disintegration moraine (Fulton, 1967) and a perceived lack of recessional moraines across the plateau surface (Tipper, 1971a). However, reconstructions of nearby glacioisostatic tilt in the Thompson Basin (Fig. 1.1; Johnsen and Brennand, 2004) and lateglacial ice-marginal lake evolution on the southern Fraser Plateau suggest a systematic southeast to northwest pattern of ice margin retreat, accompanied by regional thinning (chapters 2 and 3).

To date, beyond basic mapping of esker distribution (crestlines) from topographic data and aerial photographs (Tipper, 1971b; Bednarski, 2009; Huscroft, 2009; Plouffe, 2009a, b), detailed investigation of eskers on the southern Fraser Plateau has been limited to the exploration of two genetically different eskers: the Chasm esker (Burke et al., 2012b), and the Young Lake esker-like ridge (chapter 2), both recording formative meltwater flow to the southeast.

Chasm esker

Chasm esker (Fig. 5.1B) is a ~32 km long, discontinuous (93% continuity, with relatively long gap lengths generally >100 m), low sinuosity (1.09), largely round-crested, single ridge with long upslope segments, set within a broader canal (meltwater corridor) landsystem (Burke et al., 2012 a, b; Network 29 Appendix H). This landsystem is interpreted to have formed subglacially during a glacial lake outburst flood (GLOF) with esker formation occurring as flow collapsed from broad canal flow into a smaller ice tunnel (Burke et al., 2012a, b). Round-crested segments of the esker are interpreted to have formed largely through vertical and headward accretion at local depocentres (separated
by zones of non-deposition) under pressurized flow in the absence of a free water-plane (Burke et al., 2012b). The presence of ridge-scale macroforms and overall lack of fine materials within round-crested segments indicates rapid deposition of esker materials associated with high magnitude flows. Multi-ridged segments are genetically associated with zones of glaciological structural weakness, likely where canal roof collapse resulted in the formation of significant crevassing (Burke et al., 2012b). Flat-crested segments dominated by vertical accretion suggest that local un-roofing of the ice tunnel occurred within thin ice in the final stages of esker formation (Burke et al., 2012b). The majority of faulting and slumping observed was at landform flanks and likely relates to post-depositional processes (flank collapse as ice walls melted). A few short (<1 km) ridges connect to the trunk ridge at acute angles. Most are likely discontinuous multi-ridged segments rather than true tributaries.

Young Lake esker-like ridge

Young Lake esker-like ridge (Fig 1B) is a short (5.1 km long), continuous (100% continuity), medium sinuosity (1.11), ice-contact deposit composed of proglacial lake, braided outwash and deltaic sediments within a topographic basin (Bonaparte River valley) where long-upslope segments are absent (Chapter 2; Network 3, Table Appendix H). Morphologically it is dominantly flat-crested (88%), with the largest proportion of the ridge classified as single-ridge widenings (38%). It mainly formed by vertical and downflow accretion of subaerial outwash between and/or on dead ice (a thin remnant ice mass unconnected to the active ice sheet (c.f. Fulton, 1967)) that had become stranded within the basin in front of a regionally-backwasting ice margin (Chapter 2). Water was likely supplied from multiple sources including ice-marginal channels, subaerial run-off (supplied from precipitation, snowmelt, and glacial meltwater), as well as direct melt of glacial ice.

Methods

Landform mapping and characterization

Esker ridges on the southern Fraser Plateau were mapped and analyzed from stereopair aerial photographs (≤ 1:40 000) and a digital elevation model (DEM) (25 m
horizontal grid cells, 1 m vertical resolution (90% of measurements within 5 m of true elevation), Geobase®).

**Esker ridge mapping, terminology and ridge segment classification**

The areal extent of esker ridge segments (and esker-like ridge segments, cf. Young Lake esker-like ridge) are manually mapped as polygons defined by slope-break (visible on topographic cross-sections, aerial photographs and DEMs). Individual ridge segments are separated by gaps. Laterally-adjacent esker ridge segments which diverge and may converge downflow are classified as primary (highest relief) or secondary (lower relief), and the approximate crestline of each are mapped as lines. Longitudinally-aligned ridge segments form individual esker ridges. Eskers are defined as longitudinally-aligned ridges plus intervening gaps (digitized as straight-line segments). Esker networks are composed of one trunk esker, which forms the spine for one or more tributaries.

Ridge segments are visually classified based upon their planform and crest type (topographic cross-profile shape). Ridge planform is classified as either: 1) single ridge; 2) multi-ridged (primary ridge plus one or more secondary ridges); or 3) widening (approximately 50% increase in ridge width compared to upflow). Crest type is classified as: 1) round-crested (resemble half-cylinders); 2) sharp-crested (resemble triangular prisms); 3) flat-crested (resemble half-hexagonal prisms); or 4) multi-crested (resemble half-hexagonal prisms with multiple high points) (Fig. 5.2, cf. Burke et al., 2012b).

**Esker ridge morphometry and geomorphic context**

Esker ridge morphometry and geomorphologic context were determined for esker trunk ridges and recorded within a geographic information system database. Because tributary ridges comprise a relatively small proportion of total esker network length (<4%), analysis of esker ridge morphometry in this study was limited to trunk ridges.
Morphometric analysis includes trunk ridge length (i.e. length of the primary ridge minus gaps), and esker continuity and sinuosity.

Esker continuity is disrupted by gaps between ridge segments, either recording non-deposition or post-depositional erosion (Banerjee and McDonald, 1975; Brennand, 1994). Trunk esker continuity is quantified as the sum of individual trunk ridge segment lengths divided by esker length and presented as percentage continuity (100% continuity being a trunk ridge with zero gaps).

Trunk esker sinuosity is calculated by dividing ridge crestline plus gap length by the straight line distance between the start and end of the esker. Two additional measures were assessed:

1. Trunk esker ridge segment sinuosity - the crestline length of individual ridge segments of esker trunks (as broken up by gaps), divided by straight lines joining their start and end points, followed by summation and averaging for each trunk esker;

2. Morphological trunk esker ridge segment sinuosity - the length of trunk esker segments parsed by morphologic classification (planform or crest type), divided
by straight lines joining their start and end points, followed by summation and averaging for each morphologic type.

Initial testing showed that the sinuosity values for all measures were not significantly different and thus the sinuosity reported in this paper is trunk esker sinuosity. Results of a Wilcoxon signed-ranks test (used because sinuosity data approach a natural limit (1) and were in two categories (entire esker trunk vs. individual trunk ridge segments for a given network) demonstrated that sinuosity is not significantly different between an entire trunk esker ridge and its respective ridge segments (probability, $p = 0.69$, 95% confidence interval (C.I.)). Furthermore, there was no significant difference in sinuosity between ridge segment planform type ($p = 0.541$, 95% C.I.) or crest type ($p = 0.541$, 95% C.I.) (a Kruskal-Wallis test was used here given that sinuosity data approach a natural limit (1) and had greater than two categories (i.e., 4 different planform types and 4 different crest types)). Interestingly, when the category of widening was removed from the crest type variable, a significant difference ($p = 0.045$, 95% C.I.) was found to exist between ridge segment crest type (round-crested, sharp-crested and flat-crested) sinuosities. This suggests that variations in morphological segment sinuosity may be helpful in determining variation in depositional mode within esker networks, but this requires further analysis that is beyond the scope of this investigation.

The geomorphic context of esker trunks, including the presence of tributaries, presence within a valley or broader meltwater corridor (cf. Burke et al., 2012a), the presence of terminal ridge widenings (potential fans that could indicate the presence of an ice margin; a lack of accessibility and exposures did not allow us to confirm the presence of terminal fans), and the presence of ridge segments on adverse slopes (“upslope segments”), is also reported. Associations between trunk ridges and bedrock type (cf. Clarke and Walder, 1994) are assessed by comparing the proportion of trunk ridge length atop each bedrock type within the field area to the proportional area of bedrock types across the study area.
**Geophysical data collection**

Sedimentary exposures adequate for stratigraphic and architectural analysis are rare in the study area. Therefore geophysical tools (GPR and ERT) were employed to delineate the subsurface sedimentary architecture and composition of ridge segments; interpretations are grounded with field observations where sedimentary exposures were available. Site accessibility (much of the area is tree-covered and lacks roads) constrained the location of geophysical grids, but careful site selection allowed data collection across the range of ridge planforms and cross-sectional shapes (Appendix E.1). Where esker ridge-width allowed, GPR surveys were conducted as pseudo-three-dimensional grids (a series of closely-spaced (~5 m) crossed lines); otherwise a two-dimensional (2D) grid of ridge crest-line surveys and more distantly-spaced cross-sectional surveys (spacing normally defined by site topographic restrictions) was collected (Appendix E.1). GPR data were collected using either a Sensors and Software Inc. pulseEKKO IV (400 V) or pulseEKKO PRO (1000 V) system (dependent on availability during field seasons, Appendix E.1). The difference between a 400 V and 1000 V system is equivalent to a six fold increase in the energy radiated from the transmitter and has the potential to increase signal penetration from 5-14% (Jol, 1995). Based on maximum penetration of radar signal in this study (~10 m), this difference in voltage equates to < 1 m difference in signal penetration (though this could also be associated with subtle variation in material type between grids). Common offset (CO) lines were collected with antennas co-polarized and arranged perpendicular-broadside to the survey line (Arcone et al., 1995). Both 100 and 200 MHz antennas were used, depending on esker height and the desired penetration depth (Appendix E.1). At least one common midpoint profile (CMP) was collected from each esker and provided estimates of subsurface velocity via semblance analysis. Subsurface velocity was used in data processing and to convert two-way travel time to depth.

At several GPR grids two-dimensional ERT surveys were also collected using an Advanced Geosciences Inc. (AGI) Super Sting R8 (eight channel) system. Surveys were arranged in a dipole-dipole array, allowing optimal combination of penetration depth, horizontal and vertical resolution, and were completed with closely-spaced take-outs (0.5-1.5 m; Appendix D.1) to maximize subsurface resolution (Samouëlian et al., 2005). GPR and ERT lines were topographically surveyed using a real-time kinematic differential
global positioning system (RTK dGPS, Leica system 500), providing decimetre positional accuracy. RTK dGPS data was post-processed (Leica Geo Office v. 7) where field conditions prevented continuous communication between the base-station and rover.

**Geophysical data processing and interpretation**

All GPR data were processed using REFLEXW v5.6. Standard processing steps included application of a static correction, ‘dewow’ filter, ‘bandpass’ filter, migration (using velocities estimated from CMP surveys), ‘background removal’ filter, gain function, and topographic correction (from RTK dGPS survey). Optimal parameters for each processing step were evaluated against a single line in each grid and then applied equally to all lines in that grid. GPR profiles were analyzed for high amplitude, continuous reflections, identified throughout multiple lines in the grid. These reflections were selected as bounding surfaces, delimiting the lower contacts of radar elements within the grids. Radar elements are differentiated based on depositional style (e.g., headward, downflow, vertical, or lateral (where identifiable in cross-lines) accretion, Table 5.1) and labelled in approximate formational order based on truncation and accretion patterns (as confidently as possible given limitations of radar grid resolution and density). Offset reflections are interpreted as faults and may be more continuous than imaged in individual GPR profiles due to fault and survey line alignment, as well as resolution issues (cf. Fiore et al., 2002; Woodward et al., 2008).

ERT data were inverted using AGI EarthImager 2D v. 2.40. Inversion sequences were set to use the finite element method as a forward model and run with a smooth-model inversion routine. All negative resistivity values and data spikes were removed. Data points were measured for goodness-of-fit following inversion and the poorest values were removed prior to the subsequent inversion. This allowed the total root-mean squared (RMS) error for the inversion process to be <6% and the least-squares (L^2) coefficient to be mainly <4 (Appendix D.1-D.3, D.11-D.17). Major changes (order of
### Table 5.1. Radar element stratigraphy, characteristics and interpretation

<table>
<thead>
<tr>
<th>Element stratigraphy</th>
<th>Res. (Ωm) $^1$</th>
<th>7.1</th>
<th>240</th>
<th>812</th>
<th>2747</th>
<th>8289</th>
<th>10 m</th>
<th>Element characteristics $^2$</th>
<th>Element interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Post-glacial deposits</strong> $^3$</td>
<td>200-800</td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td>Thin, low resistivity tabular element at surface, draping lower radar elements.</td>
<td>Aeolian deposits</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td>Internal reflections typically obscured by air and ground-waves.</td>
<td>Slump deposits</td>
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<td></td>
<td></td>
<td></td>
<td>Corresponding fine sand/silt material visible at ridge surface or in section</td>
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<td><strong>Ridge deposits</strong></td>
<td>800-10000</td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
<td>Thick, wedge or lens-shaped elements that downlap or are concordant with lower radar elements</td>
<td>Headward accretion</td>
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<tr>
<td><strong>Flow direction (X-lines)</strong></td>
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<td>Internal reflections dip upglacier at an angle conformable to or greater than that of the lower bounding surface</td>
<td>Downflow accretion</td>
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<td></td>
<td>Wedge, lens, or elongate-trough shaped, moderate resistivity elements that commonly truncate lower radar elements</td>
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<td></td>
<td>All internal reflections dip downglacier at an angle conformable to or greater than that of the lower bounding surface</td>
<td>Lateral accretion</td>
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<td></td>
<td>Wedge-shaped elements, conformable with lower radar elements</td>
<td>Vertical accretion</td>
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<td></td>
<td>Internal reflections are planar or sigmoidal and subparallel, dipping away from ridge centre in Y-lines</td>
<td>Scour and fill</td>
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<td>Lens, wedge, elongate trough or tabular-shaped, high resistivity elements that truncate underlying radar elements</td>
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<td></td>
<td></td>
<td>Internal reflections are horizontal, subparallel and continuous, except where post-depositional faulting has disturbed bedding</td>
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<tr>
<td><strong>Ridge substrate</strong></td>
<td>70-1000</td>
<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
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<td><img src="image" alt="Image" /></td>
<td><img src="image" alt="Image" /></td>
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<td>Canal floor/esker substrate</td>
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<td>Trough-shaped elements that truncate underlying radar elements</td>
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<td></td>
<td></td>
<td>Internal reflections are concave or sigmoidal and downlap lower radar elements</td>
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<td></td>
<td></td>
<td>Tabular, low to moderate resistivity elements located grid-wide at base of ridge deposits</td>
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<td></td>
<td>Often associated with radar signal attenuation (fine-grained sediments and/or water table) masking internal reflections</td>
<td></td>
</tr>
</tbody>
</table>

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$^1$ Range of resistivity values from ERT profiles collected from survey grids.

$^2$ GPR and ERT data (where available) are shown above interpreted data. (thick lines are radar bounding surfaces, thin lines are characteristic internal reflections). Where ERT data were collected as cross-lines perpendicular to GPR lines, resistivity values are shown as a column at point of intersection.

$^3$ Post-ice melt.
magnitude) in resistivity were interpreted as changes in subsurface composition. Interpretations of major resistivity changes are grounded against bounding surfaces in GPR data (where overlapping) and also against sedimentary exposures and well log data (BCME, 2011) where available. The water table was identified in ERT profiles where extremely low resistivity values occur at elevations similar to those of nearby surface water.

**Results**

New data from type 1a, 1b and 2 eskers are presented below (3, 4, and 5 in Fig. 5.1B, respectively), building upon data and previous interpretations of Burke et al. (2012b) and Chapter 2 for the Chasm esker and Young-lake esker-like ridge (type 1a and 3, respectively, in this paper). Together, these new and previously published data allow us to explore the range of morpho-sedimentary relationships and geneses of esker ridges mapped across the southern Fraser Plateau.

**Esker ridge geomorphology and distribution on the southern Fraser Plateau**

Over 124 km (crestline measurement) of primary esker ridge segments are mapped in the study area (Fig 1B). Transects parallel to the retreating ice margin (Chapters 2, 3 and 4) and through esker ridges on the southern Fraser Plateau (see Fig. 5.1B) show that esker trunks are separated by a mean distance of 20 km. Esker trunks (n = 34) have a mean linear orientation of 113°/253° (circular variance = 0.25), suggesting they are generally oriented WNW-ESE, roughly perpendicular to the retreating ice margin. Trunk esker sinuosity was found to be significantly different across esker types (p = 0.004, 95% C.I.; Kruskal Wallis test).

