Subglacial Processes, Glacier Dynamics, and Deglacial Processes and Patterns Associated with the Cordilleran Ice Sheet around Okanagan Valley, British Columbia

by

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Abstract

This thesis explores subglacial processes, glacier dynamics, and deglacial processes and patterns associated with the Cordilleran Ice Sheet (CIS) in Okanagan Valley and the neighbouring Thompson Plateau in southern British Columbia. Reconstructions of subglacial processes in an area of streamlined bedforms (drumlins) on Thompson Plateau reveal that sediments within drumlins and in intervening areas record evidence of lodgement, deformation, poreflow, conduit flow, debris flows, and suspension settling of fines within a network of subglacial cavities. These subglacial cavities developed within a regional bedrock basin, and these sediments demonstrate that substrate deformation is not a dominant and pervasive process recorded within the drumlins. Based on i) drumlin morphology and pattern, including the presence of stoss-side crescentic troughs and *en echelon* arrangement, ii) drumlin composition consisting of sediment and bedrock, it is argued that drumlins on Thompson Plateau have an erosional origin. Further, regional spatial associations between drumlins and tunnel valleys (including some in bedrock) on Thompson Plateau and in Okanagan Valley, and arguments relying on the conservation of eroded sediments suggest that erosion by subglacial meltwater underburst(s) best explains the range of observations. Underbursts may have been associated with development and drainage of a subglacial ‘catch lake’ in Okanagan Valley. High geothermal heat flux in Okanagan Valley could have favoured subglacial lake development. Subglacial volcanic eruptions may have acted as triggers for lake drainage. Lastly, deglaciation of the CIS led to development of a proglacial lake in Okanagan Valley (glacial Lake Penticton - gLP). Sediment delivery to gLP occurred via tributary valleys and possibly from an ice tongue along the valley axis. Regional delta correlations record a highstand of gLP at 500-525 m asl followed by a single drainage event which eroded a portion of the lacustrine valley fill to produce distinctive remnant valley-side benches (‘White Silt’ terraces) of lacustrine sediments. No evidence of glacioisostatic tilting could be discerned within the gLP basin, this absence is tentatively ascribed to lithospheric conditions favouring rapid crustal rebound, and to thin ice with uniform thickness in the gLP basin.

Keywords: Cordilleran Ice Sheet; Subglacial Processes; Deglaciation; glacial Lake Penticton; Okanagan Valley; British Columbia

Dedication

Pour Cédric et Aksel… enfin la fin…pour un nouveau début.

Acknowledgements

The seed of this thesis was planted long ago. I'm glad it has grown into something tangible. This has not been a solitary effort however and many thanks are due.

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I have had the pleasure of sharing office space with a number of interesting characters. Thanks to Dr. Tim Johnsen, Kerry Ayers, Dr. Christina Neudorf, Dr. Chris Bone, and the late Colin Wooldridge for their enthusiasm and interesting discussions.

Above all, I thank my friends and family (immediate and extended) for their unwavering support and encouragement. My wife Leah 'lived' this thesis as much as I did and I'm glad we managed the journey together. Lastly, my sons Cédric and Aksel provided much needed motivation and inspiration at crucial moments. I couldn't have done this without all of you. Thank you!

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1. Introduction

Ice sheets play critical roles in the global climate system and the cryosphere. Their size is sufficient to affect atmospheric circulation. They are significant reservoirs of fresh water and play key roles in sea level variationsassociated with climatic changes. Proper understanding of glacial processes is critical to understanding linkages between climate, ice sheet behaviour and development of glacial landscapes. Although studies of existing ice sheets shed important light on modern processes, they are limited by two factors: a) a relatively narrow temporal window of study (years to decades), dwarfed by the much longer-lived evolution of ice sheets and their associated glacial landscapes (decades to millennia); b) the inaccessibility of the glacier bed which precludes observation of active processes. Improved remote sensing techniques (e.g., seismic, ice radar) have increased imaging of existing glacier beds and knowledge of temporal evolution of glacial bedforms (e.g., King et al., 2009). However, the vastness of modern ice sheet beds remains a limiting factor on understanding regional-scale landscape evolution and glacial processes.

In contrast, studies of deglaciated landscapes offer more widespread access to former ice sheet beds and their constituent landsystems (landforms and sediments), from which process inferences can be made. Therefore, modern and paleoglaciologic studies are complimentary endeavours, each having the potential to inform the other and to improve understanding of (paleo) ice sheet dynamics. Although this thesis examines paleo-ice sheet landsystems explicitly, the general methodology follows the complimentary approach described above: process inferences are made from landsystems. The validity of inferences is evaluated in light of current understanding of glacier processes, which integrates the knowledge base derived from paleoglaciology, modern glaciologic theory, and observations from modern ice sheets.

In comparison to the Laurentide Ice Sheet, paleo-ice sheet landsystems of the Cordilleran Ice Sheet (CIS) have received relatively limited attention. Deglacial sediments dominate the exposed stratigraphic record within valleys. These sediments have received the most attention in southern British Columbia (cf. Fulton, 1965, 1969; Shaw, 1977; Shaw and Archer, 1978, 1979; Johnsen and Brennand, 2006) though the timing of deglacial events remains poorly constrained (§ 1.2). Subglacial processes and glaciodynamics operating during CIS build-up and during glacial maximum are largely unknown in southern British Columbia. This thesis aims to address existing knowledge gaps of CIS paleoenvironments and glaciodynamics in southern British Columbia by focusing on; i) subglacial paleoenvironments and processes during a time period likely associated with CIS build-up and climax, and ii) deglacial processes and patterns associated with CIS decay.

1.1. Study areas

This thesis examines paleo-ice sheet landsystems in southern British Columbia, Canada. Study areas examined lie within the former footprint of the CIS (Fig. 1.1). Chapter 2 focuses on a portion of Thompson Plateau (Fig. 1.1). Study areas examined in Chapter 3 are centred on Okanagan Valley and the surrounding Thompson Plateau (Fig. 1.1). However, some regional assessment of landscape development spans an area including Okanogan Valley south to the Columbia River in northern Washington State, USA, and areas north of Thompson River. Chapter 4 examines an area centred on Okanagan Valley and the neighbouring valley containing Kalamalka Lake (Fig. 1.1). The field area is bounded to the north by the city of Vernon and to the south by the city of Oliver.

1.2. Chronology of Quaternary events in southern British Columbia and Okanagan Valley

There are limited chronologic controls on the timing of late-Quaternary events in southern British Columbia. The majority of chronologic data is associated with the deglacial phase of the Late Wisconsinan (Fraser) glaciation and mostly, though loosely, constrains the timing of glacial lakes.

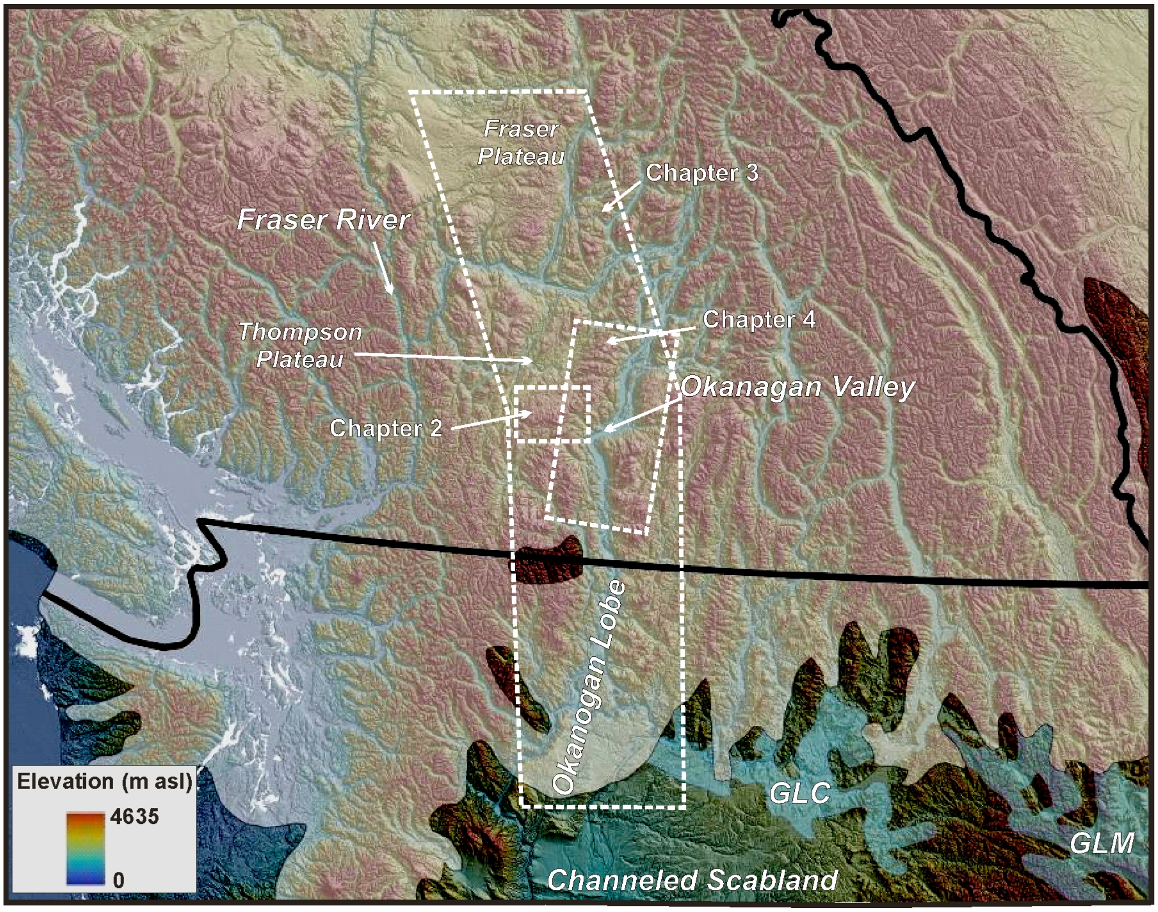


Figure 1.1. Hillshaded digital elevation model of southern British Columbia. Partial transparency represents the maximum Late Wisconsinan extent of the Cordilleran Ice Sheet along its southern margin. Lobes of ice (including Okanogan Lobe) dammed glacial lakes (GLC - glacial Lake Columbia, and GLM – glacial Lake Missoula) in the Channeled Scabland. Field areas of thesis chapters centre on Okanagan Valley and Thompson Plateau (outlined by dashed boxes). Chapter 3 extends to the Columbia River and onto Fraser Plateau bounded by the Fraser River. Data: SRTM 30 m digital elevation model (USGS). CIS margin from Dyke et al. 2003. Datum: WGS 84.

Fulton and Smith (1978) identified a stratotype at Okanagan Centre, British Columbia (Fig. 1.1) and established that sediments exposed there recorded environmental changes as far back as ~100 k a BP (MIS 5). In-situ MIS 3 (non-glacial) sediments have been identified in tributary valleys to Okanagan Valley (Fulton and Smith, 1978) near Lumby (Fig. 1.1). Similarly, organic material dated between 65-23 k cal a BP have been identified within the fill of Okanagan Valley (Fulton, 1972) and within the fill of Kalamalka Valley (Paradis et al., 2010; Thomson, 2010). In many cases, some of these organics may be detrital and may therefore record maximum ages of encasing sediments. However, some in-situ organics (wood debris) may be preserved along the southern flanks of Ellison Ridge north of Kelowna (Thompson, 2010). Preservation of this organic material suggests a complex valley fill history which conflicts with lake-based (Eyles et al., 1990, 1991) and land-based (Vanderburgh and Roberts, 1996) seismic surveys that reveal a simple valley fill architecture in Okanagan Valley, attributed to the Fraser glaciation. Preservation of such organics (and associated sediments, highlights the complex and varied spatial patterns in erosion between Okanagan Valley and its tributaries. Re-examination of the chronology of environmental changes at the Okanagan Centre stratotype (Lesemann et al. in review) mostly supports the original interpretations of Fulton and Smith (1978) up to the end of the Penultimate glaciation. This recent re-examination suggests that a substantial portion of the record of the Late Wisconsinan glaciation is missing from the stratigraphic record at Okanagan Centre.

Onset of the Fraser glaciation and ice sheet growth to its maximum extent is broadly ascribed to the period between 15-25 k cal a BP (Clague, 1980, 1989) Although the Last Glacial Maximum extent of the CIS is established through landform distribution (e.g., terminal moraines), the timing of ice dispersal from spreading centres was likely asynchronous between northern and southern sectors (Clague, 1980; Porter and Swanson, 1998). Maximum ice cover over southern British Columbia occurred within the 15-20 k cal a BP interval. The terminal position of the Okanogan Lobe along the southern CIS margin was reached by 15 K cal a BP (Clague, 1989). Rapid deglaciation characterized CIS retreat. Deglaciation of plateaus and highlands in the southern Interior was nearly complete by 10.5-11 k cal a BP (Lowdon and Blake, 1973; Clague, 1989; Fulton 1991). Deglacial lake sediments have yielded no datable material. Radiocarbon dates from bogs along Okanagan Valley suggest that the main deglacial lake in Okanagan Valley (glacial Lake Penticton; Nasmith, 1962) had drained before ~9.4 ± 0.1 k cal a BP (8.41 ± 0.1 k ^{14}C a BP) (Alley, 1976). Recently, this date has been refined to 11 ± 1 k a BP using optically-stimulated luminescence techniques (Lesemann et al., in review).

1.3. The Cordilleran Ice Sheet (CIS)

The Cordilleran Ice Sheet (CIS) engulfed much of British Columbia during the last glacial interval. An existing conceptual model of CIS evolution (Kerr, 1934; Davis and Mathews, 1944; Fulton, 1991) summarizes late-Wisconsinan ice sheet growth and decay (Fig. 1.2). This model postulates that ice sheet growth at the onset of glaciation began from high elevation cirques along the eastern and western CIS boundaries. As cirque glaciers grew into valley glaciers, they coalesced and filled deeply-incised valleys such as Okanagan Valley. At this stage, flow was strongly controlled by topography. In southern British Columbia, the CIS is generally assumed to have flowed southward toward its margin in Washington State, USA. Ongoing growth of the Cordilleran glacier is thought to have covered most valleys and mountain ranges, leading to an ice sheet phase where local domes developed and where regional flow was unconstrained by topography (Dawson, 1881; 1891; Kerr, 1934; Mathews, 1955; Fulton, 1967; Flint, 1971). This last phase may have persisted for short periods of time and may have never been attained in places (Tipper, 1971; Clague, 1989; Fulton, 1991).

At its maximum, ice thickness within the CIS was highly variable: thick ice filled valleys (up to 2500-3000 m thick in some valleys) and comparatively thin ice (few 100s m, Fulton, 1991) covered surrounding plateaus and accumulation areas (Fig. 1.2). Deglaciation patterns and processes reflect this uneven ice thickness and it is postulated that plateaus deglaciated before valleys (Davis and Mathews, 1944; Fulton, 1991). Thick valley ice led to stagnation of ice masses and development of extensive deglacial lakes (Flint, 1935b; Mathews, 1944; Nasmith, 1962; Fulton, 1965; Shaw, 1977; Shaw and Archer, 1978, 1979; Ward and Rutter, 2000; Johnsen and Brennand, 2006). Consequently, in southern British Columbia, deglaciation is considered to have occurred largely by regional stagnation (Davis and Mathews, 1944; Fulton, 1991).

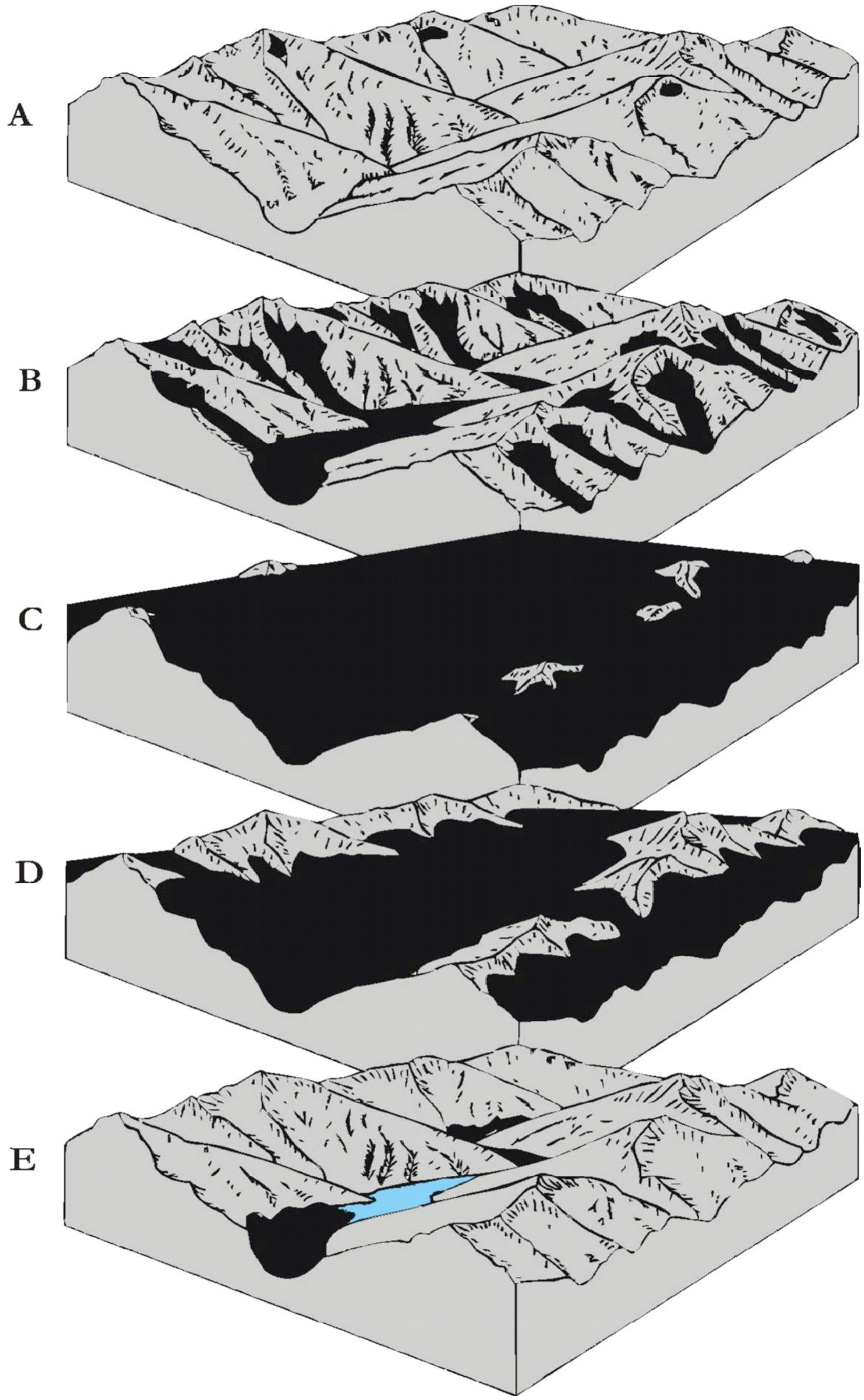


Figure 1.2. Conceptual model of Cordilleran Ice Sheet growth and decay (modified from Fulton, 1991) based on Kerr (1934), Davis and Matthews (1944).

Although this model of the CIS provides a simplified and highly conceptual depiction of glaciation, it is in fact based on limited field data, particularly in southern British Columbia. Studies have helped refine this model through smaller-scale (usually regional) studies focused on process and landscape reconstructions. Some studies support the existing model (e.g., regional stagnation of ice in/near lake basins, Ward and Rutter, 2000) whereas others have suggested more limited ice-sheet periods (McCuaig and Roberts, 2002) and refinements to the existing model suggest the possibility of coeval ice masses on plateaus and in association with lakes in the valley (Johnsen and Brennand, 2006). Consequently, local studies capture a level of complexity in CIS dynamics that can be integrated within the synoptic representation of the existing CIS model and lead to refinement of process understanding for the CIS. In southern British Columbia, and particularly in Okanagan Valley, detailed evaluation of the prevailing CIS model based on regional integration of landforms and sediments has only been summarily performed, focusing mainly on the record of deglacial lakes (Flint, 1935b; Nasmith, 1962; Fulton, 1967; 1969; Shaw, 1977; Shaw and Archer, 1979) where stagnant valley ice and deglaciated plateaus form the paleoenvironmental context for deglacial lakes. By comparison, much less is known about subglacial processes and glaciodynamics on plateaus surrounding Okanagan Valley.

1.3.1. The CIS as a modern ice sheet analogue

Unlike the Laurentide Ice Sheet that developed over the generally subdued topography of the Canadian Shield and Interior Plains, the CIS is considered a montane ice sheet, developing on the variably rugged topography of British Columbia, the Yukon and in the northern portions of Washington State, Idaho and Montana, USA. The pronounced bed topography affected ice distribution, flow patterns and meltwater routing. In turn, these variables exerted great control on glacial and deglacial processes and patterns. The montane character of the CIS is most similar to modern ice sheets in Greenland and Antarctica, both of which rest on a bed with considerable topography. As such, the CIS may well be the most appropriate paleo-ice sheet analog to modern ice sheets (Lian and Hicock, 2010). As a result, reconstructions of CIS paleoenvironments and processes have the potential to shed light on the current and future dynamics of modern ice sheets.

1.4. Subglacial processes and ice sheet dynamics

Subglacial processes operating at the ice-bed interface strongly affect the dynamics of glaciers and ice sheets (cf. Bennett, 2003 and references therein). The production, deposition, and deformation of subglacial sediments is affected by subglacial hydrology which modulates abrasion and plucking, friction leading to lodgement, and sediment dilatancy. As well, meltwater storage and release over various temporal and spatial scales has repercussions on glacier flow and ice sheet behaviour. For example glacier behaviour such as surging (Kamb, 1987) and some types of ice streaming (Alley, 1989; Kamb, 1987' 1990, 2001; Fricker et al. 2007) are associated with episodic reorganizations of subglacial drainage systems.

1.4.1. Reconstructions of CIS subglacial processes

If the prevailing CIS growth-decay model addresses ice sheet-wide ice distribution, it does not provide significant information on ice sheet processes and associated glaciodynamics *per se*. Glaciodynamics along Okanagan Valley have been inferred by Mathews (1991) who postulated the presence of an ice stream in Okanagan Valley and suggested that the valley acted as a major outlet for ice, draining the interior plateaus of southern British Columbia. However, this inference relies largely on physiographic similarities between Okanagan Valley (i.e. an overdeepened bedrock valley) and some types of ice streams found in Greenland (*Isbrae*-type ice streams, Truffer and Echelmeyer, 2003). Kleman and Glasser (2007) also postulate the operation of an ice stream in and around Okanagan Valley, and complex temporal switches in subglacial thermal regime from regional bedform patterns. In these cases, proposed process inferences depend on assumptions of landform genesis and are largely unconstrained by regional integration of field-based landform and sediment observations that could shed light on bed processes of the CIS (see Lian and Hicock, 2000, Lian et al., 2003 for exceptions). Thus, there are extensive knowledge gaps regarding processes controlling the behaviour of the CIS and the resulting landscape evolution.

1.5. Thesis scope and objectives

This thesis begins to address existing knowledge gaps in CIS glaciodynamics in southern British Columbia through:

1. an evaluation of subglacial processes on montane plateaus (Thompson Plateau) adjacent to Okanagan Valley (chapter 2); and
2. a regional reconstruction of landscape-forming events along a portion of the southern CIS margin (chapter 3); and
3. a regional study of deglacial dynamics associated with a glacial lake (glacial Lake Penticton, Nasmith, 1962) in Okanagan Valley (chapter 4).

These themes form the basis of the three substantive chapters (2-4) presented in this thesis. Detailed chapter objectives and rational are further elaborated below.

1.5.1. Subglacial processes and substrate genesis on a portion of Thompson Plateau, BC

Cordilleran valleys contain extensive sediment fills (cf. Fulton, 1965, 1972; Eyles and Clague, 1991; Vanderburgh and Roberts, 1996; Ward and Rutter, 2000; Lian and Hicock, 2001; Johnsen and Brennand, 2006). Consequently, studies have preferentially focused on these sediment sinks to perform environmental and process reconstructions. Some studies focused on landforms and sediments in areas or more subdued relief outside of major Cordilleran Valley systems have shed light on subglacial hydrology and glacier bed conditions of the CIS. For example, Lian and Hicock (2000) and Lian et al. (2003) reconstructed CIS thermal regime evolution and inferred fast glacier flow from till sedimentology, stone fabric and surface wear data and geologic structures for small valleys in the interior of British Columbia. They reported evidence for ductile and brittle sediment deformation related to temporal changes in bed thermal regime and substrate water content. Broster and Clague (1987) analyzed geologic structures in tills near Williams Lake (Interior Plateau) to infer that basal ice was at least temporarily frozen to the bed, forming brittle deformation structures (faults, joints and fractures). Infilling of these structures with sorted sediment suggested the presence of meltwater as a result of basal temperature reaching pressure melting point. A converse switch in thermal regime is indicated near Big Creek, BC, where sediments record a sequence of ductile, brittle and semi-brittle deformation as ice advanced over saturated sediments before freezing to the bed (Huntley and Broster, 1993).

In contrast, plateaus have received comparatively little attention (cf. Broster and Clague, 1987; Huntley and Broster, 1993 for some examples), despite the fact that they host extensive suites of glacial landforms and sediments. This is particularly true for Thompson Plateau where exploratory studies of landforms and sediment have focused on deglacial processes and patterns (e.g., Fulton, 1967). Little is known about glacier bed processes and subglacial conditions on Thompson Plateau. In addition, spatial continuity of landforms (valley networks, drumlins) between Thompson Plateau and Okanagan Valley raises the possibility of examining the connectivity of regional landsystems associated with major Cordilleran Valleys and possible glaciodynamics such as ice streams (cf. Matthews, 1991; Kleman and Glasser, 2007). For these reasons, and the availability of accessible sediment exposures within a portion of a regional drumlin swarm, detailed work is fosused on a portion of Thompson Plateau.

1.5.1.1. Chapter objectives

Chapter 2 examines a suite of glacigenic sediments exposed on southern Thompson Plateau, BC (Fig. 1.1). The area lies within an extensive tract of drumlins described in Chapter 3. Sedimentary exposures within drumlins, within inter-drumlin areas, and in non-drumlinized zones are used to reconstruct subglacial processes.

Specifically, the research objectives are:

1. to reconstruct paleo-CIS bed environments and (sub)glacial processes within an area of streamlined bedforms; and
2. to explore the implications of inferred bed conditions on our understanding of paleo-CIS dynamics.

1.5.2. Regional reconstruction of subglacial hydrology and glaciodynamics along the southern margin of the CIS

Former ice sheet beds contain a record of subglacial processes responsible for sediment production, movement and deposition, and the genesis of subglacial bedforms. Consequently, many reconstructions of paleo-ice sheet dynamics rely on interpretations of bedforms (e.g., drumlins, Rogen moraine, tunnel valleys, etc) and substrates (Lian and Hicock, 2000; Lian et al. 2003; Evans et al. 2006). Conflicting interpretations of landform genesis lead to varying conceptions of paleo-ice sheet dynamics. For example, sediment deformation may be responsible for producing a wide array of glacial bedforms (drumlins, Rogen moraine, some tunnel valleys) (Boulton, 1987; Boulton and Hindmarsh, 1987). However, the same landforms may also result from the erosional and depositional action of turbulent meltwater (Shaw 1996, 2002). Consequently, glaciodynamic reconstructions based on glacial bedforms alone remain equivocal. Regional landscape reconstructions may help shed light on subglacial processes by integrating inferred genetic processes for suites of glacial landforms. This approach is taken in order to reconstruct landscape genesis on Thompson Plateau and surrounding areas.

1.5.2.1. Regional erosional events in Okanagan Valley and surrounding areas

Streamlined bedforms and tunnel valleys eroded in bedrock and sediments occur within Okanagan Valley, on the valley sides leading to the plateaus, and on the plateaus themselves. In places, streamlined bedforms and tunnel valleys form continuous networks between the plateaus and Okanagan Valley. Because they occur predominantly in bedrock, tunnel valleys likely record erosion by subglacial meltwater flow. Similarly, morphologic elements associated with drumlins (e.g., crescentic scours), and the occurrence of bedrock drumlins amongst sediment drumlins suggest an erosional origin for these landforms. However, the erosional agent(s) for these drumlins is not immediately clear and requires examination.

As well, valley fill characteristics in Okanagan Valley suggest complex erosional and depositional events during glacial periods. For example, early seismic surveys of Okanagan Valley hinted at a complex valley fill (Fulton, 1972) spanning multiple glacial and interglacial periods (Fulton, 1972; Fulton and Smith, 1978). In contrast, more recent lake-based and land-based seismic surveys (Eyles et al., 1990, 1991, Vanderburgh and Roberts, 1996), suggest instead that the valley fill is relatively young and largely Late Wisconsinan in age. An important implication of these interpretations is that significant sediment erosion has taken place during glacial periods to remove the majority of the valley fill and to subsequently refill the valley with younger sediments. Eyles et al. (1990,

1991) and Vanderburgh and Roberts (1996) attribute some of this erosion and valley filling to pressurized subglacial meltwater erosion and deposition.

The array and interconnectivity of erosional landforms, and the inferred genesis of the Okanagan Valley fill raises the possibility that regional subglacial erosional events may be responsible for landscape genesis. Such events have important implications for understanding of glacial landform genesis, glacial landscape evolution, and understanding of subglacial processes and associated glacier dynamics.

1.5.2.2. Chapter objectives

Chapter 3 builds on the results of Chapter 2 and examines regional landform patterns, and landform-sediment assemblages in the Okanagan Valley-Thompson Plateau area of British Columbia, Canada and in northern Washington State, USA in order to produce a regional reconstruction of glacier bed processes and glaciodynamics.

Specifically, the research objectives are:

1. to evaluate landscape development, regional glaciodynamics and hydrologic conditions along the southern margin of the CIS;
2. to evaluate hypotheses of landform genesis and present an event sequence of landscape evolution; and
3. to investigate the implications of landform genesis and landscape evolution for CIS hydrology, glaciodynamics and geometry.

1.5.3. Paleogeographic and paleoenvironmental reconstruction of glacial Lake Penticton

The rugged topography of the CIS, coupled with the potential for preservation of stagnant ice masses within valleys, led to development of extensive glacial lakes within the deeply-incised valleys of southern British Columbia (e.g., Flint, 1935b; Nasmith, 1962; Fulton, 1965; Shaw, 1977; Shaw and Archer, 1978). The largest of these lakes (glacial Lake Penticton (gLP), Nasmith, 1962) developed in Okanagan Valley. Today, the evidence of this former lake consists of perched lacustrine sediment benches flanking the valley sides (‘White Silt’ terraces, Flint, 1935b).

The exact timing of gLP inception is unknown due to a paucity of datable material in its glaciolacustrine sediments. Regional assessment of stratigraphic relationships allows for relative timing of paleo-lake basins in southern British Columbia. In many valleys of southern British Columbia, lake sediments occur near the top of valley fills and make-up the majority of Quaternary sediments. Lake sediments typically overlie Fraser glaciation till (cf. Fulton and Smith, 1978). In turn, Holocene terrace gravels and eolian sediments, mostly dated using tephrachronologic (younger than Mt. Mazama set O tephra, 7.6 k cal a BP) and optical dating techniques (younger than 11 k a BP, Lesemann et al., in review) typically overlie lake sediments (Roberts and Cunningham, 1992; Lian and Huntley, 1999). In Okanagan Valley, only eolian sediments overlie lacustrine sediments.

The duration of gLP is equally cryptic as only minimum ages for lake existence are available (drainage by 11 k a BP, Lesemann et al., in review)). However, late-glacial lakes had potentially short lifespans. Fulton (1965) used rhythmites (interpreted as varves) in glaciolacustrine sediments of the South Thompson Valley to suggest that lake duration may have been as short as ~80 years.

Although there is agreement on the presence of a glacial lake in Okanagan Valley, varied interpretations of its paleogeography and paleoenvironment have been put forth. Key characteristics of the gLP basin are summarized below and provide context for thesis research objectives.

1.5.3.1. GLP paleogeography

Various reconstructions of the basin-wide extent of gLP have been proposed, ranging from valley-wide extent (Nasmith, 1962) to more limited development in certain portions of Okanagan Valley (Fulton, 1965; Shaw, 1977; Shaw and Archer, 1978). Areally-limited reconstructions mainly reflect constraints on field areas and study scope. For example, Fulton (1967, 1969) focused on the northern portion of the gLP basin, and consequently his reconstruction reflects mapsheet boundaries under study during his surficial mapping mandate. Vanderburgh and Roberts (1996) also report data from the area around Vernon due to their focus on land-based seismic surveys. Shaw (1977) and Shaw and Archer (1978, 1979) focused on development of gLP in the southern portion of Okanagan Valley (between Squally point and Okanagan Falls, Fig. 1) where the White Silt benches are most prominent. Silt benches and localized silt plains occur throughout Okanagan Valley and in Kalamalka Valley. Lake- and land-based seismic surveys have revealed up to 730 m of lacustrine sediments below Okanagan Lake (Eyles et al., 1990, 1991) extending beyond the modern lake boundaries where they make up a significant portion of the valley fill in the northern extent of the gLP basin (Vanderburgh and Roberts, 1996) and along the southern extent of Kalamalka Valley around Kelowna (Paradis et al., 2010) (Fig. 1.1).

Nasmith (1962) proposed that gLP occupied an isostatically-depressed basin and was dammed by ice and sediment within a valley constriction located between Okanagan Falls and McIntyre Bluff (Fig. 1.1). He envisaged drainage of gLP through steady-state breaching of the decaying ice and sediment dam, punctuated by still stands, though the mechanics of these punctuated drainages were not elaborated.

If lake sediments record the presence of a lake in Okanagan Valley, they only offer minimum estimates of lake extent. Water plane reconstructions are required to refine reconstructions of lake extent. Fulton (1969) expanded on Nasmith's (1962) work and proposed four lake stages based on tributary mouth deltas. Many of the proposed lake stages are based only on a single delta, rather than on regionally correlated deltas. Furthermore, these reconstructed water planes were developed within the spatial limits of Fulton's (1969) study (see above) and assumed to be representative of the entire gLP basin. Their extrapolation beyond the boundaries of the map sheet under study is not constrained by water plane indicators. There is therefore some uncertainty associated with lake extent, and the number of, and controls on, proposed gLP stages.

1.5.3.2. GLP paleoenvironment

Studies of glacial lake sediments in Okanagan Valley (Shaw 1977, Shaw and Archer 1978) and in the South Thompson Valley (Fulton 1965; Johnsen and Brennand, 2006) have established that high sedimentation rates were common in these glacial lakes (Fulton, 1965; Shaw, 1977; Shaw and Archer, 1978; Eyles et al., 1991). In the Thompson valley, several tens of metres of sediment were deposited in a few years in some of these lakes (as little as ~80 years, Fulton 1965). Turbidity currents transported and deposited a large proportion of these sediments within the lake basins (Shaw, 1977; Shaw and Archer, 1978; Johnsen and Brennand, 2006). White Silt benches exhibit upward thinning and fining trends suggestive of decreasing sediment volumes and increasingly distal depositional conditions with time. This was attributed to diminishing meltwater delivery in response to decaying ice during ice retreat (Fulton 1965, 1969; Shaw 1977). However, if the elements listed above summarily describe some general paleoenvironmental conditions within the gLP basin, the location of decaying ice masses within the gLP basin is less clear and has lead to various conceptions of lake type and sediment sources (§ 4.7.1). Ice distribution within the gLP basin is closely linked to the origin and distribution of White Silt benches in Okanagan Valley (§ 4.8.4) and bears on paleoenvironmental reconstructions.

1.5.3.2.1. Lake type and sediment sources of gLP

Proposed lake types for gLP include ice-lateral (ribbon) lake (Flint, 1935b; Nasmith, 1962), supraglacial lake (Fulton, 1969; Shaw 1977; Shaw and Archer, 1978), and proglacial lake (Eyles et al. 1990, 1991; Vanderburgh and Roberts, 1996). Flint (1935b) and Nasmith (1962) envisaged that gLP developed between the bedrock valley walls and a stagnant ice tongue along the axis of Okanagan Valley, which received sediment from tributary valleys to Okanagan Valley. Flint (1935b) further suggested that these ice-lateral lakes could merge into a supraglacial lake as the lake expanded over stagnant ice. In contrast, Fulton (1969) and Shaw (1977) envisaged sediment delivery by melting ice in Okanagan Valley. In this context, gLP developed mainly as a proglacial lake (with an initial ice-lateral component), including a supraglacial component developed over stagnant Okanagan Valley ice. Using seismic data from Okanagan Lake, Eyles et al. (1990, 1991) envisaged a proglacial lake developed over ice blocks (of unknown extent) grounded on the lake floor. Lastly, Vanderburgh and Roberts (1996) proposed a proglacial lake receiving sediment from tributary valleys and possibly from an actively retreating valley ice tongue. Elucidating lake type and the location of sediment sources for gLP bears directly on our understanding of ice distribution in and around the lake basin. In turn, this has the potential to help refine understanding of CIS deglacial processes and patterns.

1.5.3.2.2. Origin of the White Silt benches

Based on various lake type interpretations, explanations have been advanced to account for the distribution of White Silt benches. For example, in an ice lateral setting, the distribution of White Silt benches is rationalized through localized deposition of lake bottom sediments on the valley walls (and non-deposition along the valley axis) (Flint, 1935b; Nasmith, 1962). Additionally, White Silt benches have been ascribed to collapse of a supporting ice tongue/blocks associated with a supraglacial lake (Flint, 1935b; Shaw, 1977; Eyles et al., 1990, 1991). Erosion of lacustrine sediments along the valley during the later stages of deglaciation is the most plausible explanation for a proglacial lake basin with little to no stagnant ice. Yet, this possibility has received little consideration, save for Kvill (1976) who tentatively raised this possibility without elaborating on the exact processes.

Thus, despite a general understanding of gLP paleoenvironment, various and potentially conflicting interpretations of the gLP paleoenvironment are currently proposed. A regional synthesis, critically evaluating existing hypotheses and datasets, combined with new paleogeographic and paleoenvironmental data from the gLP basin is required in order to refine understanding of gLP's paleoenvironment, and regional understanding of deglacial processes and patterns.

1.5.3.3. Chapter objectives

This chapter provides a paleogeographic and paleoenvironmental reconstruction of gLP in order to better constrain deglacial processes and patterns within this glacial lake basin.

Specifically, the research objectives are:

1. to document gLP paleogeography by focusing on the geographic extent of gLP in Okanagan Valley through regional water plane reconstruction;

2. to assess the evidence for glacioisostatic rebound in Okanagan Valley and examine the glaciodynamic conditions responsible for glacioisostatic rebound patterns;

3. to examine the spatial evolution of gLP, focusing on its volume and the temporal changes in lake extent; and

4. to reconstruct paleoenvironmental conditions in and around the gLP basin by examining water and sediment delivery pathways to gLP in order to better constrain lake type, the distribution of ice in and around the gLP basin, and the style of lake drainage and its effects within the lake basin.

1.6. Thesis format

This thesis is written in paper format. As such, each substantive chapter is a stand-alone article. The content of some chapters sets the stage for subsequent chapters. Consequently, there is some necessary repetition of information between chapters. Nonetheless, each substantive chapter addresses specific research questions and presents original datasets. Three substantive chapters are presented. Chapter 2: *A record of linked cavity operation beneath a portion of the Cordilleran Ice Sheet, southern Interior, British Columbia, Canada* discusses subglacial conditions and substrate genesis on Thompson Plateau, British Columbia. Chapter 3: *Regional reconstruction of subglacial hydrology and glaciodynamic behaviour along the southern margin of the Cordilleran Ice Sheet in British Columbia, Canada and northern Washington State, USA,* examines landscape genesis and the genesis of streamlined bedforms (drumlins) and other glacial landforms in Okanagan Valley and Thompson Plateau. Chapter 4: *Paleogeographic and paleoenvironmental evolution of glacial Lake Penticton during decay of the Cordilleran Ice Sheet, British Columbia, Canada* examines the paleoenvironment and paleogeography of glacial Lake Penticton. Finally, Chapter 5 summarizes the salient points of this research project and presents some future avenues of research.

1.6.1. Published thesis chapter

A version of Chapter 3 is a published article: Lesemann J-E, Brennand TA. 2009. Regional reconstruction of subglacial hydrology and glaciodynamic behaviour along the southern margin of the Cordilleran Ice Sheet in British Columbia, Canada and northern Washington State, USA. Quaternary Science Reviews 28: 2420-2444.

I was responsible for all data collection, analysis, and synthesis. I was also responsible for manuscript writing and associated figure production. My co-author contributed to analysis and manuscript writing through to conceptual discussions, organization of concepts and article structure, and editing.

2. A record of linked cavity operation beneath a portion of the Cordilleran Ice Sheet, Southern Interior, British Columbia, Canada

Abstract

Subglacial bed conditions of the former Cordilleran Ice Sheet (CIS) remain largely unexplored. This paper examines substrate genesis and subglacial bed conditions under part of the CIS on the southern Thompson Plateau, British Columbia, Canada. Two diamicton lithofacies (*consolidated* and *poorly consolidated*) and a group of waterlain lithofacies (*sorted and stratified sediments*) are identified in a ~600 km^2 topographic basin bounded by a prominent bedrock ridge along a portion of its perimeter. *Consolidated diamicton* occurs on the flanks of this bedrock ridge. Its macroscale sedimentology, geologic structures, and stone fabrics and surface wear features suggest that it is a hybrid or a subglacial traction till (Evans et al. 2006) recording multiple formative processes, including lodgement, ploughing and varying degrees of ductile and brittle deformation, due to fluctuations in substrate porewater pressure. In contrast, *poorly-consolidated diamictons* occur within the regional basin. They are characterized by low strength and low silt and clay content in the diamicton matrix. They are well to poorly stratified and contain soft-sediment rip-up clasts. These diamictons are likely reworked tills with a complex polygenetic history dominated by deformation, remobilization by debris flows, melt-out and possibly elutriation of fines from the till matrix by poreflow and in association with dilatancy. *Sorted and stratified sediments* consist of laminated and bedded silt, clay and sand. They are interbedded with poorly-consolidated diamictons which suggests coeval subglacial deposition of these lithofacies within water bodies. These water bodies could have been proglacial ponds and/or subglacial cavities A conceptual model emphasizes the operation, evolution and reorganization of proglacial ponds and subglacial water-filled linked cavities and accounts for the documented sedimentary facies through episodic reorganization of cavities. The operation of linked cavities may have facilitated ice transfer, yet need not have resulted in regional fast flow.

2.1. Introduction

Processes operating at the ice-bed interface exert significant control on the dynamics of glaciers and ice sheets (cf. Bennett, 2003 and references therein). In particular, meltwater production, storage, and drainage affects glacier flow by controlling sediment dilatancy, and episodes of enhanced flow such as surging (Kamb, 1987) and ice streaming (Alley, 1989; Kamb, 1987, 1990, 2001; Fricker et al., 2007) during reorganizations of subglacial drainage systems. Subglacial water storage within cavities occurs at a range of scales, including alpine-scale linked-cavity systems (m-10s m scale) and subglacial lakes beneath ice sheets (100s m- km scale) (Siegert and Bamber, 2000; Gray et al., 2005; Fricker et al., 2007). However, there are significant knowledge gaps concerning depositional processes within subglacial cavities of various dimensions under modern ice sheets (including subglacial lakes) and in paleo-ice sheets. Exploratory seismic studies have shed light on basin-scale architectural elements of modern subglacial lakes such as Lake Vostok (Christner et al., 2006 and inferred subglacial lakes of the last Laurentide ice Sheet (Christofferson et al., 2008), both studies reveal the presence of numerous parallel reflections suggestive of rhythmic sedimentation. Integration of understanding of subaqueous depositional processes in glacial settings has led to elaboration of a theoretical framework for recognition of paleo-ice sheet subglacial lakes (Livingstone et al., 2012). Such a framework offers important guiding conceptual elements but remains largely speculative in the absence of corroborating field data and regional integration of landform evidence for subglacial lakes.

Recognition of active and dynamic subglacial hydrologic systems within modern ice sheets also implies that subglacial water may be stored in cavities smaller than those currently documented (e.g. Fricker et al., 2007; Gray et al., 2005). Many such cavities may not be easily detected due to limitations in the resolution of remote sensing and geophysical instruments typically used to identify sites of subglacial meltwater storage. As a result, there is currently no direct knowledge of the existence of metre to 100's metre-scale cavity systems that may exist under modern ice sheets. However, subglacial cavities of similar-scale are known to affect glacier dynamics within alpine glacial systems (e.g. Kamb, 1987) and they potentially play a role in the dynamics of larger glaciers.

Knowledge gaps about the existence of such cavities under modern ice sheets also imply that little is known about depositional processes and environments within these types of subglacial cavities. Examining this scale of water storage beneath paleo-ice sheets may shed light on modern subglacial ice-sheet processes at spatial scales not currently resolved by studies of modern ice sheets.

Limited understanding of subglacial conditions under the former Cordilleran Ice Sheet (CIS), particularly on plateaus of southern British Columbia., stems from the exploratory nature of most existing studies on subglacial landforms and sediment there, and a focus on deglacial processes and patterns (e.g., Fulton, 1967). Notable exceptions include Lian and Hicock (2000) and Lian et al. (2003) who reconstructed CIS thermal regime evolution and inferred fast flow from till sedimentology, stone fabric and surface wear data and geologic structures for Jesmond Valley (Fig. 2.1 inset) in the interior of British Columbia. They reported evidence for ductile and brittle sediment deformation related to temporal changes in bed thermal regime and substrate water content. Broster and Clague (1987) analyzed geologic structures in tills, inferring the dynamics and flow direction of the ice sheet near Williams Lake (Fig. 2.1 inset). They found that during advance of the ice sheet, basal ice was at least temporarily frozen to the bed, forming brittle deformation structures (faults, joints and fractures). At some point, subglacial meltwater infilled these structures with sorted sediment, suggesting that the temperature reached pressure melting point. A converse switch in thermal regime is indicated near Big Creek (Fig. 2.1 inset) where sediments record a sequence of ductile, brittle and semi-brittle deformation as ice advanced over saturated sediments before freezing to the bed (Huntley and Broster, 1993).

This paper aims to reconstruct paleo-CIS bed conditions and examines the role of subglacial hydrology and its controls on subglacial processes of southern Thompson Plateau (Fig. 2.1). This study area is chosen for two reasons: (1) little is known about subglacial processes on plateaus flanking major valley systems of the southern Cordillera, as most studies have focused on deglacial patterns and processes (Fulton, 1967); (2) exposures are available within drumlins forming an extensive regional tract.

Examination of sedimentary exposures on this portion of Thompson Plateau provides complementary data to a regional study of the Thompson Plateau landsystem (Chapter 3, Lesemann and Brennand, 2009). Together, these data allow for a regional assessment of subglacial processes. Furthermore, the spatial continuity of landforms (valley networks, drumlins – Chapter 3, Lesemann and Brennand, 2009) between Thompson Plateau and Okanagan Valley raises the possibility of examining the connectivity of regional landsystems associated with major Cordilleran Valleys and possible glaciodynamics such as ice streams (cf. Matthews, 1991; Kleman and Glasser, 2007). Process reconstructions and paleo-CIS bed conditions are reconstructed from glacigenic sediments exposed within drumlins (elongation ratios between 1.5:1 - 8.1:1, Chapter 3, Lesemann and Brennand, 2009) that have an inferred erosional origin based on (1) their similar morphology yet variable composition—including bedrock, (2) the presence of truncated strata, and (3) landform association and sediment conservation (Chapter 3, Lesemann and Brennand, 2009). The regional sedimentology, stone fabric and surface wear features, and geologic structures at sites reported in this study (which include exposures in drumlins, between drumlins and in non-drumlinized zones within the drumlinized tract) support this interpretation.

2.2. Study area

The study area, on the southern portion of Thompson Plateau in the southern Interior of British Columbia, Canada (Fig. 2.1), lies ~200 km north of the southern Late Wisconsinan CIS margin (located in northern Washington State, USA) (Figs. 1.1, 2.1). Synoptic ice flow reconstructions suggest generally southward flow over the area toward this Late Wisconsinan margin (Clague, 1989). The Fraser River and Okanagan valleys bound the Thompson Plateau to the west and east respectively. Thompson River and Similkameen River valleys mark its northern and southern limits, respectively (Holland, 1964). Smaller north-south and east-west trending valleys (e.g. Nicola Valley) dissect Thompson Plateau. Within the study area maximum elevations reach 2300 m asl on isolated bedrock summits but generally the land lies between 1200 m asl and 1900 m asl. Topography varies along an extensive (~ 250 km long and ~80 km wide), north-south trending swarm of streamlined bedforms (mainly drumlins) terminating at the Similkameen River (Fig. 2.2; Lesemann and Brennand, 2009). Near the southern swarm margin, flow deviates to the southeast (Fig. 2.2) and crosses a regional topographic depression forming a basin on the eastern edge of Thompson Plateau (Fig. 2.3). This basin rises gradually toward a prominent bedrock ridge (Trepanege Ridge) marking the south-eastern perimeter of the regional depression (Fig. 2.3). Bedrock geology within the study area consists of Triassic to Jurassic age granodiorite outcrops (Monger and Lear, 1989a, b). Trepanege Ridge contain localized outcrops of Triassic volcanic and sedimentary rocks (argilite, sandstone and conglomerate), and Eocene volcanic rocks (Monger and Lear, 1989a, b) (Fig. 2.4). Fulton (1975) performed reconnaissance-level mapping of surficial material in the area. He recognized the presence of drumlinized ‘morainal deposits’ (till with minor amounts of sand and gravel, and silt, Fulton, 1975). These sediments cover bedrock surfaces, save for prominent topographic highpoints where bedrock crops out. Mean regional sediment thickness in this basin is unknown but localized observations indicate that sediment thickness varies from a few centimetres near outcrops to over 10 m in places. Sediments described herein occur within this basin and along the flanks of Trepanege Ridge. Late-Wisconsinan synoptic ice-flow reconstructions in this area depend mainly on drumlin and striae orientations — the inferred regional ice-flow direction for Thompson Plateau being ~north-south (Clague, 1989), and northwest-southeast more locally near Trepanege Ridge (Fulton, 1967, 1975).

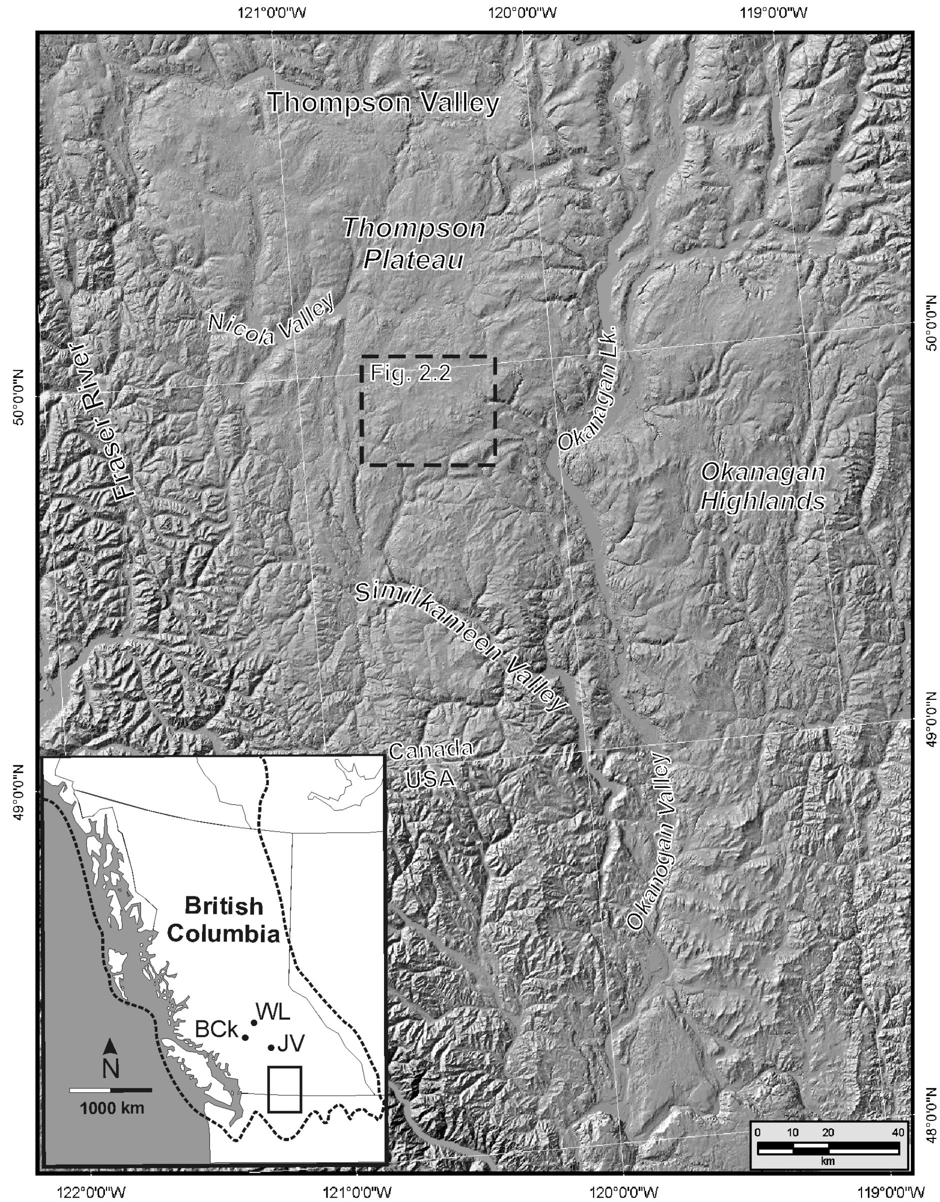


Figure 2.1. Thompson Plateau and surrounding areas, British Columbia, Canada. Hillshaded digital elevation model reveals the major physiographic features of south-central British Columbia. Inset map: BCk: Big Creek, WL: Williams Lake, JV: Jesmond Valley. Dashed rectangle is field area shown on Figure 2.3. (Digital Elevation Model: 25 m grid resolution, TRIM I dataset, British Columbia Albers Conical Equal Area projection, NAD 83 datum, © GeoBC, by permission).

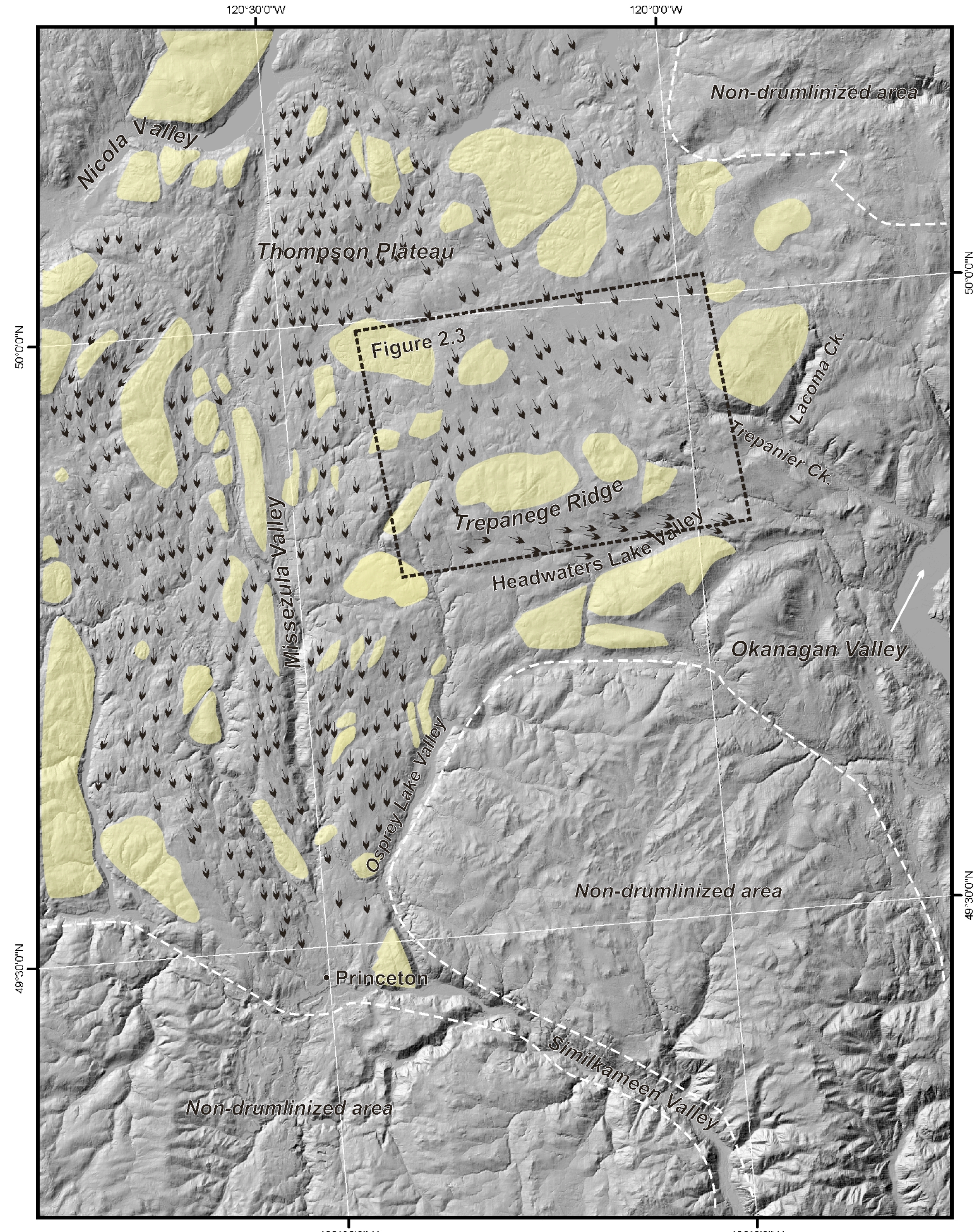


Figure 2.2: ***Flowline map of Thompson Plateau based on mapping of prominent drumlins (see Chapter 3 for mapping methodology). An extensive N-S-trending drumlin swarm bifurcates SE toward Trepanege ridge and crosses a regional topographic basin bounded by Trepanege Ridge. (Digital Elevation Model: 25 m grid resolution, TRIM I dataset, British Columbia Albers Conical Equal Area projection, NAD 83 datum, © GeoBC, by permission).***

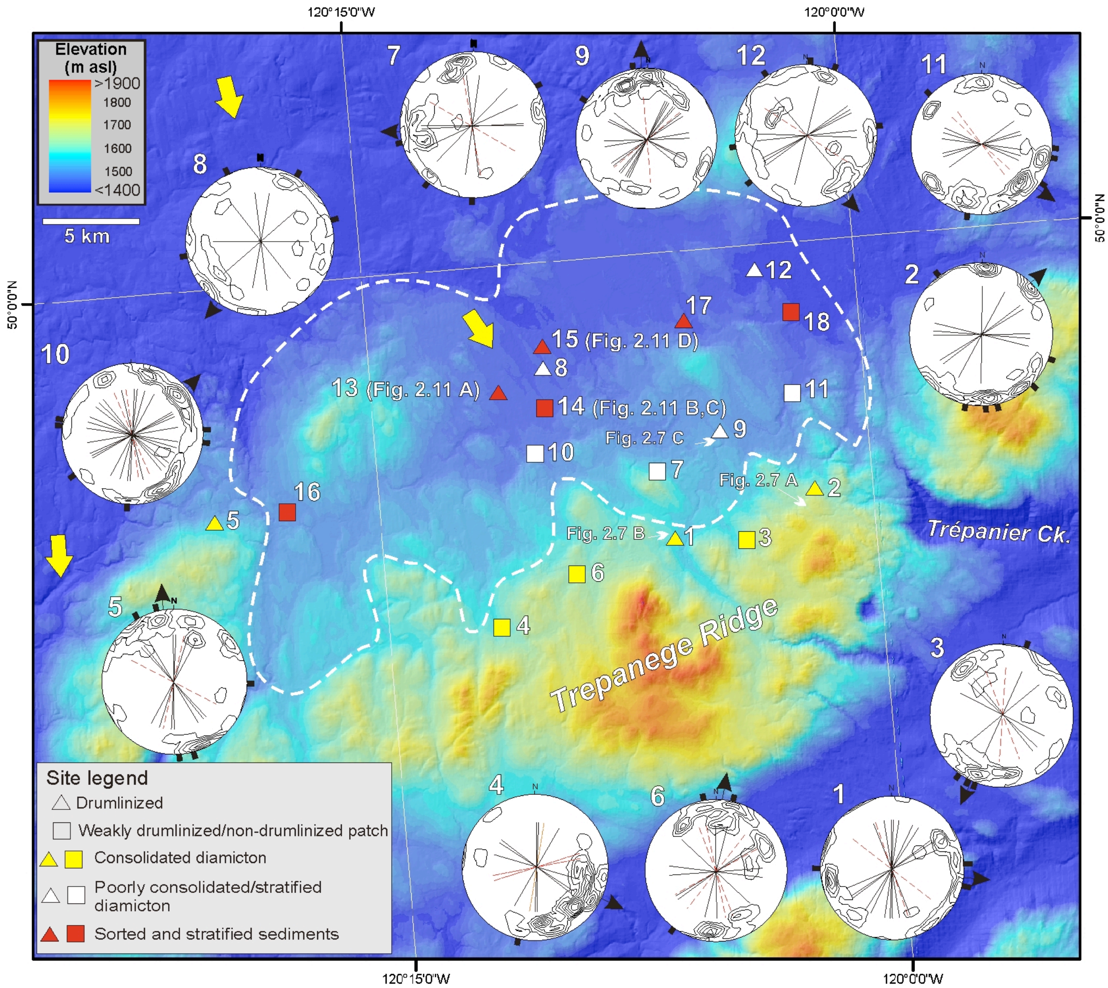


***Figure 2.3:** Regional topographic basin (white dashed line) on Thompson Plateau (refer to Fig. 2.1 for location) bounded by Trepanege Ridge to the south and west, and by more subtle bedrock highpoints to the north and east (Digital Elevation Model: 25 m grid resolution, TRIM I dataset, British Columbia, © GeoBC, by permission). Albers Conical Equal Area projection, NAD 83 datum. Synoptic Late Wisconsinan ice flow directions (Fulton 1975) indicated by yellow arrows. Field sites are numbered and categorized according to their geomorphological and sedimentological characteristics. Stereo plots highlight regional till clast fabric patterns in relation to topographic features (refer to Figs. 2.5 and 2.6 for symbol explanations). Site details are given in Appendix A.*

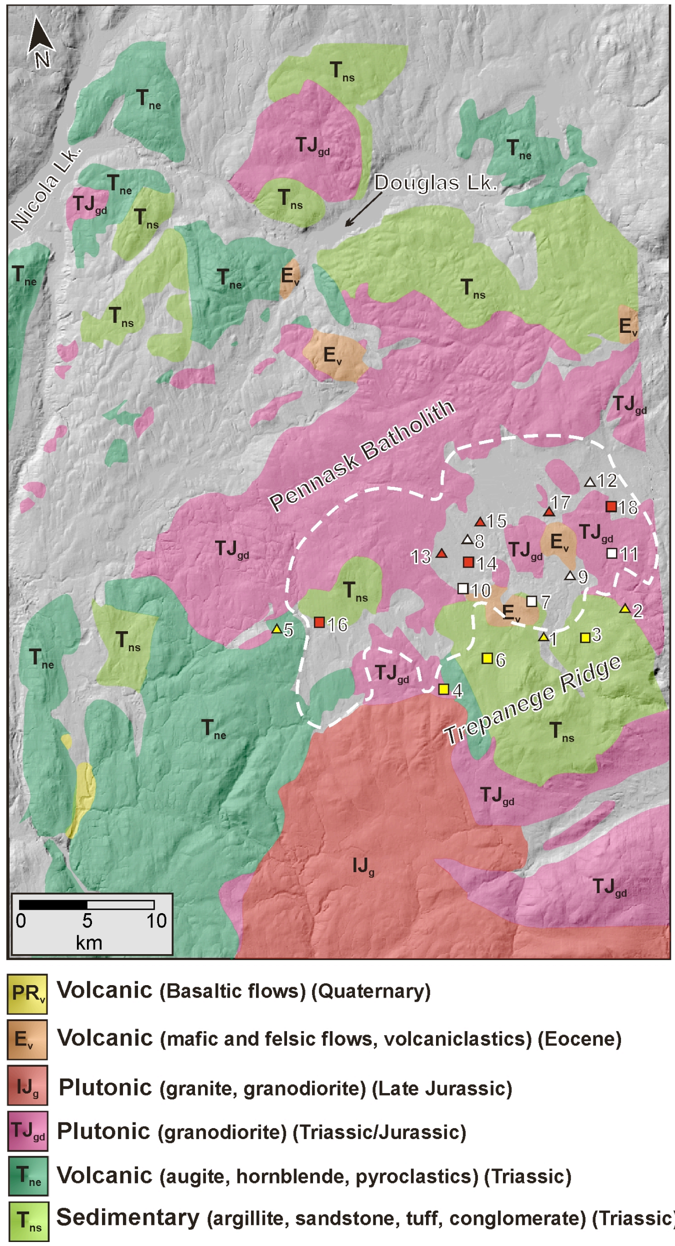


***Figure 2.4:** Bedrock map for the southern portion of Thompson Plateau highlighting the distribution of plutonic rocks (Pennask Batholith) at the northern margin of the regional basin (dashed white line) and south of Trepanege Ridge. Plutonic, volcanic and sedimentary rocks outcrop between Douglas Lake and the northern margin of the topographic basin. These rocks also outcrop within the basin and on the northern flanks of Trepanege Ridge. Refer to Fig. 2.3 for site symbology. Geology from Monger and Lear (1989 a, b). Base map: Digital Elevation Model: 25 m grid resolution, TRIM I dataset, British Columbia, © GeoBC (by permission). Albers Conical Equal Area projection, NAD 83 datum.*

2.3. Methodology

2.3.1. Landform mapping and sedimentologic analysis

Landform mapping focused mainly on regional-scale patterns of drumlins and prominent channels. Individual drumlins and groups of drumlins were mapped from 25 m pixel resolution digital terrain models and from aerial photographs. This forms the basis for producing generalized flowline maps (Fig. 2.2). Field truthing was used to verify some of the mapping (local-scale mapping). Data sources used for mapping depended on the intended purpose of the map. For example, regional-scale landform domains were identified using DEM. Within these domains, flowlines were established from mapping prominent drumlin crests in order to extract landform orientation. Landform direction was inferred from drumlin morphology where appropriate. Not every drumlin was mapped within each domain since the goal was to build a synoptic-scale representation of landform patterns

Sedimentary processes are reconstructed by combining stone fabric and surface wear features with macroscale sedimentology and geologic structures. Additional information on subglacial processes could be gleaned from microstructures and microfabrics extracted from thin sections (e.g., van der Meer 1993; Menzies 2000). However, such analysis lies beyond the scope of this article, which relies on macro-scale evidence to infer subglacial processes. Three lithofacies were recognized: consolidated diamicton, poorly-consolidated diamicton, and sorted and stratified sediment. Consolidation class is used as a first-order classifier of material assigned on the basis of the relative ease by which the diamicton could be excavated with a trowel and shovel: poorly-consolidated diamicton was friable, required little effort to excavate, and frequently collapsed under its own weight when undercut. In contrast, consolidated diamicton was more consolidated and compact, easily resisting collapse when undercut. Sites were documented using standard sedimentary descriptions (texture, sedimentary structure, bed thickness, nature of bed contacts, etc.). Sediment samples were obtained from road cuts which were cleared and cleaned with shovel and trowel. Sedimentary and other geological structures were finely cleaned with a knife and soft brush. Dry and wet sieving and Sedigraph analysis provided grain size data for the diamicton matrix (grains<2mm) (Appendix B) from samples taken adjacent to fabric sites. Diamicton textures are described according to the USDA textural classification system (Soil Survey Staff 1999). The modifiers stony and pebbly for diamictons are used for clast content > 20%.

2.3.2. *Till clast fabrics and wear features*

Within diamictons, the trend and plunge of at least 30 pebble a-axes within an area of 0.4-0.7 m^2 were measured with a compass (including clinometre). The type, nature and location of striae, keels and plucked ends on these stones (cf. Lian et al., 2003) were recorded (Tables 2.1, 2.2). Only pebbles with a:b axis ratios between 2:1-3:1 were measured, and as much as possible, clasts in the vicinity of cobbles and boulders were avoided (cf. Kjaer and Krüger, 1998). Linear stone a-axis data were plotted on Schmidt nets as lower hemisphere, equal area projections (Figs. 2.5, 2.6). Fabric data is contoured at multiples of a random distribution (1.5 σ from an assumed random distribution based on the method of Kamb (1959). Clast symbols identify modes or outliers on scatter plots (Hicock et al., 1996) (Figs. 2.5, 2.6). Fabric data were analysed using the orientation tensor method (Mark 1973) (Figs. 2.5, 2.6; Tables 2.1, 2.2). A solid arrow denotes the principal eigenvector for the bulk sample, and open arrows denote statistically significant modal eigenvectors at the 97% confidence level (calculated using the Woodcock and Naylor (1983) method) (Figs. 2.5, 2.6). Clast wear features (see § 2.4 for definitions) are also plotted on the stereograms and include: the orientation of the youngest, most delicate striae on clast tops, keels on clast bottoms (i.e. in a ploughing position, cf. Krüger (1984)), and the direction of angular plucked ends (Figs. 2.5, 2.6). The strike and dip of planar geologic structures (e.g., fractures, shears) were measured and plotted as poles to planes on lower hemisphere Schmidt nets (Figs. 2.5, 2.6).

Clast wear features have been used to constrain interpretations of till depositional history and to refine subglacial process reconstructions. Clast wear features are defined in order to establish their use in process reconstructions. *Striae* are abrasion marks produced by clast-clast contact and prolonged or intermittent dragging of tools against a bedrock surface or a neighbouring clast. In the context of till genesis, striae may result from abrasion due to sliding of debris-rich ice and/ or due to dragging of clasts within basal sediments. As part of lodgement processes, plucking and abrasion are expected is response to a glacier sliding over its bed. Clasts can develop up-glacier-dipping tapered ends in response to abrasion (so-called *bullet-shaped* clasts). Clasts may also exhibit fractured ends on the upper-side of down-glacier-dipping surfaces (the term *plucked end*, although interpretive, is also used to describe fractured ends, e.g. Lian, 1997; Liand and Hicock, 2000) develop as a result of plucking. If prolonged ploughing and clast dragging occurs, clast irregularities may be smoothed and a boat-shaped bottom surface (*keel*) may develop (Lian, 1997; Lian and Hicock, 2000). Because abrasion accompanies clast-dragging, flow-parallel striae may develop on keel surfaces (Lian, 1997, Lian and Hicock, 2000). If substrate deformation occurs following clast lodgement, clast may be remobilized and reoriented. Consequently, wear features may occur in locations of positions that are no longer consistent with those expected of lodgement processes. For example, *upturned keel*s consist of boat-shape clast surface occurring in a non-ploughing position (Lian and Hicock, 2001). Brittle deformation structures associated with lodgement till and the stress of overrifing ice include fissility, down-glacier-dipping tension fractures, and upglacier-dipping shear planes (cf. Hicock et al., 1996).

The type, nature and location of striae, keels and fractured ends (cf. Lian et al., 2003) were recorded to assess evidence of substrate deformation. Clast wear features plotted on Schmidt plots include the orientation of the youngest, most delicate striae on clast tops (assumed to record the latest events, Lian, 1997), keels on clast bottoms (i.e. in a ploughing position, cf. Krüger (1984)), and the direction of angular fractured ends (Figs. 2.5, 2.6). Additional information on subglacial processes could be gleaned from microstructures and microfabrics extracted from thin sections (e.g. van der Meer 1993; Menzies 2000). However, such analysis lies beyond the scope of this article, which relies on macro-scale evidence to infer subglacial processes.

Table 2.1 The character and genesis of consolidated diamicton

Site	Stone fabric data[1] and landform context	Unit sedimentology and stone surface wear features	Inferred genetic processes[2]
1	*n* =30 *S_1=0.63, S_3=0.01* Spread unimodal, 1 outlier	Fissile loamy diamicton (~10-15% clast content) containing fractures and numerous contorted and deformed silt stringers between more massive, clast-rich diamicton intraclasts (~20% clast content). Attenuated silt laminations around clasts and diamicton intraclasts. Fractures cross-cut stringers but do not extend into intraclasts.	*Lodgement, sediment deformation within a dilatant bed and its subsequent collapse including porewater escape. Compression against Trepanege Ridge. Fracturing due to porewater pressure drop.*
	Drumlin tail, east of crestline	Keels, plucked ends and some striae aligned with spread mode.Plucked ends exhibit no preferred direction. Multiple sets of a-axis parallel striae on stones. Single b-axis parallel striations on mode-transverse outlier.	
2	*n* =30 *S_1=0.65, S_3=0.03* Spread unimodal, 1 outlier	Massive loamy diamicton (~25% clast content) with minor textural heterogeneities. Contains angular to subangular pebbles and cobbles. Most consolidated of all sites.	*Possible lodgement, overprinting by deformation processes due to compression against Trepanege Ridge.*
	Drumlin nose, west of crestline	Most striae aligned with clast a-axis but some aligned with b-axis and some curve around clast edges. Plucked ends transverse to principal eigenvector with most pointing to the SE. No keels observed.	
3	*n* =30 *S_1=0.62, S_3=0.01* Spread bimodal	Fissile loamy diamicton containing intraclasts of more consolidated diamicton with subangular to subrounded pebbles (~15% clast content). Heterogeneous diamicton bodies often attached to angular cobbles. Matrix contains discontinuous laminated fine sand and silty clay. Edges of heterogeneous bodies/clasts intrude into and deform the finer laminations.	*Lodgement, sediment deformation within a dilatant bed and its subsequent collapse including porewater escape. Compression against Trepanege Ridge. Fracturing due to porewater pressure drop.*
	Weakly drumlinized patch on slight topographic rise	Striae aligned with both modes. Keels aligned with primary mode. Most plucked ends point in the direction of the primary mode. Some striae curve around clast edges. Two keels in non-ploughing position.	

Site	Stone fabric data[1] and landform context	Unit sedimentology and stone surface wear features	Inferred genetic processes[2]
4	*n* =32 *S_1=0.72, S_3=0.06* Spread unimodal Low flanks and trough of weakly drumlinized patch downflow of local bedrock ridge	Massive sandy diamicton (~ 15% clast content) containing subrounded clasts and more angular locally-derived clasts. Matrix dominated by coarse sand to granule-size grus from deeply-weathered local outcrops. Silt and clay less abundant than in the other consolidated diamictons. This diamicton is the least consolidated of all consolidated diamicton sites. Fractures are prominent. Some fissility may also occur. Striae on subrounded clasts aligned with spread mode and some clasts exhibit curved striae around edges. Few striae on angular, locally-derived clasts. Keels and plucked end aligned transverse to spread mode. Single keel in non-ploughing position.	*Lodgement, ductile deformation, some brittle deformation due to coarse texture and compression against Trepanege Ridge.*
5	*n* =30 *S_1=0.74, S_3=0.04* Spread bimodal Drumlin flank toward nose	Compact and fissile loamy diamicton (~20-25% clast content). Weak shears with orientations consistent with shearing force from NNW (i.e. regional ice flow direction). Striated clasts with keel-like bottoms and plucked ends mainly aligned with primary mode; plucked ends show no preferred direction. Weak secondary transverse mode trending ENE-WSW. No plucked ends or striae aligned with secondary mode.	*Mainly lodgement, ductile deformation, shearing as bed drained.*
6	*n* =30 *S_1=0.65, S_3=0.09* Spread unimodal Weakly drumlinized patch	Compact loamy diamicton (~15-20% clast content) containing pebbles and cobbles. Some localized fissility. Discontinuous (5-12 cm long) but generally undisturbed silt and silty fine sand stringers occur within matrix-rich areas. Low-angle fractures filled with fines occur near the base of the exposure. Striated and keeled clasts, some with plucked ends. Orientation of wear features reflects spread in fabric. Single keel in non-ploughing position.	*Lodgement, sediment deformation within a possible dilatant bed and its subsequent collapse including porewater escape. Compression against Trepanege Ridge.Fracturing due to porewater pressure drop.*

[1]n=number of measured stones, S=normalized eigenvalues for all clasts in sample (modality calculations not presented), modality terminology after Hicock *et al.* (1996). [2]Refer to text for explanation.