Eskers on the southern Fraser Plateau were qualitatively categorized into 3 types and 2 subtypes based on ridge morphology and geomorphic context (Figs 5.1B and 5.3, Tables 5.2 and Appendix H). Type 1 eskers are relatively low sinuosity, largely round-crested, single ridges and are often set within valleys or broad meltwater corridors (Burke et al., 2012a). Differences in length, continuity and the presence of upslope
Figure 5.3. Characteristics of esker types. Sample size varies: type 1 = 25 (type 1a = 2, type 1b = 23), type 2 = 7, type 3 = 2. Length % (Y-axis) in A, B and F excludes gaps (i.e. trunk ridge length). A) Trunk ridge segment planform by esker type. B) Trunk ridge segment crest type by esker type. C) Mean trunk esker continuity and sinuosity by esker type (refer to text for derivation). D) Geomorphic context of trunk eskers by esker type. E) Mean trunk ridge length and mean elevation by esker type. F) Bedrock type (Massey et al., 2005) underlying trunk ridges as a proportion of trunk ridge length by esker type; percent areal coverage of bedrock type for the entire study area also shown.

segments resulted in this type being split into two subtypes. Type 1a eskers (n = 2; Table 5.2; Fig. 5.3; e.g. Green Lake esker, Figs 5.4-5.7) are relatively long ridges with relatively low continuity (though a high proportion of gaps are likely related to post-depositional incision). They are set within broad meltwater corridors on the plateau surface that end in deeply-incised canyons (e.g. Chasm esker, Fig 1B, Burke et al., 2012a); they have
relatively low mean elevation. This subtype has long upslope segments, and may include minor multi-ridged reaches and flat-crested segments. Up to 40% of type 1a esker trunk length is characterized by widenings. Type 1b eskers \((n = 23; \text{Table 5.2; Fig. 5.3; e.g. Canoe Creek esker, Figs 5.1B and 5.8-5.10})\) are shorter, highly variable in orientation, may be located in small valleys on the plateau surface, and often terminate in flat-crested widenings which may record fans (access did not allow this to be clarified). Ridge planform, crest type and sinuosity are similar to type 1a but continuity is higher, long upslope segments are rare, and short tributary ridges are sometimes present. Type 2 eskers \((n = 7; \text{Table 5.2; Fig. 5.3; e.g. Hooke Road esker, Figs 5.11 and 5.12})\) are mid-length, dominantly single, sharp-crested, continuous ridges (Fig. 5.3A-E) which may contain tributaries. They exhibit sharp \((> 90^\circ)\) bends resulting in high sinuosity \((\text{mean} = 1.3, \text{but with segments as high as 1.8})\), and often initiate at the terminus of an ice-marginal channel (though are not usually set within valleys themselves; e.g., Hooke Road esker, Fig. 5.1B), occasionally ending in widenings that morphologically resemble fans. Long upslope segments are not present and mean esker elevation is considerably higher than the study area mean; these eskers are normally found on the hills surrounding the plateau at the edges of the study area, rather than on the plateau itself. Type 3 eskers \((n = 2, \text{Table 5.2; Fig. 5.3; e.g. Young Lake esker-like ridge, Fig. 5.1B})\) are approximately equally divided into single ridge, single ridge widening and multi-ridged segments. They are normally flat-crested, highly continuous, mid-length ridges of low sinuosity; upslope segments are absent. They occur within topographic basins, resulting in a mean elevation considerably lower than the study area mean, and terminate in low-gradient widenings morphologically resembling fans.
Table 5.2  A morphogenetic classification of eskers on the southern Fraser Plateau

<table>
<thead>
<tr>
<th>Esker Type</th>
<th>Examples</th>
<th>Dominant planform</th>
<th>Dominant crest type</th>
<th>Geomorphology</th>
<th>Sedimentary architecture</th>
<th>Meltwater channel type</th>
<th>Water source</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a (n=2)</td>
<td>Green Lake, Chasm'</td>
<td>Single ridge (60%)</td>
<td>Round-crested (62%)</td>
<td>Long (&gt;10 km); relatively discontinuous (80% &amp; 93% continuity); relatively long gap lengths (typically 100's m); low sinuosity (1.09, 1.10); set within broad meltwater corridors; long-upslope segments; rare short (&lt;1 km) tributaries.</td>
<td>Dominantly vertical accretion; presence of ridge-scale macroforms; flat-crested segments associated with tubular, vertical accretion.</td>
<td>Subglacial ice tunnel (perhaps near-basal englacial conduit segments) within broader meltwater corridors under/in a thin ice sheet. Ridge-scale macroforms represent localized depocentres with headward and vertical growth under thin ice. Ridge anabranching associated with crevassing. Thermo-mechanical ice-tunnel growth may result in conduit unroofing and late-stage open-channel flow.</td>
<td>Mainly point source: ice-dammed lake drainage (GLOF); possible addition of supraglacial melt for flat-topped esker segments.</td>
</tr>
<tr>
<td>1b (n=23)</td>
<td>Canoe Creek</td>
<td>Single ridge (66%)</td>
<td>Round-crested (88%)</td>
<td>Short (&lt;6 km, mean = 1.9 km); typically discontinuous (71-100% continuity; mean 94%); variable gap lengths (0-1000 m); typically low sinuosity (1.02-1.30, mean -1.10); often set within valleys (70%); long-upslope segments rare (&lt;4%); occasional (22%) short (&lt;1-1.5 km) tributaries; fan-shaped termini common.</td>
<td>Segments may include initial headward and downflow accretion (i.e., macroforms) with later vertical accretion and late-stage scour and fill.</td>
<td>Subglacial ice tunnel under and near the margins of an ice sheet.</td>
<td>Mixed sources: likely moraine-dammed or ice-dammed lake drainage (GLOF), but possible contributions from glacier melt and subaerial runoff from adjacent slopes (precipitation, snow-melt, dead ice melt).</td>
</tr>
<tr>
<td>Esker Type</td>
<td>Examples</td>
<td>Dominant planform</td>
<td>Dominant crest type</td>
<td>Geomorphology</td>
<td>Sedimentary architecture</td>
<td>Meltwater channel type</td>
<td>Water source</td>
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</tr>
<tr>
<td>2 (n=7)</td>
<td>Hooke Road</td>
<td>Single ridge (68%)</td>
<td>Sharp-crested (82%)</td>
<td>Short (&lt;7 km, mean = 3.3 km); fairly continuous (96-100%); short gap length (typically &lt;100 m); high sinuosity (1.13-1.38; mean = 1.3); long-upslope segments absent; occasional, well developed, short (&lt;2 km) tributaries. Often connected to upflow valleys and may terminate in widenings morphologically resembling fans.</td>
<td>Vertical accretion with some elements of scour and fill; high frequency of faulting.</td>
<td>Supraglacial channel or high englacial conduit atop or within an ice sheet and downflow of lateral ice-sheet margins.</td>
<td>Mixed sources: subaerial run-off (precipitation, snow &amp; ice melt) from adjacent slopes and through ice-marginal channels; glacier melt.</td>
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| 3 (n=2)    | Young Lake esker-like ridge<sup>2</sup> | Multi-ridded (37%) | Flat-crested (84%) | Short (<7 km); continuous (100% continuity); low sinuosity (1.08 & 1.11); set within valleys (100%); long-upslope segments absent. | Vertical and downflow accretion with minor elements of scour and fill. | On-ice channel or subaerial ice-walled canyon on or in a detached (from ice sheet) dead ice mass. | Mixed sources: subaerial run-off (precipitation, snow & ice melt); glacier melt. |

<sup>1</sup>Refer to Burke et al., 2012b for full data and grid interpretation.

<sup>2</sup>Refer to Chapter 2 for full data and grid interpretation.
Figure 5.4. Green Lake esker. A) Esker morphology overlain on an orthophoto (Province of British Columbia, 2010; Geobase®). Black boxes indicate locations of geophysics grids G1 and G2 (Figs 5 and 6). Black line (Z to Z’) indicates esker ridge crestline and topographic long-profile in ‘B’. B) Topographic long profile (Z-Z’) of ridge crestline and floor of meltwater corridor (elevation data from Geobase®, see ‘A’ for long profile path). Esker geomorphology indicated by coloured bars (refer to legend in ‘A’ for classification); non-coloured spaces represent gaps in esker continuity.
Figure 5.5. Character and stratigraphy of the Green Lake esker. A) Processed GPR data overlain on ERT for line Y1 at Green Lake Esker grid G1 (see Fig. 5.4 for grid location and Fig. 5.6A for full extent of radar lines in this grid). Double arrowed lines indicate where GPR cross-lines (X#) intersect the ERT profile. Resistivity units labelled RU1-3. Radar elements labelled in order of deposition. Water table elevation is ~1075 m asl (estimated from surface elevation of nearby ponds). B) Lithostratigraphic unit interpretations with associated resistivity units for line Y1, grid G1. C) Esker gravel (RU2) within exposure at the end of line X3, grid G1. Scale bar is marked in decimetre increments. D) View from line X3, grid G1 looking north. Fine-grained postglacial sediment deposited on ridge surface visible in foreground. White arrow shows enclosed depression within this multi-crested segment of the esker. E) Silt and fine sand at surface overlying gravel deposits, characteristic of thin RU3 cover over ridge surface outside of depressions. Notebook is 13 x 7.5 cm.
Figure 5.6. Processed and interpreted GPR lines from grid G1 (A) and grid G2 (B) collected from the Green Lake esker (only representative lines for G2 shown here; refer to Appendix E.8-E.9 for all lines of two dimensional data). Radar bounding surfaces (bold lines) enclose radar elements (labeled A-P (Grid 1) and A-K (Grid 2), approximate depositional order). Paleoflow directions are interpreted from radar element geometry and internal reflections (see text for full explanation). Locations where grid lines intersect are shown by labelled double-headed arrows. Dashed lines mark offset reflections interpreted as faults. GPR lines are shown with no vertical exaggeration. Inset schematic maps show geophysics grids G1 and G2. Arrowed lines indicate relative position and direction of data collection for GPR and ERT (see Fig. 5.5) surveys. Refer to Figure 5.4A for grid locations.
Figure 5.7. Processed and interpreted GPR lines from grid G1 (A) and grid G2 (B) collected from the Green Lake esker (only representative lines for G2 shown here; refer to Appendix E.10 for all lines in 2-dimensions). Radar bounding surfaces (bold lines) enclose radar elements which are numbered based on approximate depositional order (1-11, grid G1; 1-9, grid G2) and coloured according to broad depositional style (based on the interpretations in Table 1). Paleoflow directions are interpreted from radar element geometry and internal reflections (see text for full explanation). Locations where grid lines intersect are shown by labelled double-headed arrows. Dashed lines mark offset reflections interpreted as faults. GPR lines are shown with no vertical exaggeration. Refer to Figure 5.4A for grid locations and to Figure 5.6 for schematic map of grid layout.
Eskers are mainly deposited atop diamicton (based on existing surficial geology maps: Bednarski et al., 2009; Huscroft et al., 2009; Plouffe et al., 2009a, b) and their distribution appears to be unrelated to bedrock type (Fig. 5.3F). There is no significant difference (Wilcoxon signed ranks test: \( p = 1.0 \), 95% C.I.) between the areal percentage of bedrock by type under all types of esker ridge (gaps not included) and the areal percentage of bedrock in the study area.

**Esker ridge stratigraphic elements**

Interpretation of both GPR and ERT data on representative ridge segments resulted in the identification of distinctive radar elements recording esker ridge depositional style (headward accretion, downflow accretion, lateral accretion, vertical accretion, scour and fill), and pre- (ridge substrate) and post-ridge (slumping, aeolian) deposition (Table 5.1).

**Ridge substrate**

Ridge substrate was identified in ERT profiles associated with the Green Lake and Chasm eskers as a tabular and grid-wide resistivity unit. In GPR profiles the upper bounding surface of these units are associated with rapid signal attenuation, consistent with a reduction in resistivity and/or a function of their depth (weakness of signal penetration, higher likelihood of intercepting the water table). For the Green Lake esker this unit returned low to moderate (70-1000 ohm-m) resistivity measurements (e.g., RU-1, Grid G1, Fig. 5.5A) and the upper elevation of the unit corresponds to the elevation of terrain adjacent to the esker ridge (but still within the canal walls). In burrow holes and degraded road cut sections, the unit appears to be composed of fine-grained diamicton. For the Chasm esker resistivity values were high (1500-13000 ohm-m). For the cases where it was identified in the Green Lake and Chasm eskers, ridge substrate is interpreted as the canal floor (either till or canal fill (i.e., gravel), respectively), deposited prior to the formation of the esker ridge, based on its sedimentary character and stratigraphic position underlying the esker ridge.
Figure 5.8. Canoe Creek esker. A) Esker morphology overlain on an orthophoto (Province of British Columbia, 2010; Geobase®). Black box indicates location of GPR grid C1 (Fig. 5.8). Black line (X to X’) indicates path of ridge crestline topographic long-profile in ‘C’. Ice-marginal channels connect to the upflow end of the esker. Lateglacial lake spillways (Chapter 3) have incised the esker near its centre. B) Topographic long profile (X-X’) of esker crestline and topography immediately adjacent to esker ridge (elevation data from Geobase®, see ‘A’ for long profile path). Esker morphology indicated by colored bars (refer to legend in ‘A’ for classification); non-coloured spaces represent gaps in esker continuity.
Figure 5.9. Canoe Creek esker. A) Processed and interpreted GPR grid C1 collected from Canoe Creek esker (refer to Appendix E.14-E.15 for non-skewed and un-interpreted processed lines). Radar bounding surfaces (bold lines) enclose radar elements (labeled A-ZA). Paleo-flow direction is interpreted from radar element architecture and internal reflections (see text for explanation). Locations where grid lines intersect are shown by labeled double-headed arrows. Dashed lines mark offset reflections interpreted as faults. GPR lines are shown with no vertical exaggeration. Inset schematic map shows geophysics grid C1. Arrowed lines indicate relative position and direction of data collection for GPR surveys. Refer to Figure 5.8A for grid location.
Ridge deposits

Ridge deposits occur within the esker ridges and are characterized by moderate to high resistivity radar elements (800-10 000 ohm-m), consistent with the presence of sand and gravel as seen on the landform surface, in burrow holes and limited sediment exposures. Radar elements are classified into depositional styles based upon element geometry and internal reflection characteristics (Table 5.1), following the approach adopted by Burke et al. (2012b):

- **Headwardly accreting** radar elements are 20-70 m long, up to 6 m thick, wedge or lens-shaped and conformably overlie lower radar elements. They contain moderately continuous, downlapped or concordant, parallel reflections with apparent upglacier dips of 5-30° in flow-parallel lines and sub-parallel reflections in flow-perpendicular lines (e.g., RE-C, Grid G1, Fig. 5.6A; RE-C, Grid C1, Fig. 5.9A; cf. headwardly accreting gravel sheets, Wooldridge and Hickin, 2005)

- **Downflow accreting** radar elements are composed of parallel reflections that dip downflow 9-30° in flow-parallel lines and are sub-parallel in flow-perpendicular lines (e.g., RE-E, Grid G1, Fig. 5.6A; RE-J, Grid C1, Fig. 5.9A; cf. low-angle downstream accretion deposits, Wooldridge and Hickin, 2005). They typically have a lens-shaped geometry (3-5 m thick, 12-105 m long) and lower bounding surfaces that truncate deeper radar elements.

- **Laterally accreting** radar elements are similar in geometry to downflow accreting elements, but are composed of across-flow dipping reflections that appear sub-horizontal in flow-parallel lines (e.g., RE-B, Grid G2, Fig. 5.6B).

- **Vertically accreting** radar elements are composed of continuous, sub-horizontal internal reflections onlapped onto lower bounding surfaces in both flow-parallel and flow-perpendicular lines. Radar element geometry is varied and they may form wedge, lens, elongate-trough or tabular units (1-5 m thick, 20-150 m long) that unconformably overlie lower radar elements (e.g., RE-H, Grid C1, Fig. 5.9A).