Table 2.2. The character and genesis of poorly-consolidated diamicton

Site	Stone fabric data[1] and landform context	Unit sedimentology and stone surface wear features	Inferred genetic processes[2]
7	n=30 S_1=0.55, S_3=0.08 Spread bimodal[3]	Gently inclined, weakly stratified pebbly sandy diamicton beds (~5-10% clast content), separated by discontinuous silty sand beds (<1cm thick). Bed contacts are gradational and often obscured by protruding outsized clasts from overlying and underlying beds.	*Lodgement. Possible local bed dilatancy and redeposition from gravity flows mobilizing illutriated till.*
	Non-drumlinized patch, ~500 m upflow and lateral to weakly drumlinized topopgraphic high.	Striated subangular to subrounded clasts. Keels, plucked ends and striae generally aligned with clast a-axes in both modes. Two striae aligned with clast b-axis in primary mode and single b-axis aligned striae in secondary mode.	
8	n=33 S_1=0.57, S_3=0.11 Girdle-like Drumlin crestline, halfway along drumlin length.	Massive to weakly stratified stoney sandy silt diamicton (~5-10% clast content). Clear beds are not visible but 1-3 cm-thick, discontinuous zones of better sorted sand create faint stratification. Stratified zones grade into the sandy diamicton. Partially deformed and attenuated intraclasts (rip-ups or 'till pebbles') of laminated sandy silt and silty clay occur within the diamicton. More coherent bodies of better sorted silty fine sand also occur but do not show extensive attenuation.	*Lodgement, elevated porewater pressure leading to bed dilatancy accompanied by washing and illutriation of till. Resedimentation by gravity flows.*
		Clasts are dominantly angular with few striae and plucked ends. Striae and plucked ends show no preferred alignment. No keels observed.	
9	n=30 S_1=0.53, S_3=0.07	Weakly stratified sandy diamicton with ~20% clast content. Stratification is only visible due to slight textural variations near bed boundaries. A normal fault occurs through the majority of the sediment thickness.	*Lodgement. Possible local bed dilatancy and redeposition from gravity flows mobilizing washed and illutriated till. Brittle deformation following bed dewatering.*
	Spread bimodal[3], 3 outliers Drumlin crestline, one third along drumlin length.	Angular and subrounded clasts with striae curved around worn edges. Keels, plucked ends and most striae aligned with a-axis of clasts of primary mode. Four striae aligned with two outliers, including two striae aligned with b-axes. Two keels in non-ploughing positions.	

Site	Stone fabric data[1] and landform context	Unit sedimentology and stone surface wear features	Inferred genetic processes[2]
10	n=31 S_1=0.43, S_3=0.05 Girdle-like	Steeply inclined beds of pebbly sand diamicton interstratified with gravely sand diamicton (~20-25% clast content). Gravely sand diamicton contains cobbles and pebbles exhibiting slight imbrication in clast clusters. Silty clay concentrations occur at bed boundaries.	*Lodgement. Possible local bed dilatancy and redeposition from gravity flows mobilizing washed and illutriated till.*
	Weakly drumlinized patch	Angular to subangular clasts exhibiting abundant striae, some keels and plucked ends. Surface wear features show no preferred alignment.	
11	n=30 S_1=0.58, S_3=0.1 Spread unimodal	Interstratified pebbly sandy silt diamicton and pebbly silty clay diamicton beds (~ 20% clast content). Sandy silt diamicton beds contains fewer clasts and have irregular, gradational contacts with finer diamicton beds. Silty clay diamicton is fissile.	*Lodgement. Possible local bed dilatancy and redeposition from gravity flows mobilizing washed and illutriated till.*
	Non-drumlinized zone marginal to weakly drumlinized patch	Both diamicton types contain subangular to subrounded pebbles and small cobbles. Keels and plucked ends aligned with principal eigenvector. Most fractured ends point in the direction of the principal eigenvector. Striae oriented transverse/oblique to principal eigenvector.	
12	n=30 S_1=0.58, S_3=0.15 Girdle-like	Massive to very poorly stratified partially homogenized sand and silty sand diamicton (~5-10% clast content). Texture is mainly sandy but includes mm-scale sandy silt bodies dispersed throughout. Deformed intraclasts of laminated silt and sandy silt occur within diamicton. Small raft edges exhibit deformation and partial homogenization with surrounding diamicton matrix. Larger intraclasts retain internal laminated structure. Fractures.	*Lodgement and eluviation of fines followed by localized deformation possibly involving gravity flows. Brittle fracturing as bed dewaters.*
	Flute superimposed on crestline of drumlinized patch	Subrounded clasts exhibit striae, plucked ends and one keel. Surface wear features show no preferred alignment.	

[1]n=number of measured stones, S=normalized eigenvalues for all clasts in sample (modality calculations not presented), modality terminology after Hicock *et al.* (1996). [2] Refer to text for explanations. [3] Second mode not statistically-significant at 97% confidence interval (calculated using method of Woodcock and Naylor (1983)).

2.4. Criteria for till classification and premise of process reconstructions

2.4.1. Criteria for till classification

Subglacial tills have traditionally been classified based on the characteristics of the final deposit and within a tripartite scheme anchored by distinct end-members (lodgement, deformation, meltout (Dreimanis,1989). However, it is increasingly recognized that the vast majority of tills result from processes occurring along continua between these end-members. The use of this classification requires that observations be made and compared to the sedimentologic and clast morphologic characteristics predicted by each end-member in order to adequately represent the complex origins of tills and the wide-ranging interplay of processes responsible for producing the macro- and micro-scale characteristics of these deposits (e.g. Dreimanis, 1989, Hicock 1993, Hicock et al., 1996, Evans et al. 2006). Consequently, resulting reconstructions highlight the fact that the vast majority of tills are polygenetic and time-integrated products of subglacial processes. To address some of the practical difficulties of classifying tills within existing schemes, Evans et al. (2006) proposed the term ‘traction till’, which shares some conceptual notions with the concept of ‘hybrid till’ (Hicock, 1993). Both schemes recognize the “co-existence of various subglacial processes that act to mobilize and transport sediment, and deposit it as various end-members” (Evans et al. 2006).

If till sedimentology and clast wear characteristics are to be used to gain better understanding of till genesis (and resulting classifications) and the complexity of subglacial processes, then there is also a need to recognize the potential equifinality of some subglacial processes. For example, observed clast wear features such as striae aligned with the clast a-axis and bullet–shapes, plucked ends, and keels in lodged position, combined with structural data give rise to a lodgement interpretation (Lian and Hicock, 2000). However shear zones in plastically-deforming tills within ring-shear devices produce a-axis parallel striae (Thomason and Iverson, 2008). Similarly, Benn (1995) argued that many of the wear features associated with lodgement can be produced within a deforming bed setting. These results highlight the difficulty in

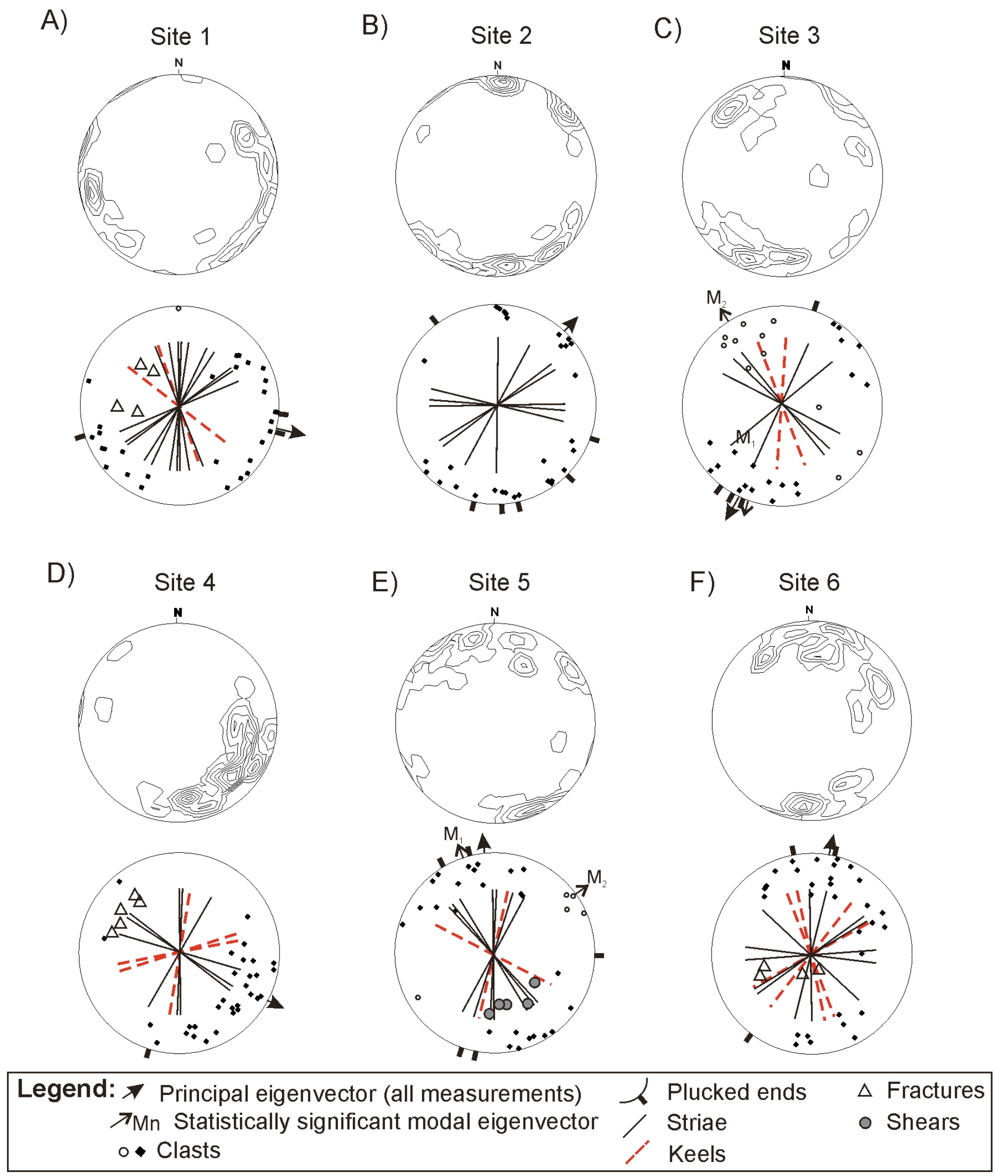


Figure 2.5. Equal-area lower hemisphere Schmidt plots of shears, fractures, and stone fabric and surface wear features from consolidated diamictons. At each site, contoured and scatter plots are provided. Only the most recent (and/or delicate) striae on clast tops, keels on clast bottoms, and angular plucked ends are displayed on point plots. Shears and fractures are plotted as poles to planes. Distinct clast symbols (open and closed circles) are used to distinguish visually separated modes and outliers (cf. Hicock et al., 1996). Separate arrows denote modal eigenvectors (Mn) where modes are statistically significant at 97% confidence level (calculated using Woodcock and Naylor (1983) method). Contour interval: 1.5 σ from an assumed random distribution (method of Kamb 1959). Fabric statistics (after Marks 1973) and additional details on clast shape and wear features are given in Table 2.1. Data are presented in Appendix C.

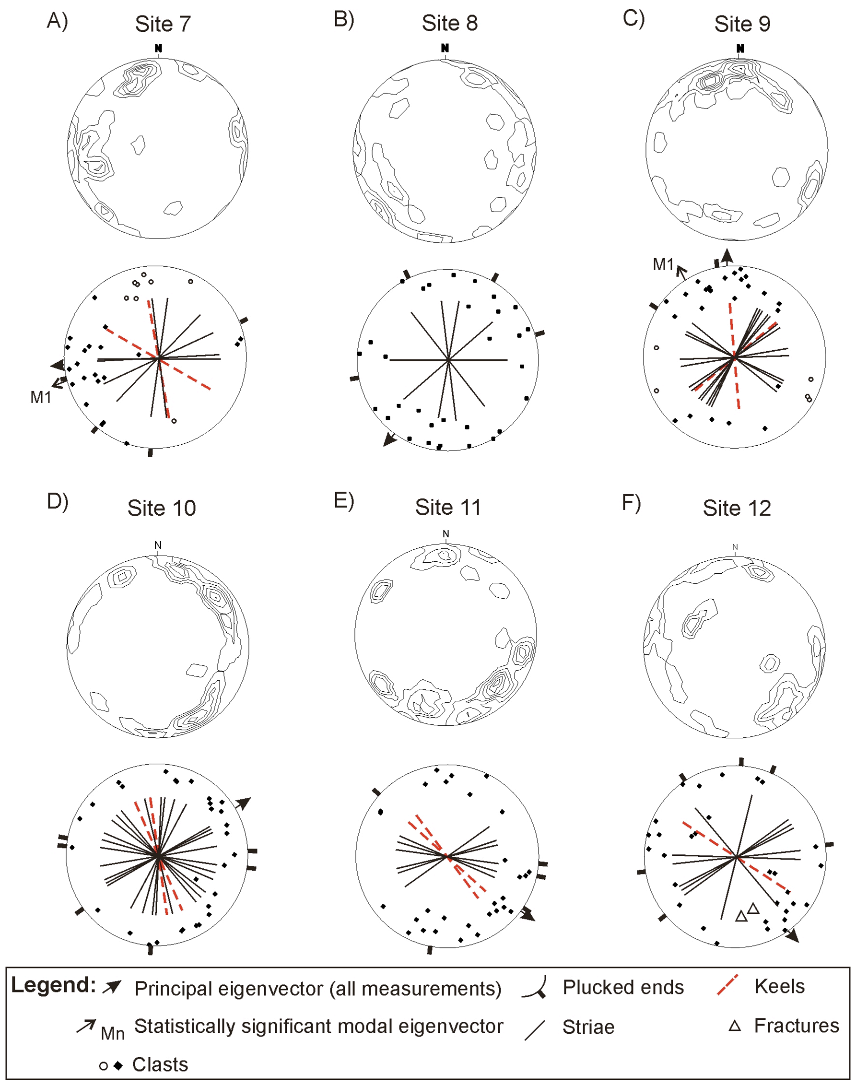


Figure 2.6. ***Equal-area lower hemisphere Schmidt plots of fractures, and stone fabric and surface wear features from poorly consolidated diamictons. At each site, contoured and scatter plots are provided. Only the most recent (and/or delicate) striae on clast tops, keels on clast bottoms, and angular plucked ends are displayed on point plots.Only the most recent (and/or delicate) striae on clast tops, keels on clast bottoms, and angular plucked ends are plotted. Fractures are plotted as poles to planes. Distinct clast symbols (open and closed circles) are used to distinguish visually separated modes and outliers (cf. Hicock et al., 1996). Separate arrows denote modal eigenvectors (Mn) where modes are statistically significant at 97% confidence level (calculated using Woodcock and Naylor (1983) method). Contour interval: 1.5 σ from an assumed random distribution (method of Kamb 1959). Details on the characteristics and location of striae and keels are given in Table 2.2. Data are presented in Appendix C.***

associating clast wear features with a singular process and raise the question of which (if any) features are definitively diagnostic of a particular process.

If till sedimentology and clast wear characteristics are to be used to gain better understanding of till genesis (and resulting classifications) and the complexity of subglacial processes, then there is also a need to recognize the potential equifinality of some subglacial processes. For example, observed clast wear features such as striae aligned with the clast a-axis and bullet–shapes, plucked ends, and keels in lodged position, combined with structural data give rise to a lodgement interpretation (Lian and Hicock, 2000). However shear zones in plastically-deforming tills within ring-shear devices produce a-axis parallel striae (Thomason and Iverson, 2008). Similarly, Benn (1995) argued that many of the wear features associated with lodgement can be produced within a deforming bed setting. These results highlight the difficulty in associating clast wear features with a singular process and raise the question of which (if any) features are definitively diagnostic of a particular process.

2.4.2. Premise of process reconstructions

During CIS build-up, it is suspected that clasts at the ice base may have ploughed into the substrate and eventually lodged. Prolonged ploughing and lodgement lead to clast a-axis alignment with the dominant shear stress direction and also to clast modification whereby bullet-shaped morphology develops by abrasion and plucking. Similarly, keels and facets form during prolonged ploughing.

The style of subglacial sedimentation and deformation may also change in response to changes in basal thermal regime such that lodged material may have undergone modification by various glaciotectonic processes (Hicock, 1990, 1992, Evans, 2006, Lian and Hicock, 2010). Because striae can form within deforming sediments and along shear planes, striae alone are not sufficient evidence to infer lodgement. Such inference must rely on observed wear features in specific locations on clasts (i.e. stoss-lee fractures, striae aligned with a keel, a-axis striae on clast surface, etc.), defining a characteristic clast morphology for lodgement. However, Benn (1995) adds complexity to the matter by proposing that striae, bullet-shapes, and fractures form at shear planes. This possibility is discounted for two reasons: 1) this possibility conditioned largely by an *a priori* assumption that sediments were deforming (Boulton and Hindmarsh, 1987), and that clast morphologic characterisitics developed along a shear plane contemporaneously with the deformation of this material. The main argument associating clast shaping with shear planes is the development of facets at near-parallel angles to the shear plane (Benn, 1995). However, the possibility of inheritance of clast morphologic elements must also be considered. Benn (1995) clearly recognizes this possibility as one explanation for the clast morphology. The association between facets and shear planes may therefore reflect clast reorientation along the shear plane and does not necessarily imply significant clast abrasion/ morphologic modification; 2) In this regard, though weak striae may form along shear planes, other morphologic features such as keels and stoss-lee forms likely represent much longer periods of abrasion and would require prolonged shearing along a single plane. This is considered to be an unlikely situation given the fact that new shear planes will develop, even over short time scales (e.g. diurnal) (Boulton et al. 2001), with changes effective stress controlled largely by fluctuations in porewater pressure. Short, anastomosing shear planes in till also suggest the unlikelihood of prolonged abrasion at shears.

Thus, in the absence of observations suggesting association of shaped clasts with shear planes, combined wear features on a single clast are associated with lodgement. It is also recognized that the latest events are most likely to be preserved on clast surfaces and may overprint, though likely not erase, the initial evidence of lodgement (Evans et al., 2006, Lian and Hicock, 2010). This allows us to make inferences regarding the relative timing and evolution of processes affecting the sediments.

2.5. Results

2.5.1. Drumlin morphology and composition

Drumlin morphology and composition varies within the study area, reflecting an altitudinal zonation of drumlin materials (Figs. 2.4, 2.7). At the highest elevations (~1600-1750 m asl), incipient drumlins occur exclusively in bedrock on the flanks of Trepanege Ridge (Figs. 2.4, 2.7 A; Lesemann and Brennand, 2009). Drumlins in bedrock and consolidated diamicton lie between ~1500-1600 m asl, (Figs. 2.4, 2.7 B). At the lowest elevations (1400-1500 m asl.), on the floor of the topographic basin, drumlins occur exclusively in sediment (Figs. 2.4, 2.7 C). Within the topographic basin, these sediments also occur in the intervening troughs between drumlins, and extend to non-drumlinized areas (Table 2.2; Fig. 2.4).

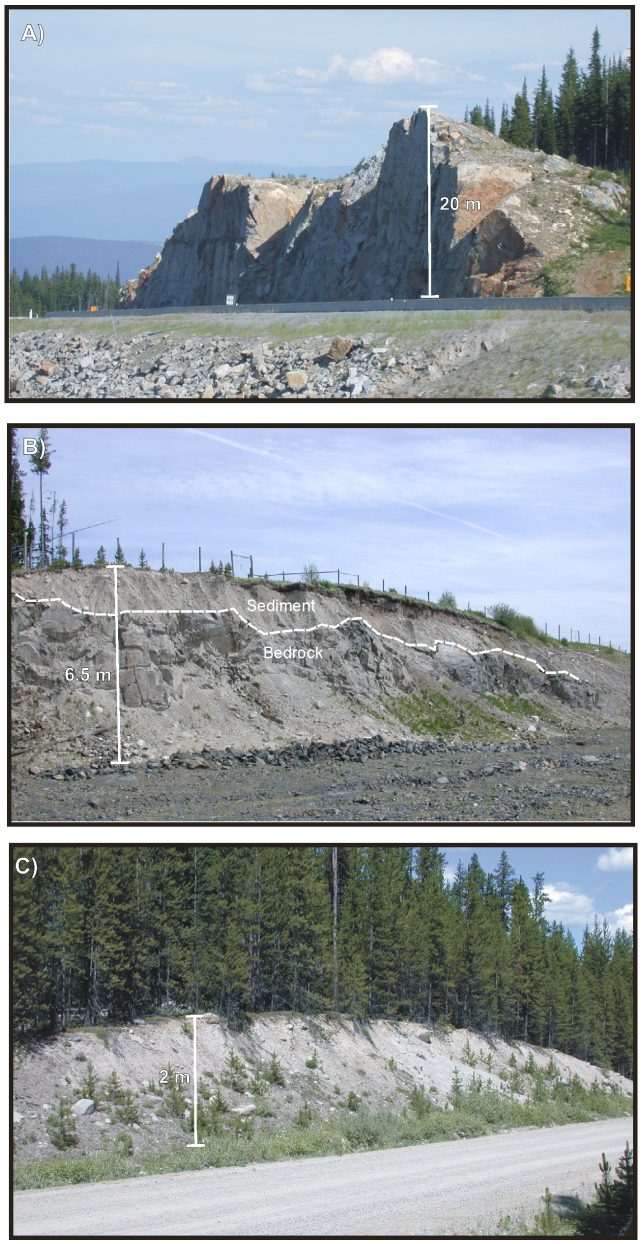


Figure 2.7. Drumlin composition on Thompson Plateau. A) Bedrock drumlins, B) bedrock-sediment drumlins, and C) sediment drumlins (See Fig. 2.4 for locations).

2.5.2. Consolidated diamicton

2.5.2.1. Sedimentology and structure

Consolidated diamicton onlaps the bedrock flanks of Trepanege Ridge; it occurs within drumlins, in weakly drumlinized areas and extends into non-drumlinized zones between drumlins (sites 1-6, Fig. 2.4, Table 2.1). This lithofacies consists mainly of consolidated, poorly sorted, massive to fissile (Fig. 2.8) matrix-supported, clay to sandy clay loam diamicton (Fig. 2.9), with 5-25% pebbles and infrequent cobbles and boulders (Table 2.1). A notable textural exception occurs at site 4 (Figs. 2.8 D, 2.9) where coarse sand dominates the matrix (sand loam). Boulder concentrations sometimes appear at the land surface where consolidated diamicton crops out. Consolidated diamicton exhibits fractures extending for 0.17-0.45 m and sometimes filled with massive clay and silt with some fine sand (Fig. 2.8 A, E, F; Table 2.1). Fractures are mainly oriented with fabric modes (Fig. 2.5). In some cases, more consolidated diamicton intraclasts occur within the lithofacies (Fig. 2.8 A, C). These intraclasts consist of angular pebbles and cobbles in a silt loam matrix. Silty stringers and laminated sand and silty clay occur around these diamicton intraclasts and within the surrounding diamicton. Stringers and laminations are commonly deformed around edges of protruding stones and/or diamicton intraclasts (Fig. 2.8 A, C).

2.5.2.2. Stone fabric and surface wear

Stone fabrics from consolidated diamicton typically show *spread unimodal* to *spread bimodal* distributions (cf. Hicock et al., 1996) (Fig. 2.5, Table 2.1) with principal eigenvalues varying between 0.62-0.74. Most fabrics exhibit strong transverse modes or spreads (when compared to the inferred synoptic ice flow direction of the last CIS (cf. Fulton, 1967, 1975; Clague, 1989)). For example, samples at sites 1 and 3 exhibit a transverse principal eigenvector and primary mode respectively, and samples at sites 2, 4, 5, and 6 exhibit a transverse spreading (smearing) of a more ice-flow parallel primary mode (Fig. 2.5) (cf. Lian and Hicock, 2000).

Figure 2.8. Sedimentology of consolidated diamictons within drumlins (sites 1, 2, 5) and in weakly drumlinized zones (sites 3, 4, 6). A) Site 1: fissile sand clay loam diamicton containing fractures (dash-dotted white lines) and contorted silt stringers (dotted white lines) between more massive diamicton intraclasts (outlined by dashed black lines). Knife is 0.2 m long. Site located in drumlin tail, east of crestline. B) Site 2: massive stony sand clay loam diamicton. Shovel is 1.1 m long. Site located in drumlin nose, west of crestline. C) Site 3: clay loam diamicton containing silty clay and fine sand laminations deformed by intraclasts. Site located in weakly drumlinized patch on slight topographic rise. D) Site 4: fractured consolidated sandy clay loam diamicton. Trowel is 0.25 m long. Site located on low flanks and trough of weakly drumlinized patch downflow of local bedrock ridge. E) Site 5: shears (dashed white lines) in stony clay loam diamicton overlain by glaciofluvial material (above dashed black line). Site located on drumlin flank, toward *nose. F) Site 6: compact fissile and fractured (dash-dotted white lines) clay loam diamicton on the high flanks of Trepanege Ridge. Site located on weakly drumlinized patch. See Figure 2.4 for site locations and Table 2.1 for additional sediment and site characteristics. Textural terminology from USDA soil classification (Soil Survey Staff, 1999).*

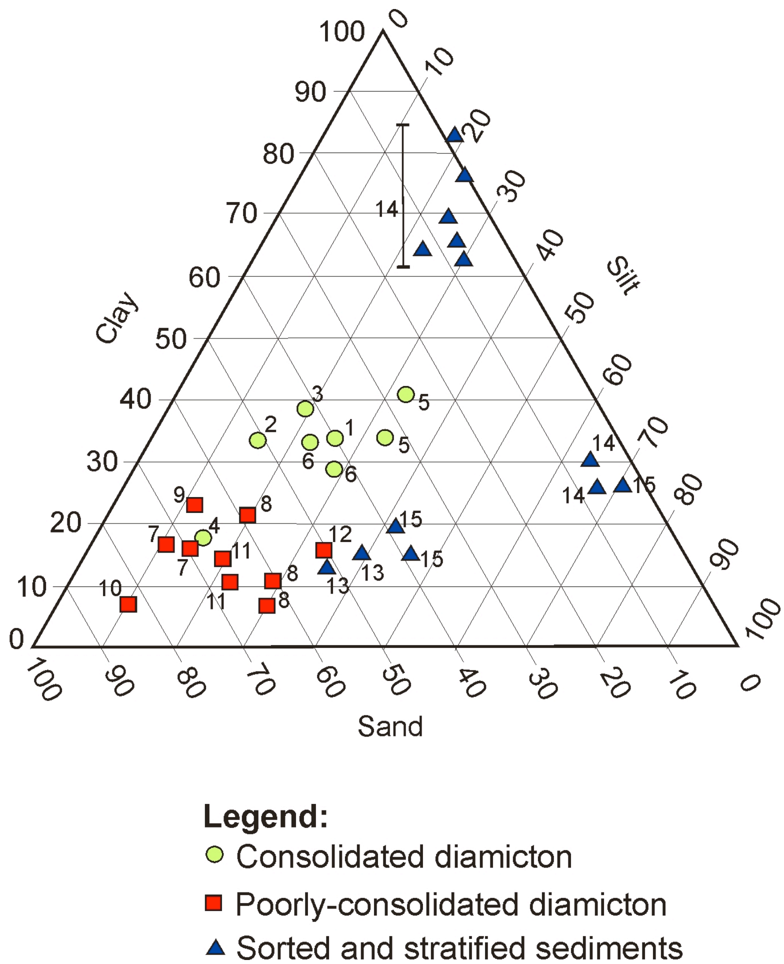


Figure 2.9: Ternary diagram of matrix texture (grains <2mm) in consolidated diamicton, poorly-consolidated diamicton, and beds of sorted and stratified sediments. Consolidated diamicton is mainly characterized by a clay loam and sandy clay loam matrix. Poorly-consolidated diamicton has a sand loam or sand matrix with low silt and clay constent. The clustering of data points for sorted and stratified sediments reflect the presence of laminae and beds of silty clay, clayey silt and silty sand. Samples do not capture the full range of observed textures. Textural terminology from USDA soil classification (Soil Survey Staff, 1999).

Most of the youngest striae on clast tops are parallel with one or more fabric modes, though some are oblique or transverse to such modes. Keels show some deviation from the primary modes, though many align with the densest clusters. Although some angular, plucked ends also align with modes, most show little directional preference. Multiple generations of striae occur on the majority of clasts larger than pebble size. Clast shape and density of striae are highly variable. Most angular clasts exhibit few striae and have lithologies similar to those of outcrops in the vicinity. Most striae are parallel to a-axes, while some are parallel to b-axes, and some curve around clast edges (Table 2.1).

2.5.2.3. Interpretation of consolidated diamicton

Based on their sedimentology, structure, and stone fabric and surface wear features, consolidated diamictons are interpreted as tills showing evidence of both lodgement with clast ploughing (Clark and Hansel, 1989) and deformation (see below). Although this hybrid interpretation is applicable to all sites, the degree of deformation is variable between sites and distinctions are made based on the degrees of sediment remobilization following initial lodgement. These differences and their controlling conditions are discussed below.

2.5.2.3.1. Evidence for lodgement and deformation processes

Evidence for lodgement includes abraded clasts with faceted upper surfaces exhibiting some a-axis parallel striae, keels on clast bottoms, and plucked ends (§ 2.4.2) (Table 2.1). Alignment of principal and primary fabric modes, striae, plucked ends and keels (e.g., site 5, Fig. 2.5 E, Table 2.1) strongly suggest clast ploughing and lodgement (Krüger, 1984; Lian et al., 2003). Clast lodgement generally produces strong unimodal fabrics and may produce weak transverse modes due to clast collisions and/or compression (Hicock et al., 1996) (e.g., site 5, Fig. 2.5 E, Table 2.1). Multiple generations of cross-cutting striae, striae curving around clast edges, and misalignment between fabric modes, striae, keels and plucked ends all suggest deformation following lodgement (cf. Lian et al., 2003).

Different processes may explain the presence of striae curving around the edges of clasts depending on the rheology of subglacial sediments. For example, if the till had a viscous rheology, curved striae may result from Jeffery/Taylor-type clast rotation (rotation around b-axis and a-axis respectively) (Lian et al., 2003) controlled by strain magnitude (Carr and Rose, 2003). Laboratory experiments (Iverson and Iverson, 2001; Rathburn et al., 2008) and interpretations of field data (Larsen and Piotrowski, 2003; Piotrowski et al., 2004; Tulaczyk, 2006) suggest that many tills had a Coulomb plastic rheology. In plastically-deforming substrates striae could form by matrix advection during March-type rotation (cf. Hooyer and Iverson, 2000), by grain-grain contact as clasts roll within a shear zone, and possibly during development of 'pressure bridges' (cf. Iverson and Iverson, 2001). In the latter case, linear clast clusters develop and bear increasing pressure during deformation. Once a pressure threshold is reached, these clusters collapse with transient redistribution of grain stresses resulting in clast movement (Iverson and Iverson, 2001). Upturned keels on the other hand give clear evidence of clast rotation. Combining various lines of evidence may lead to a consensus on substrate genesis.

At least one set of curving striae may be inherited from initial lodgement rather than from episodes of post-lodgement deformation. The number, location, and orientation of curved striae are dependent on the initial attitude of clasts as they came in contact with the bed. Iverson et al. (1994) report that strain weakening occurs during ploughing in a plastically-deforming medium. Clasts within the prow of deforming material (on the stoss end of the ploughing clast) might rotate and be deflected to the sides of the prow, leading to clast-clast contacts and possible production of curved striae. Few striae around a limited number of edges might be produced in this manner, though this has not been verified experimentally or in the field. Addition of further striae can be partly discounted where, in a cross-cutting set, the orientation of curved striae is similar to that of straight striae, and where upturned keels further suggest clast rotation during substrate remobilization. Varying directions of plucked ends, rather than unidirectional alignment with a fabric mode, also support clast remobilization following initial lodgement. However, 'double plucked-ends' may result from ploughing and lodgement processes (Krüger, 1984) and would produce roughly bidirectional fractured-ends on lodged clasts.

Diamicton sedimentology also reveals evidence of substrate deformation. Low effective stresses are required for ductile deformation. Under these conditions, overburden pressure is transmitted to the bed via clasts or through direct contact by the ice. Shear stress may be transmitted from the ice by clasts ploughing into the substrate (Tulaczyk et al., 2001), or by irregularities in the ice bed (Evans et al., 2006). Porewater pressure is an efficient modulator of effective stress and can lead to ice-bed decoupling if porewater pressure exceeds overburden pressure. Discontinuous silty sand stringers and laminated sediments are evidence of ice-bed decoupling with water flow at the interface (Piotrowski and Tulaczyk, 1999; Piotrowski et. al., 2005). Decreasing water pressure leads to recoupling of the glacier to its bed; effective pressure increases and stress is once again transferred to the substrate. Deformed fine-sand stringers and laminated silt and clay record these transient conditions (sites 1 and 3, Fig. 2.8 A, C, Table 2.1). In addition, fractures and shear planes are evidence of deformation (sites 1, 5, 6, Fig. 2.8 A, E, F, Table 2.1), indicating that localized brittle failure followed bed dewatering (Lian and Hicock, 2000; Lian et al., 2003). Therefore, the combined record of lodgement, its associated deformation, and more widespread remobilization of consolidated diamictons reflects a complex process history controlled by fluctuations in porewater pressure (Lian et al., 2003; Piotrowski et al., 2004, 2005).

2.5.2.3.2. Sediment characteristics controlled by local site conditions

Examining individual site characteristics within consolidated diamicton reveals that stone fabric modality and orientation, porewater pressure conditions, and the degree of post-lodgement remobilization are closely linked to the location of sites on the flanks of Trepanege Ridge and to the effects of local topography. The majority of fabrics record varying degrees of compression against Trepanege Ridge exemplified by flow-transverse orientation of primary or secondary fabric modes (sites 1 and 3, Figs. 2.2, 2.3), or a flow transverse spreading of the primary fabric mode (sites 2, 4, 6, Figs. 2.2, 2.3) (Lian and Hicock, 2000; Lian et al., 2003; Neudorf, 2008).

The topographic location also controlled porewater pressure and determined the degree to which ductile or brittle processes dominated deformation. For example, stone fabrics at sites 1, 2 and 3 (Fig. 2.5) are spread unimodal to spread bimodal with most primary modes oriented parallel to Trepanege Ridge (i.e. transverse to the inferred synoptic ice flow direction during the last glaciation). Weak ice-flow-parallel stone orientations occur as part of the modal spread or as a weak secondary mode (site 3, Fig. 2.5). Striae, keels, and plucked ends exhibit similar patterns. These fabrics record varying degrees of clast reorientation. Site details are presented below out of numerical order but in a sequence illustrating changes from a dominance of lodgement processes to increasing evidence for remobilization and more intense deformation.

At site 5, alignment of a strong primary fabric mode, striae, keels and most plucked ends aligned with with the synoptic ice flow direction supports initial lodgement during the extending flow during ice incursion. The lack of a preferred direction to the plucked ends, and the presence of a weak transverse secondary mode suggest a limited degree of clast re-alignment possibly due to clast collision during lodgement or as post-lodgement remobilization (Hicock et al., 1996). Alternatively, the weak transverse mode may record local compression against Trepanege Ridge during ice build-up. Shear planes record brittle deformation as the bed drained and became increasingly stiff. Drainage is favoured by the elevated topographic position of the site. Good drainage and stiff material could also explain why a limited number of clasts were re-oriented upon compression. Thus, at site 5 many initial lodgement characteristics are preserved, but brittle deformation, controlled by site drainage, overprints this record.

At site 2, macroscale sedimentology, stone fabric and surface wear features record evidence of substrate deformation with possible preservation of evidence of lodgement. The spread unimodal fabric at Site 2 suggests a rotation of clasts from flow parallel to flow transverse in response to compressive flow (Hicock et al., 1996) against Trepanege Ridge. Evidence of clast re-orientation consists of a-axis parallel striae on clasts oriented parallel to Trepanege Ridge (i.e. transverse to regional NW-SE ice flow). Plucked ends on clasts aligned with regional ice flow may suggest evidence of lodgement processes if they were associated with other wear features diagnostic of lodgement (§ 2.4.1). The notable absence of keels on clasts makes a lodgement interpretation equivocal. Curved striae and some b-axis parallel striae (site 2, Table 2.1) were produced during and following clast realignment under a compressional regime within ductile deformation. Conceptually, compressive flow (forming flow-parallel fabrics) is expected during ice build-up and encroachment on Trepanege Ridge. However, compression may have been limited at site 2 because it is located on the margin of a broad trough crossing Trepanege Ridge (Fig. 2.4), and upflow from a deeply incised valley leading from the edge of Thompson Plateau (Trépanier Creek, Fig. 2.4). It can be envisaged that this trough acted as an outlet for ice and modulated pressure build-up against Trepanege Ridge. Similarly, this outlet may also have favoured bed drainage and prevented porewater pressure build-up, accounting for the lack of evidence for ice-bed decoupling and more ductile deformation as is the case elsewhere (e.g., site 1), ultimately leading to partial clast re-orientation in the vicinity of site 2.

Site 4 displays the coarsest and most fissile consolidated diamicton documented in this study (Fig. 2.8 D). The coarse texture results mainly from friable grus incorporated from deeply weathered outcrops located upflow. Keels, one plucked end, and some striae aligned with these morphologic features suggest an episode of lodgement. However, the occurrence of these features on the fringes of the modal spread suggest some deformation with clast re-orientation following lodgement. This interpretation gains further supported by the presence of an up-turned keel (Table 2.1). However, alignment of most striae with the principal eigenvector or within the modal spread suggests that clast re-orientation was limited. Partial clast re-orientation could result from clast collisions within a deforming layer (Hicock et al. 1996). Given the location of site 4 on the flanks of Trepanege Ridge, it is also conceivable that reorientation took place due to compressive flow against Trepanege Ridge (as inferred for site 2). Additionally, limited clast reorientation and brittle fractures might reflect the coarse texture of the diamicton matrix. Enhanced drainage through the coarse matrix would impede build-up and maintenance of elevated porewater pressure, leading to limited deformation and clast re-orientation compared that expected for a fine. Compressive stress may then be manifest as fractures, rather than as clast re-orientation.

The clast fabric at site 6 mainly records ice-flow-parallel clast alignment during ice advance. Although the fabric is oblique to the inferred synoptic ice-flow direction, local flow deflection by the topography (e.g., the bedrock knob upflow of site 6, Fig. 2.4, and the flanks of Trepanege Ridge) account for the fabric orientation and corresponding alignment of keels, plucked ends and most striae. Fabrics at sites 1 and 3 are spread unimodal and spread bimodal respectively. They record increasing degrees of clast re-orientation related to flow against Trepanege Ridge. Sites 1, 3, and 6 record similar conditions. For example, clasts exhibit striae, plucked ends and keels, all indicative of lodgement processes. The remaining observations can be interpreted in various ways. For example, discontinuous laminated sediment stringers could be interpreted as the record of episodic ice-bed decoupling. However, documented stringers due to ice-bed decoupling tend to be continuous over many 10s cm length (Piotrowski and Krause, 1997). At site 1 and 3, ductile deformation must have followed ice-bed decoupling in order to account for deformation of stringers around diamicton intraclasts. Within this process sequence, clast rotation during deformation could explain the presence of striae curving around clast edges.

Dilatancy and collapse of a deforming bed (Evans et al. 2006) might offer an alternative, and perhaps more holistic process explanation for these observations. In this model, a dilatant sediment column becomes partitioned into a dilated active horizon (A horizon) with little to no strength in the matrix framework due to elevated porewater pressure, grading to a transitional zone (A/B horizon) where the matrix framework is patchy, and ultimately leading to a solid-state diamicton (B horizon) (Evans et al. 2006). This model predicts that A horizon sediments should be largely massive due to pervasive deformation and large cumulative strain that fully overprints evidence of partial sediment deformation. However, within the A/B horizon, decreasing porewater pressure leads to development of the matrix framework, capable of deforming (and preserving this deformation). Fluctuations in porewater pressure will create spatio-temporal changes in the degree of dilatancy of the A/B horizon, and in the position of this transitional horizon within the sediment column. During collapse of a dilatant layer, porewater pressure will increase and water will escape upward and between grains and intraclasts formerly in the dilatant/dilating material. This consolidation will be accompanied by deformation of the sediment. This event sequence may be recorded at sites 1, 3, and 6. In this case, diamicton intraclasts might be akin to 'till pebbles' (van der Meer, 1993, Evans et al. 2006) and their irregular shape might in part result from edge deformation while in a partially solidified state. Sand stringers represent fluid escape pathways and their discontinuous nature results from both the heterogeneous nature of the diamicton matrix, and its variable state of consolidation during till collapse. This model also predicts that as the till dewaters and returns toward a solid state, a component of simple shear could lead to brittle deformation. This could be recorded by the presence of fractures (shears?) at sites 1 and 6.

Lastly, lodgement processes might also accompany this dilatancy-solidification continuum. In particular, clasts could be lodged in two locations: 1) at the base of the A/B horizon as clasts plough into the consolidated diamicton near the B Horizon interface, and 2) at the ice-bed interface (top of the A horizon) as clasts frozen into the ice base are dragged through the till. These options could account for the recorded evidence of lodgement at the field sites. However, this possibility must be carefully examined, in particular the possibility of creating clast wear features likely to be identified in the field. For example, in the first case, the thickness and position of the A/B horizon is expected to migrate with fluctuations in porewater pressure. As a result, the shear strength of the matrix framework should vary accordingly and this should have a direct impact on the mobility of clasts: the matrix framework has to be sufficiently solidified to transfer stress and plough clasts through the substrate. Thus, this process is essentially faced with the same limitations as the proposed model of clast shaping along shear planes (Benn, 1995, § 2.4) where fluctuations in porewater pressure will change the behaviour of the material, and prolonged ploughing might be difficult to maintain. Therefore, this process might produce striae but probably not keels and plucked ends, considered to be diagnostic of lodgement processes (when occurring together on a clast – § 2.4). In contrast, clasts attached to the ice base ploughing the top of the A horizon might remain in their position sufficiently long to produce the diagnostic suite of clast wear features associated with lodgement. Alternatively, lodged clasts would be in an optimal situation for abrasion because the relative velocity between the clast and the ice would be at a maximum. Both these possibilities challenge the assumption of lodgement occurring during ice advance as stated in the starting premise (§ 2.4). However, the recognition that substrates might undergo multiple cycles of dilatancy-consolidation (Evans et al., 2006) would imply that clast lodged at the end of a dilatancy-consolidation cycle might be remobilized during a subsequent cycle(s). In such cases, it would seem very difficult to substantiate the timing of the lodgement events and the controls on its occurence (ice advance vs. post-dilatancy/collapse). Only lodgement processes and sediment remobilization could be reasonably inferred. At sites where dilatancy and collapse are inferred, the controls on lodgement events cannot be made and it is conceivable that mutiple lodgement events can take place. Though sediment deformation may reflect site-specific conditions, the similarity in fabrics and inferred substrate processes can be explained by a model of regional ice advance, flow compression, and ice build-up against Trepanege Ridge.

2.5.3. Poorly-consolidated diamicton

2.5.3.1. Sedimentology and structure

Poorly-consolidated diamicton occurs in the sedimentary basin and on the lowest flanks of Trepanege Ridge (sites 7-12, Fig. 2.4). It occurs within drumlins, in weakly drumlinized terrain and in non-drumlinized zones (Fig. 2.3; Table 2.2). It is overlain by consolidated diamicton and sorted and stratified sediments (§ 2.5.3), suggesting an environmental association between the three lithofacies. Poorly-consolidated diamicton is by far the most abundant lithofacies in the sedimentary basin on southern Thompson Plateau.

Poorly-consolidated diamicton is matrix supported, and composed of a sand or sandy loam matrix (Figs. 2.9, 2.10) with 5-25% pebble-cobble-sized clasts (Table 2.2). Clay and silt contents in the matrix typically vary between 20-30%, and are ~10-25% lower than for consolidated diamictons (Fig. 2.9). Poorly-consolidated diamictons are sometimes massive but more often weakly stratified. Better stratification and localized grading, and clast clusters are present in places (sites 7, 9, 10, 11, Fig. 2.10 A, C, D, E).

Clay and silt occur in thin discontinuous beds and laminations and as soft-sediment clasts (e.g., site 8, Fig. 2.10 B). Where the diamicton is stratified, matrix characteristics vary between individual beds (site 11, Fig. 2.10 E). Discontinuous silty sand beds often demarcate poorly-defined bed boundaries in weakly-stratified material (site 7, Fig. 2.10 A). Some beds include attenuated and soft-sediment intraclasts of laminated sand, sandy silt and silty clay (sites 8, 12, Fig. 2.10 B, F). Stones are angular, subangular or subrounded. Occasional fractures (site 12, Fig. 2.10F), single faults, and weak fissility (site 11, Fig. 2.10E) were observed within this lithofacies.


A)
B)
C)
D)
E)
Stony silty clay diamicton
Stony sandy loam diamicton
Stony silty clay diamicton
F)
Clast
Partially homogenized
sandy loam diamicton
Laminated silt

Figure 2.10. Sedimentology of poorly-consolidated diamictons within drumlins (sites 8, 9, 12), in weakly drumlinized terrain (site 10) and non-drumlinized zones within the drumlin swarm (sites 7, 11). A) Site 7: weakly stratified (dotted white lines) sand loam diamicton beds containing pebbles. Trowel is 0.25 m long. Site located in non-drumlinized patch upflow and lateral to weakly drumlinized topographic high. B) Site 8: partially deformed and attenuated sandy silt and silty clay intraclasts (dashed black outline) alongside gravel clasts (black arrow) in weakly stratified sandy diamicton. Knife blade is 0.12 m long. Site located on drumlin crestline, halfway along drumlin length. C) Site 9: weakly stratified (dotted white lines) stony sandy clay loam diamicton. A portion of the diamicton unit thickness is normally faulted (solid white line with displacement direction arrows) and overlain by glaciofluvial gravel (above black dashed line). Striped staff is 1 m long. Site located on drumlin crestline, one third along drumlin length. D) Site 10: steeply inclined, clast-rich loamy sand diamicton beds showing weak imbrication and clast clusters (outlined by dashed black lines). Trowel handle is 0.1 m long. Site located in weakly drumlinized patch. E) Site 11: interbedded silty clay diamicton and sandy loam diamicton. Site located in non-drumlinized zone, marginal to weakly drumlinized patch. F) Site 12: homogenized and fractured (dash-dotted white lines) sandy loam diamicton. Deformed intraclasts of stratified sand and sandy silt (dashed black outlines) are preserved within the partially homogenized diamicton. Site located in flute superimposed on crestline of drumlinized patch. See Figure 2.4 for site locations and Table 2.2 for additional sediment and site characteristics.

2.5.3.2. Stone fabric and surface wear

Stone fabric modalities range from spread bimodal to girdle shaped to spread unimodal (Fig. 2.6, Table 2.2). Principal eigenvalues range between ~0.43-0.58 (Table 2.2). Stones exhibit striae, some curving around clast edges and occurring on plucked ends. Striae generally parallel clast a-axes, save for some rare cases with b-axis parallel striae (Table 2.2). Striae and fractured ends show no systematic alignment or preferred direction save for site 11 where most fractured ends align with the primary mode (Fig. 2.6). Keels on clast bottoms (i.e. in ploughing position) are aligned with the preferred directions of some clast clusters (site 7, 9, 11, Figs. 2.6, 2.10 D). Some rare upturned keels are also present (site 9, Table 2.2).

2.5.3.3. Interpretation of poorly-consolidated diamicton

Poorly-consolidated diamicton described in this study is interpreted to be a subglacial sediment based on its stratigraphic context and character: it is overlain by consolidated diamicton, the latter recording clast ploughing during lodgement and post-lodgement remobilization. It contains stones exhibiting striae, keels and plucked ends consistent with lodgement. Whether poorly-consolidated diamicton is a till or not requires examination of site specific conditions. Below, individual site interpretations are presented then integrated to suggest that poorly-consolidated diamicton results from two sets of processes overprinting original glacigenic processes: 1) remobilization of till by subaqueous debris flows, and 2) washing and elutriation of till.

The presence of striae, facets, plucked ends and keels record lodgement processes (Lian et al., 2003; Lian and Hicock, 2010). However, as was the case for consolidated diamicton, the sediment has undergone varying degrees of post-lodgement modification, resulting in stratification, fracturing and faulting of the sediments, and production of upturned keels. The presence of curved striae is more equivocal as these surface wear features may be associated with post-lodgement clast remobilization (Lian and Hicock, 2003) or they may be produced during clast ploughing (Thomason and Iverson, 2008).

At sites 7, 9, 10, and 11 (Fig. 2.10 A, C, D, E) well developed stratification and localized clast imbrication and clusters (Fig. 2.10 D) may reflect (re)deposition by gravity flows. Although light and variably-oriented striae can result from gravity flows (Atkins, 2004), the abundance of striae and their alignment with keels and plucked ends suggest that they were not produced in this manner. Instead, the striae are probably glacigenic, suggesting that gravity flows remobilized till. Squeezing of dilatant sediments by glacier ice can lead to failure and mobilization as a debris flow (e.g. squeeze flow till, Dreimanis, 1989). Clast fabric modalities (mainly girdle-like or spread, Fig. 2.6, Table 2.2) are also consistent with documented debris flow fabric data. Many reported debris flow stone fabrics show weak to strong clast alignment parallel to flow direction (Lawson, 1979), though girdle distributions with low plunge angles are also observed (Lindsay, 1968). Fabric shape is dependent on whether the flow is in an extensive or compressive phase, the transport time/distance, and the rheology of the material at various stages during the flow. Our measured fabrics are comparable to those from experimental studies (Lindsay, 1968) and from some field studies (Eyles and Kocsis, 1988) of debris flows. Principal eigenvalues (Table 2.2) are in the range of those reported for subaerial debris flows (Mills, 1984; Eyles and Kocsis, 1988) and subaqueous debris flows (Surlyk, 1984). These interpretations are somewhat equivocal given the limited process inferences that can be made from fabrics alone (Bennet et al., 1999; Evans et al., 2006). This is particularly true when inferring debris-flow genesis given the temporal evolution of debris flow rheology, and some process convergence between till undergoing deformation due to elevated porewater pressure, and gravity flows (Evans et al. 2006). Therefore, debris-flow interpretations do not rely exclusively on clast fabric data. Rather, they are based on the integration of macro-scale sedimentology, clast fabrics and the sedimentary/geomorphic context.

Within poorly-consolidated diamictons, processes responsible for till remobilization and the production of laminated intraclasts require examination. At sites 8 and 12 (Fig. 2.10 B, F; Table 2), intraclasts (sediment rafts) within diamicton matrix may be produced during episodes of bed dilatancy and collapse (Evans et al. 2006). As porewater pressure fluctuates, these intraclasts could be incorporated within the more dilatant sediments of the A Horizon where the material can have a slurry-like behaviour.

Weak stratification suggests incremental deposition by sediment flows. Squeezing of dilatant sediments by glacier ice can lead to failure and mobilization as a debris flow (e.g. squeeze flow till, Dreimanis, 1989). If dilated sediments were squeezed, then intraclasts from the A/B Horizon could be incorporated within the gravity flow. Alternatively, bed shear by the debris flow itself could conceivably incorporate newer clasts within the flow. Both processes could also have occurred sequentially. Fractures and faults in poorly-consolidated diamicton units (e.g., site 9, Fig 2.10 C) suggest that brittle deformation locally overprinted ductile processes. Overprinting by brittle deformation is most simply explained by shearing of consolidated sediment and is consistent with the terminal cycle of a dilatant bed.

2.5.3.4. Lithological influence on diamicton characteristics

Textural differences between consolidated diamicton and poorly-consolidated diamicton (Fig. 2.9) are a key defining characteristic of these sediments. These textural differences potentially reflect subglacial processes, though lithological variations might also account for some differences. These possibilities are evaluated here.

Outcrops of volcaniclastic and sedimentary bedrock occur on the adverse slope rising from Douglas Lake toward the regional basin. The northern edge of the regional basin occurs ~5 km downflow of a lithological boundary: plutonic rocks of the Pennask Batholith dominate within the basin. Volcanic lithologies occur in portions of the basin and also form parts of Trepanege Ridge (Figure 2.4). If volcaniclastic rocks are responsible for supplying finer material than plutonic rocks, then some clay would be expected within the regional basin due to the extensive volcaniclastic rocks up-ice of the basin. It can be envisaged that, as the glacier advanced over the adverse slope rising from Douglas Lake, the glacier base would have been charged with sediments from volcaniclastic outcrops. As the ice crossed the lithological boundary of the Pennask Batholith, erosion of these plutonic rocks would conceivably produce coarser sediment than where volcaniclastic rocks were eroded. Nonetheless, some deposition of volcaniclastic fines would be expected within the basin.

Consolidated diamicton sites occur within areas underlain by sedimentary rocks (sites 1, 5, 6), and areas near/at transitions to igneous bedrock (sites 2, 4). Lithology at site 1 is not mapped but is likely sedimenary (Fig. 2.4). Sites 1, 3, 5, 6 exhibit greater clay and silt content than the poorly-consolidated diamictons and this could reflect the greater erodibility of the local sedimentary rocks. This effect would require abrasion, comminution and dispersal of the crushed material at the ice base within 10s m to several 100s m of the lithological transition on the flanks of Trepanege Ridge. Conversely, poorly-consolidated diamicton sites occur within areas underlain by plutonic rocks and are generally coarser than their consolidated counterparts. Lithological variations may provide a first-order control on matrix texture. However, this is unlikely to be the only controlling factor. For example, incorporation of soft-sediment clasts of laminated silt and clay within weakly-stratified poorly-consolidated diamictons requires removal, transport and concentration of fines prior to incorporation within the poorly-consolidated diamicton. Possible depositional environments for these fines could include mm- to cm-scale sorted stringers produced during decoupling at the ice-bed interface (e.g. Piotrowski et al. 2005). Larger-scale proglacial and/or subglacial depositional settings could also trap these fines. These possibilities are further discussed in § 2.6. However, irrespective of the exact depositional setting, all require prior removal of fines from the matrix. Dilatancy cycles could offer an effective mechanism for elutriation and removal of fines from the till matrix. This could occur near the end of a dilatancy cycle when upward flow of water and fines could accompany collapse of the deforming layer.

Repeated dilatancy cycles could conceivably elutriate fines from the matrix. As well, dewatering of debris flows could lead to sieving of fines from the matrix, generating dilute muddy flows or plumes. A proportion of these fines were transported and then deposited as rhythmic laminae and beds of sorted and stratified sediments (§ 2.6.4).

In conclusion, the poor consolidation and coarseness of the poorly-consolidated diamicton likely reflects primary lithologic controls, and secondary sedimentary processes near/within the glacier bed recording episodes of bed dilatancy, and elevated porewater pressure for removal of fines from the diamicton matrix by poreflow. This may have been facilitated by the regional bedrock basin which would favour regional porewater pressure build-up. Some evidence of pipeflow or conduit flow in sediment (e.g., canals) (Walder and Fowler, 1994) may be recorded by laminated intraclasts, though they may also result from melt-out over subglacial cavities (Shaw, 1982). Stick-slip glacier motion might result from periods of fluctuating porewater pressure where consolidated sediments might dilate (and/or decouple from the ice if the porewater pressure is sufficiently high) and cause sliding. Episodic fluctuations in the strength of basal coupling offer an appropriate mechanism for squeezing of dilatant sediments leading to redeposition by debris flows forming poorly-consolidated diamicton.

2.5.4. Sorted and stratified sediments

2.5.4.1. Description

Sorted and stratified sediments are overlain by poorly-consolidated diamictons. They occur within the depositional basin, within drumlin flanks and non-drumlinized areas, on topographic highs, and in areas without connections to meltwater valleys. These sediments consist of a group of lithofacies sharing common depositional environments.

The majority of sorted and stratified sediments consist of massive or thinly-bedded clayey silt and silty sand (Figs. 2.9, 2.11). Massive silty sand units can be 0.05-1 m thick and the thickest units are interbedded with clayey silt beds up to 0.15 m thick. Bedded silty sand also occurs as normally-graded, poorly to well-developed beds. The tops of silty sand beds are sharp and irregular where they underlie poorly-consolidated sandy diamicton beds (Fig. 2.11 A). Thin units (0.2-5 cm thickness) of rhythmically-bedded and laminated clay and silty clay form a subordinate component of the sorted and stratified sediment suite (Figs. 2.9, 2.11 B, C). These beds and laminae are normally graded from medium sand to clay (Fig. 2.11 B). The base of medium sand beds is sharp and contains some clay from the underlying bed so that bed texture changes upward from clayey medium sand, to medium sand, to clay (Fig. 2.11 C). Medium sand beds may be massive, planar laminated or exhibit ripple formsets in the case of the thickest beds (Fig. 2.11 B, C). Syndepositional deformation structures are common and consist of millimetre-scale flame structures and non-directional loading features. In places, multiple beds have been syndepositionally convoluted (Fig. 2.11 B): convoluted beds are overlain by planar, normally-graded laminations and beds. Silty sand units may contain soft sediment clasts of clayey silt (Fig. 2.11 D). These clasts occur as isolated ~0.1 m diameter rounded fragments showing some faint smearing at their perimeter (Fig. 2.11 D). Alternatively, they can occur as millimetre- to centimetre-diameter subangular to subrounded fragments dispersed randomly within the coarser sand (Fig. 2.11 D). In the former case, they are difficult to identify and can only be seen momentarily during section cleaning due to slight differences in moisture content where the surrounding sand dries more quickly than the finer silty sand rip-ups.

2.5.4.2. Interpretation

Rhythmic clayey silt beds and laminae suggest still-water conditions dominated by suspension settling in a pond or other small water body. Slightly more energetic conditions are recorded by sand beds with erosive lower contacts punctuating suspension deposition. Sorting and normal grading of these beds and the presence of ripple formsets suggest repeated turbidity flows with bedload transport followed by suspension settling. These turbidity currents may result from episodic meltwater transfer into a pond or from local failures of the pond margins. Syndepositional deformation structures associated with poorly-consolidated diamictons may record more widespread deformation and have important implications for interpretations of the depositional setting and development of these ponds. These implications are further discussed in § 2.6.3.

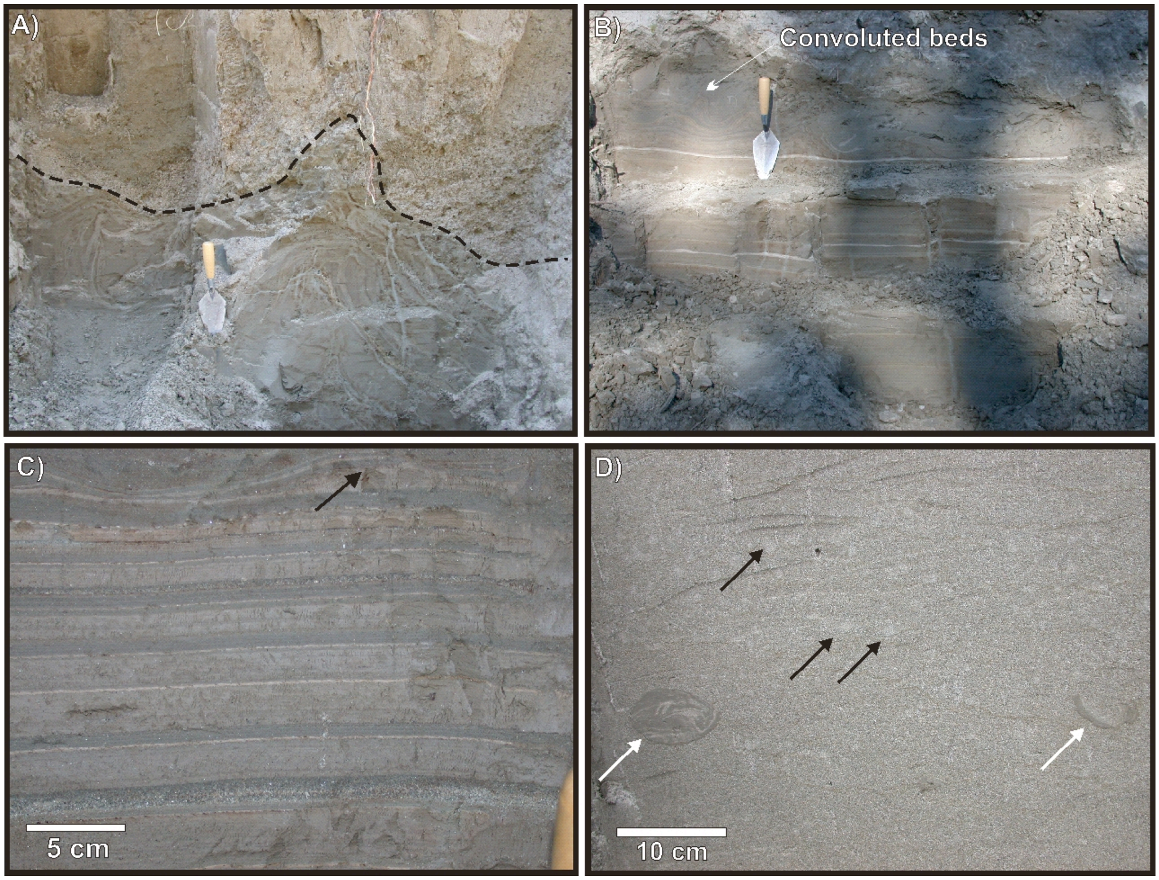


Figure 2.11. Sedimentology of sorted and stratified sediments found within drumlins (sites 13, 15), in non-drumlinized areas (site 14), and interbedded with the poorly-consolidated diamictons. Refer to Figure 2.4 for site locations. A) Site 13: massive to poorly-stratified silty fine sand loaded by overlying poorly consolidated diamicton (above dashed black line). Trowel is 0.25 m long. B) Site 14: rhythmically graded beds of sand to clay. Some beds are convoluted. C) Site 14: detail of rhythmically graded beds. Ripple form sets are visible in upper part of photograph (arrow). D) Site 15: rounded rip-up clasts of clayey silt (white arrows) and cm-scale rip-ups of silty sand (black arrows) within sand beds.

Rhythmic deposition of clayey silt beds could suggest seasonal hydrological controls on sedimentation. However, rhythmicity within lacustrine sediments may form under other circumstances. For example, laboratory experiments on settling and consolidation of clays (Hawley, 1981) reveal that cohesive clay beds can develop in a few hours and do not require prolonged (i.e. seasonal) quiescent conditions to form individual beds. Similarly, freshwater clay flocculation in glacial subaqueous settings may lead to sub-seasonal clay deposition (Hodder and Gilbert, 2007; Hodder, 2009). Thus, rhythmic sediments record episodic inputs of meltwater and sediment and their timing reflects fluctuations in the (glacio)hydraulic system which need not be seasonal. An important implication of this relative timing is that ponds recorded on Thompson Plateau may be short-lived depositional basins.

2.6. Conceptual model of substrate development

Documented lithofacies of Thompson Plateau record a broad range of (sub)glacial processes and highlight a complex polygenetic history for the sediments (and constituent clasts). The following section develops a conceptual model of substrate development incorporating the range of inferred subglacial processes (Fig. 2.12).

2.6.1. Genetic hypotheses for sorted and stratified sediments

A first-order interpretation of the depositional setting of sorted and stratified sediments suggests that they mainly record subaqueous deposition within ponded water bodies. These water bodies likely developed near or in contact with ice and in topographic irregularities in the bedrock and the bed. Three depositional settings are possible and are examined below: 1) deposition during deglaciation, 2) proglacial deposition during ice advance, and 3) subglacial deposition. Settings 2) and 3) could be transitional if ice advances over proglacial ponds. Therefore, some elements of these settings are discussed together.

2.6.1.1. Deposition during deglaciation

Fulton (1967, 1975) used the existence of large meltwater channels (10s-100s m scale in width and length) to argue that deglaciation of portions of Thompson Plateau proceeded largely by stagnation of local ice masses. If this model is applied to the study area, then sorted and stratified sediments could record deglacial processes. Within this scenario, deposition during deglaciation could involve two possibilities: deposition within subaerial ponds in contact or marginal to decaying ice masses, or deposition associated with meltwater channels leading to small ponds. Within either of these models, rhythmic, laminated and bedded clayey-silty sediments would record suspension settling of sediments in subaerial ponds delivered by episodic meltwater input from decaying ice. They could conceivably be varves. Coarser sand beds with ripple formsets would record episodic higher energy events. The fine-grained texture of these sediments suggests quiescent conditions, likely distal to meltwater sources. Deposition within an ice-contact/ice-marginal setting would likely involve episodic high energy events depositing sand, gravel, and diamicton (e.g. Clayton et al. 2008). At site 13 (Fig. 2.11 A), this could be recorded by the presence of poorly-consolidated diamicton deforming the laminated fines. Poorly-consolidated diamictons are interpreted as subglacial sediments. Their occurrence in a deglacial pond requires subaerial remobilization, possibly by small debris flows, or pushing and slope failure near decaying ice.

The strongest argument against sedimentation in deglacial ponds is the occurrence of sorted and stratified sediments within drumlins (typically drumlin flanks, § 2.5.4), though not all sorted and stratified sediments occur within drumlins. Furthermore, syndepositional deformation of pond sediments by poorly-consolidated diamicton, interpreted as subglacial sediments increases the likelihood that sorted and stratified sediments are not deglacial in origin.

2.6.1.2. Proglacial sedimentation during ice advance

Southward ice advance over Thompson Plateau could lead to proglacial ponding of meltwater. This is particularly true as ice encroached over the Pennask Batholith (Fig. 2.4) and entered the regional basin. Sorted and stratified sediments could be associated with this proglacial ponding. Within this depositional setting, sorted and stratified sediments would record sedimentation in contact with, or near, the ice margin.. Deformation of sorted and stratified sediments by poorly-consolidated diamictons would likely result from debris flows triggered by ice-push of proglacial material or from debris flows of sediments near the ice front. This interpretation implies that poorly-consolidated diamictons were not deposited regionally prior to deposition of sorted and stratified sediments. A corrollary implication is that the texture of poorly-consolidated diamictons mainly reflects primary processes of erosion and entrainment of igneous bedrock sources, instead of subglacial elutriation of fines, § 2.6.2.3) since little to no diamicton would be expected within the basin during ice advance. A further implication of this proglacial ponding model is that cavities and ponds should occur near the base of the glacial stratigraphic sequence. Although these sediments are observed near the land surface today, the inferred erosional origin of drumins on Thompson Plateau (Lesemann and Brennand, 2009; Chapter 3) implies that an unknown thickness of sediment has been eroded and that near-surface sediments may record ice advance. A paucity of data on sediment thickness in the regional basin limits our ability to assess the stratigraphic position of the sorted and stratified sediments. Better undestanding of sediment thickness within this basin would help test this hypothesis of proglacial deposition. Proglacial deposition of sorted and stratified sediments would predate drumlin erosion and can therefore be rationalized with the occurrence of these sediments within drumlins. A perplexing element of this depositional setting is related to the preservation of sorted and stratified sediments. These sediments might be expected to show evidence of deformation due to glacier advance and overriding, especially if proglacial pushing of sediment occurred during ice advance, as suggested by one model of emplacement of poorly-consolidated diamictons in a proglacial setting. Within this depositional model, glacigenic wear features would record subglacial. Clasts bearing these features could occur in an ice-marginal setting if they were sheared to the surface at the glacier front and redeposited by debris flows near the ice front. Alternatively, it is conceivable that sediments within ponds developed in topographic lows or in sheltered areas (e.g. downflow of a protective outcrop) might survive glacier overriding.

2.6.1.3. Subglacial deposition within cavities

A third genetic possibility for sorted and stratified sediments is linked to deposition within subglacial cavities. Within this depositional setting, cavities would develop subglacially within irregularities in the bedrock or in the till sheet. This till sheet would have been either quasi-impermeable or water-saturated, with water supply likely exceeding porewater flow (Darcian flow) or conduit flow in order to allow ponding. Such cavities were rimmed by elevated areas acting as glacier pinning points and as cavity seals. Within this depositional setting deformation of sorted and stratified sediments by poorly-consolidated diamictons would likely result from debris flows triggered subglacially at cavity margins where stress was preferentiall transferred. This interpretation implies that poorly-consolidated diamictons were emplaced prior to cavity development. A corollary implication is that the texture of poorly-consolidated diamictons could reflect secondary processes of subglacial elutriation and poreflow, as well as erosion and entrainment of igneous bedrock sources. A further implication is that subglacial cavities could develop throughout the glacial stratigraphic sequence. Again, better constraints on sediment thickness and the stratigraphic architecture of the regional basin could help test these hypotheses and refine understanding of glacial processes on Thompson Plateau. As was discussed for the proglacial depositional hypothesis, subglacial deposition of sorted and stratified sediments would predate drumlin erosion and can therefore be rationalized with the occurrence of these sediments within drumlins. The survivability of these sediments in a subglacial setting is also a perplexing element. Although ice-marginal sediment pushing processes do not apply in this setting, subglacial sediment failure near cavity margins does raise question about the absence of pervasive deformation of sorted and stratified sediments.

In conclusion, both the ice-advance proglacial depositional setting and the subglacial depositional setting can be rationalized with processes emplacing overlying poorly-consolidated diamictons, and with the occurrence of these sediments within drumlins and in substrates between drumlins. Consequently, given data constraints and unresolved questions regarding sediment thickness and the characteristics of the basin stratigraphy, both depositional settings offer viable options for deposition of sorted and stratified sediments.

2.6.2. Nature of ice flow and ice-bed coupling during substrate development

The characteristics of sorted and stratified sediments (whether deposited proglacially or subglacially) raise questions about the nature of ice flow and ice-bed coupling within the regional basin. For example, proglacial deposition during glacier advance would likely be followed by glacier overriding. Similarly, subglacial cavity sediments could be affected by the flow of overlying ice. Deformation could ensue in both cases. Yet, preservation of primary sedimentary structures, an absence of pervasive deformation and homogenization, and a dominance of syndepositional deformation suggests that sorted and stratified sediments underwent limited strain following their deposition. Two explanations may account for this preservation:

i) If proglacial ponds survived glacier overriding, they could conceivably have become subglacial cavities. Sediments could be preserved if water pressure within subglacial cavities remained sufficiently elevated to act as a ‘slippery spot' locally lowering the strength of ice-bed coupling and limiting deformation. This local effect could have been enhanced by the regional basin which would favour regional development of elevated porewater pressure. However, both diamicton types on Thompson Plateau record evidence of lodgement processes, implying strong ice-bed coupling, at least locally. This suggests that the glacier bed may have consisted of strongly- and weakly-coupled patches, controlled by subglacial hydrology (Menzies, 1989; van der Meer et al., 2003, Piotrowski et al., 2004; Evans et al., 2006). Inferences of substrate deformation following lodgement (§ 2.5.3.3) further speak to the spatio-temporal variability of ice bed coupling.

ii) The preservation of sorted and stratified sediments might reflect the rheology of deforming subglacial sediments. Depth-limited deformation is associated with a plastic bed rheology (Iverson et al., 1997, 2001; Piotrowski et al., 2001; Rathburn et al., 2008) and this could account for preservation of sorted and stratified sediments despite glacier overriding. Interpretations of poorly-consolidated sediments as deformed lodgement till (§ 2.5.3.3) implies a viscous bed rheology and could contradict the inference of plastic deformation. However, uncertainties remain concerning associations between clast wear features and subglacial processes (and hence bed rheology) (e.g. Benn, 2002) (§ 2.4.1). Furthermore, spatial changes in bed rheology and the co-occurrence of both viscous and plastic rheologies has been proposed in some settings (Benn, 2002; ,Fowler, 2003) and may be compatible with filed observations on Thompson Plateau.

2.6.3. Till emplacement and deformation episodes

A suite of lodgement processes (including abrasion, plucking, ploughing and lodgement of clasts (Fig. 2.12 A) is inferred during the early stages of substrate development as ice advanced over Thompson Plateau. This phase is best recorded at sites along the flanks of Trepanege Ridge (Fig. 2.4). A precursor episode of sediment incorporation into the ice must also precede deposition. This would have likely included freezing-on of plucked and abraded material. The rising adverse slopes leading from Nicola Valley to the regional basin and Trepanege Ridge (Fig. 2.1) also offer the possibility of sediment accretion by glaciohydraulic supercooling (e.g. Alley et al., 1998). It is difficult to assess to what degree the CIS quarried local bedrock or simply recycled interglacial sediments and/or its own outwash. Some clast surface wear features may have been derived from recycled till. Subrounded to rounded clasts together with other, more angular clasts and with lithologies similar to nearby bedrock outcrops, suggest a number of sources. Irrespective of source, clast wear features such as striae and keels are associated with abrasion during clast ploughing. Plucked ends and surface abrasion aligned with the clast a-axis are associated with abrasion and lodgement (Lian et al., 2003; Lian and Hicock, 2003, 2010). Spatio-temporal variations in porewater pressure allowed for deformation and remobilization of the lodged material. Clast rotation, and varying degrees of ductile and brittle deformation resulted from these subglacial hydrologic changes.

2.6.4. Sedimentation in subglacial cavities

Following bedrock erosion and emplacement of sediments at the glacier bed (Fig. 2.12 A1), cavities are hypothesized to have developed at the ice-bed interface and ponded by ice at pinning points around cavity margins (Fig. 2.12 B1). Sedimentation takes place in subglacial cavities (Fig. 2.12 C1) by a range of processes including: 1) possible elutriation of fines from till by groundwater flow, and possibly pipe flow and conduit flow. Fines are deposited from suspension (1, 4, Fig. 2.12 C1); 2) melt-out from the ice base (4, Fig. 2.12 C1); 3) localized deformation and destabilization of subglacial sediments at cavity pinning points forming debris flows (2, 3, Fig. 2.12 C1); with rip-up clasts triggered by either squeeze flow (Dreimanis et al., 1987) due to high shear stress at cavity margins and/or cavity reorganization through glacier movement; and 4) meltwater inflows (5, Fig. 2.12 C1). In a subglacial cavity (Fig. 2.12 C1), sediments enters the cavity through washing and elutriation of fines (1), localized deformation at cavity pinning points (2), subaqueous debris flows (3), meltout and suspension settling of fines (4), and stratified and sorted sediment deposited by meltwater (5).

2.6.5. Sedimentation in proglacial ponds

In a proglacial setting (Fig. 2.12 A2), ponds develop in front of the glacier within bedrock depressions and/or pre-existing sediments (possibly proglacial) (Fig. 2.12 A2).

Within proglacial ponds (Fig. 2.12 B2), deposition occurs due to inflows from groundwater (1), supraglacial and ice frontal debris flows (2), and/or through ice-push processes and sediment squeezing (Dreimanis, 1987) close to the cavity margin (3). Fines are deposited from suspension (4). Lastly, meltwater inflows could deliver sediment (5) (Fig. 2.12).

These summaries of ponded water sedimentation encompass the range of processes and lithofacies that have been documented in the regional basin on southern Thompson Plateau. *Every cavity* would not necessarily have contained *every lithofacies* documented here. Indeed some processes (meltwater inflows) account for a minor portion of sedimentary facies, while others (debris flows, suspension deposition) were much more abundant.

2.6.6. Cavity expansion and reorganization

If proglacial ponds persisted through glacier overriding, then some could conceivably have operated as subglacial cavities. Their further evolution could therefore resemble the development of subglacial cavities (Figure 2.12 C1). Consequently, further discussion will only deal with subglacial cavities (Figure 2.12 D). Subglacial water-filled cavities are reorganized in space and time as they expand, collapse (sometimes partly), and are displaced in response to coupled glacier hydrology and movement (Fig. 2.12 D). Such hydrologic reorganization is consistent with that inferred in Antarctica (Gray et al., 2005; Fricker et al., 2007) where cavities (larger than those inferred here) decant wholly or partially into neighbouring ones without necessarily triggering an outburst at the ice margin. Periodic cavity reorganization shifts the locus and process(es) of sedimentation and creates a patchwork of lithofacies. For example, as water is forced to drain from one cavity to a neighbouring one, cavity pinning points shift, water is transferred to other cavities via channels, and shear stress is transferred to material that was previously deposited *within* the former cavity. Consequently, some deformation of cavity sediments can occur locally to produce diamictic material exhibiting some or no evidence of pre-existing structures (e.g. Fig. 2.8 B, F). The degree of deformation likely reflects effective pressure at the ice bed interface, controlled overburden pressure, porewater pressure, and hydraulic conductivity of cavity sediments. Thus, local sediment deformation could

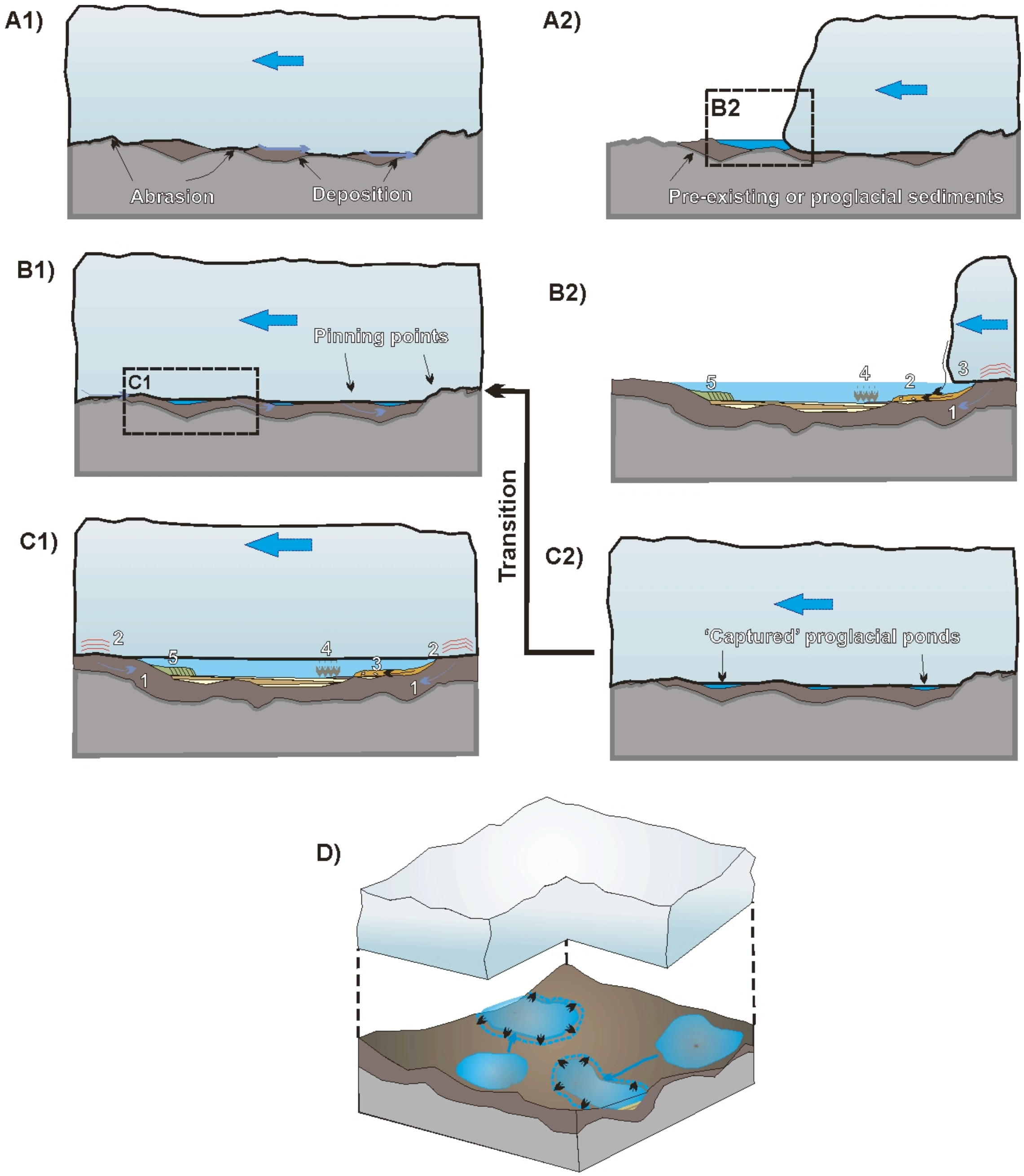


Figure 2.12. Conceptual model of subglacial sediment genesis. A1) Emplacement of subglacial till and water movement at the ice bed interface. A2) Proglacial sediment deposition and pond development; B1) Subglacial cavity inception. B2) Sedimentation within proglacial ponds. C1) Sedimentation within subglacial cavities. C3) Capture of proglacial ponds and possible transition to a subglacial cavity. D) Cavity expansion and reorganization.

occur and deformed sediments could be redeposited within a new cavity by a subaqueous debris flow (with possible rip-up clasts, weak stratification, etc.).