- **Scour-and-fill** radar elements are trough-shaped, up to 1-3 m thick, 5-20 m long, have lower bounding surfaces that truncate deeper reflections, and are composed of concave or sigmoidal internal reflections regardless of line orientation (e.g., RE-I, Grid G1, Fig. 5.6A; RE-W, Grid C1, Fig. 5.9A; RE-J, Grid H1, Fig. 5.12A; cf. cut and fill, Sambrook Smith et al., 2006).
Figure 5.10. Processed and interpreted GPR grid C1 collected from Canoe Creek esker (refer to Appendix E.16 for non-skewed and un-interpreted processed lines). Radar bounding surfaces (bold lines) enclose radar elements which are labelled based on approximate depositional order (1-21) and broad depositional style (colour). Paleoflow direction is interpreted from radar element architecture and internal reflections (see text for explanation). Locations where grid lines intersect are shown by labelled double-headed arrows. Dashed lines mark offset reflections interpreted as faults. GPR lines are shown with no vertical exaggeration.
Figure 5.11. Hooke Road esker. A) Esker morphology overlain on an orthophoto (Province of British Columbia, 2010; Geobase®). Black box indicates location of GPR grid H1 (see Fig. 5.10). White dot locates the exposure discussed in the text. Black line (Y to Y’) indicates path of esker crestline and topographic long-profile in ‘B’. B) Topographic long profile (Y-Y’) of ridge crestline and topography immediately adjacent to the esker ridge (elevation data from Geobase®, see ‘A’ for long profile path). Esker morphology indicated by colored bars (refer to legend in ‘A’ for classification); non-coloured spaces represent gaps in esker continuity.
Figure 5.12. Processed and interpreted GPR grid H1 (A) collected from the Hooke Road esker (refer to Fig. S8 for un-interpreted processed lines). Radar bounding surfaces (bold lines) enclose radar elements (labeled A-N). Paleoflow directions are interpreted from radar element architecture and internal reflections (see text for explanation). Locations where grid lines intersect are shown by labelled double-headed arrows. Dashed lines mark offset reflections interpreted as faults. GPR lines are shown with no vertical exaggeration. Inset schematic map shows geophysics grid H1. Arrowed lines indicate relative position and direction of data collection for GPR surveys. Refer to Figure 11A for grid location. B) Radar elements are labelled based on approximate depositional order (1-9), and coloured according to broad depositional style (based on the interpretations in Table 1).
Postglacial deposits

Ridge flanks and substrate adjacent to ridges may be mantled by low resistivity (200-800 ohm-m), asymmetric, trough-shaped elements. These elements are distributed unevenly along ridge flanks and within basins between ridges and they are composed of discontinuous or faulted, sinuous or chaotic internal reflections (e.g., RE-N, Grid H1, Fig. 5.12A). The position and character of these elements are similar to slump deposits along the flanks of other eskers on the Fraser Plateau (cf. Burke et al., 2012b), and therefore we interpret them as slump blocks generated due to removal of lateral and/or buried ice support. Thin (< 1 m) low resistivity units that mantle the ridge crest (e.g., ENE end of RU3, Grid G1, Fig. 5.5, RE-Q, Grid G1, Fig. 5.6A), and are composed of fine sand as observed at the landform surface, are interpreted as aeolian deposits (often obscured within the air and ground-wave of GPR lines; Lian and Huntley, 1999).

Green Lake esker (type 1a)

Geomorphology

Green Lake esker is 12.6 km long (including gaps), up to 15 m high, and lies within a mainly NW-SE-oriented, till-walled meltwater corridor (cf. canal, Burke et al., 2012a) that is up to 2 km wide in places (Fig. 5.4; Network 11 Appendix H). The esker terminates where the meltwater corridor opens into a wide canyon at the southern edge of the Fraser Plateau that is incised into volcanic bedrock and exhibits a knickpoint with plunge pool (Fig. 5.1B, Burke et al., 2012a). The esker is dominated by a single, low sinuosity (1.1) ridge (93% of length), but also exhibits several widenings (80-400 m in width), and two short multi-ridged segments near its terminus (Fig. 5.4A and B). The esker is largely round-crested (68% length) but has two multi-crested segments (17% length) in its upflow reach (0-8 km, Fig. 5.4B) and two flat-crested segments (15% length) in its downflow reach (8-12.6 km, Fig. 5.4B; Appendix H). Hectometre-scale long-profile undulations are present at round-crested segments. Long gaps (~500 m average) occur between ridge segments (80% continuity), mainly within the upflow reach. No fans or tributaries were observed along the length of the esker ridge.
Stratigraphy and sedimentary architecture

Two grids of geophysical data were collected from Green Lake esker (Fig. 5.4). Grid G1 is on a multi-crested, single ridge widening in its upflow reach, and is located on an adverse slope. Grid G2 is on one flat-crested ridge within a multi-ridged segment in its downflow reach, and is located on a gently descending slope. Grid G1 includes > 700 m of GPR data (Figs 5, 6, Appendix E.8-E.10), and an overlapping, flow-perpendicular line of ERT data (Y1, 166.5 m, Fig. 5.5). Grid 2 includes 510 m of GPR data (Figs 6, Appendix E.11-E.13), but no ERT data due to site access issues.

Grid G1

Three resistivity units (RU1-3) were identified from the ERT data, and align with major bounding surfaces in the overlapping GPR profile (Grid G1, line Y1, Fig. 5.5A). RU1 (Fig. 5.5A) is at least 30 m thick (no lower contact visible) and contains low, but gradually increasing resistivity values upwards through the unit (~70 Ωm near the centre and base of the profile to ~800 Ωm at the top of the unit, Fig. 5.5A). Signal attenuation in the GPR data coincides with decreasing resistivity near the top of RU-1 and suggests an increase in water content or changes in sediment texture, or both (cf. Neal, 2004). The downward decrease in resistivity begins close to local water table (1075 m asl, based on the elevation of a nearby lake). The resistivity values within the top of RU-1 match those for fine-grained diamicton interpreted from elsewhere on the plateau (Burke et al., 2012a, b; Chapter 2) and with published ranges for such sediments (cf. Samouëlian et al., 2005). Till is visible in the walls of the meltwater corridor within which the esker resides and so it is likely that RU1 lies mainly below water table and records the eroded till boundary of the meltwater corridor (cf. Burke et al., 2012a).

RU2 is a discontinuous, 3-5 m thick, moderate to high resistivity unit (800-3500 Ωm), limited to esker ridges and containing RE-A to RE-K (Fig. 5.5A). Such high resistivity values likely record rounded pebble to cobble gravel similar to that observed in the limited esker exposures (Fig. 5.5C). It comprises the majority of this esker ridge segment (Fig. 5.5B). RE-A and RE-B cover much of the grid, are composed of sub-parallel reflections recording widespread vertical accretion (Fig. 5.7A). The lower bounding surfaces of RE-C and RE-D truncate deeper reflections and they are composed of reflections that dip upflow at 10-20° (Fig. 5.6A), consistent with headward accretion of gravel sheets (Table
5.1). These elements record a phase of upglacier landform growth in a northwestward direction (Fig. 5.7A) that is followed by downglacier landform growth of RE-E to RE-H (Fig. 5.6A). These downglacier stacked radar elements have internal characteristics consistent with both downflow (RE-E and RE-F) and vertical (RE-G and RE-H) accretion and lower bounding surfaces that truncate deeper reflections (e.g., 60-70 m, and 145-158 m, Line X2, grid G1, Fig. 5.6A). The truncation of lower radar elements suggests these elements were emplaced upon an erosional surface and their dipping reflections suggest they were deposited as prograding avalanche beds (Table 5.1), similar to slip-face accretion deposits described by Wooldridge and Hickin (2005). RE-I is a solitary radar element, which is composed of downflow dipping reflections that are onlapped onto a concave-up lower bounding surface that truncates deeper reflections (Fig. 5.6A). These characteristics are consistent with scour and fill (Fig. 5.7A, Table 5.1). The remaining radar elements that make up RU2 (RE-J and RE-K) are stacked in a downglacier direction to record further downglacier landform growth of vertically (RE-J) and downflow (RE-K) accreting elements.

RU3 is a 1-5 m thick, low to moderate resistivity (70-800 Ωm) unit that occurs largely within the ground-wave of overlapping GPR data along ridge crests, but thickens in inter-ridge depressions (Fig. 5.5A). It contains RE-L to RE-Q. RE-L to RE-P are located in inter-ridge depressions. They have a trough-shaped geometry and contain discontinuous and chaotic reflections (Fig. 5.6A). They are interpreted as slump deposits developed after primary esker ridge deposition because they are only observed in flow-perpendicular lines and on ridge flanks, have a low resistivity (RU3, Fig. 5.5), and internal reflections are discontinuous/chaotic (Table 5.1, c.f. Burke et al., 2012b). Though composed largely of reworked esker fill, the low resistivity of these elements likely reflects a change in drainage efficiency (and therefore higher water content) associated with compaction, decreased pore space (Samouëlian et al., 2005) and the possible addition of some fine-grained windblown sediment from the ridge surface during slump activity. The stratigraphically-highest radar element (RE-Q) is partially obscured within the ground- and air-waves of the GPR data, but massive fine sand observed at the surface of the esker ridge (Fig. 5.5E), and the low resistivity of RU 3 (Fig. 5.5A) suggest that this element likely records an aeolian veneer (cf. Lian and Huntley, 1999) capping the esker ridge (Table 5.1).
Offset reflections are identified throughout the grid regardless of line orientation, but are most common at locations in close proximity to land surface depressions and slumps (e.g., at ~80-100 m on Y1, Fig. 5.6A), dipping away from the ridge apex. Offset reflections likely record faulting (Fiore et al., 2002) due to post-depositional adjustment in response to ice meltout and loss of ice-support.

In summary, the sedimentary architecture of the Green Lake esker at grid G1 indicates that the southeastward flow of water deposited sheets of gravel and sand through a combination of vertical, headward and downflow accretion and minor phases of scour and fill (Fig. 5.7A). Initial widespread vertical landform growth (stages 1-2, Fig. 5.7A) was followed by more localized landform growth in an upglacier (stage 3, Fig. 5.7A) and downglacier (stages 4-8, Fig. 5.7A) direction. This evolution defines the development of composite macroforms and is consistent with the sedimentary architecture reported at similar ridge widenings elsewhere and attributed to flood flows in ice tunnels (Brennand 1994; Burke et al., 2008, 2010, 2012a and b). The fact that Green Lake esker at grid 1 also displays moderate faulting associated with the melting of buried ice and is located within a broad till-floored meltwater corridor with an adverse slope, suggest that it was deposited within an ice tunnel flowing full of water at or just above the ice-bed interface (a closed, near-basal englacial conduit, potentially connected upflow and downflow to a fully subglacial ice tunnel) (Table 5.2). The enclosed depression within the ridge widening (forming the multi-crested morphology) coincides with a gap in RE-A in line Y1 (Fig. 5.7A). Therefore, the enclosed depression is interpreted as an area of non-deposition caused by the presence of an ice-block around which flow bifurcated before re-merging downflow (i.e., near the vicinity of line Y2; Fig. 5.7A). The presence of an ice block here is readily explained by crevassing (cf. Gulley and Benn, 2007; Gulley, 2009; Gulley et al., 2009) in relatively thin ice as canal-wide flow through Green Lake meltwater corridor collapsed into a smaller, esker-forming ice tunnel (cf. Burke et al., 2012a and b). Such dynamic, flood-scale flow indicates the sudden release of a point-source water reservoir (e.g., ice-dammed lake) (Table 5.2). Post-depositional modification of the esker included slumping of ridge flanks and deposition of a thin, discontinuous aeolian veneer.
Grid G2

GPR profiles in grid G2 (Figs 6B, S3, S4, S5) record the sedimentary architecture of a flat-crested ridge within a multi-ridged segment of Green Lake esker (Fig. 5.4). Although 510 m of GPR line were collected, only part of lines X1 and Y1 are presented (Fig. 5.6B) because these are representative of the full grid (Figs S3-S5). The esker crest rises 6 m above surrounding lake level (~1044 m asl) and lake bathymetry is unknown. GPR lines imaged to 6 m depth, the signal being attenuated below the water table. Consequently, the earliest esker sedimentation may not be recorded in the geophysical data at grid 2. The deepest imaged esker sedimentation is recorded by RE-A to RE-E forming tabular elements composed of reflections consistent with vertical (RE-A, C, F and G), downflow (RE-E), and lateral (RE-B and D) accretion stacked in a downglacier direction and recording downglacier landform growth (stages 1-6, Figs 7B and S5). An unconformity separates these lower elements from those above, where similar vertically accreting elements (RE-H to RE-K, Figs 6B and S4) are stacked in a downflow direction (stages 7-9, Figs 7B and S5).

In summary, the sedimentary architecture of grid G2 is dominated by vertically-accreting radar elements (with minor downflow and lateral accretion, Figs 7B and S5). However, these elements are typically stacked in a downglacier direction indicating a dominance of downflow (southeast), rather than vertical, landform growth. This architecture within a flat-topped, low-sinuosity ridge is consistent with sedimentation in a braided outwash stream, confined within an ice-walled channel (Chapter 2) – either an ice tunnel at atmospheric pressure under thin ice (cf. Hooke 1984) or an unroofed, ice-walled canyon (cf. Burke et al., 2012a; Chapter 2). The flat-crest and gently descending path of the esker and its location within a broad meltwater corridor at grid 2, combined with a lack of pervasive faulting in flow-parallel radar lines confirm deposition on land rather than ice, and the likely evolution of the ice-walled channel from an antecedent subglacial ice tunnel (or near-basal englacial conduit) (Table 5.2). Meltwater supply would have been largely from the same sources as for Grid 1 (i.e., ice-dammed lake; Table 5.2), but may also have included contributions from nearby supraglacial melt draining into the ice-walled canyon.
Canoe Creek esker (type 1b)

Geomorphology

Canoe Creek esker is 5.8 km long (including gaps), up to 24 m-high (Network 30, Appendix H) and incorporates hectometre-scale long-profile undulations. Its path follows a relatively steep (~5°) descending slope downflow from an ice-marginal channel (one of a series of nested ice-marginal channels above the Canoe Creek basin) before crossing relatively flat topography and continuing into the pre-existing topographic low at the head of the Chasm meltwater corridor (Figs 1B and 8A; cf. Burke et al., 2012a). Though apparently continuous with meltwater channels both upflow and downflow, for most of its distance the esker itself is not located in a well-defined valley. The esker is a low sinuosity (1.05), dominantly single ridge upflow (0-2 km, Fig 8C) and multi-ridged downflow (2-6 km, Fig. 5.8B). Multi-ridged segments exhibit a primary ridge ~5 m higher than parallel, secondary ridges. The esker ridge has a generally high continuity except for a zone of post-depositional spillway incision (Fig. 5.8) associated with later ice-marginal lake drainage (Chapter 3); this results in an artificially low continuity value (85%). The majority of Canoe Creek esker is round-crested (85% of total length); a flat-crested segment (15% of total length) with steep flanks (~10°) occurs at ~1.6-2.3 km (Fig. 5.8). Other esker ridges north of the Canoe Creek ridge (Fig. 5.8A) were not considered tributaries of this esker because they are discontinuous with the main esker ridge and are of different size and ridge shape than the main Canoe Creek esker.

Sedimentary architecture

One 2D grid of radar data (grid C1, Figs 8A, 9, S6) was collected from a round-crested, multi-ridged segment in the downflow reach of Canoe Creek esker. This grid includes > 500 m of radar data, penetrating ~5-10 m below the ground surface (Fig. 5.9). Lines X1 and Y1 were surveyed across a gravel pit floor, imaging esker material at depth, whereas lines X2 and Y2-Y4 were surveyed across the primary landform surface. Given that signal attenuation is fairly consistent at depth below the surface it is most likely the result of system power rather than a change in material.

The lowest recorded radar elements within grid C1 are trough-shaped RE-A, B and D (Fig. 5.9), which contain sub-horizontal reflections characteristic of vertical accretion.
and consistent with vertical landform growth (stages 1-2, Fig. 5.10). Stacked on top of these elements are upflow dipping (5-30°) reflections within RE-C and RE-E, which record headwardly-accreting elements indicative of headward landform growth (stages 3-4, Fig. 5.10). Headward growth was followed by a return to vertical accretion elements (RE-F to H) containing largely sub-horizontal reflections with discrete elements arranged to indicate generally downflow landform growth (Stages 5-6, Fig. 5.10). Downflow accretion elements (RE-I to RE-L) containing parallel, downflow dipping reflections and downflow dipping lower bounding surfaces are present at the downflow edge of the vertical accreting elements (Stages 7-10, Fig. 5.10) and record downflow landform growth. Vertical accretion was re-established with deposition of trough-shaped elements RE-M to N (Fig. 5.9 and Stages 11-12, Fig. 5.10). Downflow and upflow of ridge crestline topographic highs, are short (5-20 m long), trough-shaped elements RE-O to RE-Y (Fig. 5.9). These elements have concave-up lower bounding surfaces that truncate underlying reflections, consistent with scour and fill (cf. cut and fill, Sambrook Smith et al., 2006) during the final stages of esker deposition (stages 13-19, Fig. 5.10). Scour and fill processes would be consistent with reduced accommodation space associated with waning flows. Visible in flow-perpendicular lines only are trough-shaped elements (RE-Z to RE-ZC) with curved, concave-up lower bounding surfaces indicative of post-depositional slumping (stages 20-21, Fig. 5.10). Offset reflections interpreted as faults are limited in flow parallel lines and are common within the slump units of cross-lines.