Regional controls and feedbacks on the development and duration of linked cavities over this portion of Thompson Plateau may also be important. For example, linked cavity systems have been implicated as a bed condition favouring rapid ice flow (e.g., Kamb, 1987). Subglacial water-filled linked cavities in the sedimentary basin of the southern Thompson Plateau likely decreased bed friction, consistent with the required conditions for preservation of cavity sediments (§ 2.6.3) and could have promoted sliding (e.g., Lliboutry, 1983). Such effects may have facilitated enhanced flow toward Trepanege Ridge where ice would have thickened. In turn, the presence of Trepanege Ridge may have favoured porewater pressure build-up in the basin by pressure melting against the ridge and by impeding subglacial drainage.

We also consider the phase of cavity development and evolution to have been temporally limited as cavity development and subsequent collapse would lead to smoothing of bed irregularities (e.g. Boulton, 1982) and potentially greater sediment homogenization. This cannot be shown to have occured in the sedimentary basin. There are several possible explanations for this: 1) the hydrologic system inferred from the sedimentary record did not persist for a sufficiently long time, 2) elevated porewater pressure and cavity development suppressed erosion by decreasing the effective pressure over basin substrates, leaving insufficient sediment for infilling, and also protecting subglacial substrates from prolonged and pervasive deformation by the ice base, and (or) 3) the latter part of the sedimentary record is missing due to erosion (Lesemann and Brennand, 2009, Chapter 3).

2.7. Conclusions

This chapter has documented paleo-bed conditions in and around a topographic basin on the southern Thompson Plateau in British Columbia. Documented lithofacies include: 1) a till (consolidated diamicton) recording a range of subglacial processes including abrasion and plucking associated with clast ploughing and lodgement, and the variable influence of sediment deformation. The main controls on till characteristics are fine-grained volcaniclastic bedrock lithologies, fluctuations in porewater pressure and the regional influence of a bedrock ridge (Trepanege Ridge) which acted as a barrier and led to flow compression; 2) Poorly-consolidated diamicton exhibits textural characteristics that potentially reflect primary lithologic controls from plutonic lithologies, and secondary processes of subglacial elutriation of fines in association with episodes of sediment dilatancy. Poorly-consolidated diamicton was remobilized by debris flows into ponded water bodies 3) Sorted and stratified sediments were also deposited within ponded water bodies A conceptual model of substrate development implicating the development and evolution of proglacial ponds and or subglacial cavities is proposed. Some proglacial ponds may have evolved into subglacial cavities where they could have operated as a network of linked-cavities. Regional porewater pressure build-up and meltwater ponding may have favoured development of cavities in the topographic basin. These subglacial water-filled cavities were metres to 10s metres in diameter and decimetres to possibly metres deep. Linked cavity evolution resulted in the patchwork of lithofacies preserving waterlain sediments, debris flow sediments and tills during the last glacial cycle of the Cordilleran Ice Sheet.

3. Regional reconstruction of subglacial hydrology and glaciodynamic behaviour along the southern margin of the Cordilleran Ice Sheet in British Columbia, Canada and northern Washington State, USA

Abstract

Subglacial landsystems in and around Okanagan Valley, British Columbia, Canada are examined to evaluate landscape development, subglacial hydrology and Cordilleran Ice Sheet dynamics along its southern margin. Major landscape elements include drumlin swarms and tunnel valleys. Drumlins are composed of bedrock, diamicton and glaciofluvial sediments; their form truncates the substrate. Tunnel valleys of various scales (km – 100s km length), incised into bedrock and sediment, exhibit convex longitudinal profiles, and truncate drumlin swarms. Okanagan Valley is the largest tunnel valley in the area and is eroded >300 m below sea level. Over 600 m of Late Wisconsin-age sediments, consisting of a fining-up sequence of cobble gravel, sand and silt fill Okanagan Valley. Landform-substrate relationships, landform associations, and sedimentary sequences suggest that the landscape is best explained by meltwater erosion and deposition during regional ice sheet underbursts.

During the Late-Wisconsin glaciation, Okanagan Valley functioned as part of a subglacial lake spanning multiple connected valleys (few 100's km) of southern British Columbia. Subglacial lake development started either as glaciers advanced over a pre-existing sub-aerial lake ('catch lake') or by incremental production and storage of basal meltwater. High geothermal gradients, geothermal springs and/or subglacial volcanic eruptions contributed to ice melt, and may have triggered, along with priming from supraglacial lakes, subglacial lake drainage. During the underburst(s), sheetflows eroded drumlins in corridors and channelized flows eroded tunnel valleys. Progressive flow channelization focused flows toward major bedrock valleys. In Okanagan Valley, most of the pre-glacial and early-glacial sediment fill was removed. A fining-up sequence of boulder gravel and sand was deposited during waning stages of the underburst(s) and

bedrock drumlins in Okanagan Valley were enhanced or wholly formed by this underburst(s).

Subglacial lake development and drainage had an impact on ice sheet geometry and ice volumes. The prevailing conceptual model for growth and decay of the CIS suggests significantly thicker ice in valleys compared to plateaus. Subglacial lake development created a reversal of this ice sheet geometry where grounded ice on plateaus thickened while floating valley ice remained thinner (due to melting and enhanced sliding, with significant transfer of ice toward the ice sheet margin). Subglacial lake drainage may have hastened deglaciation by melting ice, lowering ice-surface elevations, and causing lid fracture. This paper highlights the importance of ice sheet hydrology: its control on ice flow dynamics, distribution and volume in continental ice masses.

3.1. Introduction

Glacier flow results from the interplay between ice creep, sediment deformation, and the configuration and operation of the glacier hydrologic system. Beneath glaciers, porewater pressure within sediment modulates deformation, with repercussions on glacier flow velocity. Thus, sediment deformation is implicated in both 'normal' (non-streaming) flow and episodes of rapid ice flow (ice streaming) in ice sheets (Alley et al., 1987 Bennett, 2003). Rapid ice flow may be further enhanced if porewater pressure exceeds ice overburden pressure, leading to localized ice-bed decoupling and enhanced sliding over a thin water film (Kamb, 2001; Kamb and Engelhardt, 1991). Some subglacial meltwater is stored in various reservoirs. Extensive subglacial lakes (100s km^2) and smaller water-filled cavities (km^2-10s km^2) have been identified under the Antarctic ice sheets (Gray et al., 2005; Siegert et al., 2005; Wingham et al., 2006; Bell et al., 2007; Fricker et al., 2007). Some lakes are located at the heads of ice streams (Siegert and Bamber, 2000), others are located under ice streams (Fricker et al., 2007). Water transfer occurs between lakes and cavities (Wingham et al., 2006; Fricker et al., 2007), and may or may not be accompanied by an outburst at the ice margin. Some subglacial lakes may drain via channels eroded in the glacier substrate (Evatt and Fowler, 2007), others may drain as sheets (Flowers et al., 2004). There is likely a complex and dynamic interplay between meltwater production, storage, release, and glacier dynamics operating at a variety of spatial and temporal scales. However, the full implications of subglacially-stored water on ice sheet behaviour remains unclear (Bell et al., 2007) and is hampered by the inaccessibility of modern ice sheet beds.

Former ice sheet beds offer a landsystem record of past subglacial sediment transport and concomitant glacier hydrology and dynamics (e.g., Evans, 2006). Many reconstructions of paleo-ice sheet dynamics rely on process-based interpretations of bedforms (e.g. drumlins, Rogens, tunnel valleys, etc) and substrates. Conflicting interpretations of landform genesis lead to varying conceptions of paleo-ice sheet dynamics. For example, sediment deformation, in part controlled by subglacial hydrology, is invoked as a mechanism capable of producing a wide array of glacial bedforms (drumlins, Rogens, some tunnel valleys) (Boulton, 1987; Boulton and Hindmarsh, 1987, Ó Cofaigh et al. 2010). However, the same landforms are also attributed to the erosional and depositional action of turbulent meltwater sheets (Shaw 1996, 2002). Consequently, glaciodynamic reconstructions based on glacial bedforms remain equivocal.

Cordilleran valleys contain extensive sediment fills (cf. Fulton, 1965, 1972; Eyles and Clague, 1991; Vanderburgh and Roberts, 1996; Ward and Rutter, 2000; Lian and Hicock, 2001; Johnsen and Brennand, 2006). Consequently, studies have preferentially focused on these sediment sinks to perform environmental and process reconstructions. Some studies focused on landforms and sediments in areas of more subdued relief outside of major Cordilleran Valley systems, have shed light on subglacial hydrology and glacier bed conditions of the CIS (cf. Lian and Hicock, 2000; Lian et al. 2003). In contrast, plateaus have received comparatively little attention (cf. Broster and Clague, 1987; Huntley and Broster, 1993 for some examples), despite the fact that they host extensive suites of glacial landforms and sediments. This is particularly true for Thompson Plateau where exploratory studies of landforms and sediment have focused on deglacial processes and patterns (e.g., Fulton, 1967, 1975). Consequently, little is known about regional subglacial processes on Thompson Plateau.

Valley fill characteristics in Okanagan Valley suggest complex erosional and depositional events during glacial periods. For example, early seismic surveys of Okanagan Valley hinted at a complex valley fill (Fulton, 1972) spanning multiple glacial and interglacial periods (Fulton, 1972; Fulton and Smith, 1978). In contrast, more recent lake-based and land-based seismic surveys (Eyles et al., 1990, 1991, Vanderburgh and Roberts, 1996), suggest instead that the valley fill is relatively young and largely Late Wisconsinan in age. An important implication of these interpretations is that significant sediment erosion has taken place during glacial periods to remove the majority of the valley fill and to subsequently refill the valley with younger sediments. Eyles et al. (1990, 1991) and Vanderburgh and Roberts (1996) attribute some of this erosion and valley filling to pressurized subglacial meltwater erosion and deposition. The array and interconnectivity of erosional landforms, and the inferred genesis of the Okanagan Valley fill raises the possibility that regional subglacial erosional events may be responsible for landscape genesis. Such events have important implications for understanding of glacial landform genesis, glacial landscape evolution, and understanding of subglacial processes and associated glacier dynamics.

3.2. Objectives

This chapter examines regional landform patterns, and landform-sediment assemblages in the Okanagan Valley-Thompson Plateau area of British Columbia, Canada and in northern Washington State, USA in order to produce a regional reconstruction of glacier bed processes and glaciodynamics. Specific research objectives aim to:

1. evaluate landscape development, regional glaciodynamics and hydrologic conditions along the southern margin of the CIS;

2. evaluate hypotheses of landform genesis and present an event sequence of landscape evolution; and

3. investigate the implications of landform genesis and landscape evolution for CIS hydrology, glaciodynamics and geometry.

3.3. Field area

The area described in this paper encompasses part of the Interior Plateau of south central British Columbia, Okanagan Valley, Okanagan Highlands and a portion of northern Washington State (Fig. 3.1). The Interior Plateau is subdivided into a series of smaller regional plateaus that include Thompson Plateau, Okanagan Highlands and Shuswap Highlands with elevations ranging between ~900-1400 m asl (Holland, 1964). These areas are bounded by the Coast Mountains to the west and a series of linear mountain ranges (Monashee Mountains, Purcell Mountains) to the east with summits reaching 2500 m asl. In comparison to the bounding mountain ranges, topography on the plateaus is generally subdued. Most altitudinal variation results from prominent valleys dissecting the plateau surfaces. For example, Okanagan Valley is a north-south trending valley separating the eastern edge of Thompson Plateau from Okanagan Highlands.

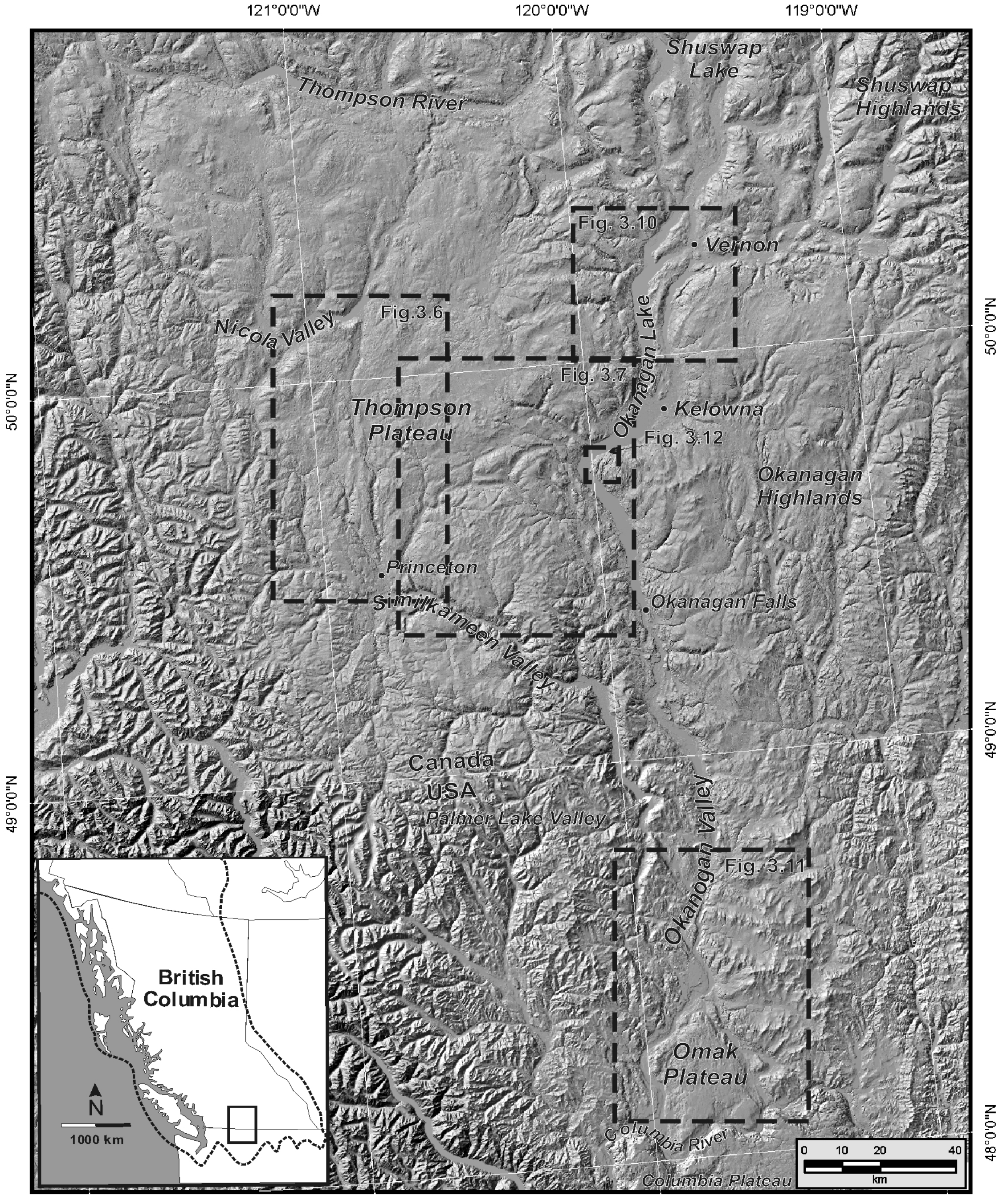


Figure 3.1. ***Hillshaded digital elevation model of the field area highlighting major physiographic elements of the British Columbia southern Interior Plateau discussed in the text. Dashed rectangles show locations of figures. Inset situates map within British Columbia. Elevation data for Canadian regions: BC Government TRIM I dataset, 25 m resolution (© GeoBC). Data for American regions: USGS 10 m DEM reprojected to BC Albers, NAD 83.***

It is a fault-controlled valley extending over 300 km from Shuswap Lake to the Columbia River in Washington State,[1] USA. There is a ~900 m elevation difference between Thompson Plateau and the modern Okanagan Valley floor at ~330-340 m asl. Smaller valleys, generally oriented ~east-west, dissect Thompson Plateau. These include the Thompson, Nicola, and Similkameen valleys. For these valleys, elevation differences between the plateaus and valley floors are generally ~300 m. The subglacial landsystems of Thompson Plateau and Okanagan Valley are spatially interlinked; they are described separately below for organizational simplicity.

3.4. Landform mapping used in regional reconstructions

Regional reconstructions within this chapter rely on process inferences from landforms including drumlins and bedrock valleys. Mapping of individual drumlins and groups of drumlins is used to recognize regional landform patterns and is the basis for generalized flowline maps (Figs. 3.2, 3.3). As well, drumlin morphology (Table 3.1) is used to help constrain genetic processes and landscape evolution. Drumlins were mapped using a combination of aerial photographs (1:15 000 scale), digital terrain models (DEM) (25 m pixel resolution), and field observations. Data sources used for mapping depended on the intended purpose of the map. For example, regional-scale landform domains were identified using DEM. Within these domains, flowlines were established from mapping prominent drumlin crests in order to extract landform orientation. Landform direction was inferred from drumlin morphology where appropriate. Not every drumlin was mapped within each domain because the goal was to build a synoptic-scale representation of landform patterns (Figs. 3.2, 3.3). In contrast, groups of drumlins were examined in more detail on the Thompson Plateau and in Okanagan Valley in order to assess morphometry. Length and width were assessed from DEM and verified with aerial photographs. Length was measured along drumlin a-axes. Width was taken as maximum measurement of the b-axis. The maximal lateral extent ('footprint') of

[1] Spelling of Okanagan Valley varies in Canada and the USA. In the USA, the second 'a' is replaced with 'o' giving Okanogan Valley. In this text, we use the Canadian spelling to describe areas north of the 49th parallel as well as the whole valley when the regional context is implied. American spelling is used when describing specific areas south of the 49th parallel.

drumlins (constraining a- and b-axis measurements) was determined from aerial photographs and roughly corresponds to the slope break near the drumlin base and the surrounding areas. Drumlin heights were measured using topographic maps (for the largest) and verified in the field (where possible) with tape measures and levels. Not all drumlins were measured. Visual assessment of the range of forms (in the field and using aerial photographs) was used to target height end-members in order to represent the range of drumlin sizes. Drumlin internal composition was first assessed using aerial photographs (e.g. sediment vs. bedrock drumlins) and verified in the field where possible (Fig. 3.4). Most compositional data reflect surface materials (Table 3.1). Some compositional data are drawn from existing literature (e.g Kovanen and Slaymaker, 2004). Descriptions of drumlin morphology reflect the established terminology of Shaw and Kvill (1984) (Fig. 3.5).

3.5. Description: Landform suites of Thompson Plateau

3.5.1. Drumlins

The landscape of Thompson Plateau and some cross-cutting valleys is ornamented by a continuous swarm of north-south oriented drumlins for a distance of ~110 km (Figs. 3.1, 3.2, 3.3). The Thompson Plateau swarm forms the southern extremity of the much larger Fraser-Thompson-Okanagan (FTO) swarm (175 km long and 10-65 km wide) extending between ~51°N and 48°N latitude in British Columbia and Northern Washington State (Figs. 3.1, 3.2).

Topography has little effect on the orientation of the Thompson Plateau swarm: drumlins occur on plateaus, in some valley-bottoms, and on adverse slopes leading out of valleys. The swarm maintains its orientation across many valleys (Fig. 3.3). Locally, swarm orientation does reflect topography: smaller drumlin fields bifurcate around some bedrock highs. These high points are not streamlined and standout on digital elevation models by their non-lineated texture (Fig. 3.3).

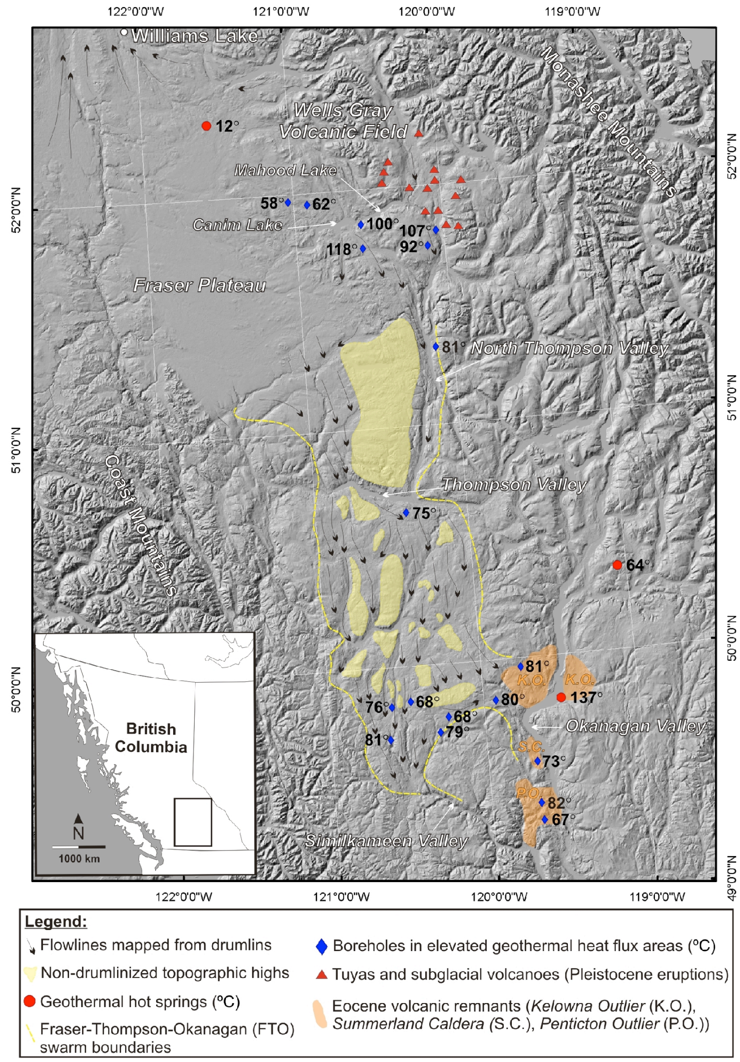


Figure 3.2. Generalized flowline map of the Interior Plateau within British Columbia (inset map). Flowlines are mapped from drumlins in the Fraser-Thompson-Okanagan (FTO) swarm, beginning on Fraser Plateau and extending southward across Thompson Plateau toward Okanagan Valley and Similkameen Valley. Maximum swarm width reaches 65 km. A north-trending drumlin swarm (partially shown) also extends from Fraser Plateau toward Williams Lake. Along the southern swarm, drumlins form a network of interconnected corridors, separated by non-drumlinized topographic highs. High geothermal heat flux occurs along the FTO swarm. The Wells Gray volcanic field is characterized by numerous tuyas and other subglacial volcanoes that erupted during the Pleistocene. Eocene volcanic remnants create high geothermal heat fluxes measured in boreholes. Borehole measurements on Thompson and Fraser Plateau also show high geothermal fluxes. Geothermal hot springs occur along Okanagan Valley and on Fraser Plateau. Elevation data: BC Government TRIM I dataset, 25 m resolution (© GeoBC). BC Albers projection, NAD 83. Geothermal flux data: Fairbank and Faulkner, 1992; Gratsby and Hutcheon, 2001.

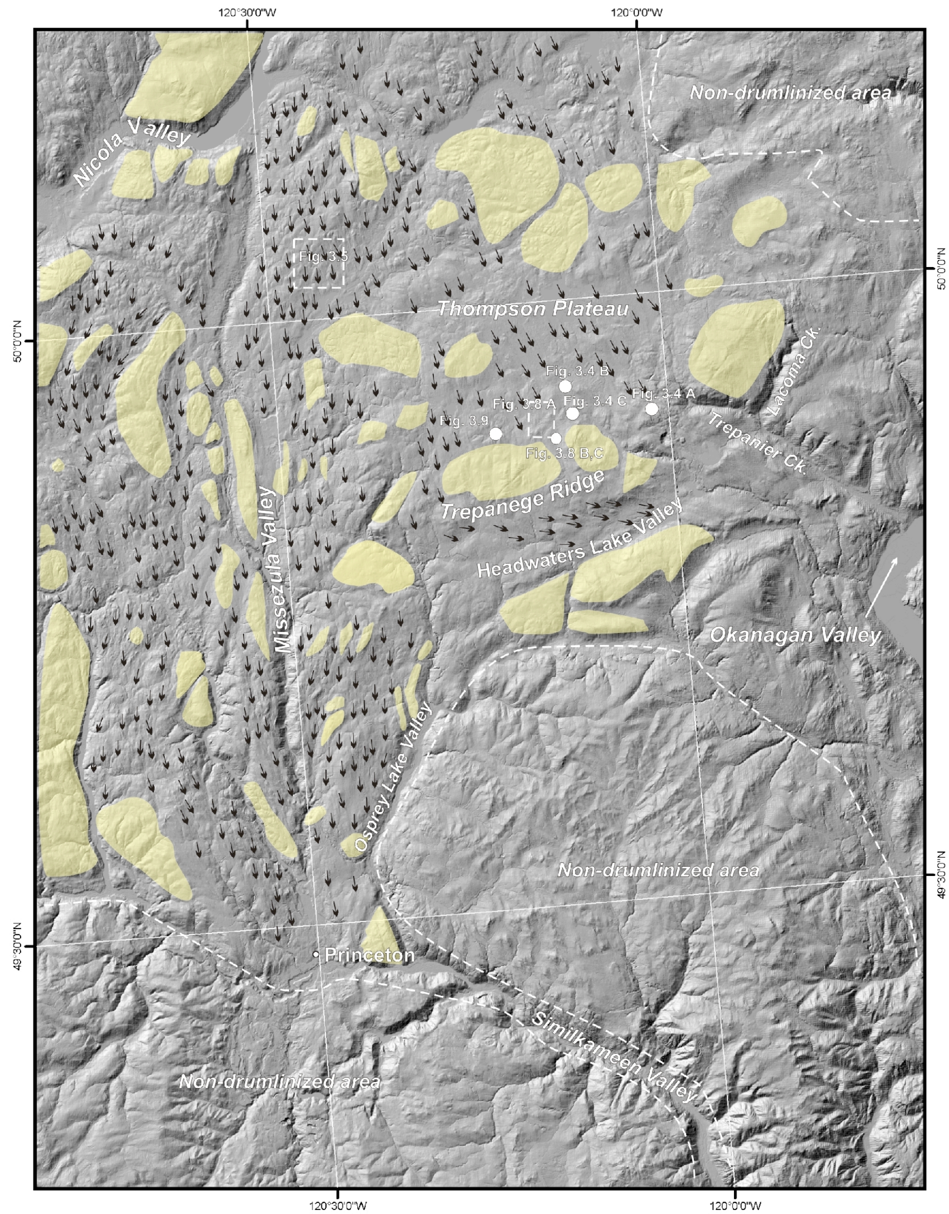


__Figure 3.3.__ ***Hillshaded digital elevation model of Thompson Plateau. Drumlins are pervasive on Thompson Plateau and maintain consistent orientations across valleys. Refer to Figure 2 for symbol legend. Drumlin flowlines are generally north-south. Notable flowline bifurcations occur near Trepanege Ridge and in Headwaters Lake Valley. Missezula Valley is the largest bedrock and sediment-walled valley present on Thompson Plateau. It connects Nicola Valley and Similkameen Valley and exhibits a convex longitudinal profile (Fig. 6). Elevation data: BC Government TRIM I dataset, 25 m resolution (© GeoBC).BC Albers projection, NAD 83.***

Table 3.1: Summary characteristics of drumlins on Thompson Plateau and in Okanagan Valley

	Location			Single Drumlin Morphometrics				Morphology	
	Drumlin Field Name	Coord. [a]	Area (km^2)	Length (m)	Width (m)	Height (m)	Elongation Ratio	Planform; Longitudinal Profile [b]	Composition
Okanagan Valley (Canada)	**Swan Lake**	50° 17' N. 119° 17' W	23	350-2000	150-300	40-100	2.3:1-6.5:1	Classical morphology, crescentic troughs; asymmetrical	Bedrock, glaciolacustrine drape
	Ellison Ridge	50° 10' N. 119° 22' W	13.5	150-700	60-250	15-60	2.4:1-6:1	Rhomboidal forms in space-filling pattern, controlled by bedrock structure, flute clusters downflow of obstacles	Bedrock, glaciofluvial sediment in troughs, thin sediment cover on some
	Wood Lake	50° 03' N. 119° 22' W	10	400-800	20-45	8-15	14:1-20:1	Spindle, crescentic stoss-side troughs, overprinted glacial lake shorelines	Bedrock, stratified and massive diamicton
	Kelowna	49° 57' N. 119° 25' W.	22	200-400	65-140	15-40	2.8:1-3.1:1	Isolated forms, classical morphology, crescentic troughs; asymmetrical	Bedrock knobs, glaciolacustrine veneer
	Squally Point	49° 45' N. 119° 42 W.	40	300-1500	150-400	15-80	2:1-4:1	Symmetrical and asymmetrical forms, flanked by channels; undulating crestline	Bedrock
Okanogan Valley (USA)	**Tonasket**	48° 44' N. 119° 31' W.	30	380-700	180-350	10-40	1.8:1-2.2:1	Classical morphology; symmetrical and assymetrical longitudinal profiles	Bedrock, glaciofluvial sand and gravel onlapping flanks, sand drape over crests
	Omak Plateau	48° 13' N. 119° 30' W.	105	250-2800	85-400	unknown	3:1-7:1	Classical morphology in bedrock, spindles in sediment	Bedrock, gravel, sand, diamicton
	Columbia Plateau [c]	47° 53' N. 119° 30' W.	550	100-2000	100-500	14-30	3:1-8:1	Classical assymetrical drumlins, parabolic, spindle	Diamicton, stratified sand and gravel, bedrock knobs at heads[c]
Thompson Plateau	**South Nicola Valley**	50° 02' N. 120° 25' W.	280	100-500	60-300	5-30	1.6:1-8.3:1	Classical morphology, *spindle* drumlins overprint larger *transverse asymmetrical* forms, *parabolic* drumlins, crescentic troughs, preferential development at positive steps, *en echelon* arrangement	Diamicton, bedrock knobs at heads of some forms
	South Trepanege Ridge	49° 58' N. 120° 10' W.	390	90-200	50-120	10-35	1.6:1-1.8:1	Classical symmetrical morphology	Diamicton, stratified sand and gravel
	Missezula Valley	49° 50' N. 120° 32' W	200	100-400	60-200	35-70	1.5:1-2:1	Classical asymmetrical morphology, crescentic stoss-side troughs	Bedrock, diamicton, glaciofluvial sediments

[a] Coordinates are approximate and refer to a point close to the centre of a designated drumlin group. b Morphological terminology aster from Shaw and Kvill (1984)
[c] Some drumlin composition data from Kovanen and Slaymaker (2004)

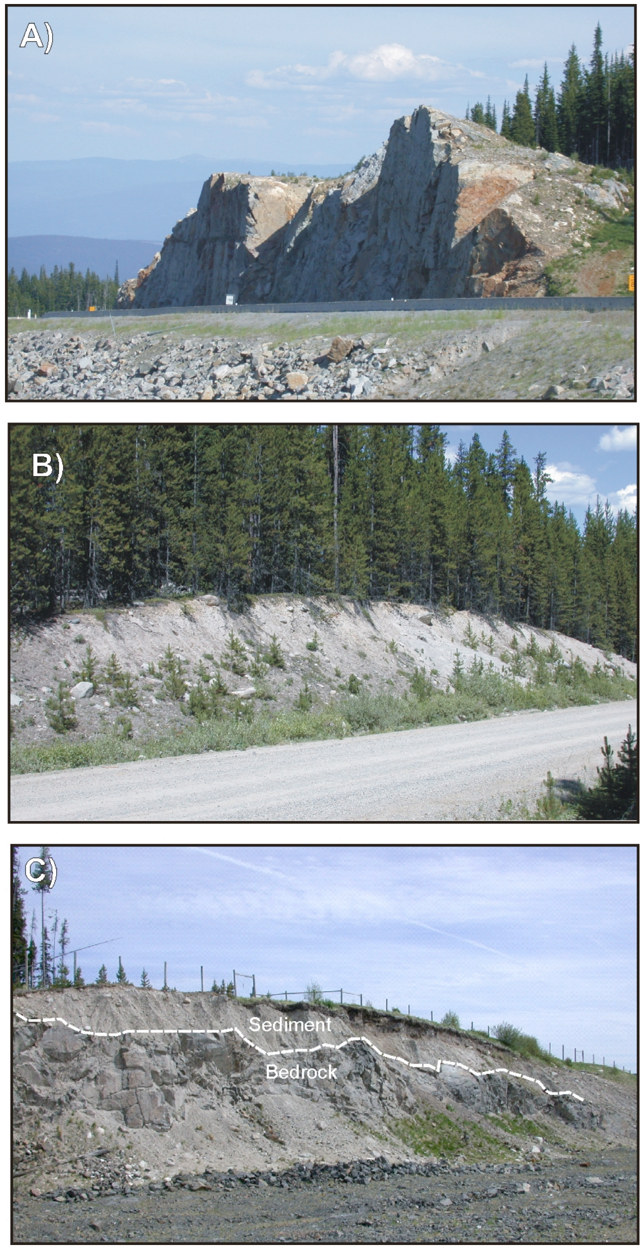


Figure 3.4. Thompson Plateau drumlins composed of A) bedrock (exposure height: 20 m), B) stratified pebbly sand diamicton and discontinuous silty sand beds interpreted to record deposition within a linked cavity (exposure height: 2 m), and C) combinations of bedrock and consolidated loamy diamicton interpreted to record lodgement and compression against Trepanege Ridge (exposure height: 6.5 m). Refer to Figure 3.3 for locations.

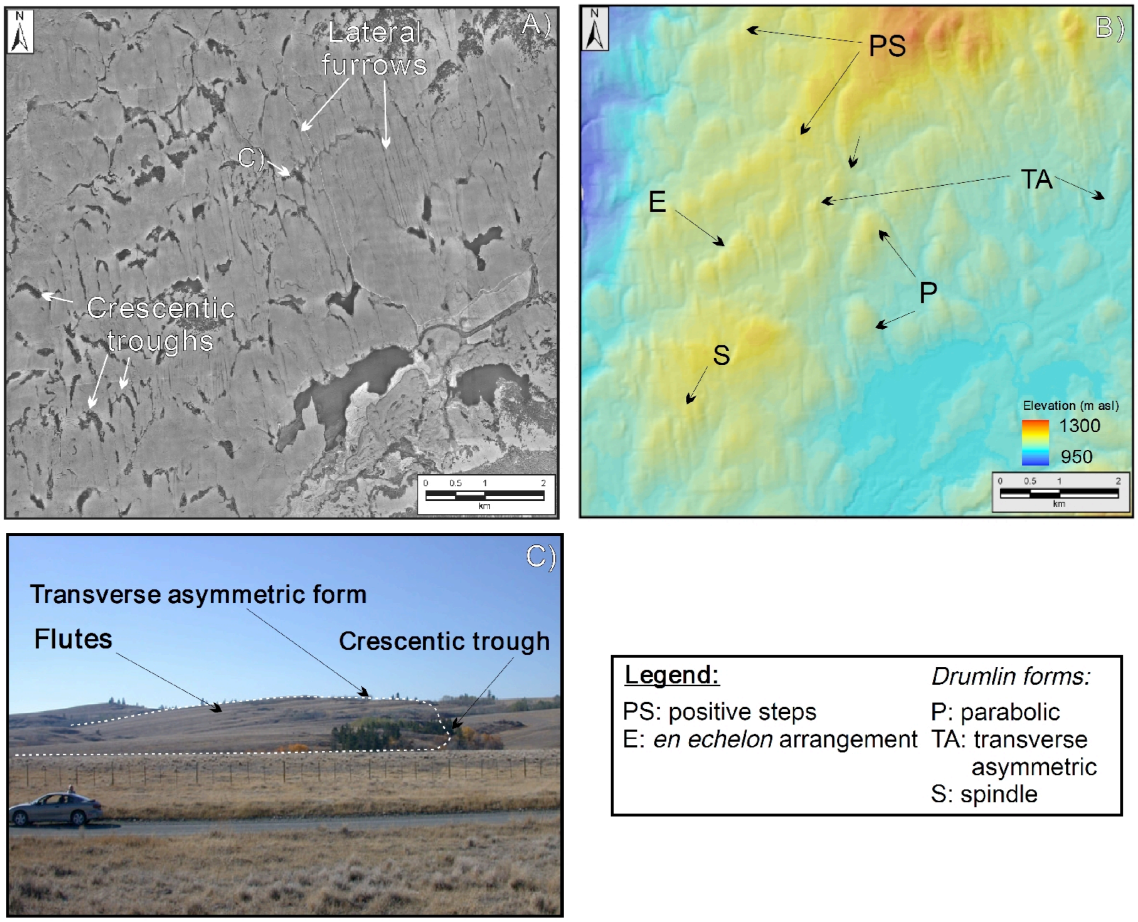


Figure 3.5. Morphological characteristics of drumlins on Thompson Plateau (refer to Fig. 3.3 for location). A) Drumlins exhibit stoss-side crescentic troughs sometimes occupied by lakes, and lateral furrows BC Aerial photograph #15 BC96057-88 (copyright British Columbia Government, by permission). B) Drumlins develop preferentially over positive steps and have en echelon arrangement. Drumlin shapes include parabolic forms, transverse asymmetric forms, and spindle forms. C) Field photograph of bedrock flutes superimposed on larger transverse asymmetric forms (dotted line; partially visible). A prominent crescentic trough occurs at the head of this composite form. Height of transverse asymmetric form is 6 m.