In summary, formation of the type 1B Canoe Creek esker occurred subglacially, based on the apparent dynamic feedback between esker deposition and channel enlargement (see below) and the largely intact nature of primary architecture (with minor faulting in radar cross-lines indicating some removal of ice support and associated minor slumping and faulting of ridge flanks (stages 20-21, Fig. 5.10, Fiore et al., 2002; Woodward et al., 2008)) (Table 5.2, cf. Price 1966). Elements of macroform development are present, including headward and downflow accretion, indicating high-magnitude discharges, likely associated with flood-scale flows and rapid delivery of meltwater from a point-source (Brennand, 1994; Burke et al., 2008). This indicates a dynamic depositional environment (cf. Brennand, 1994; Burke et al., 2008, 2010, 2012b) in which initial flow velocity and associated shear stress were varied because of tunnel morphology (e.g., constrictions), leading to initial deposition in vertically-accreting elements and headwardly accreting
elements, recording upflow landform growth (stages 1-4, Fig. 5.10). Thermomechanical melting of tunnel roof and walls resulted in downflow tunnel enlargement; the corresponding increase in accommodation space allowing vertical (stages 5-6, Fig. 5.10) and downflow (stages 7-11, Fig. 5.10) accretion within the esker ridge. With waning flow, the ability of thermomechanical melting to counter conduit in-filling (or creep-closure, though this is less likely given nearby flat-topped segments that indicate relatively thin-ice conditions) resulted in reduced accommodation space and late-stage scour and fill, especially around ridge topographic highs (stages 13-19, Fig. 5.10). The flood-scale discharges implicated by the sedimentary architecture require a relatively large amount of water to be available in a short period of time and suggest the conduit was connected to an ice-dammed meltwater reservoir. The short length of type 1b eskers in comparison to type 1a is likely associated with a position closer to the ice margin. Shorter transport distances would account for more limited sediment supply (restricting size) and would require smaller ice-dammed reservoirs to support drainage to the ice margin. Upflow connections to ice-marginal channels suggest that some meltwater delivery may also have been from subaerial or supraglacial sources (e.g., a dead ice- or moraine-dammed lake to the NW).

**Hooke Road esker**

**Geomorphology**

The Hooke Road esker network includes one trunk ridge and two tributary ridges (Fig. 5.11A). These ridges begin at the downflow ends of ice-marginal meltwater channels on the slopes of the Marble Range (Tipper 1971a) (Figs 1B, 11A; Network 18 Appendix H). They initially follow a northeast, downslope path, then take a ~90° turn to the southeast, following a path largely parallel to slope elevation contours. The trunk ridge is 6.7 km long (including gaps), undulatory in long-profile, largely continuous (98%, with one gap near its downflow end), and is one of the most sinuous of all eskers measured on the southern Fraser Plateau, with a trunk esker sinuosity of 1.38 (and a maximum segmental sinuosity of 1.86). The esker path is largely downslope (based on topography immediately adjacent to the esker ridge); minor, apparently upslope segments (5-10 m high) are within the overall height range of the esker itself (20-25 m). The trunk ridge is a single, sharp-crested ridge (5-35 m high) for 82% of its length (Fig. 5.11A). Exceptions include a short
(500 m) multi-ridged segment (Fig. 5.11A) with a small (150 x 400 m), kettled (minor depressions < 3 m deep), flat-crested widening ~2.5 km downflow from the esker head, and a multi-crested, hummocky widening (possibly a fan) at the esker terminus (Fig. 5.11A).

Composition and sedimentary architecture

An exposure within the upper 2-3 m of the sharp-crested ridge (~200 m upflow from grid H1, Fig. 5.11A) reveals planar-bedded, clast-supported, moderately to poorly-sorted, subrounded pebbles and cobbles in a coarse sand matrix. Rare lenses (0.5 m wide, 0.2 m high) of well-sorted coarse sand are present near the top of the exposure. Extensive normal and reverse faults extend through the exposure and planar beds dip to the NE and SW on esker flanks, recording post-depositional failure. At the land surface, along the esker crestline, pebbles and cobbles are commonly exposed.

One, 140 m long, 2D grid of GPR data (grid H1, Fig. 5.12) was collected along the primary sharp-crested ridge in a multi-ridged segment of Hooke Road esker (Fig. 5.11). It is dominated by tabular radar elements (RE-A to RE-I) composed of largely planar, sub-horizontal internal reflections. In flow-parallel lines, lower bounding surfaces are typically conformable to underlying reflections; occasionally they truncate lower radar elements (e.g., RE-D truncates RE-A from ~30-40 m in line X1, Fig. 5.12A). These elements are dominated by vertical accretion and are composed of sub-horizontal reflections in flow parallel lines. Subtle downflow stacking of RE-A to RE-I (visible from the relationships between lower bounding surfaces) suggests a tendency for downflow growth of the esker (stages 1-7, Fig. 5.12B). Apparent dipping of some reflections is linked to offset reflections (faults) and likely indicates post-depositional rotation of the element. Offset reflections, interpreted as faults, are common and cross element boundaries in both flow-parallel and flow-perpendicular lines, and increase in frequency near lows in ridge topography (e.g., 50-95 m, line X1, Fig. 5.12A). RE-J and RE-K are 10-20 m long, lenticular elements along the esker crestline with concave-up lower bounding surfaces that truncate underlying reflections. These elements record the final stage of esker deposition and are consistent with localized scour and fill (stage 8, Fig. 5.12B). RE-L to RE-N are lenticular elements along the esker flank, and are only visible in flow-perpendicular lines. Their curved, convex-up lower bounding surfaces, inclusion of common offset reflections that dip away
from the ridge crestline, and location on ridge flanks is consistent with post-depositional slumping (stage 9, Fig. 5.12B). In summary, sedimentation at grid H1 was dominated by vertical accretion with extensive post-depositional modification by faulting and slumping. Combined with the high sinuosity, sharp-crested morphology of the ridge, the apparent absence of composite macroforms, the absence of long-upslope sections that exceed esker thickness, lack of relationship between the ridge and a containing meltwater corridor, and its connection to ice-marginal channels, the trunk ridge was most likely deposited in a supraglacial channel (cf. Price, 1966 and high sinuosity values for supraglacial streams in Marston (1983)) or high (near ice surface)-englacial conduit (cf. Gulley and Benn, 2007; Gulley, 2009; Gulley et al., 2009) and later let down through melt-out of underlying ice (Table 5.2). The well-preserved pattern of vertical accretion post melt-out suggests ice was likely thin or melt was slow and uniform. The absence of composite macroforms and the presence of scour and fill along with the high frequency of flow parallel faulting indicates gradual deposition under open-channel flow in a braided stream environment on top of ice. Late stage scour was likely accomplished following a reduction in sediment supply while discharge remained relatively high. Meltwater and subaerial runoff from the slopes of the Marble Range could have sustained a consistently low discharge in support of gradual sedimentation (Table 5.2). The slumping, combined with the high amounts of faulting in both line orientations (increasing in frequency at topographic lows in ridge crestline), and slightly dipping beds in flow-perpendicular orientation indicates collapse following the removal of both lateral and buried ice support.

Discussion

**Toward a morphogenetic classification of eskers**

Previous studies have made use of esker morphology, specifically crest-type and longitudinal slope, to directly infer subglacial hydrological conditions (e.g., Shreve, 1985; Hooke, 1984). However, our data do not universally support previously proposed relationships. A multi-crested, single ridge widening (grid G1, Fig. 5.4A) in Green Lake esker is located on an adverse slope and contains sediments recording flow in a broad subglacial ice tunnel. Faults between ridge-crests indicate ice meltout and subsequent sediment collapse. This data supports Shreve’s (1985) hypothesis that multi-crested or
broad-crested ridges form on upslope gradients, though not his explanation of multi-crested ridge evolution in terms of flow slippage down the sides of an initial ridge. However, not all upslope ridge segments of eskers formed in subglacial ice tunnels are multi-crested widenings (e.g., 4.5-5 km, Green Lake esker, Fig. 5.4B; 8-10 km, Chasm esker, Burke et al., 2012b). In addition, a multi-crested widening is observed on a relatively flat slope in an esker formed in a supraglacial channel, and sharp-crested ridges are only observed in eskers formed in supraglacial channels (Hooke Road esker, Fig. 5.11A) on the southern Fraser Plateau.

Whereas Shreve’s (1985) hypothesis relating esker morphology to genesis does not explain our data, we propose modifications that do. By combining the morpho-sedimentary relations in esker ridges with their geomorphic context robust inferences of meltwater channel type and water source are possible. When geomorphic context is first used to constrain meltwater channel type, similarities in morpho-sedimentary relations and derived genetic interpretations suggest a morphogenetic classification for eskers on the southern Fraser Plateau (Figs 1, 3; Table 5.2; statistical methods (e.g., cluster analysis) were not applied due to inadequate sample sizes (study area restrictions) and difficulties incorporating categorical data). The general principles from this classification may be applicable to eskers elsewhere; some characteristics are related to local conditions (e.g., thin ice, presence of the Marble Range).

**Type 1 (subglacial) eskers**

Type 1 eskers (e.g., Green Lake, Chasm and Canoe Creek eskers) are mainly formed in subglacial ice tunnels or low (near ice-bed contact)-englacial conduits flowing full of water (closed conduit flow) (Table 5.2, Fig. 5.13). They are dominantly round-crested, relatively discontinuous single ridges with low sinuosity (similar to the mean sinuosity of 1.06 calculated for esker ridges across the Keewatin (Storrar et al. 2014)) set within valleys or broad meltwater corridors or located down flow from them (Fig. 5.3). Their association with valleys and meltwater corridors favours subglacial formation. They typically contain ridge-scale macroforms, and exhibit limited faulting in the flow-parallel direction. Thus, hectometre-scale long-profile undulations and widenings, and non-depositional gaps between ridge segments, mainly reflect the depositional dynamics in a non-uniform ice tunnel with adequate sediment supply (e.g., Brennand, 1994; Burke et al., 2012b; Burke
some gaps are the result of post-depositional erosion in the region (e.g., lake spillways, Fig. 5.8). Faulting and slumping are mainly associated with the removal of lateral ice support. Multi-ridged segments (>20% of ridge length, Fig. 5.3) enclose kettle holes (e.g., Chasm esker) and reflect exploitation of structural weaknesses in ice-tunnel walls where anabranched tunnels developed in areas of reduced accommodation space during high flow (Burke et al., 2012b). Occasional multi-crested segments (e.g., Green Lake esker) are associated with the meltout of buried ice blocks. Consequently, some deposition may have occurred within low-englacial conduits. Flat-crested reaches (8% of ridge length, Fig. 5.3) are dominated by vertical accretion, consistent with later open channel flow (Russell et al., 2001) and local unroofing of the ice tunnel under the thin ice conditions on the southern Fraser Plateau at the time of esker formation (cf. Burke et al., 2012b). Relatively low sinuosity (range: 1.01-1.5), likely related to deposition within low sinuosity subglacial channels (e.g., Burke et al., 2012a; Atkinson et. al., 2013), and rare faulting below esker crestlines (Figs 5.5 and 5.9; cf. Burke et al., 2012a, Fig. 5.11) suggest deposition directly on the glacier bed or over thin ice in a low-englacial position. Relatively low ridge continuity is the combined result of zones of non-deposition or erosion within subglacial conduits (Brennand, 2000) and significant post-depositional incision (e.g., lake drainage spillway incision associated with the Canoe Creek esker). Type 1 eskers are subdivided into 2 subtypes based on length, continuity and geomorphic context.

Type 1a (subglacial GLOF) eskers

Type 1a esker networks (Green Lake and Chasm eskers) exhibit single trunk ridges located in broad meltwater corridors terminating in large recessional bedrock cataracts (Burke et al., 2012a). Trunk ridges are >10 km long and lack segmental or terminal fans (Table 5.2). The meltwater corridors were formed by GLOFs, likely associated with supraglacial lake drainage based on their geomorphology, geomorphic context and the sedimentary composition and architecture of their fill (Burke et al., 2012a). The length of type 1a networks is attributed to the location of the ice-dammed lake and their formation in pre-existing meltwater corridors with a ready supply of esker-building sediment (Burke et al., 2012a). This interpretation is also consistent with the presence of ridge-scale macroforms in the eskers (Fig. 5.7; cf. Chasm esker, Burke et al. 2012b Fig. 5.12) recording the waning stage of the GLOF (e.g., Burke et al., 2010; Burke et al., 2012b). The apparent lack of segmental fans suggests esker sedimentation was
synchronous rather than time-transgressive (e.g., Brennand 2000). Low continuity is likely the combined result of areas of non-deposition due to ice-tunnel constriction associated with increased velocity and flow competence, as well as post-depositional incision.

**Type 1b (subglacial, undivided) eskers**

Type 1b eskers (e.g., Canoe Creek esker) are <6 km long and terminate in flat-crested widenings (likely terminal fans, though this could not be confirmed). Their short length and fan-like terminations indicate they likely formed close to the ice margin. The presence of composite macroforms (Fig. 5.9A) implies these eskers may have been formed by flood flows (e.g. GLOFs); however, they are not as well developed as those in type 1a eskers and the lack of well-defined meltwater corridors indicates that floods were likely of lower magnitude, or sediment supply was lower (Burke et al., 2015) than for type 1a eskers. Furthermore, the dominance of vertical accretion in late-stage sedimentary architecture implies less dynamic feedback than is recorded in type 1a eskers (though late stage scour and fill may indicate some interplay between accommodation space and esker sedimentation, or late reduction in sediment supply while discharge is still relatively high). This apparent lack of dynamism also supports lower magnitude flows (associated with smaller ice-dammed reservoirs), impacting sediment transport and supply (Burke et al., 2015). Greater ridge continuity than type 1a eskers may be attributed to shorter tunnel length resulting in less opportunity for tunnel constrictions to produce areas of higher competence and non-deposition, or location near a thin ice margin, resulting in less creep-closure restricting flow. Furthermore they may be susceptible to less post-depositional incision. For example, shorter ridges were less likely to act as significant dams exposed to post-glacial meltwater incision.

**Type 2 (supraglacial) eskers**

Type 2 eskers (e.g., Hooke Road esker) are mainly formed in high-englacial or supraglacial channels, fed by ice-marginal streams or supraglacial melt (Table 5.2, Fig. 5.13). They are relatively short, highly sinuous and continuous, mainly single ridges (Fig. 5.3) dominated by vertical accretion and post-depositional faulting (Table 5.2). Ridges are not associated with valleys and do not exhibit long-upslope sections that exceed esker thickness, indicating a supraglacial or subaerial origin. Ridges are dominantly sharp-crested, continuous and highly sinuous indicating derbis being let down to the bed from a
supraglacial channel (cf. Price, 1966 and high sinuosity values for supraglacial streams in Marston (1983)) or high (near ice surface)-englacial conduit (cf. Gulley and Benn, 2007; Gulley, 2009; Gulley et al., 2009). The dominating presence of planar beds and sedimentary architecture that supports open-channel flow in a braided stream environment indicates non-flood discharges. Extensive post-depositional modification evidenced by faulting in flow parallel and perpendicular directions, and ubiquitous slumping of ridge flanks indicates substantial melt-out of underlying ice (Table 5.2). The continuity of type 2 esker ridges may reflect relatively thin ice, and thus a limited height of let-down. Type 2 esker tributaries (< 2 km long) appear to represent a structurally-controlled rectangular drainage pattern more than a typical dendritic subglacial esker network (e.g. Röthlisberger, 1972). They are consistently downflow of ice-marginal channels, and take paths consistent with topographic and inferred ice slopes (e.g., Fig. 5.11A). The path of a supraglacial channel or high-englacial conduit with water flow under atmospheric pressure is theoretically controlled by ice-surface slope (Shreve, 1972), but in reality may be disrupted by glaciological structural weaknesses (e.g., crevassing; cf. Hooke, 1984; Gulley, 2009; Gulley et al. 2009). On the flanks of the Marble Range, where most of the type 2 eskers are found, the ice surface likely sloped down towards the southeast, because regional ice margin retreat was northwestward (Chapter 3). The rectangular drainage pattern of ice-marginal channels and esker networks is best explained by initial water flow through an ice-marginal channel parallel to the slopes of the Marble Range. Where this channel encountered an ice-marginal (lateral) crevasse, flow was diverted into this ice-walled channel, then a supraglacial channel or high englacial conduit that eventually forced water flow to the southeast following ice surface slope towards the southeast-most ice margin.
Figure 5.13. Three-dimensional scatterplot of measured esker variables, demonstrating esker groups based on morphology and geomorphic context. Long round-crested ridges have generally lower continuity and sinuosity, likely representing deposition during the waning stages of catastrophic drainage events (type 1A). Shorter round-crested ridges exhibit tight groupings around 100% continuity, especially for shorter ridges and represent smaller (catastrophic?) drainage events closer to the ice margin (type 1B). Sharp-crested ridges are of the highest sinuosity indicating supraglacial or high-englacial deposition (type 2). They are highly continuous, perhaps reflecting a significant sediment supply and deposition over thin ice (see text for further comment). Flat-crested ridges are high-
continuity, moderately long and low sinuosity, representing deposition in subaerial, ice-walled channel environments (type 3).

**Type 3 (ice-walled canyon) eskers**

Type 3 eskers (e.g., Young lake esker-like ridge, chapter 2; Fig. 5.1B) are formed in ice-walled channels or re-entrants where water supply is likely contributed from subglacial, supraglacial and ice-marginal sources (Table 5.2, Fig. 5.13). They are found in deeply-incised valleys and are surrounded by enclosed depressions (kettle holes) recording the melting of detached ice blocks (ridge flanks contain significant faulting and slumping). The ridges are relatively short (2-6.5 km long), largely flat-crested (84%) and highly continuous, with low trunk esker sinuosity (~1.10) that mimics that of the valley. They terminate in flat to low gradient segments resembling fans or deltas (Chapter 2). Sedimentary architecture includes largely tabular, vertically-accreting beds (with minor components of downflow accretion and scour and fill) recording open-channel deposition. In the study area, flat-crested segments at the ridge terminus are commonly found in association with fine-grained glaciolacustrine sediment (e.g., Young Lake esker-like ridge, Chapter 2) suggesting they formed where channelized meltwater flows encountered proglacial lakes. In the case of the Young Lake esker-like ridge, the downflow end is interpreted as an esker deposited within an unroofed, ice-walled canyon terminating in a delta, and the upflow end is characterized by zones of extended hummocky and flat deposits which surround enclosed depressions, classified as kames deposited overtop and adjacent to stagnating ice (Chapter 2). Therefore the Young Lake esker-like ridge formed in both a supraglacial, and partially ice-walled channel. It is likely that other type 3 eskers formed in similar situations based on their position within topographic basins where meltwater from the surrounding plateau would have flowed overtop of downwasting, stagnant ice into the proglacial environment.

**Implications for esker mapping**

**Controls on esker distribution**

Esker formation is primarily dependent on deposition of sediment by flowing water within an ice-walled channel (Bannerjee and McDonald, 1975). Previous work has suggested that esker distribution is therefore dependent on the spacing of subglacial ice tunnels, which in turn is dependent on water pressure and transmissivity within subglacial
sediment (e.g., Boulton et al., 2009) or the pre-channelized distributed meltwater system (e.g., Hewitt, 2011). These models make two assumptions: 1) that eskers form in subglacial ice tunnels; and 2) that eskers form over multiple seasons from perennially redeveloped R-channel networks supplied largely by water derived from the glacier base (though Hewitt (2011) does include a term for englacially and supraglacially derived water). On the southern Fraser Plateau, it has been shown that although some esker forming channel systems were subglacial; others were supraglacial or ice-walled canyons within dead ice. Transects across the study area aligned roughly perpendicular to esker path and parallel to the reconstructed ice margin (cf. chapter 2) indicate a mean spacing of 20 km for all esker trunks mapped in the study area (Fig. 5.1). This is comparable to calculated and experimentally-derived esker spacings (Boulton et al., 2009; Hewitt, 2011), which are based on the assumption that subglacial ice tunnels are charged by perennial groundwater flow. However, esker formation in a combination of supraglacial channels (type 2) and ice-walled canyons (type 3) as well as subglacial ice tunnels is not consistent with the assumptions of these models. Furthermore, some subglacially-formed eskers were likely deposited from GLOFs (type 1) rather than perennial meltwater flows. Even if all type 1b eskers are presumed to have formed from steady-state flows (i.e., potentially formed in perennially redeveloped ice tunnels) they still maintain an average spacing of 52 km, a figure much larger than that predicted by existing models of channelized hydrology (cf. Hewitt, 2011).

Type 1a eskers (e.g., Green Lake, and Chasm eskers) likely formed within meltwater channels associated with GLOF’s from supraglacial lake sources (Burke et al., 2012a, b) (Table 5.2). It is unlikely that the volumes of water available through groundwater sources alone could be responsible for the discharges necessary to form the ridge-scale macroforms observed in these eskers. The linkage between some type 2 eskers and ice-marginal drainage networks on the Marble Range indicate that water supply for these esker networks was subaerial (i.e. direct runoff from the Marble Range, precipitation, snowmelt, melting of dead ice, or drainage from moraine-dammed lakes) or from direct melt of the glacier margin (Table 5.2). Water supply for type 3 eskers could have been from melt on, in, or under dead glacial ice, off of the active ice front or subaerial sources, and likely included all of these (Table 5.2). However, the presence of these ice-walled canyons in topographic basins and deeply-incised valleys, in and around stagnating ice.
(Chapter 2) suggests that active ice processes may have been some distance away and subaerial water supply was likely dominant. The existence of supraglacial channels, GLOF-supplied subglacial ice tunnels and the contributions of meltwater sources beyond those subglacially-generated indicate that many of the eskers on the southern Fraser Plateau were formed in channels carrying water generated outside the subglacial environment. Although in other situations, sediment supply may be an issue in esker formation, and consequently esker spacing (cf. Brennand, 2000; Burke et al., 2015), the presence of a near ubiquitous surficial coverage of clast-rich till and outwash (Lian, 1997; Plouffe et al., 2011) on the southern Fraser Plateau would indicate this was not a major controlling factor in esker formation here.

**Implications for ice sheet modelling**

Eskers on the southern Fraser Plateau record a variety of water sources and ice-walled channel types; they may not be ubiquitously interpreted as the casts of R-channels under the CIS. Therefore, numerical models used in the reconstruction of CIS retreat must accommodate multiple water sources to deliver water to the glacier hydrologic system (e.g., Hooke and Fastook, 2007; Hewitt et al., 2012) and not be concerned solely with water generated at the base of the ice sheet. This paper proposes a morphogenetic classification of eskers built on an analysis of morpho-sedimentary relationships and geomorphic context, which, if verified by further testing, will allow maps classifying esker genesis to be produced. Such maps will improve the use of eskers as inputs into, or verification of, numerical ice sheet models including channelized drainage.

**Conclusions**

As eskers formed in supraglacial channels, englacial conduits, and ice-walled channels as well as subglacial ice tunnels, the availability of basal meltwater only reflects part of the hydrological system responsible for esker formation on the southern Fraser Plateau. Therefore esker spacing is likely not an accurate proxy for groundwater transmissivity or geothermal heat-flux (though downflow routing of meltwater, once in a subglacial position, may still be controlled by the low hydraulic potential provided by pre-existing conduits, cf. Burke et al. 2012a) for the study area. The evaluation of ridge sedimentary architecture plays a critical role in the classification of esker morphogenetic
types. It is clear that local conditions (e.g., local topography, potential for point source water inputs, sediment supply) are also important controls on ridge location and morphology. Therefore any study invoking esker morphology to interpret esker forming environments and infer characteristics of glacier hydrology needs first to establish the general limits of esker forming environments, based on integrated morpho-sedimentary investigation of esker types (Fig. 5.13). Further testing of this morphogenetic classification on larger datasets and the successful application of statistical clustering methods are necessary to objectively evaluate the proposed classification; however, should this morphogenetic classification prove robust, it will invaluably enhance the efficiency of remote-predictive mapping (reducing the time and expense of fieldwork) and will improve the use of eskers as verification for numerical ice sheet models that include channelized meltwater flow.

References


6. Conclusions
Through refining our understanding of deglacial pattern and style of the last CIS across the southern Fraser Plateau in south-central British Columbia this thesis has tested the hypothesis that the lateglacial landform and sediment record supports the large-scale regional stagnation of the CIS over south-central BC (cf. Fulton 1991). Detailed reconstruction of lateglacial ice-marginal lakes, and glaciotectonic and glaciofluvial ridges indicate a pattern of retreat from southeast to northwest and a style of deglaciation dominated by active ice-marginal retreat with localized ice stagnation. Below each research objective and its related conclusions are reviewed.

**Detailed mapping of lateglacial landforms**

A new 1:175 000 scale (1:40 000 mapping scale) map of lateglacial landforms (eskers, moraines, fans and deltas, channels) and paleolakes was produced (Appendix A and associated digital files in the attached digital repository (Appendix J)). Hundreds of newly mapped glaciotectonic ridges on the plateau refute previous indications that moraines are absent in the study area. Their orientation indicates that the last CIS ice margin retreated from southeast to northwest. This direction of retreat is confirmed by the pattern of ice-marginal lake evolution in areas where the concentration of glaciotectonic ridges is lower. Morphosedimentary relationships indicate that fields of non-overridden glaciotectonic ridges formed under active ice conditions at or near the retreating ice margin. Detailed mapping reveals esker networks formed in multiple environments on the southern Fraser Plateau, including on the lower slopes of the nearby Marble Range, across the low relief plateau proper, and in valley systems deeply incised into the plateau surface. Morphosedimentary classification of esker networks demonstrates varied water sources contributed to glacier hydrology, including ice-dammed lake drainage, water delivered from subaerial sources, and direct glacier melt. Ice-dammed lake drainage formed long esker networks in broad meltwater corridors that stretch >30 km across the plateau indicating large areas of the plateau were deglaciating at the same time and that the equilibrium line was high. The presence of ice-dammed lakes (likely supraglacial) indicate a low ice-surface slope, and subaerially-formed (flat-topped) esker segments confirm that the ice was relatively thin at the time of their formation.
Ice-marginal lake reconstruction and implications for deglacial pattern and style

Ice-marginal lake systems provide the opportunity for reconstructing ice margin position based on the elevation and position of the lake surface. A total of eight glacial lakes (with multiple stages) and four non-ice contact paleolakes were identified within four interconnected paleolake systems on the southern Fraser Plateau (chapters 2 and 3). Lake geography and extent was reconstructed through the identification of lake level indicators and indicators of ice margin position (e.g., gilbert-type deltas, paleo spillways, ice-marginal channels, lake-bottom sediments, subaqueous fans, and grounding-line moraines). These lakes, developed on the relatively flat plateau surface, were generally shallow, but extensive. Based on limited sediment accumulations they were likely short-lived.

Within each lake system, lakes with multiple stages are identified by their dominant spillways. Spillways were operational starting at high elevation and progressing to lower elevations as the ice margin retreated and lake level gradually lowered. The spatial pattern of lake evolution within each lake system clearly demonstrates a southeast to northwest development of lake extent. Furthermore, inter-basin relationships, including lake system elevation and cross-cutting lake drainage networks reinforce the general northwestward pattern of lake evolution. The evolving position of the ice margin, as it formed an effective dam for these lakes, was located on the west-northwest side of the lakes as they developed (chapter 3). This general pattern is in agreement with ice retreat in the Young Lake basin (chapter 2) and inferences from glacioisostatic tilt from the Thompson (Johnsen and Brennand, 2004) and Nicola (Fulton and Walcott 1975) basins to the south.

Attempted reconstruction of glacioisostatic tilt from the plateau paleolakes was only possible in one lake system where primary water plane indicators (e.g., gilbert style deltas) were available for analysis. However, in this lake system the primary water plane indicators were linearly distributed, preventing reliable reconstruction of a tilted planar surface. The application of regionally reconstructed tilt values (e.g., Johnsen and Brennand 2004, Fulton and Walcott 1975) produced lake surfaces which would have been unlikely given spillway elevations and ice-dam locations. Therefore zero tilt was assumed in the resultant modelling of the glacial lake systems on the plateau. Zero tilt has been
previously inferred for lake systems reconstructed in the Fraser Basin immediately west of the study area (Huntley and Broster 1994) and may be appropriate here if most glacioisostatic rebound was complete prior to lake formation. This would indicate relatively thin ice conditions during lateglacial lake formation on the plateau, and that basin lake systems to the south with relatively high modelled rebounds (e.g., in the Thompson and Nicola systems) formed earlier than plateau systems, again reinforcing a regional northwestward retreating ice margin.

Regional reconstructions of moraine and esker genesis: implications for deglacial pattern and style

A total of 432 individual moraine ridges with a cumulative crest-line length of just under 500 km were mapped on the southern Fraser Plateau. Three types and four subtypes of moraine ridge were classified based on morphosedimentary relationships and geomorphic context. Type 1a, discrete glaciotectonic moraines (overridden), are long, broad-crested, flat-topped ridges aligned roughly transverse to putative ice advance onto the southern Fraser Plateau (Plouffe et al., 2011) and containing evidence of thrusting and stacking of stratigraphic units. Type 1b, discrete glaciotectonic moraines (recessional), are short to long, sharp or round-crested sinuous ridges containing evidence of internal thrusting and stacking of stratigraphic units. Type 1c, glaciotectonic ribbed terrain (subglacial), are fields of long, broad-crested ridges overprinted by streamlined forms and incised by subglacial channels. Ridge forms are interrupted by enclosed depressions. The ridge-depression sequence resembles a hill-hole pair and the stacked internal stratigraphy implies a glaciotectonic origin. Type 1d, glaciotectonic moraine fields (recessional), are fields of short to long, sharp- or round-crested sinuous ridges containing evidence of shearing and stacking of diamicton with intervening units of subglacial or proglacial outwash indicating episodic ice-margin fluctuations during deglaciation. Type 2, grounding line moraines (recessional), are short, convex-downflow, sharp- to round-crested, kettled ridges, normally located in valley-bottom locations. They contain moderately to well-sorted gravel and lacustrine sediments with minor amounts of diamicton. Type 3, crevasse-fill ridges, are low-relief, round-crested ridges developed in a rectilinear pattern. They contain stratified, moderately-sorted gravel indicating these are likely crevasse-controlled eskers formed at a decaying ice margin or dead, detached ice mass.
Relative timing of moraine ridge formation and the alignment of these ridges with respect to the ice margin allows reconstruction of ice-flow directions over time. During ice advance, glaciotectonic (type 1a) moraines were formed or modified and subsequently overridden by early ice advance. These moraines indicate that ice advanced into the study area from the northeast, flowing towards the southwest. Subglacially-formed glaciotectonic ribbed terrain (type 1b) is composed of ridges largely oriented eastnortheast-westsouthwest indicating a shift in ice-flow direction towards the southeast. Discrete, sharp-crested moraines (type 1c) and moraine fields (type 1d) have not been overridden and therefore likely formed during ice retreat. They are generally oriented northeast-southwest indicating ice retreat towards the northwest (though some type 1c moraines reflect more random orientation, likely associated with local, topographically controlled ice flow. Individual grounding line moraines (type 2) are restricted to basins where ice-marginal lakes were formed. Their orientation is affected by the orientation of the basin at the location of formation. Crevasse-fill ridges (type 3) are more randomly oriented and likely reflect formation under a decaying ice margin heavily influenced by local topography or a dead, detached ice mass.

Implications for lateglacial ice-sheet pattern and style drawn from moraine and ribbed terrain genesis relate to subglacial and ice-marginal conditions necessary for ridge formation. The formation of subglacial ribbed terrain (type 1b) requires a shift in thermal regime to explain its hill-hole-like pair formation (subglacial freeze-on, thrusting and stacking), and later streamlining and incision by subglacial meltwater channels. This is consistent with previous interpretations of a transition from cold-based to warm-based ice on the plateau as the ice sheet thickened (Lian and Hicock 2000). Recessional glaciotectonic moraines (type 1d) are sharp-crested indicating they have not been overridden; however morphosedimentary architecture (including compressive fabrics and shears) indicates glaciotectonic shear was responsible for ridge building. The concentration of type 1d moraines on and around the plateau surrounding the largest ice-marginal lake, glacial Lake Dog, is conspicuous and may be related to the potential for flotation of the ice-margin and a redistribution of subglacial stress to local topographic highs. This pattern implies active ice-margin retreat over a >600 km² section in the northwest of the study area. The presence of an active ice margin is consistent with computed glacioisostatic tilts to the northwest (Fulton and Walcott 1975, Johnsen and
Brennand 2004) which suggest the presence of an ice divide and ice-surface slope rising to the northwest of the study area. Indeed, grounding-line (type 2) moraines across shallow plateau valleys confirm the presence of ice-dammed lakes and a relatively orderly sequence of lake drainage and evolution following ice retreat to the northwest across the Plateau. In addition, this information is significant for drift prospecting wherein lateglacial ice flow directions can explain mineral anomalies and facilitate location of ore bodies. Localized stagnation, represented by the formation of crevasse-fill ridges (type 3), occurred in some areas of the plateau as deglaciation progressed.