Some valleys trending oblique to the main swarm direction also contain smaller drumlin fields with alignments deviating from the regional (N-S) trend. Examples of this occur near the plateau edge, bordering Okanagan Valley. There, a ~20 km-wide field of bedforms diverges from the main swarm and is oriented SSE toward Trepanege Ridge and Okanagan Valley (Fig. 3.3). Along the easternmost margin of the swarm, bedforms terminate abruptly at the edge of the plateau. Near its southern margin, the Thompson Plateau drumlin swarm converges toward the Similkameen Valley and ends abruptly at the city of Princeton. Drumlins are absent immediately south of Princeton but re-occur within Okanogan Valley between the 49th parallel and the Columbia River (Fig. 3.1).

Thompson Plateau drumlins contain a range of materials including bedrock, unconsolidated sediments (diamictons and glaciofluvial/glaciolacustrine sediments) and combinations of these materials (Fig. 3.4; Table 3.1) (Chapter 2). They exhibit parabolic, spindle and transverse asymmetric planforms (Fig. 3.5) (Shaw and Kvill, 1984). More classic lemniscate forms (Chorley, 1959) also occur, but are less frequent. Drumlins exhibit greater relief and asymmetric long profiles on adverse slopes and over bedrock steps where they form en echelon clusters (Fig. 3.5). Where topography is more subdued, drumlin heights decrease, and their planforms become more elongate and symmetrical. Bedrock knobs occur at the heads of some drumlins. Crescentic depressions and crescent-shaped lakes are present around the head of the majority of drumlins on adverse slopes and at positive bedrock steps. These depressions often extend along drumlin flanks to form lateral furrows, defining the margins of individual forms (Fig. 3.5). At the heads of en echelon clusters, coalesced crescentic depressions in places form sinuous channel-like troughs that may be water-filled (Fig. 3.5 A).

3.5.2. Valleys

Valleys crossing Thompson Plateau are eroded into bedrock and sediment at two scales. The largest valleys are as long as 100 km with variable depths reaching a maximum of 150 m (Missezula Valley) (Figs. 3.3, 3.6; Table 3.2). They lie parallel to the regional drumlin swarm. Some traverse the entire plateau and connect transverse bounding valleys such as the Nicola and Similkameen valleys. Drumlins occur the valley walls and shoulders. Modern rivers have incised into a discontinuous sediment fill along most of their length.

Smaller valleys occur in groups and traverse non-drumlinized bedrock highs such as Trepanege Ridge (Fig. 3.7). Their length varies between 10-15 km and their depth can be as much as 50 m (Table 3.2). These valleys begin among drumlins and are eroded in sediment at their heads and terminations. They are bedrock-walled along their elevated middle sections. Sorted and heterogeneous gravel deposits partially fill the valleys, and eskers are present on the floors of low-lying valley segments (Fig. 3.8 A). In all cases, valley incision is greatest along sediment reaches and shallower along bedrock reaches (along their elevated mid-section). Some bedrock valleys form anabranched networks (Garnet and Osprey Lake valley networks, Fig. 3.7). Many large and small valleys exhibit convex longitudinal profiles (constructed from digital elevation data), gaining at least 450 m of elevation for Missezula Valley and up to 200 m for smaller valleys (Figs. 3.6, 3.7; Table 3.2). These valleys terminate near Princeton where they decrease in elevation and merge with the bedrock valley of Similkameen River (Fig. 3.6).

3.5.3. Bedrock outcrops

Bedrock outcrops on Thompson Plateau are concentrated around Trepanege Ridge and occur as knobs near the head of some drumlins. These bedrock surfaces are often striated and their topography is undulating. Up to 1200 m asl, sediment of variable thickness (3-15 m) overlies bedrock. These sediments consist mainly of diamictons with some glaciofluvial deposits. Drumlin forms truncate these sediments and some drumlins contain bedrock (Chapter 2). Above 1200 m asl, sediment cover is absent and only bedrock is exposed. These higher surfaces show incipient drumlinization.

Bedrock valleys on Trepanege Ridge leave upstanding bedrock knobs as interfluves showing poorly developed streamlining (Fig. 3.7). Other interfluves exhibit patches of deeply weathered bedrock (Fig. 3.9). These patches have limited areal extent (100's m^2). The estimated depth of weathering from exposures is at least 7 m and may be greater (Fig. 3.9). Surfaces of deeply weathered outcrops contain rounded to sub-rounded boulders up to 4 m in diameter. Boulders rest either directly on the bedrock surface or on/in a thin mantle (1-2 m thick) of diamicton over bedrock. Whether or not this diamicton is in situ is difficult to determine as most exposures of weathered bedrock result from road construction over Thompson Plateau and sediments have been disturbed near exposures. In some outcrops, partly excavated, rounded corestones surrounded by saprolite remain anchored in the more coherent bedrock below (Fig. 3.9).

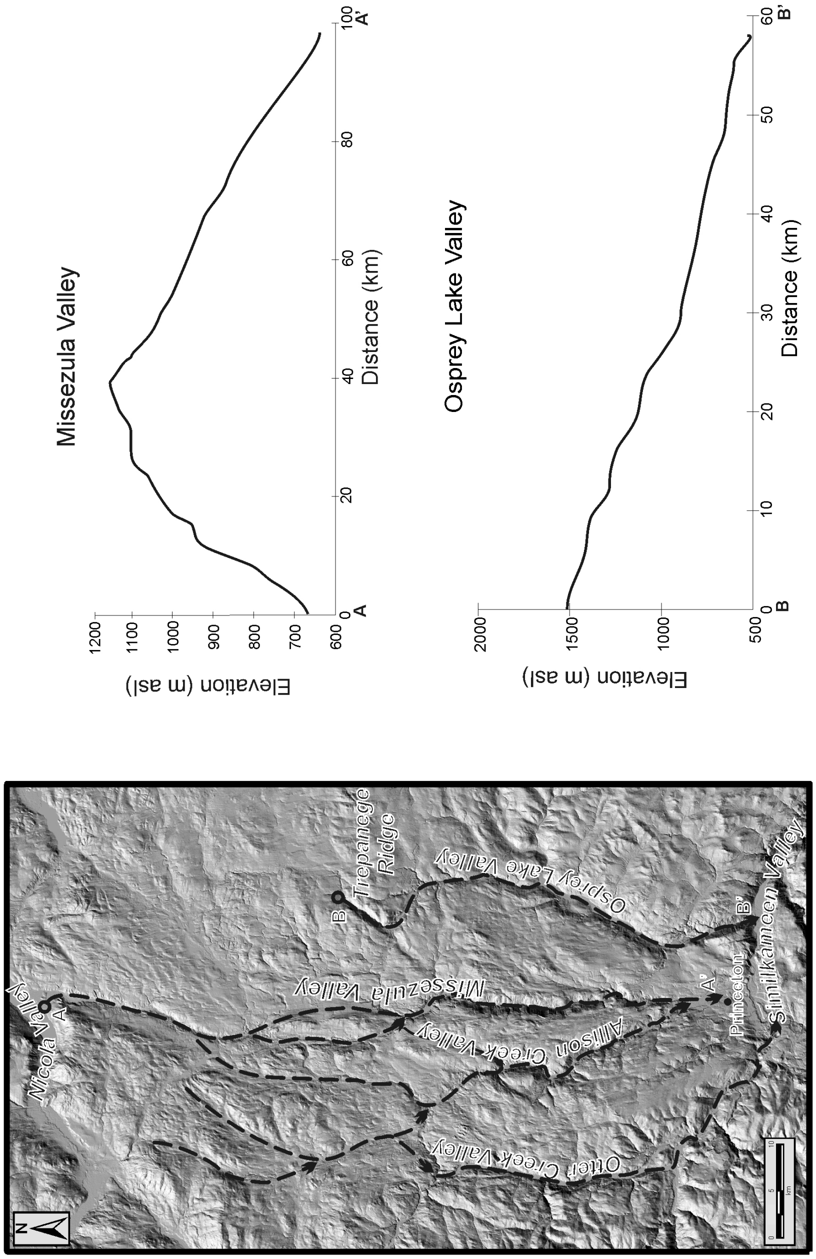


Figure 3.6. ***Hillshaded digital elevation model of mainly bedrock valleys (refer to Fig. 3.1 for location). Longitudinal profiles of Missezula and Osprey Lake valleys. Missezula valley gains over 400 m along its length and has an anabranching pattern near its crest. It forms part of a larger network of incipient valleys that progressively become better-defined toward Similkameen Valley (Otter Creek, Allison Creek and Osprey Lake valleys). All valleys are flanked by drumlins along their length and terminate abruptly at the Similkameen Valley. Elevation data: BC Government TRIM I dataset, 25 m resolution (© GeoBC, by permission). BC Albers projection, NAD 83***

Table 3.2. Summary characteristics of prominent bedrock and sediment-walled valleys on Thompson Plateau and in Okanagan Valley

	Name	Coordinates [a]	Length (km)	Width (m)	Depth (m)	Elevation gain, loss (m) [b]	Channel Boundary; Fill
Thompson Plateau	**Missezula Valley**	49° 54' N, 120° 34' W	100	300-200	60-150	+100, -350	Bedrock; glaciofluvial floor
	Osprey Lake Valley	49° 40' N, 120° 20' W	58	150-1000	80-125	+50, -800	Bedrock; glaciofluvial floor
	Trepanege A-A'	49° 53' N, 120° 10' W	10.2	60-100	35-100	+150, -400	Sand and gravel upflow and downflow ends; bedrock middle section
	Trepanege B-B'	49° 52' N, 120° 12' W	9.7	60-100	25-80	+200, -250	
	Trepanege C-C'	49° 53' N, 120° 14' W	9.9	60-100	25-80	+300, -300	"
	Trepanege D-D'	49° 53' N, 120° 14' W	8	60-100	25-65	+350, -500	"
	Trepanege E-E'	49° 53' N, 120° 14' W	9.8	60-100	25-80	+100, -150	"
	Trepanege F-F'	49° 53' N, 120° 15' W	7.9	60-100	25-80	+250, -400	"
	Trepanege G-G'	49° 53' N, 120° 15' W	8.4	60-100	30-70	+100, -350	"
Okanagan Valley	**Cougar Canyon**	50° 10' N, 119° 17' W	9.5	200-700	60-85	+50, -160	Bedrock; unknown fill
	Goode Creek	49° 45' N, 119° 41' W	7.8	75-125	35-40	+250, -550	Bedrock; glaciofluvial floor
	Wildhorse Canyon	49° 45' N, 119° 42' W	7.6	75-125	35-45	+200, -275	Bedrock; glaciofluvial floor
Okanogan Valley	**Omak Lake Valley**	48° 17' N, 119° 24' W	42	600-5000	80-200	+325, -300	Bedrock; sand, boulder gravel and lacustrine silt fill
	Soap Lake Valley	48° 14' N, 119° 37' W	25	400-3000	80-130	+430, -500	Bedrock; no fill

[a] Approximate midpoint valley coordinates

[b] Most valleys have convex longitudinal profiles rising to a crestpoint before dropping. Elevation gain refers to maximum elevation increase along upflow segment. Elevation drop referes to maximum elevation decrease along downflow segment.

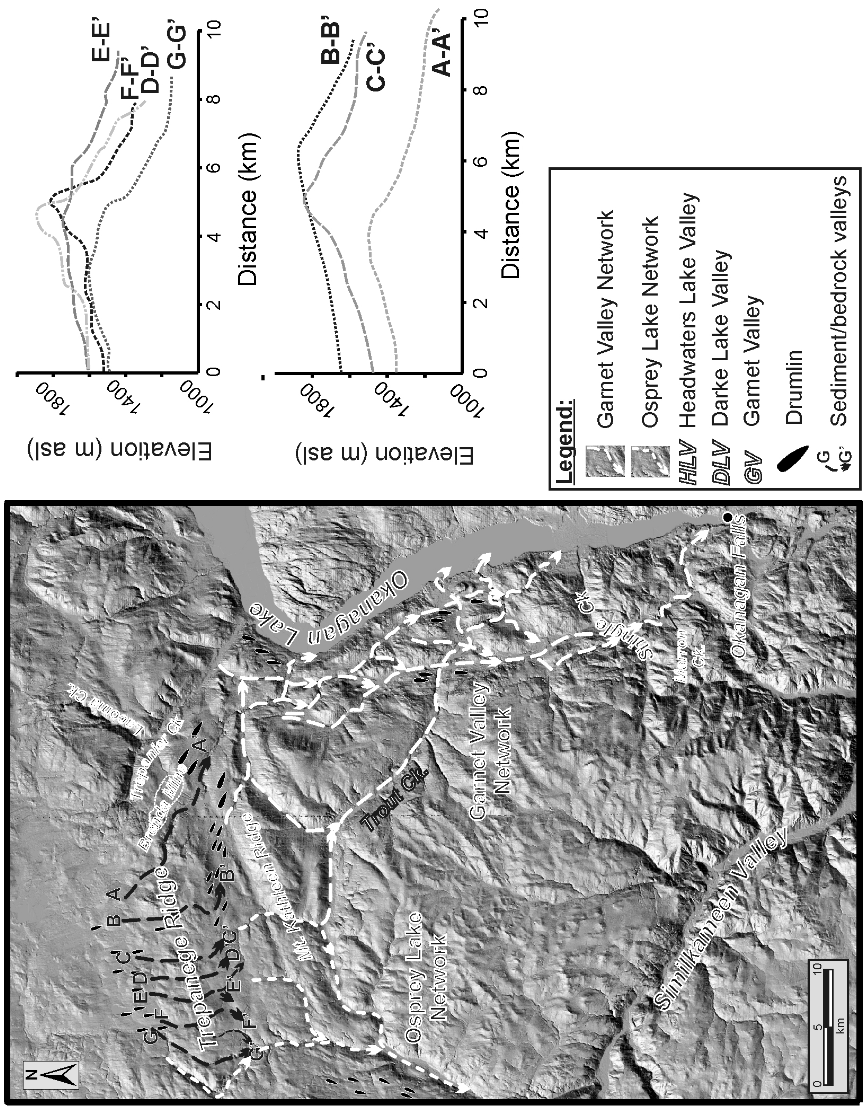


Figure 3.7. ***Mainly bedrock valleys along the west side of Okanagan Lake (see Fig. 4.1 for location). Valleys (A-G) traversing Trepanege Ridge exhibit convex longitudinal profiles (constructed from digital elevation data: BC Government TRIM I dataset, 25 m resolution, © GeoBC, by permission). Valley walls are composed of sediment in their headward and downflow reaches and bedrock along elevated middle sections across Trepanege Ridge. Valleys initiate among drumlins upflow of Trepanege Ridge. A secondary drumlin field occurs in Headwaters Lakes Valley, where bedrock valleys (A-G) debouch. A single channel crosses Mt. Kathleen Ridge and connects to Okanagan Valley via Darke Lake Valley and Garnet Valley, forming the anabranched Garnet Valley Network, and to Similkameen Valley via the Osprey Lake network. Selected drumlins occurring near the transition to/from bedrock valleys highlight the close spatial relationship between these landforms.***

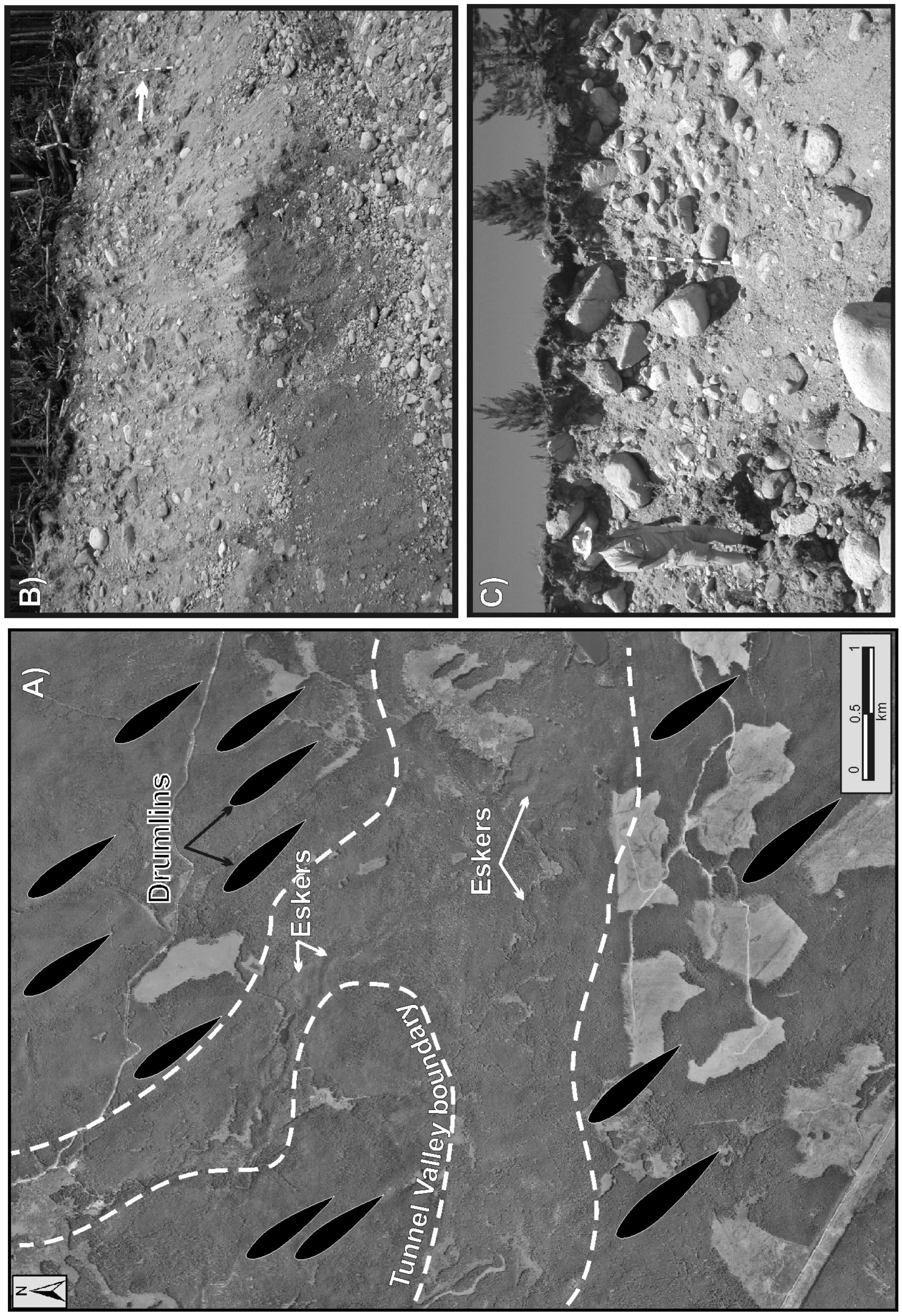


Figure 3.8. ***A) Aerial photograph showing a segment of a sediment-walled tunnel valley cross-cutting the Thompson Plateau drumlin swarm. Eskers occupy the valley floor. B) Sorted and imbricate cobble gravel partially fills valleys and creates low-relief undulations. C) Heterogeneous gravel also occurs within valleys and forms irregular mounds (rod is 1 m long). BC Aerial photograph # 15BCB96056-185 (copyright British Columbia Government, by permission). Refer to Figure 3.3 for locations.***

3.5.4. Landform associations

There is close spatial association between drumlins and valleys on Thompson Plateau. This is exemplified near Trepanege Ridge (Fig. 3.7). This ridge exhibits an altitudinal gradient in drumlin composition and landform types. At the lowest elevations, drumlins are composed of sediments. With increasing elevation on the ridge flanks, drumlins are composed of both sediment and bedrock until only incipient bedrock drumlins and non-drumlinized bedrock occur (above 1200 m asl.). Valleys begin in sediment among drumlins (Fig. 3.8 A) on the north side of Trepanege Ridge and continue through the transition in drumlin composition. Above the drumlin termination (on the northern ridge flank), bedrock valleys cut across the ridge (Fig. 3.5). Ridge elevation

Figure 3.9. Deeply-weathered bedrock outcrop on Trepanege Ridge partially covered by diamicton (0.3-1.5 m thick). Some large corestones are partially excavated and remain anchored in saprolite. Exposure height is 4 m. Refer to Figure 3.3 for location.

decreases gradually to the west where drumlins curve around the western ridge flanks toward Headwaters Lake Valley (Fig. 3.7). There, they form a small swarm within the confines of the southern flanks of Trepanege Ridge and Mt. Kathleen Ridge (Fig. 3.6). Glaciofluvial sediments (4-12 m thick) overlying diamicton and bedrock floor this valley. Within Headwaters Lake Valley, most drumlins are composed of glaciofluvial sediments and diamicton. Valleys traversing Trepanege Ridge merge with the small drumlin swarm in Headwaters Lake Valley (Fig. 3.7). This small swarm terminates abruptly against the north slope of Mt. Kathleen Ridge. Steep-sided and narrow bedrock valleys cut across this ridge.

3.6. Descriptions: landform suites in Okanagan Valley

3.6.1. Drumlins

Drumlins occur within the confines of Okanagan Valley. They are mainly composed of bedrock and exhibit variable planforms (Table 3.1). In some rare cases, drumlins are composed of sediment with protruding bedrock knobs at their heads or along their length. Six locations illustrate these characteristics: three fields of drumlins in the Okanagan Valley between Vernon and Kelowna (Fig. 3.10), and three fields in the Okanogan Valley between the Canada-USA Border and the Withrow Moraine (Fig. 3.11). The Swan Lake drumlin field (Fig. 3.10 A, B; Table 3.1) occurs on a prominent bedrock ridge rising above the modern valley floor (Fig. 3.10 A). These bedrock drumlins exhibit stoss-side crescentic depressions. A discontinuous, decimetre-thick veneer of glaciolacustrine silt onlaps these drumlins. A more prominent bedrock interfluve separating Okanagan Lake from Kalamalka Lake and Wood Lake preserves the Ellison Ridge drumlin field, southeast of Vernon (Fig. 3.10 C; Table 3.1).

These drumlins are composed entirely of bedrock and occur on the crest and western flank of the interfluve. They exhibit rhomboidal planforms with a space-filling arrangement controlled by bedrock joints (Fig. 3.10 C). A discontinuous veneer (0.2-0.5 m) of diamicton caps some drumlins. Finally, more elongate, spindle drumlins and flutes occur on the east side of Wood Lake (Fig. 3.10 D; Table 3.1). These forms are more subdued than other drumlins in this valley. They are composed of massive and stratified diamictons with some bedrock knobs protruding at their heads and in places along the forms. Crescentic troughs occur at the heads of drumlins. However, meltwater channels draining toward the modern valley at angles oblique to drumlin orientations, have reoccupied portions of these troughs and dissected them (Fig. 3.10 D). Shorelines from extensive deglacial lakes also overprint the forms (Nasmith, 1962).

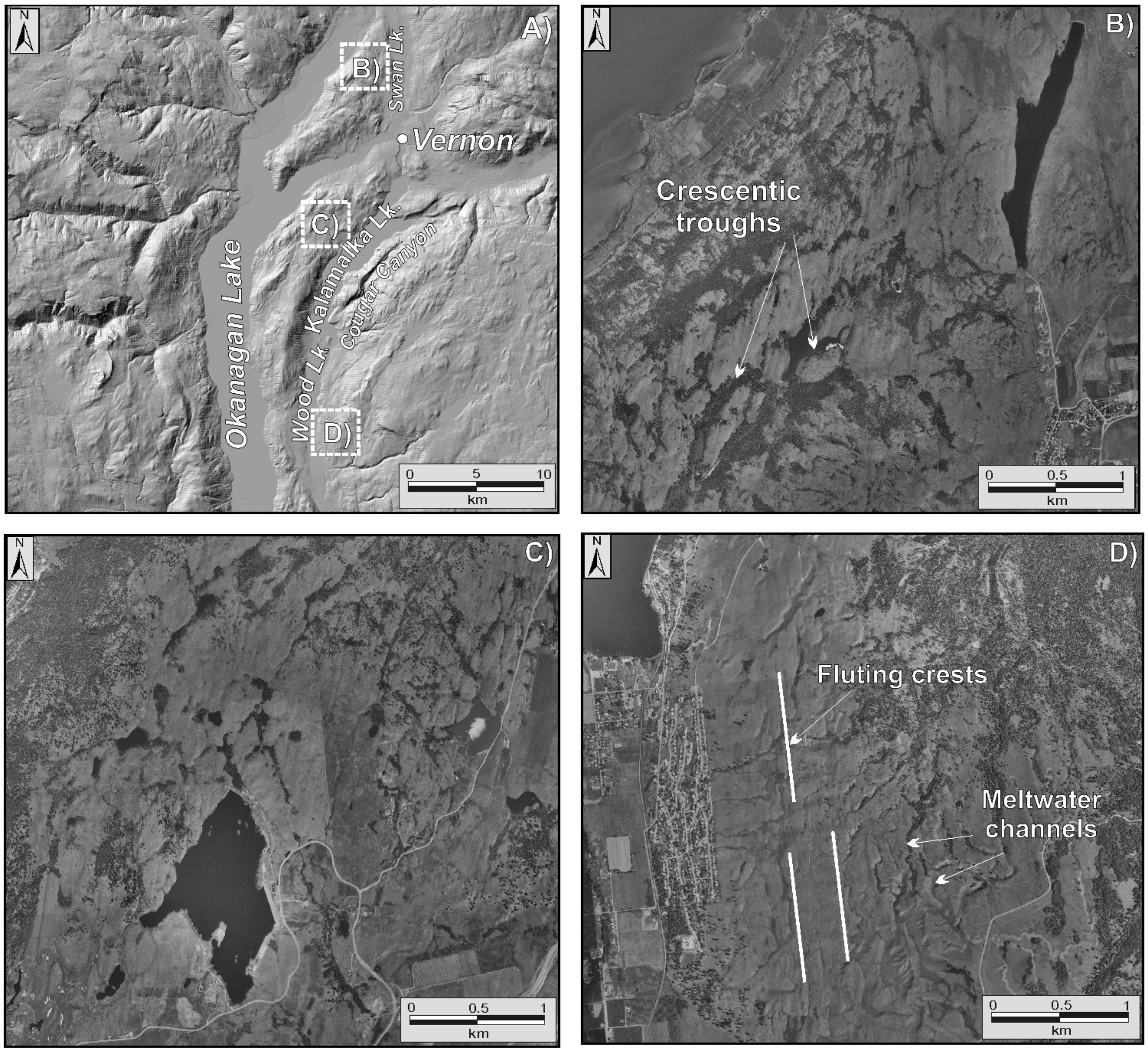
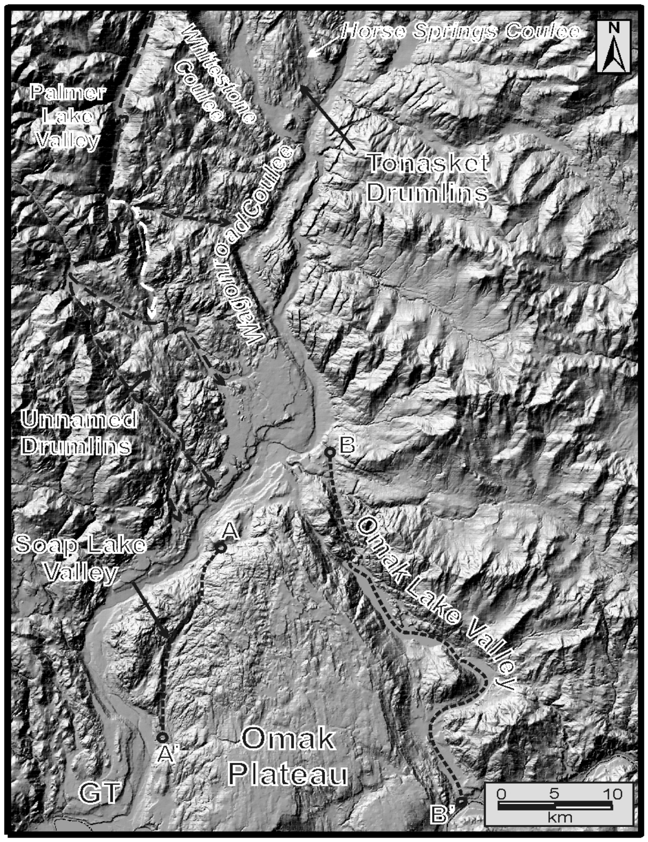
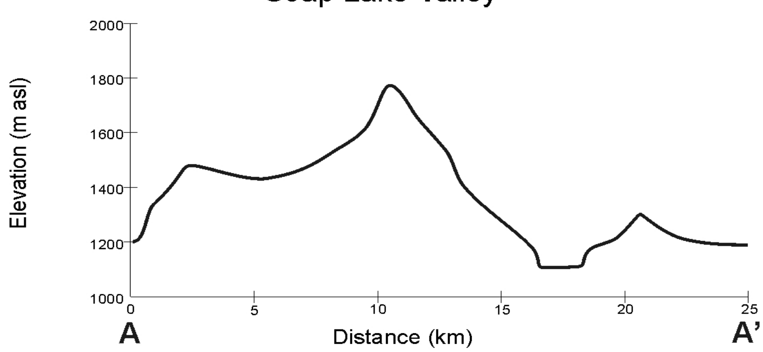


Figure 3.10. A) Bedrock drumlins and tunnel valleys in Okanagan Valley (see Fig. 1 for location). Cougar Canyon is a prominent bedrock valley on the east side of Kalamalka Lake. B) Swan lake drumlins exhibit stoss-side crescentic troughs. C) Ellison Ridge drumlins are arranged in a 'rhomboidal' pattern dictated by bedrock structure. D) Wood Lake drumlins and flutings (crests highlighted by white lines) where crescentic troughs and meltwater channels coalesce and form short channels draining toward Wood Lake valley floor. Elevation data: BC Government TRIM I dataset, 25 m resolution. BC Albers projection, NAD 83. BC Aerial photographs # 15BC97023-126, 15BC97027-43, 45 (copyright British Columbia Government, by permission).



Soap Lake Valley



Omak Lake Valley

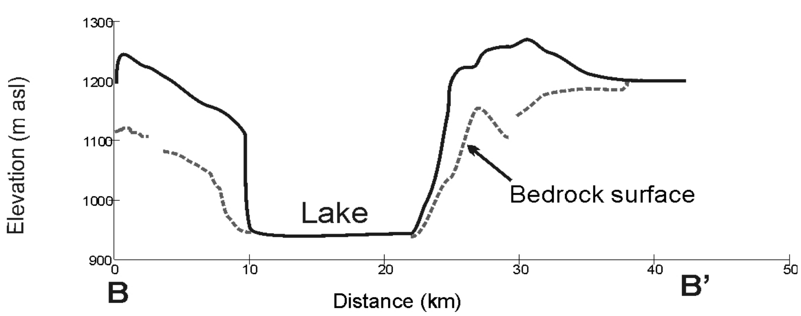


Figure 3.11. Hillshaded digital elevation model illustrating drumlins, coulees and valleys of Okanogan Valley, USA (see Fig. 3.1 for location). Drumlins are composed of bedrock (Tonasket group) and sediment (Omak Plateau group). Omak Plateau surface has a 'scabbed' appearance characterized by patchy removal of surficial sediments exposing bedrock. Prominent bedrock valleys flank drumlin outcrops. Soap Lake and Omak Lake valleys have convex longitudinal profiles (constructed from digital elevation data: USGS 10 m DEM). For Omak Lake valley, this is the case for both the modern channel floor and the underlying bedrock surface (determined from water well logs, source: Washington State Department of Ecology). Tonasket drumlins are flanked by Whitestone and Horse Springs Coulees oriented with downsloping gradients. The southern extension of Palmer Lake Valley (yellow dashed arrows) flanks an unnamed group of bedrock drumlins. Wagonroad Coulee is a hanging bedrock valley parallel to the Okanogan Valley. It does not presently carry water. The Great Terrace (GT) at the confluence of Okanagan Valley and the Columbia River is composed of sand and gravel.

In Okanogan Valley, bedrock drumlins occur near Tonasket, WA, in small fields on interfluves between valleys (bedrock coulees) (Fig. 3.11). More extensive fields occur on Omak Plateau (Fig. 3.11) and on the Columbia Plateau (Fig. 3.1) where they are composed of sorted and stratified sediments (based on water-well logs) and diamictons, respectively (Kovanen and Slaymaker, 2004). On Omak Plateau, individual drumlins are continuous across sediment-bedrock outcrops. This drumlinized surface also exhibits a ‘scabbed’ appearance (Fig. 3.11) characterized by patches of exposed bedrock.

3.6.2. Valleys

Bedrock valleys of various scales are present in the Okanagan Valley corridor. The most prominent of these is Okanagan Valley itself. The longitudinal bedrock profile of Okanagan Valley contains high points near Vernon and Okanagan Falls (Fig. 3.1) (Eyles et al., 1990; 1991; Vanderburgh and Roberts 1996). However, bedrock remains 300 m and over 275 m below the modern valley floor for Vernon and Okanagan Falls respectively. Lake-based and land-based seismic surveys have revealed a V-shaped bedrock valley cross-section, and a long profile characterized by overdeepenings (reaching as much as 300 m below sea level) (Eyles et al., 1990, 1991).

Smaller-scale (5-10 km length) bedrock valleys occur on the sides of Okanagan Valley. Prominent examples include Cougar Canyon east of Kalamalka Lake (Fig. 3.10 A), Wildhorse Canyon and Goode Canyon traversing Squally Point (Fig. 3.12), Palmer

Lake valley west of the Tonasket drumlins (Figs. 3.11, 3.13), and a number of bedrock coulees near Tonasket and Omak, Washington and on the flanks of Omak Plateau (Fig. 3.11). These valleys share many similar morphological characteristics including convex longitudinal profiles (gaining as much as 300-400 m elevation, based on digital elevation data) and overdeepening(s) along their length (Figs. 3.11, 3.12; Table 3.2). Sediments of variable thickness partially fill some channels (Table 3.2). In some cases, convex longitudinal profiles describe the bedrock surface as well as the modern valley floor (Fig. 3.11).

A >25 km-long network of anabranching bedrock valleys, herein named the Garnet Valley network, occurs west of Okanagan Lake and south of Mt. Kathleen Ridge (Fig. 3.7). This valley network connects the Thompson Plateau and Okanagan Valley landsystems via segments of Trout Creek, Shingle Creek and Marron Creek. Valleys are oriented oblique to Okanagan Valley and become increasingly parallel as they debouch along Trout Creek, Shingle Creek and ultimately at Okanagan Falls through a series of deeply incised, steep, narrow and sinuous bedrock gorges. Lakes occupy some longitudinal depressions along the bedrock floors. Glaciofluvial and glaciolacustrine sediments partially fill some bedrock valleys and form modern valley floors. In many places, modern streams have excavated the sediment fill, in some cases to bedrock.

3.6.3. Bedrock outcrops and forms

Between drumlin fields (described above), extensive portions of Okanagan Valley (up to 70 km in length) have few to no drumlins. Instead, the valley walls are generally sediment-free between 400 m and 750 m asl. Sediments are concentrated in tributary valleys and in benches of lacustrine sediments resting directly on bedrock between ~342-410 m asl along Okanagan Valley. Till is absent from the valley sides. Rarely, striae are observed on valley walls, in places where the bedrock surface has been recently exhumed. Exposed bedrock is undulating (100-400 m wavelength). Pronounced lineations along structural weaknesses produce groups of linear residual hills resembling whalebacks and roches moutonnées (e.g., Squally Point, Fig. 3.12). Residual forms are defined by flanking troughs (cf. Bradwell, 2006). In some locations, troughs are smooth, sinuous and channel-like. Bedrock residuals are highly weathered. Structural lineations are present in the bedrock and are sub-parallel to ice flow direction along the valley axis.

Thin (4.8 m) fills of sand, sub-rounded and sub-angular gravel, and silt occupy some trough floors. Few troughs presently contain streams and the rare streams on Squally Point are intermittent. Sedimentation in deglacial lakes (Nasmith, 1962) associated with decay of the CIS accounts for the presence of silt in troughs and on some bedrock highs.

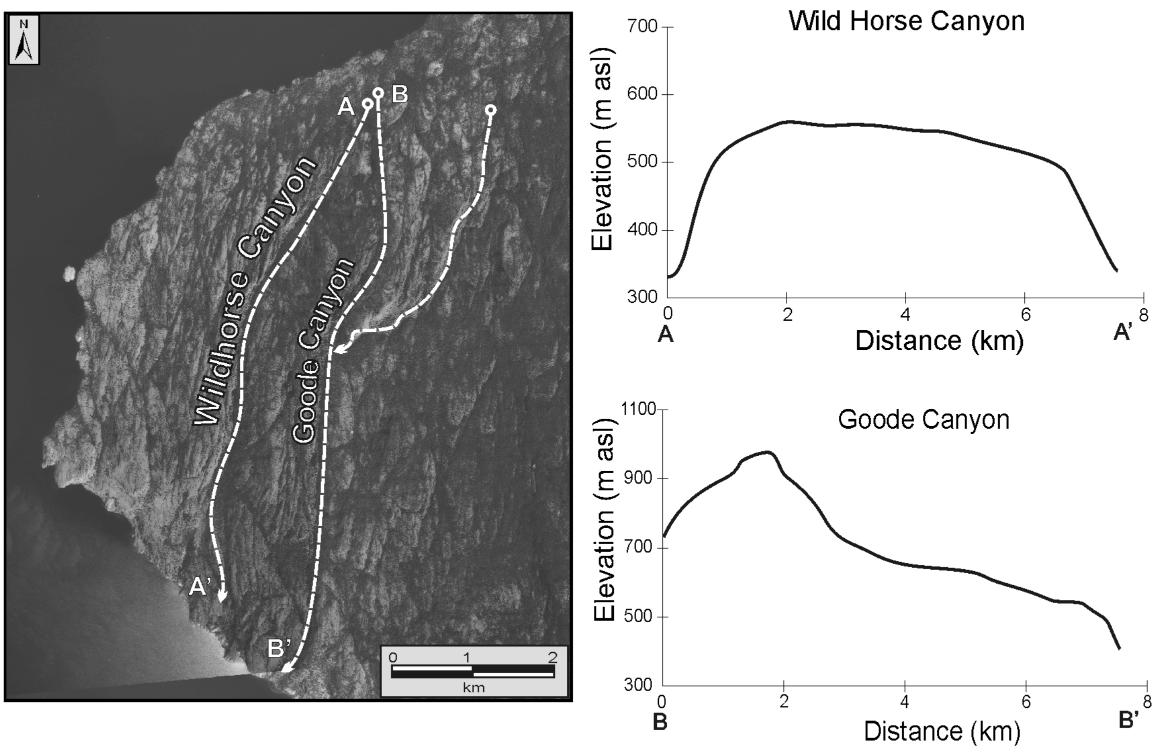


Figure 3.12. Aerial photograph of bedrock forms at Squally Point in Okanagan Valley (Refer to Fig. 3.1 for location). These forms resemble whalebacks (symmetrical) and roches moutonnées (asymmetrical). Sinuous troughs define morphology and size of the residual forms. Surfaces are devoid of surficial sediment, save for troughs that are floored with glaciofluvial sand and gravel. Goode Canyon and Wildhorse Canyon are two prominent bedrock valleys traversing Squally Point. They exhibit convex longitudinal profiles (constructed from digital elevation data: BC Government TRIM I dataset, 25 m resolution). BC Aerial photograph #BC4195-033 (copyright British Columbia Government, use by permission).