On the southern Fraser Plateau over 124 km of esker crestline associated with primary esker segments were mapped. A total of 34 individual esker trunk ridges were identified with a compiled average WNW-ESE linear trend. Based on morphosedimentary relationships and geomorphic context three types (and two subtypes) of esker trunk ridge were classified. Type 1a, long (>10 km) subglacial GLOF eskers are dominantly round-crested, relatively discontinuous single ridges with low sinuosity set within valleys or broad meltwater corridors. They typically contain ridge-scale macroforms and infrequent flow-parallel faults. Type 1b, short (<6 km) subglacial GLOF, outburst flood, or seasonal eskers are similar to type 1a but are shorter and with higher continuity. They may contain ridge-scale macroforms and are typically dominated by late-stage vertical accretion. Type 2, supraglacial or high-englacial eskers are typically short, highly sinuous and continuous, single ridges. They are not usually located within valleys, although they may occur downflow from them, and do not consistently exhibit upslope segments where slope exceeds esker thickness. Their sedimentary architecture is dominated by vertical accretion marked by common post-depositional faulting. Type 3, ice-walled open channel eskers are flat-topped, largely flat-crested, highly continuous and low sinuosity ridges found in deeply-incised valleys and surrounded by kettle holes. They terminate in low-gradient segments resembling fans or deltas. Tabular, vertically-accreted beds dominate the sedimentary architecture though minor downflow accretion and scour and fill may also be present.

Esker position with respect to the ice surface and margin and morphosedimentary relationships suggest different water sources were responsible for the formation of different esker types. Water sources for type 1a and 1b subglacial eskers were likely point sources such as ice- (marginal, supraglacial or subglacial) or moraine-dammed lakes. For
supraglacial or high englacial (type 2) eskers formed close to the ice margin and ice-walled open channel (type 3) eskers a mix of water sources including subaerial runoff (precipitation, snow and ice melt), glacier meltwater, and ice-dammed or moraine-dammed lakes likely supplied their respective conduits. Sediment supply likely also varied according to esker formational position. Type 1a and b eskers formed subglacially and therefore formative meltwater flows eroded and transported subglacial sediment (till or the gravel fill within meltwater corridors (Burke et al., 2012)). Supraglacial or high englacial (type 2) and ice-walled open channel (type 3) eskers were likely formed largely of ice-marginal, supraglacial or englacial sediment or recently exposed subglacial drift.

Previous work has attempted to link esker distribution to subglacial ice-tunnel location (spacing) and groundwater transmissivity (Boulton et al., 2009) or the pre-channelized (distributed) meltwater system (e.g., Hewitt, 2011). On the southern Fraser Plateau eskers formed in three different positions with respect to the ice surface. As a result, non-subglacial eskers were eliminated from computation of subglacial esker spacing. The average spacing of subglacial eskers is ~52 km, a figure much larger than model predictions for subglacial ice tunnels. Furthermore, this figure includes subglacial eskers that likely formed during high magnitude, low frequency ice-dammed lake drainage events, rather than more steady state basally-generated meltwater flows derived from groundwater seepage to perennially operating R-channels. These conclusions indicate that either esker spacing is likely not a good proxy for ice-tunnel spacing in this environment or mechanisms for the formation of subglacial eskers include those beyond basally-generated meltwater alone. This suggests future models of subglacial channelized hydrology for environments similar to the southern Fraser Plateau should incorporate contributions from non-steady state sources (e.g., ice-dammed lakes). They should also be flexible enough to include eskers formed in multiple positions with respect to the ice surface, including supraglacial/high-englacial positions and ice-walled channel environments.

The general WNW-ESE trend of esker ridges, combined with individual flow directions derived from the sedimentary architecture of esker ridges indicates that a hydraulic low likely existed to the ESE of most esker ridges during their formation. This places a decaying ice margin progressing from the ESE to the WNW of the study area, a pattern that aligns with ice retreat suggested by ice-marginal lake evolution (chapter 3).
and moraines (chapter 4). Supraglacial eskers downflow from ice-marginal channels on the Marble Range developed from WNW to ESE indicating the general ice-surface slope towards the ice margin. The nested pattern of these eskers and channels argues for northwestward retreat across the study area.

The picture that emerges of the last CIS retreating across the southern Fraser Plateau is one of thin ice with a low ice surface slope, supporting the formation of supraglacial lakes, GLOF eskers and ice-marginal lake systems. However, ice-surface slope must have been high enough to allow active retreat in order to form the fields of recessional glaciotectonic ridges on the higher margins of the Plateau. The length of GLOF eskers places supraglacial lake formation 20-40 km from the ice margin, indicating much of the Plateau was below the equilibrium line altitude at the time of their formation, consistent with the delivery of subaerial water off the slopes of the Marble Range supplying type 2 esker systems and ice-marginal channels. This picture of deglaciation is not all that different to the decaying Greenland Ice Sheet today, where low ice surface profiles and seasonally high equilibrium line altitudes result in the development of high numbers of supraglacial lakes, some of which are linked to the ice-margin by efficient channelized drainage networks (Hoffman et al. 2011).

Future research

As research progressed for this project, new significant avenues of research potential became apparent: Chapter 2, 3 and 4 indicate the relative timing of deglaciation as it progressed across the Plateau. Absolute chronologies suggest deglaciation of south-central British Columbia was rapid (Clague and James 2002), but these chronologies are based on a small number of minimum ages for deglaciation (largely radiocarbon ages from basal peat sources) and the area is noted for its paucity of organic material in lateglacial sediments (Lian and Hicock 2001). Future work should apply novel dating techniques (optically stimulated luminescence, terrestrial cosmogenic nuclide) to expand the inventory of deglacial ages for this region, and establish the rate at which deglaciation occurred. Chapter 4 outlined the morphosedimentary classification of moraine systems on the Plateau. The identification and classification of these ridges advances our understanding of deglacial pattern across the plateau (i.e., the absence of moraines may
no longer be used to support arguments for regional stagnation); however, this classification scheme needs to be tested on larger sample sizes and with greater investigation into sedimentary architecture to lower the probability that certain moraine types may have been missed or mis-classified. Detailed work into ribbed terrain elsewhere in the interior of British Columbia is necessary to further understanding of subglacial processes operating under the last CIS. Chapter 5 detailed esker morpho-sedimentary architecture and distribution for the southern Fraser Plateau, indicating that studies invoking esker morphology to interpret esker forming environments and infer characteristics of glacier hydrology have the potential for success at the local scale. Further testing of the morphogenetic classification system put forward in chapter 5 is needed, on larger datasets in order for more objective evaluation of its broad suitability. The application of statistical clustering methods on the classification system is something that should be worked into any larger scale study in order to further establish statistical boundaries on esker classification. Further work on the presence of subglacial lakes or potential for supraglacial lake development on the southern Fraser Plateau and environs is critical for establishing deglacial water sources. The results of chapter 4 could be extended in a similar fashion as further testing could produce broadly applicable morphogenetic classifications for moraine systems, though more quantitative evaluation of moraine morphology would likely be necessary.
References


Appendices
Appendix A.

Late-glacial landforms and paleolakes of the southern Fraser Plateau, British Columbia, Canada

The map and associated documentation in this appendix represents the mapping and classification process and graphical output for late-glacial landforms and paleolakes completed for this thesis. The map was designed to be printed on a 36’ x 48’ sheet at a scale of 1:175 000. A page-sized thumbnail of the map is included in appendix A.1. The full-size map is split over multiple pages in appendix A.2 for printing as multiple 8.5” x 11” pages. The mapping process, parameters and cartographic specifications are recorded in appendix A.3. Digital data attached to this thesis includes a high-quality .pdf version of the full-sized map and a geodatabase containing individual featureclasses for mapped landforms.
A.1 Thumbnail image of landform classification map.
A.2 Full-size page version of lateglacial landform map.
and paleolakes of the sou...
Southern Fraser Plateau, Brit

Moraine types

- Type 1a - Discrete glaciotectonic moraine (overridden)
- Type 1b - Discrete glaciotectonic moraine (recessional)
- Type 1c - Glaciotectonic moraine field (subglacial ribbed terrain)
- Type 1d - Glaciotectonic moraine field (recessional)
- Type 2 - Grounding-line moraine
- Type 3 - Crevasse fill ridges
<table>
<thead>
<tr>
<th>ker types</th>
<th>Channels</th>
</tr>
</thead>
<tbody>
<tr>
<td>1a - Subglacial long</td>
<td>[Proglacial (small)]</td>
</tr>
<tr>
<td>1b - Subglacial short</td>
<td>[Proglacial (large)]</td>
</tr>
<tr>
<td>2 - Supraglacial/high-englacial</td>
<td>[Subglacial (small)]</td>
</tr>
<tr>
<td>3 - Ice-walled channel</td>
<td>[Subglacial (large)]</td>
</tr>
<tr>
<td></td>
<td>[Ice-marginal (small)]</td>
</tr>
<tr>
<td>[fluvial deposits]</td>
<td>[Ice-marginal (large)]</td>
</tr>
</tbody>
</table>

*Spillways*
eldwork
observation
A.3 Landform classification and cartographic process.

Abstract

Retreat of the last Cordilleran Ice Sheet across the southern Fraser Plateau in south-central British Columbia, is recorded by meltwater channels, eskers, moraines, deltas and fans, which together with the distribution of lake-bottom sediments facilitate reconstruction of ice-marginal paleolakes and ice margins. Landform mapping (working scale ≤ 1:40 000, presentation scale 1:175 000) and classification across the 7842 km² study area is based on morpho-sedimentary relationships and genetic interpretation, completed through detailed aerial photograph and digital elevation model analysis, combined with field observations including sedimentology and selective application of shallow geophysics. We advance a new subclassification of lateglacial landforms on the southern Fraser Plateau that facilitates elucidation of the pattern and style of ice decay across the region and provides new insights into the dynamics of the waning ice sheet in the vicinity of a putative ice divide.
Introduction

The southern Fraser Plateau is a broad, moderately high (1200-1500 m asl), low relief surface set in the intermontane region of British Columbia, between the Coast Mountains in the west and the Columbia Mountains in the east (Fig. A.1, Holland 1976). Underlying the relatively subdued topography are successive basalt flows originating from local volcanic centres active in the late Eocene and Miocene (Bevier, 1983; Andrews and Russell, 2008). Post-depositional warping and uplift of the plateau surface established regional drainage patterns (Mathews, 1989) which are still imprinted on the landscape today. They formed antecedent drainage routes for glacial rivers and natural basins for ice-marginal paleolakes. There is evidence that the southern Fraser Plateau was occupied by the Cordilleran Ice Sheet several times during the Pleistocene (Mathews and Rouse, 1986; Lian et al., 1999). It generally thickened and advanced from alpine snowfields in the Coast and Columbia ranges, until the ice masses coalesced somewhere over the Fraser Plateau, forming an ice divide (cf. Wilson et al., 1958) and forcing ice-flow north and south (Tipper, 1971a; Clague and James, 2002). It is not clear from the geomorphic or stratigraphic record how many times a true ice sheet stage was reached.

Glacial ice last occupied the southern Fraser Plateau from ~20-10 cal. ka BP (Clague and James, 2002; Chapter 1), waning and retreating in response to rapid climate amelioration (Clague and Ward, 2011). Interpretations of deglaciation have favoured regional stagnation, based on an absence of recessional moraines and presence of kettled topography (Fulton, 1967, 1976). A rapid rise in equilibrium line altitude has been implicated to explain top-down melting of the interior ice-sheet (Fulton 1991). However, the hypothesis of regional stagnation stands in opposition to 1) glacioisostatic evidence that suggests an ice divide somewhere over the Fraser Plateau (Fulton and Walcott, 1975; Johnson and Brennand, 2004), 2) the progressive northwestward evolution of ice-marginal lakes on the Plateau that suggests northwestward retreat of the ice margin (Chapter 2 and 3) evidence of plateau-ice coeval with dwindling valley-bottom ice (Johnsen and Brennand, 2006), and 4) patterns of ice-marginal channels that suggest retreat back towards the Coast Mountains (Margold et
Figure A.1  A) Overview map overlain on multi-directional hillshaded DEM (ESRI 2015) showing area covered by detailed map of the southern Fraser Plateau. Ice advanced into the study area from the Coast Mountains to the West and the Columbia Mountains to the East. Location of putative ice divide (pink dashed line, Wilson et al. 1958) and non-synchronous extent of the Cordilleran Ice Sheet at local last glacial maximum (thick gray line, Dyke et al. 2004) shown. B) Physiography of the study area (Holland 1976) with ice margin positions reconstructed from late-glacial ice-marginal lake boundaries (Chapter 3).
al., 2013). As deglaciation progressed lateglacial landforms were preserved, allowing reconstruction of ice sheet decay patterns and dynamics. Several authors have documented such landforms (e.g., Tipper, 1971b; Huntley 1995; Bednarski, 2009; Plouffe, 2009a, b; Huscroft, 2009; Margold et al., 2011) but their efforts have been limited in scale, geographic extent or detail of landform classification, restricting the ability to test the hypothesis of regional stagnation (cf. Chapter 2).

The motivation behind this map was to assemble a robust genetic landform database at the 1:40 000 scale in order to test the prevailing view of regional stagnation on the southern Fraser Plateau during decay of the last CIS. Regional stagnation is characterized by crevasse-fill ridges as well as an absence of large, linear glaciotectonic moraines. Active retreat is characterized by glaciotectonic moraines, and ordered, successive ice-dammed lake formation. These landforms and reconstructed environments are the focus of the map.

Methods

Data acquisition

Mapping was completed by heads-up digitizing of landforms onto orthophotographs (1 m resolution, 1:125 000 scale orthophoto mosaics, Province of British Columbia, 2010) using a combination of stereographic aerial photograph (1:10 000 to 1:40 000 scale) analyses, hillshaded DEM analysis (25 m horizontal grid cells, 1 m vertical resolution (90% of measurements within 5 m of true elevation), Geobase® hillshade with a 45° sun angle constructed into two hillshades: one illuminated from the northwest and one illuminated from the northeast in order to reduce directional bias), and field mapping (16 weeks of fieldwork). Field mapping included checking landforms for accuracy in classification (e.g., using morpho-sedimentary relationships and geomorphic context where possible), checking landform extents and identifying new landforms not captured from remote sensing observations. Oblique landscape geovisualization in Google Earth provided instructive assistance with the identification and mapping of some landforms. Landform classification was determined based on geomorphic and stratigraphic (where available) context, and morpho-sedimentary relationships of representative landforms elucidated by combining topographic surveying, sedimentology and shallow geophysics. Choices on representing landforms as points, lines or areas were made based on the
needs of later quantitative and qualitative analysis. For example, eskers were first mapped as lines, based on the crestline of the esker ridge, in order that esker spacing, orientation, continuity and sinuosity could be analyzed. Later esker planforms were mapped as areas, based on slope-break at ridge edges, and cross-sectional geometry was attributed in order for esker genesis to be more robustly inferred. For detailed information on the classification of specific landforms and environments, the reader is directed to the following publications: eskers (Burke et al., 2012b; Chapters 2 and 5), ice-marginal lake systems (Chapters 2 and 3), and moraines (Chapter 4).

**Mapped late-glacial landforms**

**Moraines**

Moraines form at or near the glacier margin by glaciotectonic stress, or as aprons of material accumulated by dumping of glacier-borne material and accumulation of glaciofluvial, and glaciolacustrine sediments (Bennett, 2001; Evans, 2007). They generally form ridges parallel or subparallel to the ice margin. They can provide information on ice margin position, ice-flow direction, and the dynamics of ice flow (Bennett, 2001; Evans, 2007).

Moraines were identified primarily based on the presence of a ridge identified from the hillshaded DEM and stereophoto pairs and were differentiated from eskers by their asymmetric cross-profiles and geomorphic context (e.g., perpendicular to meltwater channels, cross-valley orientation). Moraines may occur as discrete ridges of different shape (e.g., sharp-crested or broad-crested), scale and geomorphic context (e.g., in valleys or on topographic highs). Some were selected for detailed field investigation based on field accessibility, availability of exposures and opportunity for application of shallow geophysics – ground penetrating radar if composed of sand and gravel, or electrical resistivity tomography if composed of diamicton and/or assorted glacial drift. Observations of ridge sedimentology, structural geology and sedimentary architecture accompanied shallow geophysics at available exposures. Moraines were categorized into three types, with 4 subclasses of type 1, based on their context and morpho-sedimentary relationships (Chapter 4) (Table A.1).

**Table A.1**  
**Mapping results and landform classification**
<table>
<thead>
<tr>
<th>Landform Type</th>
<th>Type no.</th>
<th>Landform sub-type</th>
<th>N&lt;sup&gt;1&lt;/sup&gt;</th>
<th>Mean length (m)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>(overridden)</td>
<td>1a</td>
<td>Discrete glaciotectonic ridge</td>
<td>12</td>
<td>1570</td>
<td>Broad-crested (occasionally streamlined), single arcuate ridges. Thrusted and stacked internal structure.(^2)</td>
</tr>
<tr>
<td>(recessional)</td>
<td>1b</td>
<td>Discrete glaciotectonic ridge</td>
<td>20</td>
<td>645</td>
<td>Sharp, narrow-crested, single arcuate ridges. Thrusted and stacked internal structure.(^2)</td>
</tr>
<tr>
<td>(subglacial, ribbed terrain)</td>
<td>1c</td>
<td>Glaciotectonic moraine field</td>
<td>21</td>
<td>3510</td>
<td>Broad-crested (occasionally streamlined) arcuate nested ridges, occurring in fields. Thrusted and stacked internal structure.(^2)</td>
</tr>
<tr>
<td>Moraine</td>
<td>1d</td>
<td>Glaciotectonic moraine field (recessional)</td>
<td>317</td>
<td>1155</td>
<td>Sharp, narrow-crested, sinuous, nested ridges occurring in fields. Thrusted and stacked internal structure with sands and gravels deposited during ice margin oscillation.(^2)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Grounding-line moraine</td>
<td>11</td>
<td>280</td>
<td>Sharp, narrow-crested, single linear ridges normally with gentle ice-proximal and steep ice-distal slopes, located within paleo-lake basins. Moderately sorted sand and pebble to cobble gravel with minor pockets of laminated silt and fine-sand.(^3)</td>
</tr>
<tr>
<td></td>
<td>3</td>
<td>Crevasse-fill ridge</td>
<td>51</td>
<td>350</td>
<td>Low elevation, reticulate ridges. Moderately sorted, subrounded pebble to cobble gravel.(^4)</td>
</tr>
<tr>
<td>Meltwater channel</td>
<td>1</td>
<td>Subglacial</td>
<td>45</td>
<td>5610</td>
<td>Upslope grade at channel base, esker within channel. Perpendicular to reconstructed ice margin, follows topographic slope.</td>
</tr>
<tr>
<td></td>
<td>2</td>
<td>Proglacial</td>
<td>216</td>
<td>3325</td>
<td>Channel does not follow most direct topographic slope. May be nested with other channels.</td>
</tr>
<tr>
<td>Esker</td>
<td>1a</td>
<td>Subglacial (low-englacial) long</td>
<td>2</td>
<td>20295</td>
<td>Long, round-crested, relatively low continuity, low sinuosity dominantly single-ridges set within broad meltwater corridors with upslope sections.(^5) Contains ridge-scale macroforms and limited flow-parallel faulting. Flat-crested segments contain tabular sedimentary units.(^6)</td>
</tr>
<tr>
<td>Landform Type</td>
<td>Type no.</td>
<td>Landform sub-type</td>
<td>N</td>
<td>Mean length (m)</td>
<td>Description</td>
</tr>
<tr>
<td>---------------</td>
<td>---------</td>
<td>------------------</td>
<td>---</td>
<td>-----------------</td>
<td>-------------</td>
</tr>
</tbody>
</table>
| Esker         | 1b      | Subglacial short | 23 | 1900           | Short, round-crested, low-continuity, low sinuosity, ridges set within valley systems which may or may not have an upslope grade. Contains incipient macroforms, vertical accretion units, and late-stage scour and fill. 
Short, continuous, highly sinuous, sharp-crested ridges dominated by highly faulted, tabular sedimentary units. Sometimes associated with ice-marginal channels. |
<p>|               | 2       | Supraglacial (high-englacial) | 7 | 3290           | Short, continuous, low sinuosity ridges set within large valley systems without upslope grades. Depositional elements are dominantly vertical and downflow accretion with minor scour and fill. Terminate in kettled fan/delta deposits. |
|               | 3       | Ice-walled channel | 2 | 4295           | Short, continuous, low sinuosity ridges set within large valley systems without upslope grades. Depositional elements are dominantly vertical and downflow accretion with minor scour and fill. Terminate in kettled fan/delta deposits. |
| Delta (Gilbert-type) | 1 | 6 | N/A | Near flat-topped deposits at the termini of meltwater channels. Sedimentary architecture includes foresets and topsets. Normally surrounded by lower energy lacustrine sediments indicating deposition in a lake basin. |
| Lake maintenance spillway | 1 | 10 | 1705 | Bedrock or till-floored spillway maintaining a stillstand. |
| Ice-marginal paleolake landforms | 2 | Subaqueous fan | 7 | N/A | Deposits at the termini of meltwater channels with steeper slopes than the near flat-topped deltas. Foresets not observed in sedimentary exposures, but lower energy lacustrine sediments nearby indicate deposition in a lake basin. |
|               | 2       | Ice-marginal channels | 747 | 975           | Channel does not follow most direct topographic slope. May be nested with other channels. Channels formed in contact with ice and land and may grade to lake level. The elevation of the channel mouth-floor underestimates lake level. |
|               | 2       | GLOF spillway | 4 | 2365           | Ice or gravel based spillway formed by rapid glacial lake drainage. Base of spillway records a minimum lake elevation. Reconstructed top of spillway may record lake stage maximum. |</p>
<table>
<thead>
<tr>
<th>Landform Type</th>
<th>Type no.</th>
<th>Landform sub-type</th>
<th>N⁴</th>
<th>Mean length (m)</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ice-marginal paleolake landforms</td>
<td>2</td>
<td>Outburst flood spillway</td>
<td>4</td>
<td>1060</td>
<td>Gravel-based spillway rapidly breached by non-glacial lake. Base of spillway records a minimum lake elevation. Reconstructed top of spillway may record lake stage maximum.¹²,¹⁶</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

¹Number of discrete landforms mapped; ²Bennett, 2001; ³Benn, 2996; ⁴Sharp, 1985; ⁵Burke et al. 2012a; ⁶Burke et al. 2012b; ⁷Chapter 5; ⁸Chapter 2; ⁹Syverson and Mickelson, 1997; ¹⁰Price, 1966; ¹¹Lonne, 1995; ¹²Perkins and Brennand 2014; ¹³Kehew and Lord, 1986; ¹⁴Postma, 1990; ¹⁵Greenwood et al. 2007; ¹⁶Costa and Schuster 1988;

Meltwater Channels

Meltwater channels may be differentiated based on their position with respect to an ice sheet or glacier (Fountain and Walder, 1998). Supraglacial (Marston 1983), englacial (Gulley et al., 2009) and subglacial (Kehew et al., 2012) channels form at different vertical positions within the ice, but the position of former subglacial channels is observable where they have eroded the substrate (Table A.1). Proglacial channels lead away from the ice margin, whereas ice-marginal channels follow the ice-edge and are useful for indicating an ice margin position (Table A.1; Syverson and Mickelson, 2009). Nested ice-marginal channels may show the horizontal movement and general lowering of the ice margin through time. Water flow direction reconstructed from slope (for ice-marginal and proglacial channels) or estimates of hydraulic potential based on indications of ice-surface slope (in the case of subglacial channels) is represented by an arrow-head on the downflow end of the channel. Previous efforts in mapping meltwater channels on the southern Fraser Plateau (e.g., Tipper, 1971; Bednarski, 2009; Huscroft, 2009; Plouffe, 2009a, b; Margold et al., 2011) either generalized results based on the necessity of representing channels at a small scale, or do not include subglacial channels as part of their mandate.

Supraglacial and englacial channel systems are not mapped in this project, as their geomorphic signature may be equivocal or non-existent; however, their presence is invoked to explain lake-drainage events, where no other spillway type is present, or where eskers mark their former position (Perkins and Brennand, 2014). Subglacial channels were identified by convex long profiles or association with subglacial landforms (e.g.,
subglacial eskers). It is likely that their distribution is under-represented as many subglacial channels were probably re-occupied by proglacial channels during ice retreat, and thus mapped as proglacial channels. Proglacial channels were probably ubiquitous across the mapped area, but are difficult to distinguish where they intersect with modern channels. Hence, proglacial channels are conservatively recorded on the map where an extant channel does not currently support a stream, or because the current stream is largely under-fit. Discrete and nested ice-marginal channels are identified where channels do not follow the direct topographic slope, but take a path oblique to it.

Eskers

Eskers are sinuous ridges of stratified sand and gravel (and sometimes diamicton) that represent the casts of ice-walled meltwater channels (Bannerjee and McDonald, 1975; Brennand, 2000). They may be deposited by flowing water in supraglacial channels, englacial conduits, subglacial ice tunnels, or ice-walled canyons (Burke et al., 2015). Eskers were primarily identified and mapped based on their ridge form and orientation perpendicular to the ice margin. They were mapped initially as lines based on ridge-crest and then subsequently mapped as polygons based on qualitative observation of slope-break position near the ridge base. Detailed classification differentiated between primary (highest relief) ridge and secondary (adjacent, lower relief) ridges, as well as between esker trunk (the spine of an esker network) and tributaries (esker ridges connecting into a main trunk).

Integration of esker morpho-sedimentary relationships and geomorphic context allowed classification of eskers in the study area into three types and two subtypes (Chapter 5) (Table A.1).

Ice-marginal paleolakes

Ice-marginal paleo-lake systems exist where water is dammed between the ice-margin and some topographic barrier (Carrivick and Tweed, 2013). When the ice dam is removed, the paleo-lake drains, leaving a record of its existence in remnant glaciolacustrine deposits and landforms including deltas, subaqueous fans, shorelines and spillways. Ice-marginal lake systems may be useful as indicators of ice margin positions, as they can reveal successive stages associated with different positions of the
ice front (e.g., Jansson, 2003). Successive lake stages are identified from suites of landforms that develop at successively lower elevations against the receding ice margin when new outlets open. Only paleolake systems for which glaciolacustrine deposits and landforms were identified beyond modern lake boundaries are mapped; however, it is assumed that lateglacial paleolakes existed in all basins where modern lakes now exist.

Ice-marginal lake stages were iteratively reconstructed in a geographic information system by modelling a paleo-water plane on a digital elevation model from primary and secondary indicators and predicting the location of ice-dams (Chapter 3) (Table A.1). Where geomorphic evidence of terrestrial dams could not be invoked to explain lake formation, ice-dams were reconstructed. Primary water plane indicators include deltas (Postma, 1990) and lake maintenance paleo-spillways (LaRocque et al., 2003). These provide the best estimation of the actual water plane elevation. Secondary water plane indicators include subaqueous fans (Winsemann et al. 2007), glacial lake outburst flood (GLOF, via ice dam or earthen dam) or outburst flood (earthen dam) paleospillways (Costa and Schuster, 1988), and ice-marginal channels (Syverson and Mickelson 1997). They provide a minimum water plane elevation.

Gilbert-style deltas were first identified from aerial photographs and the hillshaded DEM as flat-topped deposits at the terminus of meltwater channels. The interpretations were confirmed in the field by the observation of foreset beds in available exposures or GPR profiles. Subaqueous fans were mapped where fan-shaped deposits at the terminus of meltwater channels were steeper than the near flat-topped delta systems, where delta foresets were not observed in exposures yet surrounding lake bottom sediments confirmed deposition in a lake basin. Spillways are meltwater channels formed or enlarged during lake drainage events. They are differentiated largely based on substrate. Bedrock and till are more resistant to erosion than ice or glaciofluvial materials. On the low gradient plateau surface bedrock and till-floored spillways, therefore, record gradual drainage associated with lake-stage maintenance.

Results

The number and mean length of each landform are presented in Table A.1. Moraines mainly occur in the centre and northeast of the study area and some are spatially
grouped by type. Type 1a, discrete glaciotectonic ridges (overridden), are largely found southeast of the Green Lake basin, oriented WNW to ESE. Type 1b, discrete glaciotectonic ridges (recessional), occur throughout the study area. Type 1c, glaciotectonic moraine fields (subglacial ribbed terrain), occur around the Green Lake basin and are oriented parallel to the long axis of the lake basin. Type 1d, glaciotectonic moraine fields (recessional), occur on the interfluves and slope flanks around Dog Creek basin. Grounding line moraines are restricted to valley-bottoms associated with paleo-lakes. Crevasse-fill ridges are locally found in association with other ice stagnation landforms such as kettled topography (kettled topography is present in small amounts and is not regionally connected throughout the study area, it likely represents localized pockets of ice decay). The implications of moraine distribution and genesis for the pattern and style of decay of the last CIS are discussed further in Chapter 4.

Proglacial and subglacial meltwater channels are found throughout the study area. The longest subglacial channels (>30 km long) are occupied by long subglacial (Type 1a) eskers and cross the middle of the study area from northwest to southeast. Short subglacial (type 1b) esker systems are associated with shorter subglacial channels throughout the study area. Large groups of nested ice-marginal channels occur on the NE slopes of the Marble Range, the eastern slopes of Mount Big Bar, on the slopes south of Loon Lake and north of Bonaparte Lake. On the NE slopes of the Marble Range, some ice-marginal channels are aligned with eskers (type 2) downslope. All type 2 eskers are located in the SW of the study area on or near the slopes of the Marble Range. Type 3 eskers formed in ice-walled canyons and are aligned with proglacial channels downstream, or terminating in lake basins (e.g., Young Lake basin).

Paleo ice-marginal lake systems developed largely in the west-central and northwest sectors of the study area where adverse topography likely encouraged lake formation against the retreating ice margin (e.g., glacial Lake Meadow, glacial Lake Dog). Smaller, valley-bottom lakes developed in the eastern part of the study area (e.g., glacial Lake Young, glacial Lake Brigade). All lake systems developed a pattern of higher-elevation stillstands in the east evolving to lower-elevation stillstands in the west. Classic, gilbert-style delta systems were only identified in the largest paleolake system, glacial Lake Dog, and here they only develop on the northeast side of the basin, perhaps indicating the direction from which the majority of meltwater and sediment supply was
being delivered into the lake. A hjulstrom delta was also identified in glacial Lake Young. Subaqueous fans are distributed through most of the lake systems and again are normally deposited on the north side of the lake basin in which they are mapped. Spillways mainly drained lakes to the south and west. A full discussion of lake reconstruction and evolution is presented in Chapters 2 and 3.

Discussion

Large fields of glacitectonic moraine ridges have been identified on the plateau. Many of these (type 1b and d moraines) are sharp-crested and do not appear to have been overridden, indicating they are recessional in origin (Chapter 4). Type 1d moraines occur in broad fields in the northwest part of the study area and indicate active retreat was occurring at the margin of the ice sheet in this area. Their NE-SW orientation suggests active ice margin retreat toward the northwest (Chapter 4), consistent with previous records of glacioisostatic rebound towards the northwest of the study area (Johnsen and Brennand, 2004) and the successive record of ice marginal paleolakes (Chapter 3). The orientation and slope of nested ice-marginal channels on the slopes of the Marble Range and elsewhere indicate that ice was retreating towards the northwest as the ice-surface lowered (Margold et al., 2013). Eskers (type 2) developed downflow of some of these channels as well as long subglacial eskers (type 1a) indicate water flow towards a hydraulic low in the southeast (Chapter 5). Paleo-ice marginal lake systems evolved from high-elevation still-stands in the east of each basin to low-elevation still-stands in the west indicating an overall movement of the ice margin and its role as an ice-dam from east to west across the study area (Chapter 3). Indications of ice stagnation (e.g., crevasse fill ridges (Chapter 4); ice-walled canyon eskers (Chapters 2 and 5); flat-topped esker segments (Burke et al. 2012b, Chapter 5) are present locally, but do not appear to be regionally significant. Taken together, the landform record appears to support active ice margin retreat across the southern Fraser Plateau generally from southeast to northwest; stagnation was localized. Future updates to the resolution of available digital elevation data and higher resolution aerial photograph data will likely result in the potential for more detailed mapping of this area and further identification of late-glacial landforms to round out this picture.
Conclusions

The mapping and detailed classification of deglacial landforms allows a critical leap between mapping of surficial material and constructive inference on the deglacial pattern and dynamics for the last CIS over the southern Fraser Plateau. The level of genetic detail is now present to further our interpretations on the location of the ice margin at various time intervals and the pattern of retreat. The landform record indicates that active ice retreat occurred in places on the southern Fraser Plateau (Chapter 4) with a general direction of retreat from southeast to northwest (Chapters 2, 3, and 5). Pockets of localized ice stagnation were present (Chapter 4), but do not appear to be regionally important across the Plateau.