Figure 3.13. Southward-looking oblique aerial photograph of Similkameen Valley and Palmer Lake Valley. The modern course of the Similkameen River makes a sharp east-northeast turn in a wetland area north of Palmer Lake. Palmer Lake Valley is a bedrock valley with modern underfit and intermittent streams. It may have operated as a tunnel valley during glaciation. Smaller bedrock coulees like Whitestone Coulee branch off toward Okanogan Valley.

3.6.4. Moraines

Some landforms are notable for their absence from the landscape. In particular, major end moraines are not present in Okanagan Valley or on Thompson Plateau. This absence is ascribed to deglaciation dominated by regional stagnation rather than active retreat (Fulton, 1967, 1975). The only major morainal landforms are located on Columbia Plateau, 200 km south of Thompson Plateau.

This landsystem includes the Withrow terminal moraine and smaller recessional moraines (Kovanen and Slaymaker, 2004). They are associated with the Okanogan Lobe, fed by ice from surrounding plateaus and mountains in British Columbia and

Washington and converging on the Okanagan Valley before debouching on Columbia Plateau. The distribution of drumlins on Columbia Plateau shows a splaying pattern radiating from the confines of Okanagan Valley (Kovanen and Slaymaker, 2004). They extend to the moraine margin and recessional moraines are draped over them (Kovanen and Slaymaker, 2004). The Withrow moraine remains enigmatic as it appears to contain both depositional and erosional landform elements, including eroded bedrock, and numerous channels breach the moraine.

3.7. Landsystem interpretations

We evaluate possible formative agents and processes for the glacial landsystem of the Thompson Plateau and Okanagan Valley building interpretations from individual landforms to landform associations.

3.7.1. Drumlin genesis: competing hypotheses

The most widely held current view on drumlin genesis invokes a deforming substrate as the agent of formation (Boulton, 1987). Drumlins are produced as the deforming material encounters bed obstructions such as bedrock knobs or zones of better drained (less deformable) material (Boulton, 1987). Deformational drumlins, akin to a sheath fold in the deforming sediment, occur where increasing deformation streamlines the folded mass into a drumlin (Boulton, 1987). Deforming material can accrete in low-pressure zones leeward of bed obstructions to create a depositional drumlin (Boulton, 1987). Finally, streaming deforming bed material can erode a drumlin from pre-existing substrate (Boulton, 1987). In this erosional drumlin scenario, drumlins are erosional remnants with a thin ‘carapace’ of deformation till on the drumlin surface (Boulton, 1987; Hart, 1997). These processes require pervasive sediment deformation, often to a depth equal to the height of the resulting drumlins (Boulton, 1987; Hart, 1997), though the sheath fold hypothesis involves a thinner deforming layer (Boulton 1987). Consequently, material within these drumlins should exhibit deformation over a large portion of the drumlin height, except for erosional drumlins where undeformed remnants may be preserved. A variation of the deforming bed hypothesis invokes excavational streams of deforming sediment, creating troughs between drumlins (Boyce and Eyles, 1991). The bed deformation model and its variants rely on data from experiments at Breiđarmerkurjökull, Iceland showing m-thick bed deformation (Boulton, 1987). Associations of thick deforming beds with modern ice streams rely on seismic interpretation of m-scale deforming sediment layers beneath ice stream B (Alley et al., 1987). Tulaczyk (1999), van der Veen (1999) and Piotrowski et al. (2001) have questioned the results of the Breiđarmerkurjökull experiment and the degree to which they accurately represent bed conditions beneath glaciers and ice sheets. In addition, more recent results from Antarctic ice stream beds suggest that previously inferred m-thick deforming beds overestimate the thickness of the deforming layer (Alley, 2000). Instead, deformation takes place over a thin, cm-scale layer (Engelhardt and Kamb, 1998). Finally, the extent and pervasiveness of deforming beds under paleo-ice sheets has also been questioned (Piotrowski et al., 2001, 2002) and field and laboratory studies substantiate Antarctic results of cm-scale deformation (Iverson and Iverson, 2001; Tulaczyk, 1999). Although there is little doubt that bed deformation occurs under elevated porewater pressure, new findings on the mechanics of bed deformation raise questions about its effectiveness for generating metre to decametre-scale drumlins.

Drumlin genesis has also been attributed to erosion and/or deposition by turbulent meltwater flows associated with subglacial underbursts (e.g. Shaw, 1994 a, b, 1996, 2002). In the meltwater erosion model, drumlins are upstanding residuals initially eroded by horseshoe vortices developed around bed obstacles such as bedrock knobs or other bed defects (Shaw and Sharpe, 1987; Sharpe et al., 2004). With increasing streamlining, erosion occurs by flow shear stress over the form rather than by flow separation (Shaw, 1994 a). The meltwater erosion model can account for groups of drumlins eroded into different substrates within the same drumlin field or swarm, and a land surface unconformity (Sharpe et al., 2004) as is the case on Thompson Plateau. Flow separation within an erosive, sediment-laden underburst, and vortex generation around obstacles can account for crescentic scours (recorded by crescentic lakes today, Fig. 3.5) on the upflow end of drumlins and lateral furrows on drumlin flanks (Kor et al., 1991). In addition, this hypothesis predicts preferential development of drumlins over positive (upflow-facing) steps (Pollard et al., 1996), en echelon arrangements and superimposed and nested scales of drumlins (Shaw, 1996, 2002), all of which are observed on Thompson Plateau.

3.7.1.1. Thompson Plateau drumlins

Based on individual drumlin characteristics and on regional landform associations, drumlins are predominantly erosional forms. Within the same swarm, Thompson Plateau drumlins composed of bedrock, bedrock and diamicton, and diamicton and undeformed sorted and stratified sediments and truncated at the drumlin surface (Fig. 3.4), preclude a depositional origin for these forms (Chapter 2). Sediment within drumlins record lodgement, localized deformation processes, and deposition with proglacial ponds and/or subglacial water-filled cavities. Clast fabrics record regional ice build-up and compression against Trepanege Ridge, rather than drumlin-scale deformation (Chapter 2).

Multiple lines of evidence argue against drumlin erosion by a deforming bed on Thompson Plateau. An erosive deforming bed mechanism implicates a deformation till 'carapace' overlying the drumlinized unconformity (Hart, 1997). This carapace is absent from the Thompson Plateau drumlins. If a thin deforming bed was subject to very high stresses it could be, theoretically, extremely mobile and erosive (Boyce and Eyles, 1991). Under such circumstances it is argued that the till carapace could be absent over sediment drumlins and be replaced by striae and other forms of abrasion on bedrock drumlins. However, erosion of a sediment drumlin surface implies transport of sediment. Deposition is expected if shear stress drops below a critical threshold. Therefore, the notion that sediments (absence of till carapace) would not be preserved following erosion by ice is questionable. The surfaces of some incipient bedrock drumlins in the Thompson Plateau drumlin swarm are striated. These incipient drumlins are weakly streamlined residuals between eroded bedrock troughs. It is unclear if these striae are contemporaneous with drumlin formation as bedrock may have experienced grounded, sliding ice over much of the glacial cycle. Hart (2006) interpreted striated, streamlined bedrock forms in front of alpine glaciers to be the product of abrasion by a thin mobile till sheet. For this process to operate efficiently, till thickening needs to be suppressed and the abraded and eroded material needs to be evacuated from the bed. Without sediment evacuation, such a process may be adequate for producing ornamentations on bedrock surfaces or on drumlins but it is unlikely to produce large-scale bedrock drumlins (8-15 m high). Striae may simply be created by abrasion associated with glacier sliding.

A fully erosive sediment deformation mechanism (Boyce and Eyles, 1991) for forming erosional drumlins should result in large sediment fluxes downflow of the eroded drumlins. Termination of drumlins also marks cessation of the erosive action of the deforming bed. Eroded sediments should therefore occur near the termination of drumlins. No end moraines exist at the sharply-defined margin of the swarm or within the Thompson Plateau drumlin field, and all drumlins appear to be erosional (there is no apparent downflow transition to depositional or deformational drumlins based on available data). The area around Trepanege Ridge and Headwaters Lake Valley (Fig. 3.7) further exemplifies the difficulties in accounting for the eroded sediments in this fully erosive context. Landform associations in this area show downflow transitions between drumlins and bedrock valleys crossing Trepanege Ridge (Fig. 3.7). In the eastern portion of the drumlin swarm, upon encountering Trepanege Ridge, drumlins give way to sediment-walled then bedrock valleys across the ridge, and bedrock valleys become sediment-walled valleys that taper into the Headwaters Lake Valley drumlin field downflow of the ridge. To the west, where ridge elevation is lower, some drumlins curve around the Ridge's western flanks and join the Headwaters Lake Valley drumlin field. This field terminates abruptly against the bedrock walls of Mt. Kathleen Ridge. Only a small bedrock valley cuts across Mt. Kathleen Ridge and, more importantly, no major sediment accumulations are present at the termination of the swarm. This observation is inconsistent with drumlin erosion by a mobile deforming bed. An erosive mobile bed should have excavated material between drumlins and conveyed it downflow. A transition to depositional landforms (e.g., drumlins or end moraines) downflow of the field is required to account for sediment continuity. This transition is not observed as the drumlins abut directly against bedrock walls. Consequently, erosion of the Thompson Plateau drumlins by an abrasive mobile deforming sediment sheet cannot be supported. Arguments of sediment flux conservation are developed more comprehensively in § 3.7.3.

3.7.1.2. Okanagan Valley drumlins

Drumlins in Okanagan Valley are erosional forms. Many are composed of bedrock (Swan Lake, Ellison Ridge, Tonasket and some on Omak Plateau), and bedrock drumlins preclude genesis by folding or deposition of deforming sediment sheets. Sediment drumlins occur on Omak Plateau and above Wood Lake. Some of the Wood

Lake drumlins exhibit bedrock knobs at their heads, and sediment tails. Incipient crescentic troughs, exploiting bedrock joints, occur at the head of some drumlins (Fig. 3.10 B, C). Sediment tails downflow of bedrock knobs could be indicative of lee-side sediment squeezing (Hart, 1995), or, theoretically, of erosion by deforming sediment sheets (e.g. Boulton, 1987). However, drumlin morphological characteristics can be equally well explained by meltwater erosion (Shaw, 1994b, 1996). Determining the formative agent of the Wood Lake drumlins is difficult given the paucity of sediment exposures. However, morphologic elements and the regional context of erosion constrain interpretations. For example, drumlins to the north and south of the Wood Lake forms are eroded in bedrock. If streaming excavational till sheets are invoked for drumlin erosion, there should be evidence of widespread till deposition on the valley sides and in the valley fill. No such deposits are present in these locations (Eyles et al., 1990; Vanderburgh and Roberts, 1996). Subglacial meltwater erosion can account for drumlins eroded in a range of substrates within the same swarm, consistent with drumlin characteristics in Okanagan Valley. In addition, sediments removed by meltwater may be stored within the thick valley fill of Okanagan Valley (Vanderburgh and Roberts, 1996).

Drumlins on Omak Plateau are eroded in sediment and bedrock, creating a ‘scabbed’ appearance on the plateau. Landform continuity on Omak Plateau suggests that drumlin erosion in sediment and bedrock was contemporaneous across the plateau. Scabbed patches are unlikely to result from deglacial or Holocene stream erosion as many of them do not correspond to present-day or relict stream courses, and sediment removal is widespread instead of being focused along valleys. Based on arguments integrating regional landforms and sediments for the Okanagan Valley, drumlin fields throughout the valley probably record erosion by turbulence and corrasion in subglacial meltwater flows.

3.7.2. Valleys

Tunnel valleys and/or tunnel channels are ubiquitous landforms of glaciated areas. They consist of linear depressions eroded in sediment and/or bedrock. They are widely recognized as forming subglacially in association with channelized meltwater flows, yet the specifics of their formation are debated (cf. Boulton and Hindmarsh, 1987;

Brennand and Shaw, 1994). Diagnostic morphological characteristics include convex and undulating longitudinal profiles (although this is not a necessary condition) with basins and thresholds, indicative of pressurized subglacial meltwater flow. Persistent questions of tunnel valley/channel genesis pertain to the magnitude and frequency of formative flows. Interpretations of formative events also dictate terminology used to classify these forms. For example, tunnel valleys result from flows that only partially fill valleys (Boulton and Hindmarsh, 1987). In contrast, tunnel channels form by flows at bankfull conditions (Brennand and Shaw, 1994; O'Cofaigh, 1996; Cutler et al., 2002). In addition, steady-state and more catastrophic interpretations of tunnel valley/channel genesis have been proposed (cf. Mooers, 1989; Wingfield, 1990). Tunnel valleys/channels often occur amongst drumlin swarms. As well, drumlins occur within tunnel channels and on their margins. These landform associations have been used to argue for a common genetic agent (Brennand and Shaw, 1994). .

Most valleys of Thompson Plateau and Okanagan Valley exhibit convex longitudinal profiles. They also occur in close spatial association with drumlins, even containing them. Some valley floors carry eskers. Therefore, these valleys formed wholly or partly subglacially, by pressurized meltwater flows capable of ascending adverse slopes. They are tunnel valleys. It is difficult to determine whether all of these valleys operated at bankfull conditions. Use of the term tunnel valley, rather than tunnel channel, reflects this uncertainty. However, the possibility that some were tunnel channels at least at some point in their development cannot be ruled out.

Boulton and Hindmarsh (1987) proposed that tunnel valleys form by piping and deformation of a water-saturated glacial substrate, starting at the ice margin. Where porewater pressure is sufficiently high to render the sediment weak, incremental downward and headward erosion by sapping of dilatant unconsolidated sediment can occur. Once initiated, sediments migrate toward low-pressure conduits with evacuation to the ice margin. A counter argument suggests that high-pressure meltwater canals will develop in deforming sediment beds (Walder and Fowler, 1994). Low flow velocities in canals (Walder and Fowler, 1994) likely limit clast entrainment and they are unlikely to erode large tunnel valleys. However, a new definition of canals as sediment-walled channels (Ng, 2009) implies no such limitations on flow velocities. Application of the bed deformation model to Pleistocene tunnel valleys suggests that they may result from incremental step-wise headward incision and episodic ice-margin retreat. (Mooers, 1989; Patterson, 1994). Glaciofluvial fans, often in proximity to end moraines, mark temporary positions of the ice margin during retreat (Patterson, 1994; Cutler et al., 2002).

In contrast, some tunnel valleys/channels may be formed by one or very few short-lived events where the channel conveys water over its entire length, sometimes at bankfull conditions (Wright, 1973; Wingfield, 1990; Brennand and Shaw, 1994; Fisher and Taylor, 2002; Cutler et al., 2002). Given the great length and depth of some tunnel valleys/channels, this may involve voluminous catastrophic release of subglacially and possibly supraglacially stored water (Brennand and Shaw, 1994; Burke et al., 2012). In these cases, one or several overlapping fans could occur at conduit mouths where flow expands. Sedimentation may also take place within the channel itself (Russell et al., 2003). Deposition can take place subaqueously (Russell et al., 2003) and/or subaerially (Cutler et al., 2002). These flows may be fully contained within channel boundaries (e.g. Wright, 1973) or they may start as broad unstable sheetflows that collapse to create discrete bankfull channels (Brennand and Shaw, 1994; Shaw, 1996; Fisher and Taylor, 2002). In the latter case, drumlins may form during the sheetflow phase and tunnel channels during sheet collapse (Shaw, 1996). Waning stage flows of the 1996 Icelandic jokulhlaup evolved from sheetflow (Flowers et al., 2004) to more channelized conditions allowing for tunnel valley erosion near the ice margin (Russell et al., 2007) and esker deposition (Burke et al., 2008). Tunnel valleys in Okanagan valley and on Thompson Plateau are dominantly eroded in bedrock, precluding genesis by piping and flushing of deforming sediments. Additionally, valleys have continuous courses (rather than segmented) and do not exhibit sediment fans along their course, as would be expected during time-transgressive development. These valleys likely formed during a single event.

3.7.2.1. Bedrock tunnel valleys

On Thompson Plateau and in the Okanagan Valley corridor, most valleys are eroded in bedrock. In a few cases, they start in surficial sediment but exhibit bedrock walls within a few hundred metres of their upflow ends. Bedrock perimeters preclude any formative mechanism involving incremental sapping, collapse and flushing of deformable sediment (e.g., Boulton and Hindmarsh, 1987). They do leave open the possibility that the valleys may have a polygenetic origin. Bedrock valleys similar in scale to Missezula Valley may be remnants of Tertiary-age fluvial drainage networks (Tribe, 2005). But the presence of drumlins on the valley walls clearly indicates modification during glaciation, and convex longitudinal profiles over 100 km reaches and rising 450 m are difficult to explain by subaerial fluvial processes, irrespective of valley age. Many valleys clearly conveyed subglacial meltwater during glaciation.

The Garnet Valley network exhibits an anabranching pattern similar to those formed by outburst floods in Antarctica (Sugden et al., 1991, 1995; Anderson et al., 2001; Lowe and Anderson, 2002; Denton and Sugden, 2005; Lewis et al., 2006) and under the Laurentide Ice Sheet (Brennand and Shaw, 1994; Regis et al., 2003). The anabranching pattern and convex longitudinal profiles of this valley network suggest that it operated contemporaneously and subglacially as tunnel valleys (c.f. Brennand and Shaw, 1994).

3.7.2.2. Squally Point: residual bedrock highs flanked by bedrock troughs

At Squally Point, bedrock troughs flank upstanding bedrock residuals (Fig. 3.12). These residuals resemble streamlined bedrock forms such as whalebacks and roches moutonnées. Similar bedrock forms within steep-sided, overdeepened u-shaped bedrock valleys elsewhere in British Columbia have been attributed to the operation of paleo-ice streams (Evans, 1996). This topographic setting shares similarities with that of modern Greenland ice streams (*Isbraes*) (Truffer and Echelmeyer, 2003). Evans (1996) suggested that abrasion under relatively thin, fast flowing ice resulted in roches moutonnées, while thicker ice produced whalebacks. Roberts and Long (2005) used similar arguments to reconstruct flow conditions from bedrock forms in the forefield of Jakobshavns Isbrae, Greenland. In both reconstructions, bedrock forms are attributed to glacier abrasion and plucking, both processes requiring glacier sliding. Evans (1996) and Roberts and Long (2005) did not discuss possible alternative erosive agents such as meltwater, even though some of the forms they describe bear a striking resemblance to s-forms (Sharpe and Shaw, 1989; Kor et al., 1991). For example, sinuous troughs flanking residual bedrock knobs (figs 6 B, C, D in Roberts and Long, 2005) are similar to s-forms such as cavettos attributed to meltwater erosion (e.g., Sharpe and Shaw, 1989; Kor et al., 1991; Bradwell, 2005). In addition, Roberts and Long (2005, fig. 6 E) showed extensive boulder-strewn bedrock surfaces. No other sediments appear to overlie bedrock. These boulders are similar to boulder lags documented in glaciated areas affected by large meltwater flows (Kor et al., 1991; Shaw, 1996; Rampton, 2000). Similarities of the documented bedrock forms to those observed in fluvially-eroded bedrock channels (Richardson and Carling, 2005), and the presence of a possible boulder lag on the bedrock surface, suggest that meltwater erosion may have wholly or partially sculpted some forms (i.e. they are s-forms, Kor et al., 1991).

Isbrae flow mechanisms are spatially variable. Although thermally enhanced creep and sliding are dominant flow mechanisms along the thickest portions of the isbrae (Truffer and Echelmeyer, 2003), the importance of sliding on glacier motion increases closer to the margin. Bed lubrication is required for efficient sliding and there is evidence for an established subglacial hydrological system beneath Jakobshavns Isbrae inferred from borehole measurements (Lüthi et al., 2002). The exact nature of the hydrologic system beneath the thickest portions of the isbrae is unclear and it may take the form of linked cavities or groundwater flow (Lüthi et al., 2002). Closer to the glacier terminus, plumes of sediment efflux at the floating margin (Roberts and Long, 2005) suggest that drainage is mainly channelized there. This system may periodically switch to a more distributed one in response to supraglacial water drainage to the bed, causing glacier decoupling and transient acceleration (Zwally et al., 2002). These episodes may reflect seasonal control on subglacial meltwater routing (Zwally et al., 2002), although this is not necessarily the case (cf. Truffer et al., 2005). Variable amounts of meltwater are present at the bed and potentially allow for fluvial bedrock sculpting to occur in addition to glacial abrasion.

Catastrophic outburst flood tracts from the Altai Mountains of Siberia contain further evidence that meltwater can produce forms resembling roches moutonnées and whalebacks (cf. Herget, 2005). Boulders with smooth stoss and jagged lee sides may result from corrasion by sediment transported in the turbulent flow. Irregular, jagged lee-ends form by hydraulic plucking related to kolks and cavitation (Whipple et al., 2000). Altai boulders are unequivocally water-worn. They lack striae, easily differentiating them from ice-abraded forms. However, in the subglacial environment, striae can ornament fluvially-sculpted forms where ice recouples with the bed following subglacial water flow (Kor et al., 1991). Fluvially formed striae also result from jökulhlaups (McCarroll and

Matthews, 1989) and make it difficult to distinguish formative agents. Similar issues arise in the interpretation of s-forms (cf. Munro-Stasiuk et al., 2005). Bedrock erosional forms in glaciated regions can be produced by more than one agent and require careful examination if inferences of glacial processes rely on these forms.

The Squally Point bedrock forms are probably polygenetic and palimpsest forms. It is possible that glacial abrasion played a role in their formation, however, this role is difficult to ascertain because striae were not conclusively identified on the weathered bedrock surfaces (striae may be buried or were present prior to weathering). The fact that flanking sinuous troughs define bedrock forms, the continuity and alignment of Squally Point forms with other bedrock and sediment drumlins, and the interpretations of subglacial valley scouring and deposition of a thick (~400-m) glaciofluvial valley fill (Vanderburgh and Roberts, 1996) (§ 3.7.2.3) all suggest that the Squally Point forms were at least partly formed or modified by meltwater flows. These flows had to be subglacial given the convex longitudinal profile of bedrock troughs. However, reactivation of erosional bedrock surfaces can occur periodically so that resulting forms are time-integrated products (a palimpsest) of many erosion cycles.

3.7.2.3. Okanagan Valley fill characteristics

The Okanagan Valley fill consists of up to 800 m of sediment overlying the bedrock surface (Eyles et al., 1991). Seismic stratigraphy of the valley fill below Okanagan Lake (Eyles et al., 1990, 1991) and onshore (Vanderburgh and Roberts, 1996) reveal a similar stratigraphy characterized by three thick facies. Based on sediment samples recovered during drilling though the valley fill (reaching bedrock), the valley fill is comprised of a fining-up and thinning-up sequence of boulder gravel (200 m thick), overlain by sandy silt (250 m thick) and capped by laminated silt (~100 m thick), the latter associated with deglacial lakes (Vanderburgh and Roberts, 1996). Sand and gravel facies are subglacial meltwater deposits (Vanderburgh and Roberts, 1996). Till and interglacial sediments are absent from the valley fill (Vanderburgh and Roberts, 1996), suggesting that the entirety of the fill is relatively young (Late Wisconsin age) and the majority of it was deposited subglacially by pressurized meltwater (Eyles et al., 1991; Vanderburgh and Roberts, 1996). Interglacial sediments and till were either removed from much of the valley and/or not deposited within it.

3.7.3. *Sediment flux conservation and landform associations*

The principle of sediment flux conservation is used as a test of possible landscape forming events. Any explanation of landscape evolution must account for this sediment flux. A suite of meltwater processes best explains individual landscape elements in the study area. In this section, sediment flux conservation is explored in the context of a subglacial meltwater model for landscape evolution.

Drumlins on Thompson Plateau and in Okanagan Valley are residuals of a surface layer that acted primarily as a sediment source. In terms of sediment conservation, material eroded between drumlins should be preserved somewhere downflow of these erosional landforms. Few terminal moraines have been recognized along the southern CIS margin (Richmond, 1986) and none are identified on Thompson Plateau or in the area between Thompson Plateau and the Withrow Moraine on the Columbia Plateau. Small recessional moraines occur on Columbia Plateau (Kovanen and Slaymaker, 2004). However, they drape drumlins and, therefore, post-date their formation. Their volume is far too small to account for the volume of sediment excavated along the drumlin swarms. On Thompson Plateau, some sediments are preserved in the thin fill of tunnel valleys but they fall short of accounting for the total volume removed over Thompson Plateau. Okanagan Valley may be a more adequate sediment sink. However, if drumlin alignment is indicative of the general direction of sediment flux, only a fraction of sediment transport was directed toward Okanagan Valley from Thompson Plateau. Over most of Thompson Plateau there is a discrepancy in sediment conservation as the erosional landscape does not have a local depositional counterpart. Similar sediment flux continuity questions arise for other drumlin swarm terminations near Trepanege Ridge and the Headwaters Lake Valley area, near Princeton at the southern terminus of the FTO drumlin swarm, and on Omak and Columbia Plateaus.

The proposed meltwater hypothesis of drumlin and tunnel valley formation in the study area best explains the apparent sediment flux continuity enigma. Meltwater can transport sediment far beyond the immediate termination of the drumlins and is therefore better able to account for the required sediment flux continuity. Multiple landscape elements favour the meltwater interpretation. Drumlins and tunnel valleys are commonly associated landscape elements. They form sequentially when sheetflows eroding drumlins collapse into discrete, linear flows to erode valleys (Brennand and Shaw, 1994; Fisher et al., 2005). If underbursts capable of eroding broad portions of the landscape are invoked, water discharge conservation must also be considered along with sediment flux conservation. In the case of the Headwaters Lake area, single bedrock conduits cutting through Mt. Kathleen Ridge and connecting to the Garnet Valley network (Fig. 3.7) are considered partial outlets for the sheetflows that eroded drumlins. The northern tributaries of Trepannier Creek also show a pattern of channels leading away from the termination of the drumlin swarm and connecting Thompson Plateau to Okanagan Valley via bedrock valleys such as Trepanier and Lacoma Creeks (Fig. 3.7). Flushing of meltwater and sediment toward Okanagan Valley occurred through its tributary valleys. For sediments to be drawn toward Okanagan Valley, hydraulic potential must have been lower toward the valley and accommodation space for water and sediment must have been available. Therefore, drumlin formation on Thompson Plateau was contemporaneous with the proposed subglacial meltwater scouring of Okanagan Valley (Vanderburgh and Roberts, 1996), and formation/retouching of streamlined bedrock forms in the Okanagan Valley corridor. In addition, the Garnet Valley anabranched network and bedrock tunnel valleys at Cougar Canyon, Squally Point and possibly in Washington State operated at this time. Thus, storage of sediment in Okanagan Valley may satisfy sediment flux continuity requirements for this portion of the landscape. However, most of the Thompson Plateau drumlin swarm does not immediately converge toward Okanagan Valley. Instead, it ends abruptly in the Similkameen Valley (Fig. 3.3), ultimately connecting to the Okanagan Valley corridor via Palmer Lake Valley (Figs. 3.1, 3.11, 3.13). Sediment flux continuity needs to be addressed for this area.

Missezula Valley extends over most of the length of the drumlin swarm. Other smaller-scale valleys, parallel to Missezula Valley, develop near the swarm margin (Osprey Lake valley, Allison Creek valley, Otter Creek valley). Drumlins occur within Missezula Valley and on the interfluves between valleys. There is an overall convergence of the swarm toward the Similkameen Valley and localized convergence occurs around individual valleys. These landform associations record the transition from an unstable sheetflow collapsing into discrete channels, similar to the event sequence inferred for Trepanege Ridge area.

During waning flow sediment also could have been deposited along the Similkameen Valley and in other connected valleys such as Palmer Lake Valley, Wagonroad Coulee, and Whiteswan Coulee (Fig. 3.11). Extensive elevated terraces above modern river level have been interpreted as outwash terraces formed during deglaciation (Ryder et al., 1991). This interpretation relies on the position of these terraces relative to the modern and early Holocene terraces of the Similkameen River, and their composition of coarse gravel. Neither of these criteria negates the possibility that the gravels are depositional end-members of an underburst. Similar features occur along outburst paths in high-relief landscapes (Montgomery et al., 2004). Later incision, possibly during deglaciation, could isolate these deposits as terrace remnants. Additional coarse gravel deposits are present along Okanogan Valley. Coarse basal gravels, overlain by rhythmic fine sand and silt, characterize the general fluvial stratigraphy along the valley south of the 49th parallel. Coarse gravel sometimes overlies these deposits where bedrock coulees debouch along the valley axis. The basal coarse gravel may be associated with waning of the proposed underburst. It occurs within hanging isolated fans at the mouth of bedrock coulees and tunnel valleys. Often, these valleys lack integration within the modern drainage network and show no to little evidence of post-depositional incision by rivers. In fact, many of the coulees do not presently contain streams. Therefore, coulees and associated deposits have been unaffected by post-glacial development of the fluvial system. They are more likely remnants of glaciofluvial activity during glaciation.

Considerable volumes of sediments are also stored at the confluence of Okanogan Valley and the Columbia River (Fig. 3.1). Sand and gravel deposits in the Great Terrace (Fig. 3.11) (Russell, 1898; Flint, 1935a) and in 'gulch fills' (perched deposits in small tributary valleys of the Columbia River downstream of the Okanogan confluence) may also record such underburst(s). Finally, the proposed underburst(s) may have been contemporaneous with the Late Pleistocene outburst(s) from glacial Lake Missoula (Shaw et al., 1999). The Okanagan underburst would be an additional meltwater source for the Channeled Scabland formation during the Late Pleistocene. In terms of sediment conservation, a meltwater underburst/outburst could transport sediment far beyond the ice margin (e.g., Leventer et al., 1982). Sediments from the Okanagan underburst could be stored as far as the Astoria Fan and its network of submarine canyons off the coast of Washington, Oregon and California (Zuffa et al., 2000; Normark and Reid, 2003). Provenance of sediments in the Astoria Fan indicates a source area that includes portions of southern British Columbia and possibly Okanagan Valley (Prytulak et al., 2006).

3.7.3.1. Deeply weathered outcrops

Deeply weathered outcrops in glaciated terrain may record non-glaciated regions, or relict surfaces preserved beneath cold-based ice (Kleman 1994; Kleman and Bergstrom, 1994; De Angelis and Kleman, 2006). On Thompson Plateau, the limited aerial extent (100s m^2) of deeply weathered outcrops and their juxtaposition to neighbouring landform-sediment associations argue against them as records of non-glaciated regions or as records of cold-based ice. Deeply weathered outcrops occur in close proximity to drumlins and tunnel valleys, both of which require meltwater at the bed for their development (e.g., Boulton and Hindmarsh, 1987; Shaw 1996). Glacigenic sediments contained within drumlins record further evidence of warm-bed conditions in the form of lodgement, ductile deformation and possible subglacial water-filled cavities at the ice-bed interface. (Chapter 2). The position of deeply-weathered outcrops, on the upflow (northern) flank of Trepanege Ridge favours pressure melting rather than freezing conditions.

How can the presence and distribution of small-area, deeply-weathered bedrock outcrops be rationalized with the paleoglaciology implied by associated landforms and sediments? Notably, anchored corestones and large boulders resting on bedrock surfaces (Fig. 3.9) are in situ and are effectively lag stones following the removal of saprolite. As well, detached corestone boulders resting on/in thin diamicton may also be lags. Consequently, these outcrops and associated boulders result from partial subglacial erosion (saprolite removal) of preglacially weathered bedrock. In regards to genetic agents, the presence of striae clearly indicate that ice was once grounded on bedrock. However, ice cannot be responsible for differentially eroding saprolite and leaving lag corestones on bedrock surfaces. Instead, both the stripping of saprolite and boulder entrainment are expected under the action of ice. Entrained boulders were deposited within diamicton resting on the abraded surfaces. Transport distances were short as boulders do not show evidence of striae or glacigenic shaping (though many are weathered). This material was later drumlinized. A viable complementary explanation suggests that diamicton, saprolite and small corestones were eroded by subglacial meltwater flows, while leaving open the possibility that the latter two were first eroded and transported by ice. Where boulders rest on bedrock, meltwater removed saprolite. Where boulders rest on/in diamicton, meltwater eroded diamicton leaving behind ice-transported boulders as lags. Flows were sufficiently competent to entrain clasts as large as 0.5 m diameter and any overlying diamicton on the adverse slope of Trepanege Ridge. These flows were therefore subglacial and driven by hydrostatic pressure from the overlying glacier. Appealing to Occams' Razor and process continuity in landscape evolution, this subglacial meltwater erosion was likely contemporaneous with the underburst invoked for drumlins and tunnel valley formation. The limited degree of bedrock erosion and incompleteness of saprolite removal in some locations reflects duration of erosion, a product of the time required to strip any overlying diamicton and the location of the outcrops on interfluves between tunnel valleys. In a continuity model for landscape evolution, this spatially limited erosion took place during the transition from sheetflow (drumlins) to channelized flow (tunnel valleys) over Trepanege Ridge. Finally, some bedrock striae may have formed as ice recoupled with the bed following the proposed underburst.

3.7.3.2. Non-drumlinized highs

Non-streamlined patches within streamlined tracts have been attributed to 'sticky spots' (areas of enhanced basal drag) along corridors of rapid ice flow (Stokes et al., 2006, 2007) resulting from a range of subglacial conditions that include bedrock bumps, low relief till-free patches, better-drained till areas, and zones of meltwater freeze-on (Stokes et al., 2007). Within such corridors, Rogens ('ribbed moraine') may be the geomorphic signature of compressional flow and freeze-on (Stokes et al., 2006). Alternatively, non-streamlined highs may represent zones of cold-based ice with limited erosion between fast flow corridors (Kleman et al., 1997).

Non-streamlined bedrock highs on the Fraser and Thompson plateaus occur within drumlinized terrain attributed to erosion by meltwater underbursts. During underbursts, drumlin erosion occurs under decoupled portions of the ice sheet. In this context, non-drumlinized highs record locations where the ice remains grounded.

Increases in shear stress are expected over such insular grounded zones during decoupling. Pressure melting from high localized shear stresses, not freeze-on, may be expected on bedrock highs, and localized sliding and abrasion could occur at this time. Thus, non-drumlinized highs may indeed record zones of enhanced basal drag. If rapid ice flow results from ice-bed decoupling during underbursts (e.g. Shaw 1996), these non-drumlinized highs should be viewed as 'anchoring islands' within the underburst flow path. Their effect is to moderate any rapid ice sheet flow conditioned by enhanced basal sliding.

3.7.4. An underburst landsystem

The weight of field evidence discussed above lends support to the possibility that a regional-scale subglacial meltwater underburst shaped and/or modified major landscape elements of Okanagan Valley and Thompson Plateau. On Thompson Plateau, drumlins and tunnel valleys record existence of a short-lived sheet flow collapsing to channelized flow during which the majority of drumlin erosion occured. The shape and fill characteristics of Okanagan Valley suggest voluminous and energetic subglacial meltwater flows. For example, till and pre-glacial sediments are absent from the Okanagan Valley sides and fill (Vanderburgh and Roberts, 1996). This absence, and the overdeepening of the bedrock valley, has been attributed to extensive scouring of the valley by pressurized subglacial meltwater during the last glaciation (Eyles et al., 1991; Vanderburgh and Roberts, 1996). Glaciated Icelandic valleys subject to frequent jökulhlaups are overdeepened as much as 300 m below sea level (Björnsson, 1996). Repeated jökulhlaups scour and flush sediment from these valleys. Valleys refill during waning stages and under non-jökulhlaup conditions (Björnsson, 1996). Subglacial fluvial processes excavate 30-300 times more sediment than direct glacial scouring (Björnsson, 1996). Icelandic valleys overdeepened by jökulhlaups are likely appropriate analogues for Okanagan Valley. Underbursts associated with Okanagan jökulhlaups could account for the undulating bedrock long profile and overdeepenings, and the valley fill dominated by glaciofluvial sediment (Vanderburgh and Roberts, 1996). In addition, the v-shaped bedrock valley cross section may also result from subglacial fluvial scouring. However, caution must be used in giving too much weight to the valley cross-sectional shape interpreted from unmigrated seismic data (Eyles et al., 1991). By extension, such underbursts scoured the valley sides and created the numerous tunnel valleys found throughout the Okanagan Valley corridor (Garnet Valley network, and Squally Point, Cougar Canyon, Omak Lake, Soap Lake, Palmer Lake valleys; Figs. 3.7, 3.10, 3.11, 3.12, 3.13). Bedrock and sediment drumlins (Swan Lake, Ellison Ridge, Wood Lake, Tonasket and Omak plateaus), were enhanced, and perhaps fully-formed (e.g., those exhibiting crescentic troughs) during the same underburst(s).

3.8. Implications of the Thompson Plateau and Okanagan Valley landsystem for glaciodynamics, plumbing and geometry of the CIS

3.8.1. Effects of an underburst on glacier flow over Thompson Plateau and Okanagan Valley

We attribute large parts of the Thompson Plateau and Okanagan Valley landsystems to meltwater erosion and deposition during underbursts. Such underbursts would have caused ice-bed decoupling and may have resulted in short-lived episodes of rapid ice flow. Similar but more localized behaviour occurs in modern glaciers and ice caps (Zwally et al., 2002; Bartholomaus et al., 2008).