Software

Heads-up digitizing, viewing of orthophotographs and hillshaded DEM's was completed in ArcGIS 9.3-10.2. Stereophoto pairs and Google-Earth imagery were used for augmenting landscape views and where greater detail was necessary. Landform classification was entered and stored in a personal geodatabase system (MSAccess). Final map production was completed in Coreldraw X5.

Data

Line and polygonal datasets of landforms along with their detailed classification are included in the supplementary material as a file geodatabase for display and analysis.

Map design

Map presentation scale was chosen to maximize scale (1:175 000) while still maintaining a reasonable printable size (A1). Supporting map elements were placed around the main map with efforts to preserve balance. Map projection was chosen to align with environmental and geologic map standards for the Province of British Columbia.

References


Huntley, D.H. 1995. Surficial geology of Churn Creek (92O/07), Empire Valley (92O/08), Dog Creek (92O/09) and Mount Alex (92O/10) map areas. *Geological Survey of Canada, Open File 3155*, Scale 1:50 000.


Appendix B.

Stereo-pair Aerial Photographs Consulted

The following table contains reference information for stereo-pair aerial photographs covering the southern Fraser Plateau and consulted during mapping in this thesis.

B.1 Stereo-pair aerial photographs consulted

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Appendix C.

Site Coordinates

The table below contains site-specific coordinates for significant sites discussed in this thesis.

### C.1 Site coordinates

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Appendix D.

Electrical Resistivity Tomography data

The data presented in Appendix figures D.2-D.17 were collected using parameters outlined in Appendix Table D.1. Data shown is included in this appendix because it was represented incompletely in chapter figures or was not used in any chapters but may be helpful for future investigations. When combined with data presented in the thesis chapters, the Figures presented in Appendix D complete the documentation of processed and interpreted ERT data.

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D.5 70 Mile moraine: A2 (Chapter 4)
D.6 70 Mile moraine: A3 (Chapter 4)
D.7 Loon Lake moraine (3D inverted resistivity quadrangle): B1 (Chapter 4)
D.8 Green Lake moraine: D1 (Chapter 4)
D.9 Holden Lake moraine: E3 (Chapter 4)
D.10 Machete Lake moraine: (included in appendix I.2)
D.11 Green Lake esker: Grid 1 Line Y1 (Chapter 5)
D.12 Chasm esker: Grid 1b, Line X6 (not included in this thesis, previously published in Burke et al. 2012)
D.13 Chasm esker: Grid 1b, Line Y7 (not included in this thesis)
D.14 Chasm esker: Grid 2, Line X2 (not included in this thesis, previously published in Burke et al. 2012)
D.15 Chasm esker: Grid 3, Line X1 (not included in this thesis, previously published in Burke et al. 2012)
D.16 Chasm esker: Grid 3, Line X6 (not included in this thesis, previously published in Burke et al. 2012)
D.17 Chasm esker: Grid 3, Line Y3 (Smooth-model inversion, not included in this thesis)
Appendix E.

Ground Penetrating Radar data

The data presented in Appendix figures E.2-E.18 were collected using parameters outlined in Appendix Table E.1. Data shown is included in this appendix because it was represented incompletely in chapter figures or was not used in any chapters but may be helpful for future investigations. When combined with data presented in the thesis chapters, the Figures presented in Appendix E complete the documentation of processed and interpreted GPR data.

E.1 Ground Penetrating Radar data collection parameters

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<th>Antenna separation (m)</th>
<th>Step-size (m)</th>
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\(^1\) p3D: pseudo three-dimensional grid built by interpolating grid lines ≤ 5 m apart
E.2 Young Lake esker-like ridge: Grid 1 (chapter 2, uninterpreted)
E.3 Young Lake esker-like ridge: Grid 2 (chapter 2, uninterpreted)
E.4 Dog Creek delta: (chapter 3, delta 3, uninterpreted)
E.5 Dog Creek delta: (chapter 3, delta 3, interpreted)
E.6 White Lake moraine: Grid C1 (Chapter 5, uninterpreted)
E.8 Green Lake esker: Grid G1 (chapter 5, uninterpreted)
E.9 Green Lake esker: Grid G1 (chapter 5, interpreted)
E.10 Green Lake Esker: Grid G2 (chapter 5, uninterpreted)
E.11 Green Lake Esker: Grid G2 (chapter 5, interpreted)
E.12 Green Lake Esker: Grid G2 (chapter 5, event sequence)
E.13 Green Lake esker: Grid G2b (not included in thesis, uninterpreted)
E.14 Canoe Creek esker: Grid C1 (chapter 5, uninterpreted)
E.15 Canoe Creek esker: Grid C1 (chapter 5, interpreted)
E.16 Canoe Creek esker: Grid C1 (chapter 5, event sequence)
E.17 Hooke Road esker: Grid H1 (chapter 5, uninterpreted) and H2 (not included in thesis, uninterpreted)
E.18 Hooke Road esker: Grid H1 (chapter 5, interpreted (A) and event sequence (B))
Appendix F.

Clast fabric data

Data in the tables below represent actual data behind clast fabric statistics and diagrams presented in this thesis (see appendix C for site locations). Data collected, but not presented in the thesis is also included (see F.4 for site locations).
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<th>Plunge° (A)</th>
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Appendix G.

Assessment of paleo-water plane tilt (Chapter 3).

Elevations of primary water-plane indicators within LS-3 were graphed and assessed for groups at similar elevations, or slopes that could possibly reflect glacioisostatic tilt (Fig. 3.5A, B). LS-1 and LS-2 did not have sufficient primary water-plane indicators to assess glacioisostatic tilt. From an examination of delta brink-point elevations (Fig. 3.2A, B) it is clear that two groups of deltas formed in the Dog Creek basin: upper and lower deltas (Fig. 3.2A, B). The best fit first-order plane through the three lower deltas suggests a water-plane tilt of 0.9 m km\(^{-1}\) towards 57° (Fig. 3.5B). This tilt is of a significantly different magnitude and direction from the tilt computed from paleo-shoreline deformation in the Thompson Valley nearby (1.7 m km\(^{-1}\) towards 321° (Fig. 3.5B), Johnsen and Brennand 2004). However, these deltas are linearly distributed in space – not an ideal scenario for calculating the dip of a planar surface. If the Thompson Valley water plane tilt (Johnsen and Brennand 2004) is applied to LS-3, areas not expected to have been inundated by water, based on geomorphic and sedimentological observations, are flooded (Fig. 3.5A, B). These results suggest that the Thompson Valley tilt does not fit the delta distribution in LS-3. Further, no glacioisostatic tilt is recorded for deglacial lake water planes in the nearby Fraser River basin; rather, a consistent, step-wise elevation change of water-plane indicators is observed throughout the basin (Huntley 1996). Therefore lake water surface planes are modeled as horizontal planes (zero glacioisostatic tilt) in this study (Fig. 3.5).
Appendix H.

Detailed Classification of Esker Trunks

Data contained in the following tables specifies the morphologic characteristics and geomorphic context used in individual esker trunk classification in chapter 5 of this thesis. Esker networks may be linked to GIS data in the digital data attached to this thesis.
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<sup>1</sup>Network IDs for eskers explored in this paper: Green Lake esker = 11; Chasm esker = 29; Canoe Creek esker = 30; Hooles Road esker = 18; Young Lake esker-like ridge = 3

<sup>2</sup>Length of trunk ridge compared to mean value (3408 m) and standard deviation (=5325 m).

<sup>3</sup>Segment planform in order of decreasing proportion of overall trunk ridge length.

<sup>4</sup>Segment crest cross-profile type in order of decreasing proportion of overall trunk ridge length.

<sup>5</sup>Ridge continuity (see text for derivation) – High = 100%, Medium ≥ 95% and < 100%, Low: < 95%.

<sup>6</sup>Mean gap length – Long > 400 m; Medium > 100 m and ≤ 400 m; Short ≤ 100 m.

<sup>7</sup>Sinusity (see text for derivation) – High > 1.20; Medium > 1.10 and ≤ 1.20; Low ≤ 1.10
Appendix I.
Moraine-like ridges on the southern Fraser Plateau

Several moraine-like landforms demonstrate the principle of geomorphic equifinality on the southern Fraser Plateau. These landforms have been classified as moraines in the past but are re-classified here based on new information and detailed fieldwork. They are included in this thesis to demonstrate that the field-work component of this research was a significant and necessary process in the disambiguation and classification of landforms on the southern Fraser Plateau.

I.1 Rim Lake ridges

The Rim Lake ridges are a series of NW-SE-oriented ridges perched on the northeast slope of the Bonaparte Valley (Fig. I.1a, b). Although previously mapped as moraine ridges (Plouffe 2009a), field observations indicate that they are composed of bedded argillite and limestone bedrock (Fig. I.1c), part of the Cache Creek group (Campbell and Tipper 1972). The bedrock structure strikes to the northwest and dips ≈48° to the northeast. Previous bedrock mapping (Campbell and Tipper 1972) identified several faults and sets of dipping beds (dipping away from the Bonaparte River), striking subparallel to the ridges but on the opposite side of the Bonaparte River (Fig. I.1a). Just north of the ridges in the Bonaparte Valley there is a large landslide deposit (Plouffe 2009a) within which Rim Lake is a sag pond. Based on their bedrock composition, the backtilt of the bedrock towards the valley wall, and the presence of other landslides nearby, these ridges are re-interpreted as rotational bedrock slumps.
Figure I.1. (a) Rim Lake ridges (Plouffe 2009). Image centred on coordinates 51° 4’ 6” N, 121° 27’ 47” W. Several ridges (red lines) near Rim Lake previously mapped as moraines (Plouffe 2009, crestlines are transverse to the inferred direction of ice advance Plouffe et al. 2011). A bedrock fault and dipping beds on the southwest side of Bonaparte River valley strike roughly parallel to these ridges (Tipper 1972). Dotted white line shows head scarp of bedrock slump; closed white arrow indicates slump direction. Open white arrow, ‘b’, shows view direction of photograph in (b). ‘c’ locates photograph in (c). (b) Ground view of bedrock exposure (dotted white arrow) cropping out near putative moraine ridges (black arrows) (Bonaparte River in foreground). The lower ridge and bedrock exposure are at similar elevations. (c) Bedrock ridge exposure showing back-tilted bedding (strike/dip is 281°/48°).
I.2 Machete Lake ridges

The Machete Lake ridges have been previously mapped by Bednarski (2009) as major moraine ridges (Fig. I.2a). Detailed investigation shows that they are a series of subparallel ridges with an average relief of about 10 m, located at the northeast end of Machete Lake near the western outlet of a major meltwater channel (Bednarski 2010) (Fig. I.2a) aligned parallel with the axis of Machete Lake valley. Lake bathymetry records a series of closed circular basins and reach a maximum 10 m depth (Province of British Columbia 2011). Ridge length ranges from 500 to 1000 m and cross-ridge elevation profiles demonstrate a lack of consistent pattern in slope angle from the north to the south sides of the ridges (Fig. I.2a). Exposures within ridge sediments are rare, but several shallow roadcuts display a > 1 m thick surficial unit of poorly sorted, unconsolidated, clast poor, subrounded pebble to cobble gravel in a coarse sandy matrix. An ERT profile was completed on the southernmost ridge. The profile shows 4 resistivity units. RU1 ranges from 5-10 m thick and occurs from the base of the profile upwards. It is characterized by very low resistivity values (29-60 ohm-m). RU2 is a discontinuous lens-shaped unit ranging from 4-15 m thick and has moderately low resistivity values (60-400 ohm-m). It is exposed at the surface near the southern end of the profile and appears to narrow towards the north end of the profile. RU3 is a 5-10 m thick unit of very low resistivity values (29-60 ohm-m) that is overlain by RU4, a discontinuous ~3 m thick layer present at the surface near the middle of the profile.

RU1 has low resistivity values likely associated with saturation from the local water table (Machete Lake surface elevation = 1110 m asl). The lens shape of RU2 suggest it is a discontinuous and its moderate resistivity values are similar to diamicton values found elsewhere on the southern Fraser Plateau (cf. Burke et al. 2012). These characteristics along with the position of the ERT profile at the base of a concave slope suggest that this unit may be a debris flow deposit. However, the resistivity values may also be artificially lowered by increased water content in this layer like the unit below and may therefore reflect a coarse sand or gravel unit, similar to RU3, but with larger void spaces, increasing drainage efficiency and resulting in slightly higher resistivity values. The surficial unit within the ridges is very poorly sorted and similar to stratified ice contact deposits mapped in the same valley, northwest of this location (Bednarski 2009), suggesting it is similar in origins to these deposits. The sand in the unit was wet indicating movement of groundwater through the ridge system, possibly supplied by streams coming off the high relief south side of the valley. This is supported by the information provided by the low (29-300 ohm-m) resistivity values for the surface gravel unit (RU3, Fig. I.2b). It is likely that RU4 is similar in composition, but with less moisture content, or reflects slightly higher resistivity associated with loosely consolidated surface material left over from logging activity in the area.

The enclosed, circular depressions shown within the Machete Lake bathymetry suggest the area covered by the lake was previously occupied by stagnating ice blocks, that left kettled topography after ice disappeared. This is supported by the numerous kettle holes and glacial ice-contact material mapped in the surrounding area (Bednarski 2009). The ridges themselves are not kettled, and neither sedimentary exposures nor geophysical data show evidence of pervasive internal deformation resulting from ice meltout, suggesting they were deposited in and around these disintegrating ice blocks. The poorly sorted gravel is consistent with glaciofluvial deposition characterized by short transport distances. The equivocal resistivity signatures of RU2 suggest that this unit could be interpreted as a debris flow diamicton or coarse grained gravel deposit. If interpreted as a debris flow, it is possible the material slumped off of the valley side while crevassed, stagnating ice still occupied the valley bottom. Debris material would have thickly filled the crevasses, and been spread thinner over the ice surface. Surrounding meltout of ice would have deposited a surface layer of glaciofluvial material overtop of the diamicton (RU3 and RU4), and resulted in deformation of inter-ridge deposits, with only minor deformation expected in ridge deposits. If the unit is interpreted as a coarse gravel deposit, it may be the result of inflow from a westerly flowing meltwater channel (Bednarski 2010) entering the valley, and depositing material in an ice-contact delta environment. In this interpretation the coarse grained gravel (RU2) is the avalanche face of the delta, whereas the finer grained sediment above (RU3 and RU4) are the topset material. The ridge deposits result from glaciofluvial deposition around stagnating ice blocks.
within the Machete Lake valley. Further work is required to differentiate which of these two formational mechanisms is correct for the Machete Lake ridges; however, based on their position parallel to the valley axis, between steep valley walls, surficial composition of glaciofluvial material and lack of deformation structures, it is unlikely these ridges are moraines.

Figure I.2. (a) Machete Lake glaciofluvial delta. G1 (open black arrow) indicates location and direction of ERT survey G1. R-R’ (white line) shows location of topographic profile in (b). Dashed white arrow indicates paleoflow in major deglacial meltwater channel (Bednarski 2009). ‘c’ indicates the location of exposure shown in (c). Coloured arrows denote inferred ice flow directions after Plouffe et al. (2011) (yellow = early, green = intermediate, blue = LGM). Topographic profile R-R’ cross ridge set and shown in inset. (b) Interpreted ERT profile G1 transverse to ridge. Dotted lines indicate interpreted resistivity unit boundaries (RU1-RU4). Refer to Appendix D.1 for ERT survey details and processing results. (d) Coarse sand and gravel visible in exposure 3, at the surface of one of the ridges (scale bar is 1 m high; scale card has 1 cm increments).
Appendix J.

Summary of data included in digital storage

1. Unedited list of field sites and observations
2. Geodatabase of lateglacial landforms
3. Poster sized map of Appendix I
4. Optical age data collected and processed but not presented in thesis
5. ERT Raw files
6. GPR Raw files