The dynamics of some Icelandic jökulhlaups may provide a window into the plausible dynamics of the CIS during an underburst on Thompson Plateau and in Okanagan Valley. Some Icelandic jökulhlaups appear to have drained both as broad sheetflows and channelized flows. Modeling of the 1996 Grimsvötn jökulhlaup suggests complex water exchanges between a distributed sheet-like configuration and individual channels. Channels develop from a sheet near the underburst initiation point. As a function of varying water pressure along the outburst path, water may also exit channels to maintain sheet-like configuration (Flowers et al., 2004). The Grimsvötn jökulhlaup was accompanied by 10 m of vertical displacement of the ice surface (Björnsson, 1997 in Flowers et al., 2004) and clearly shows that water pressure was sufficiently high to decouple the ice from its bed. Despite this decoupling, rapid flow did not occur. This is attributed to sufficient marginal grounding along the valley sides of the outlet glacier (Skeiðarárjökull) to prevent complete decoupling and acceleration (Björnsson, 2002). However, acceleration during underbursts is possible given sufficiently high discharges capable of valley-side decoupling (Björnsson, 2002) or, in the case of an ice sheet, sufficiently broad flows that allow for rapid flow near the decoupling centerline. Passage of the Grimsvötn flood wave fractured the ice surface due to brittle behaviour of the decoupled glacier (Roberts et al., 2000). Similar fracturing occurs in surging glaciers (Thorarinson, 1969). However, jökulhlaups do not necessarily lead to ice fracturing. Modeling results suggest that episodic drainage of subglacial lakes, where ice deforms to fill the void created by drained water, is possible without brittle collapse and disintegration of the ice surface (Pattyn, 2008). Response of the ice surface during jökulhlaups likely depends on the rate of drainage, the geometry and size of the subglacial cavity produced by drainage, and the viscosity and thickness of the ice. Thus, the effects of an underburst on Thompson Plateau and in Okanagan Valley should be spatially variable as a function of the subglacial topography and ice conditions. Thompson Plateau may have experienced localized decoupling but limited enhanced flow due to anchoring bedrock highpoints (Section 3.7.3.1).

3.8.2. Underburst operation and collapse

Critiques of hypotheses invoking underbursts as important glacial geomorphic agents have argued that inherent instability of sheetflows makes them improbable agents for subglacial landscape erosion (Walder, 1994). Despite counter arguments showing prolonged stability of turbulent sheetflows over laminar flows (Shoemaker, 1992, 1994), sheet flow for the 1996 Skeiðarárjökull outburst (Flowers et al., 2004) and the existence of a landform continuum corresponding to the collapse and channelization of sheetflows (e.g., Brennand and Shaw, 1994), meltwater hypotheses remain controversial. Landscapes of the Laurentide Ice Sheet are the basis for most existing inferences of broad subglacial meltwater underbursts. There, subdued bed relief does not create substantial topographic basins for water ponding (the Great Lakes basins (Evatt et al., 2006) and the east arm of Great Slave Lake (Christoffersen et al., 2008) notwithstanding). In contrast, the CIS developed over much more pronounced topography. Subglacial lakes (Eyles and Clague, 1991) and supraglacial ponds could have developed more readily. The Thompson Plateau-Okanagan Valley landscapes record transient sheet-like conditions and a multitude of evidence for channelized drainage, consistent with predicted collapse of unstable sheet-like flow (Walder, 1994).

There is remarkable landform continuity in the landscape of the Fraser and Thompson plateaus, and Okanagan Valley. Consequently, explanations of formative mechanisms must account for this continuity. Drumlin continuity across valleys transverse to the swarm orientation indicate that these valleys did not divert flow substantially during drumlin erosion. In fact, better-developed drumlins on adverse slopes leading away from valleys are consistent with coherent flows impinging on a positive bed step (Pollard et al., 1996) and show that flow maintained its coherence across the valleys. Along an underburst path, valleys should act as zones of flow resistance and of hydrostatic head loss, compromising sheetflow stability. If pressure losses took place through valleys of Thompson Plateau, they initially did not sufficiently divert water to significantly alter flow lines. For this to occur, the initial water volume must have been large enough to completely fill valleys along the underburst path, while maintaining a coherent sheet form despite the likely turbulence generated by the pronounced topography. Valleys could also have been partly sediment-filled, thus reducing the resistance effect of these topographic steps on flow coherence. This scenario may be difficult to achieve given the size and multitude of valleys encountered along the swarm. An alternative possibility is that sheetflows encountered successive subglacial water bodies occupying valley troughs, thus minimizing head loss along the outburst path and maintaining sheet coherence sufficiently long to erode broad drumlin swarms on Thompson Plateau.

There is little doubt that sheetflow configuration was short-lived. Sheet flow was maintained for a portion of drumlin erosion on Thompson Plateau where flow depth may have been as much as the drumlin heights (though it could be shallower if erosion occurred by progressive lowering of the sediment surface). Sheet-stage erosion led to channelized flow. Multiple landscape elements speak to diminishing sheet stage erosion toward the southern swarm margin and increasingly valley-focused flows. For example, transitions from drumlins, to incipient bedrock drumlinization, to tunnel valleys over Trepanege Ridge show the effects of this topographic obstacle on flow stability. Sheetflow was not sufficiently deep to submerge the ridge. In fact, sheet flow separated into discrete flow lines to create tunnel valleys over the non-drumlinized portion of the ridge and to form the narrower swarm curving along the low-lying western flank leading to Headwaters Lake Valley (Fig. 3.7). Similarly, bifurcation of meltwater flow toward

Okanagan Valley (evidenced by SE-oriented drumlins leading in to the bedrock valleys of Trepanier and Lacoma Creeks) illustrates decreasing sheetflow stability and the influence of lower hydraulic potential in Okanagan Valley, diverting flow toward this area.

The majority of the erosion in Okanagan Valley took place under channelized flow conditions, as evidenced by numerous scales of bedrock valleys (Figs. 3.7, 3.10 A, 3.11, 3.12). Partial and possibly complete scouring of the pre-glacial and early-glacial valley fills, and perhaps sub-sea level bedrock erosion took place at this time (some sub-sea level bedrock erosion may have occurred during prior glaciations). Smaller-scale bedrock valleys such as Cougar Canyon, Squally Point valleys, Omak and Soap Lake valleys and other bedrock coulees, and bedrock drumlins at Swan Lake, Ellison Ridge, Kelowna (Fig. 3.9), Tonasket and Omak Plateau (Fig. 3.11) were eroded and/or enhanced by the underburst. Missezula valley and the Osprey Lake Valley (Fig. 3.6) record focusing of water from the collapsing sheet flow over Thompson Plateau. It is appropriate to consider Similkameen Valley as an extension of these two channels as it likely provided an outlet for this flow. The clear continuation of the bedrock-dominated Similkameen and Palmer Lake valleys (Fig. 3.13) also indicates that subglacial flows were directed southward and joined Okanogan Valley via bedrock coulees debouching at Omak and Tonasket, WA (Figs. 3.11, 3.13). Thick basal sequences of boulder gravel in the fill of Okanagan Valley (Vanderburgh and Roberts, 1996) may represent waning-stage deposition from the underburst. The same conclusion applies to boulder gravel deposits at the mouth of hanging bedrock channels and for ‘gulch fills’ in Washington State. Substantial amounts of sediments and meltwater discharged through the Columbia River. Drumlins on Omak Plateau and the ‘scabbed’ appearance of the sediment surface on Omak Plateau suggest that erosion occurred as confined meltwater flow expanded where Okanagan Valley opens onto the Columbia Plateau. Erosion during flow expansion may seem counterintuitive. However, it is not due to the fact that flows were subglacial. Under these conditions, flow velocity over Omak Plateau would have been extremely high due to a minimal and decreasing gap width between the overlying ice and its bed. Erosion is enhanced where the flow cross-sectional area is reduced and high velocity flows could have eroded drumlins and ripped out sediment to create the ‘scabbed’ appearance. Patterns of splaying drumlins on Columbia Plateau up to the Withrow Moraine may also record this flow expansion under the Okanogan Lobe.

The Withrow Moraine is heavily dissected by channels ranging in size from 100s m to much larger features such as Moses Coulee, reaching 10s km length and over 1 km in width. These channels are eroded in both bedrock and sediment. The proposed flows would have breached the moraine and continued toward the Columbia River. Sediments carried in the flow would have been transferred beyond the ice margin, and possibly swept by the Spokane Floods as far as the Astoria fan on the coast of Oregon and California (Brunner et al., 1999; Normark and Reid, 2003).

3.8.3. Meltwater reservoir development

The proposed events for the Thompson Plateau and Okanagan Valley require subglacial and/or supraglacial production and storage of meltwater prior to underbursts. Rejections of hypotheses invoking subglacial underbursts often cite the lack of documented sources of meltwater to supply the required discharge (Clarke et al., 2005). Deep valleys of British Columbia provide favourable conditions for subglacial lake development (cf. Livingstone et al., 2012). Subglacial lakes can develop by multiple mechanisms including accumulation of basal meltwater within bedrock depressions, or within basins delimited by thermal (Cutler et al., 2002) or hydraulic potential barriers (Björnsson, 1975, Livingstone et al., 2012). Alternatively, lakes can start as subaerial systems and advancing ice can capture the water body during ice sheet growth (Erlingsson, 1994). Grounded ice around the subaerial lake advances over water, floats and thins to create a temporary coalescing ice shelf (Pattyn et al., 2004; Alley et al., 2006). The shelf grows until it grounds on a sill downflow of grounding lines from feeder glaciers (Fig. 3.14). Rudoy (1998) used the term ‘catch lake’ to describe this type of Pleistocene lake in the Altai Mountains of Siberia. It is also a mechanism by which some Antarctic subglacial lakes (Pattyn, 2003; Studinger et al., 2003; Alley et al., 2006) and other Pleistocene glacial lakes (Erlingsson, 1994) are thought to have formed. Eyles and Clague (1991) speculated on similar mechanisms to explain possible subglacial lacustrine sedimentary sequences in valleys of British Columbia. Once initiated, lakes can expand by further melting of the ice lid, by meltwater inflow from the surrounding area depending on regional head, or from supraglacial sources.

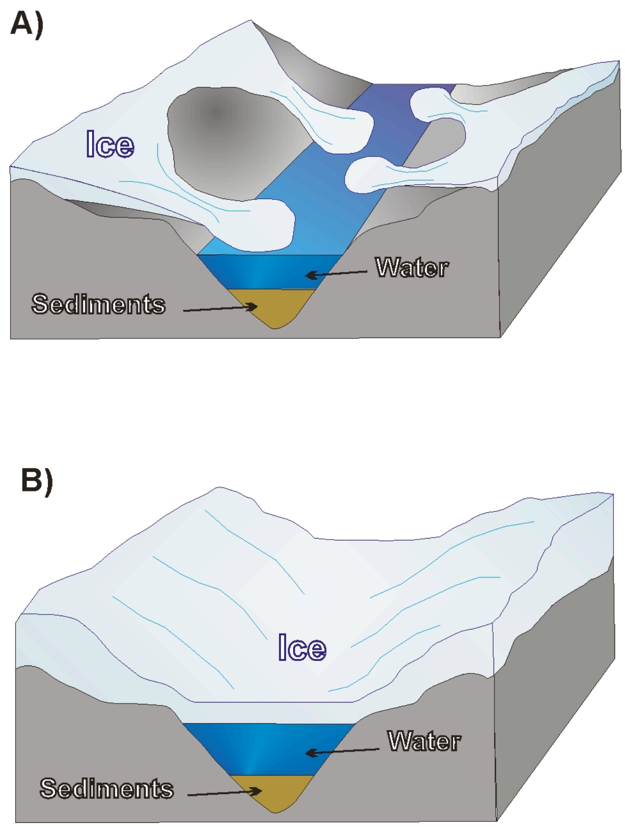


Figure 3.14. Conceptual model of catch lake development. A) Advancing glaciers in tributary valleys of Okanagan Valley ground on bedrock valley walls and develop floating margins over the existing lake. B) Lake capture occurs as floating glacier margins coalesce to form a continuous seal across the lake. If the system becomes fully sealed, pressurization begins. The subglacial lake affects ice sheet geometry by enabling rapid ice flow along the valley. This contributes to development of a buttressing seal downvalley.

3.8.3.1. Subglacial lake development in Okanagan Valley

We postulate that, during the onset of the Fraser Glaciation (Late Wisconsin-age) ice advanced over a pre-existing lake occupying the structural trough of Okanagan Valley to form a ‘catch lake’. However, nuances are required, particularly in reference to ice shelf development required to seal a subaerial lake system. Modern ice shelves presently develop under cold-based glacier conditions, which suppress marginal calving. Under warm-based conditions, calving occurs and expansion of the shelf margin is limited. Evidence for warm-based conditions around Thompson Plateau and Okanagan Valley was discussed in Chapter 2. This is potentially problematic for lake sealing. However, there are mitigating factors that still allow ice shelves to develop in Okanagan

Valley. For example, the proposed subglacial lake length (within the deep, narrow trough of Okanagan Valley) would have been much greater than its width. Length:width ratios on the order of 22:1 – 67:1 are estimated (the range addressing our uncertainty in subglacial lake extent). Subglacial lake length exceeds 200 km and could have been much greater. More importantly, subglacial lake width only varies from 3-9 km. Consequently, the shelving distance across the lake is relatively short and probably attainable under high ice fluxes in spite of a calving front. Furthermore, most modern ice shelves from which comparisons are drawn emanate from a point source and develop in a large bay. In Okanagan Valley, ice shelves could develop from multiple input points (tributary valleys) along the length of the subaerial lake. Synoptic CIS flow reconstructions (Clague, 1989) suggest accumulation centres to the North, West and East of Okanagan Valley. A nascent ice shelf would not have to expand 200 km southward from a single point source. Instead, probable coalescence of individual shelves from tributaries located along the valley axis would reduce the shelving distance and make for swifter lake capture (Fig. 3.14 A). Finally, the role of seasonal lake ice must be considered. If calving into a subaerial lake prior to capture is considered, bergs are either evacuated or remain within the basin. Following a short-lived summer melt and calving season, lake ice could trap remaining bergs and incorporate them within the ice cover. The following year, advancing shelves would produce more bergs that could add to existing ones. On a seasonal basis, lake area under ice cover would increase and gradually build toward capture. Conceivably, a subglacial lake could develop in this manner. Once sealed, overpressurization of the system begins (Erlingsson, 1994; Alley et al., 2006).

Physiographic characteristics of Okanagan Valley also occur in neighbouring valleys. Presently, not all valleys are hydrologically connected but they could be, given minor (m-scale) changes in modern basin divides. Water build-up in a subglacial lake would hydrologically connect valleys. The potential subglacial lake for southern British Columbia could extend to deep valleys of the Shuswap Lakes region. Seismic stratigraphy of Shuswap Lake reveals massive overdeepening of the bedrock trough (Eyles and Mullins, 1997) and a relatively simple sedimentary fill mimicking that of Okanagan Valley (Eyles and Mullins, 1997; Vanderburgh and Roberts, 1996). This strongly suggests similar erosional and depositional histories for these basins.

The importance of subglacial lakes on ice dynamics and as components of subglacial hydrology is increasingly recognized. For example, subglacial lakes occur in the onset zones of ice streams (Siegert and Bamber, 2000; Dowdeswell and Siegert, 2003; Bell et al., 2007). However, their role in rapid ice flow is yet to be fully understood. Beyond rapid flow, effects of subglacial lakes on ice sheet dynamics are two-fold. First, grounded ice surrounds the lake but ice flattens and remains thin over the lake surface due to reduced shear stress (Pattyn, 2003; Pattyn et al., 2004). This creates rapid changes in ice surface slope over short distances. Second, glacier flow accelerates over the floating 'shelf' areas. This faster-flowing corridor enhances mass transfer from the areas of grounded ice to an ice ridge downflow (Alley et al., 2006). In addition, shorter-lived velocity increases can occur during partial or complete supraglacial lake drainages (Zwally et al., 2002). An important consequence of enhanced mass transfer over the subglacial lake is downflow ice thickening to create a buttressing seal on bedrock highs along the length of the subglacial lake. Lake expansion occurs as long as sealing points are available and hydrostatic pressure in the reservoir does not exceed overburden pressure at sealing points (Alley et al., 2006). Catastrophic drainage can occur if water pressure rises to overburden pressure at the seal.

Antarctica contains a number of subglacial lakes (Siegert et al., 2005) and serves as a partial analogue for our invoked CIS subglacial lakes. Most Antarctic lakes occupy bedrock depressions. Their drainage was thought to be unlikely due to confinement in bedrock basins. However, it now appears that these lakes may drain partially and in the process, subglacial water transfer to neighbouring lakes occurs without triggering complete drainage (Wingham et al., 2006). Water transfers are unnoticed at the ice margin and vertical displacement of the glacier surface creates accommodation space for transferred water. Such transfers, if sufficiently large, may result in outbursts at the margin (Wingham et al., 2006). This raises the possibility that subglacial lakes in one location can feed and enlarge neighbouring ones (e.g., Fricker et al., 2007). It is entirely plausible that several subglacial lakes may have developed and grown in the numerous bedrock valleys of southern British Columbia, and these lakes may have emptied and filled in domino-like fashion. For example, an extensive subglacial lake within the Shuswap Lake Valley system could have connected to the Okanagan Valley subglacial lake as seals were broken.

3.8.3.2. Favourable conditions for regional subglacial lake development and drainage

Geothermal and volcanic activity in southern British Columbia may have facilitated expansion, maintenance and drainage of subglacial lakes under the CIS. Geothermal gradients in Okanagan Valley and southern British Columbia are very high (Fairbank and Faulkner, 1992). Geothermal springs with waters reaching 137°C emerge into Okanagan Lake near Kelowna (Grasby and Hutcheon, 2001) (Fig. 3.2). The Okanagan Valley Fault determines the location of springs active at least since the Eocene. These springs could have contributed to the initial development of a subglacial lake from a grounded ice tongue in the valley bottom and/or expansion of a catch lake.

Physiographic and geothermal conditions could also play a critical role in controlling lake drainage. Subglacial lakes lying in troughs aligned with regional ice flow direction tend to be less stable than lakes of equivalent geometry oriented transverse to regional ice flow (Pattyn, 2008). However, lake size, ice thickness, flow velocity and especially the steepness of the ice surface slope also control lake stability (Pattyn, 2008). Thus, small lakes (2.5 km diameter) under slow moving (non-streaming) and relatively thick ice (up to 4 km) with steep surface slopes (0.1°) tend to be less stable than lakes with the reverse conditions (cf. Pattyn, 2008). Ice over Okanagan Valley likely thinned and may have flowed rapidly during subglacial lake development (§ 3.8.3.1) resulting in low surface slopes over the lake. Thus the apparent instability imparted by the local physiography may be partly countered by the ice sheet geometry over the subglacial lake. It is therefore difficult to ascertain which of these variables most affected lake stability and drainage potential, especially since these results do not integrate the effects of rapid meltwater inputs capable of destabilizing lake seals. Such potential drainage triggers exist in southern British Columbia.

Beyond providing initial subglacial lake development, geothermal springs could also prove critical as drainage triggers if enhanced activity rapidly increased meltwater volumes and destabilized seals. Another possible outburst trigger lies ~150 km northwest of Okanagan Valley. Subglacial volcanoes (tuyas) occur within the drainage network of North Thompson valley (Fig. 3.2) in the Wells Gray-Clearwater volcanic field. Volcanoes at 1200 m asl erupted under at least 500 m of ice (Hickson et al., 1995). These volcanoes have dominantly basaltic compositions. The higher temperatures and lower pressure of basaltic eruptions, compared to rhyolitic eruptions, lead to greater meltwater production (> 80% of heat transferred to ice) (Höskuldsson and Sparks 1997), and generate higher magnitude jökulhlaups (Smellie, 1999; Tuffen et al., 2001). In addition, the >500 m ice thickness over eruptive centres favoured subglacial ponding rather than supraglacial overflow of meltwater (cf. Smellie, 2006), and probably increased the likelihood of producing an underburst routed through existing valleys. Eruptions occurred during the late stages of the Fraser Glaciation (Hickson, 1987), although the exact timing is uncertain (Hickson et al., 1995). Under elevated water level conditions such as in an ice-dammed lake or subglacial lake, the North Thompson valley would connect with the Shuswap Lake-Okanagan Valley network via Thompson valley (Fig. 3.2). Any meltwater produced by eruptions was temporarily ponded in a subglacial lake and underwent rapid drawdown shortly after (Neuffer et al., 2006). This could also lead to ice roof sagging over the subglacial lake, increasing the potential for supraglacial lake development, and providing an additional meltwater source and drainage trigger for the growing subglacial lake. Sudden water input to the connected subglacial lakes would dramatically increase hydrostatic pressure given elevation differences between tuyas and valleys. A domino-like cascading effect of successive lake failure could occur with initiation of the underburst along valleys due to rapid water input. Thus, geothermal and volcanic activity can be invoked for subglacial lake expansion and as possible drainage triggers.

3.8.3.3. Meltwater reservoir drainage and underburst routing

It is perhaps more problematic to explain how drainage of a subglacial lake in the Wells Gray-Clearwater volcanic field could lead to drumlin erosion on plateaus, especially in the western portion of the Fraser-Thompson-Okanagan (FTO) corridor. This question cannot be fully addressed without considering the scale of this swarm and the location of other potential water sources. The Thompson Plateau swarm is a small portion of the much more extensive FTO swarm (Fig. 3.2). Therefore, the scale of events responsible for the FTO swarm far exceeds those proposed for the Thompson Plateau and Okanagan Valley regions. Drumlin continuity within this swarm and similar morphological characteristics to those on Thompson Plateau suggest similar formative processes. Consequently, drumlins of the FTO swarm may also record subglacial meltwater erosion by a much larger underburst than the one discussed for the Okanagan Valley landsystem earlier. These more speculative regional connections are exlored.

The Wells Gray-Clearwater area may be an adequate source of water and initiation point for an underburst. However, drainage along this system is largely confined to the North Thompson valley and cannot, at first glance, account for drumlins in the western portion of the FTO swarm south of Fraser Plateau (Fig. 3.2). Therefore, two questions need to be addressed: 1) If drumlins along the FTO swarm record erosion by meltwater flows, where was the meltwater source located?, and 2) How can initially channelized drainage (e.g. Wells Gray-Clearwater drainage along North Thompson valley) lead to broad erosional drumlin swaths on Thompson Plateau? These questions are treated separately.

Tracing drumlin swarms to their starting points can help locate a hypothetical meltwater reservoir. The Fraser Plateau is located at the head of two substantial drumlin swarms, the southward-trending FTO swarm and a northward trending swarm (partially shown on Fig. 3.2). The Monashee Mountains and Coast Mountains bound the eastern and western plateau margins respectively, creating a regional topographic low over Fraser Plateau where streamlined bedforms are largely absent. This area may have been the site of a subglacial lake, especially as they preferentially develop at the base of mountain ranges (Dowdeswell and Siegert, 2003). In addition, it is also possible that this topographic low over Fraser Plateau persisted throughout glaciation if the CIS never fully formed a continuous ice dome over Interior British Columbia (Tipper, 1971). The topographic low would have taken the form of a pronounced sag in the ice surface capable of collecting and storing supraglacial meltwater. Hydraulic connections between Fraser Plateau and the Wells Gray-Clearwater volcanic complex could occur through the bedrock valley of Mahood and Canim Lakes (Fig. 3.2). Consequently, it is conceivable that a subglacial and/or supraglacial reservoir could develop on Fraser Plateau. Its drainage may have been contemporaneous with subglacial volcanic eruptions initiating drainage along the North Thompson River. At present, field evidence for this reservoir is sparse. Its location is inferred from drumlin distribution and rationalized through landform continuity arguments. A full discussion of sedimentary deposits on Fraser Plateau, possibly recording this reservoir, is beyond the scope of this paper. Southward flow along the North Thompson valley debouches into the Thompson valley before expanding onto Thompson Plateau. A prominent constriction occurs at the confluence of these valleys and could have acted as a hydraulic dam. Hydraulic damming and subsequent release from the North Thompson Valley allows flows to overtop Thompson Plateau and decreases the effects of head loss through Thompson Valley.

3.9. Effects of subglacial lake development on CIS geometry

The possibility that rapid ice flow occurred over wide areas during underbursts was largely rejected earlier (§ 3.8.1). However, rapid ice flow could still have occurred during subglacial lake development in the valleys, with repercussions on CIS geometry. The main repercussion is a reduction of ice volume in deep valleys compared to plateaus. This is accomplished in two ways: i) ice does not completely fill valleys as a partial volume is occupied by subglacial water, and ii) thin ice is maintained over subglacial lakes, a result both of capture mechanics and enhanced flow (§ 3.8.3).

Ice sheet geometry modified by the presence of subglacial lakes also affects patterns and processes of deglaciation. Thin valley ice and comparatively thicker plateau ice would lead to deglaciation of valleys before plateaus, given a postulated rapid rise in ELA during deglaciation (Fulton, 1975, 1991), especially if underbursts led to localized rapid flow and ice fracturing over the subglacial lake during drawdown and along the underburst path (Björnsson, 2002). Some fractured portions of the CIS could also be evacuated during these event(s) but a majority would be trapped within valleys at bedrock constrictions to create temporary dams (cf. Eyles et al., 1990). This emerging pattern of deglaciation is a reversal of the prevailing model of CIS decay in southern British Columbia where thick valley ice persists through deglaciation, while plateaus deglaciate prior to valleys (Davis and Mathews, 1944; Fulton, 1991). This revised deglacial model suggests that climate need not be the main driver of CIS decay. It also has important implications for the representation of CIS behaviour and the parameterization of subglacial hydrology in ice sheet models.

3.10. Conclusions

The landsystem of the Thompson Plateau-Okanagan Valley region records evidence of erosion and deposition by subglacial meltwater underburst(s). Proposed underburst(s) occur in an area previously identified as a potential corridor for rapid ice flow (Eyles et al., 1990; Mathews 1991; Kleman and Stroeven 2006; Kleman and Glasser, 2007). The prior proposition of ice streaming relied on the presence of drumlins, and physiographic characteristics such as overdeepened bedrock troughs that resemble some modern ice stream settings. Landforms include drumlins of various shapes composed of sediment and bedrock, and tunnel valleys eroded in bedrock and sediment. Prevailing genetic explanations for these landforms, focusing on sediment deformation (Boulton and Hindmarsh, 1987), are unable to account for the composition of drumlins, their morphology and spatial arrangement, the morphology and composition of tunnel valleys, landform associations and sediment flux continuity. A unifying landscape model involves erosion and deposition by subglacial meltwater underbursts. These processes account for drumlin genesis, the spatial patterns and association of drumlins with bedrock valleys, and thick sediment fills in valleys of interior British Columbia.

The meltwater model of landscape evolution proposed here raises questions about the subglacial hydrology of the CIS, particularly subglacial storage and release of meltwater. Underbursts inferred from the subglacial landsystem record require large reservoirs to store and release water. The pronounced topography of British Columbia favoured subglacial lake development in deep bedrock valleys such as Okanagan Valley. Geothermal conditions (thermal springs and subglacial volcanic eruptions) may have enhanced subglacial lake development and were possible underburst triggers. In addition, pronounced regional sags in the ice surface, possibly persisted throughout glaciation (Tipper, 1971), and could have formed over eruption sites. These sags may have been the sites for supraglacial lakes, the drainage of which could have facilitated subglacial lake growth and/or triggered subglacial lake drainage. Although the scale of these inferred processes far exceed any documented events in modern ice sheets, the hydrologic conditions and processes are consistent with our understanding of modern subglacial hydrological conditions in Iceland, Greenland and Antarctica.

In terms of glaciodynamic behaviour, the Thompson Plateau and Okanagan Valley landsystems may (indirectly) record rapid ice flow. Rapid ice flow may result in two ways: i) short-lived localized accelerations, akin to surges, occurring during underbursts (duration of these accelerations is controlled by the duration of the underburst), and ii) longer-lived accelerations during subglacial lake build-up. The latter may slow over time as ice thickens downflow of the subglacial lake. Glacier acceleration results from loss of resistance to flow, not sediment or ice deformation. Rapid ice flow is partly regulated by the number, size and location of pinning points along the underburst path. Sediment deformation, and abrasion processes are not questioned *per se*, their ability to produce the drumlins in our region is . Consequently, extensive tracts of streamlined bedforms along this portion of the CIS may not be indicative of ice flow (rapid or otherwise) driven by bed deformation. This paper highlights the ongoing need for critical evaluation of assumptions that underpin landform genesis models and many glacial landsystem models. It also raises important questions about the role of meltwater in shaping glaciated landscapes.

Interpretations of the Thompson Plateau and Okanagan Valley landsystems also have repercussions for ice-sheet modeling in terms of ice sheet geometry and ice volumes of the former CIS. Subglacial lake development and transient rapid flow can redistribute ice mass within an ice sheet. In the CIS, preferential ice flow over valleys (and the presence of lakes there) thins valley ice, influencing patterns and processes of deglaciation. Thin valley ice and comparatively thicker plateau ice leads to deglaciation of valleys before plateaus. This deglacial pattern is a reversal of the prevailing model of CIS decay in southern British Columbia. It suggests that endogenous processes may partly control CIS decay, in addition to exogenous variables like climate forcing.

4. Paleogeographic and paleoenvironmental evolution of glacial Lake Penticton during decay of the Cordilleran Ice Sheet, British Columbia, Canada

Abstract

Decay of the last Cordilleran Ice Sheet (CIS) led to development of extensive deglacial lakes in the deeply-incised valleys of southern British Columbia. These lakes and their associated landforms constitute an extensive sedimentologic and geomorphologic archive of past environmental changes associated with decay of a montane ice sheet. CIS decay in Okanagan Valley near its southern terminus, led to development of glacial Lake Penticton (gLP) as a proglacial lake impounded by an ice and sediment dam in a valley constriction near McIntyre Bluff. Regional correlation of perched deltas at tributary valley mouths and bedrock spillways around McIntyre Bluff record a single regional water plane at 500-525 m asl for gLP. An absence of perched deltas below this elevation suggest that drainage of gLP took place as a single (catastrophic?) event without intervening water plane stabilization, contrary to previous reconstructions of four distinct lake stages. Correlated perched deltas also allow for assessment of glacioisostatic rebound in southern British Columbia. Despite evidence for high glacioisostatic rebound within other glacial lake basins of southern British Columbia, no evidence of glacioisostatic rebound can be discerned within Okanagan Valley, suggesting that rebound either occurred within the time span of delta development, and/or that potentially uniform rebound occurred over the gLP basin due to thin valley ice of relatively uniform thickness. Remnant benches of lacustrine sediments on the valley sides (the 'White Silt' benches) record abundant sediment and meltwater transfer from tributary and possibly from valley-axis sources. Seismic profiles and transitions between seismic facies lend little support to previous suggestions that gLP developed as a supraglacial lake and that White Silt benches result from collapse of supraglacial sediments. An unconformity seen on seismic profiles and truncating putative glaciolacustrine sediments suggests that the White Silt benches result from erosion along the valley axis during lake drainage. Sediment delivery from tributary valleys requires meltwater sources in the surrounding uplands and plateaus coeval with a

proglacial lake in the valley axis. Receding ice may also have been present in the valley axis. However, its extent was probably minimal since it has left little evidence within the seismic record of the valley fill.

4.1. Introduction

The Cordilleran Ice Sheet (CIS) developed over the pronounced topography of the northern portions of the western Cordillera of North America, covering British Columbia and extending westward into the Rocky Mountain foothills, northward into portions of the Yukon, and southward into northern Montana, Idaho and Washington State. This topography played a key role in determining the distribution and thickness of ice over the landscape (see CIS model of Kerr (1934), and Davis and Mathews (1944) summarized by Fulton (1991) amongst others). For example, in the southern interior of British Columbia, valleys accumulated much greater volumes of ice than plateaus. The pronounced bed topography also had an effect on the production, storage and drainage of meltwater during glacier advance and retreat. Consequently, glacial lakes were ubiquitous during growth (Eyles and Clague, 1991; Huntley and Broster, 1994) and especially decay of the Cordilleran Ice Sheet (Flint, 1935b, Matthews, 1944; Naismith, 1962; Fulton, 1967, 1969; Ward and Rutter, 2000; Johnsen and Brennand, 2004). Deeply-incised valleys of British Columbia hosted a number of these lakes and their record is largely preserved as valley-side benches of glaciolacustrine sediments. Sediments and landforms from glacial lakes record environmental changes associated with the CIS and offer insights into deglacial processes and patterns of a montane ice sheet developed over pronounced topography. Studies of the CIS offer potential analogues for improved understanding of modern ice sheets (e.g. Greenland, Antarctica) developed over pronounced topography.

4.2. Objectives

This chapter focuses on paleogeographic and paleoenvironmental reconstructions of glacial Lake Penticton (gLP), in southern British Columbia. Specifically, focus is placed on:

- documenting gLP paleogeography and evolution by focusing on the geographic extent of gLP in Okanagan Valley through regional water plane reconstruction;
- assessing the evidence for glacioisostatic rebound in Okanagan Valley and examining glaciodynamic conditions responsible for glacioisostatic rebound patterns;
- examining the spatial evolution of gLP, focusing on temporal changes in lake extent and volume; and
- reconstructing paleoenvironmental conditions in and around the gLP basin by examining water and sediment delivery pathways to the gLP basin in order to better constrain lake type, the distribution of ice in and around the gLP basin, and the style of lake drainage and its effects within the lacustrine basin.

4.3. Study area and previous research

4.3.1. Study area

The study area is located around Okanagan Valley and the surrounding Thompson Plateau in south-central British Columbia. Specifically, the focus is placed on a ~ 2000 km^2 area, centred on Okanagan Lake (Fig. 4.1). Okanagan Valley is a north-south trending, structurally-controlled valley bounded to the west by Thompson Plateau and to the east by the Okanagan Highlands. Okanagan Valley formed initially by mid-Eocene extension (Templeman-Kluit and Parkinson, 1986) and has been repeatedly eroded by glacial and fluvial processes. The valley extends southward into Washington State where it meets the Columbia River. The northward extension of the valley leads to the Shuswap Lakes and the Thompson River drainage (Holland, 1964) (Fig. 4.1). Fluvial and glacial processes have eroded the valley as much as 650 m below sea level (Eyles et al., 1990, 1991; Vanderburgh and Roberts, 1996). Steep-sided, dominantly v-shaped bedrock tributary valleys traverse the valley sides and link Okanagan Valley and the surrounding plateaus (Figs. 4.2, 4.3 A, B, C). Total relief (plateau to bedrock floor) exceeds 2000 m (Eyles et al., 1990, 1991). Presently, valley floor elevation is ~340 m asl and generally decreases southward toward the Columbia River (~250 m asl at confluence). The modern valley floor is dominated by N-S-oriented lake basins. The largest of these is Okanagan Lake (~342 m asl) extending along valley for up to 120 km with a width varying between 3-5 km. The valley fill below Okanagan Lake and surrounding areas contains up to ~730 m of sediments (Eyles et al., 1990, 1991), exceeding 100 km^3 in volume (Eyles et al., 1991, Vanderburgh and Roberts, 1996).

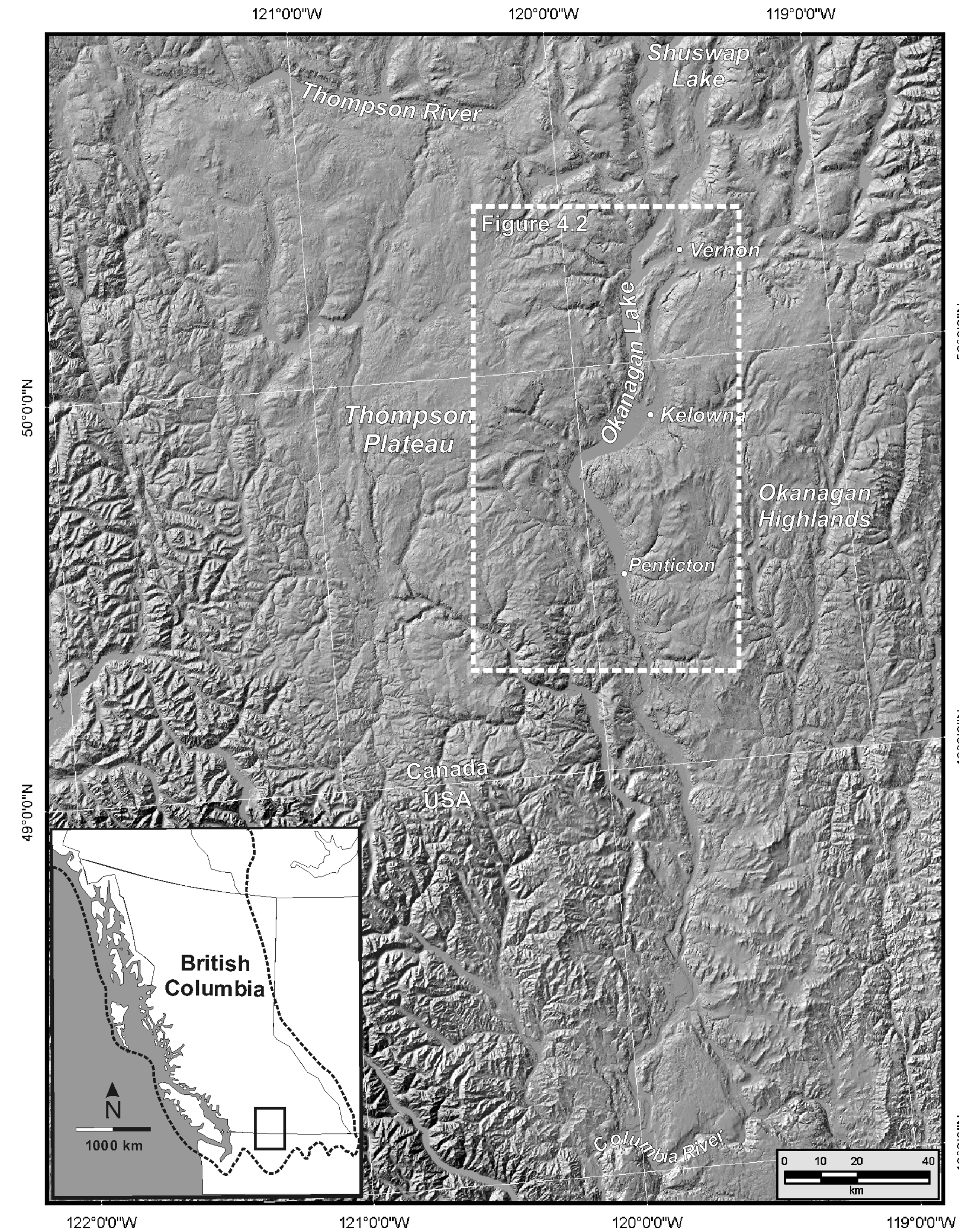


Figure 4.1. Hillshaded digital elevation model of the field area highlighting major physiographic elements of the southern Interior Plateau discussed in the text. Data for Canadian regions: BC Government TRIM I dataset, 25 m resolution (© GeoBC, by permission). Data for American regions: USGS 10 m DEM reprojected to BC Albers, NAD 83.

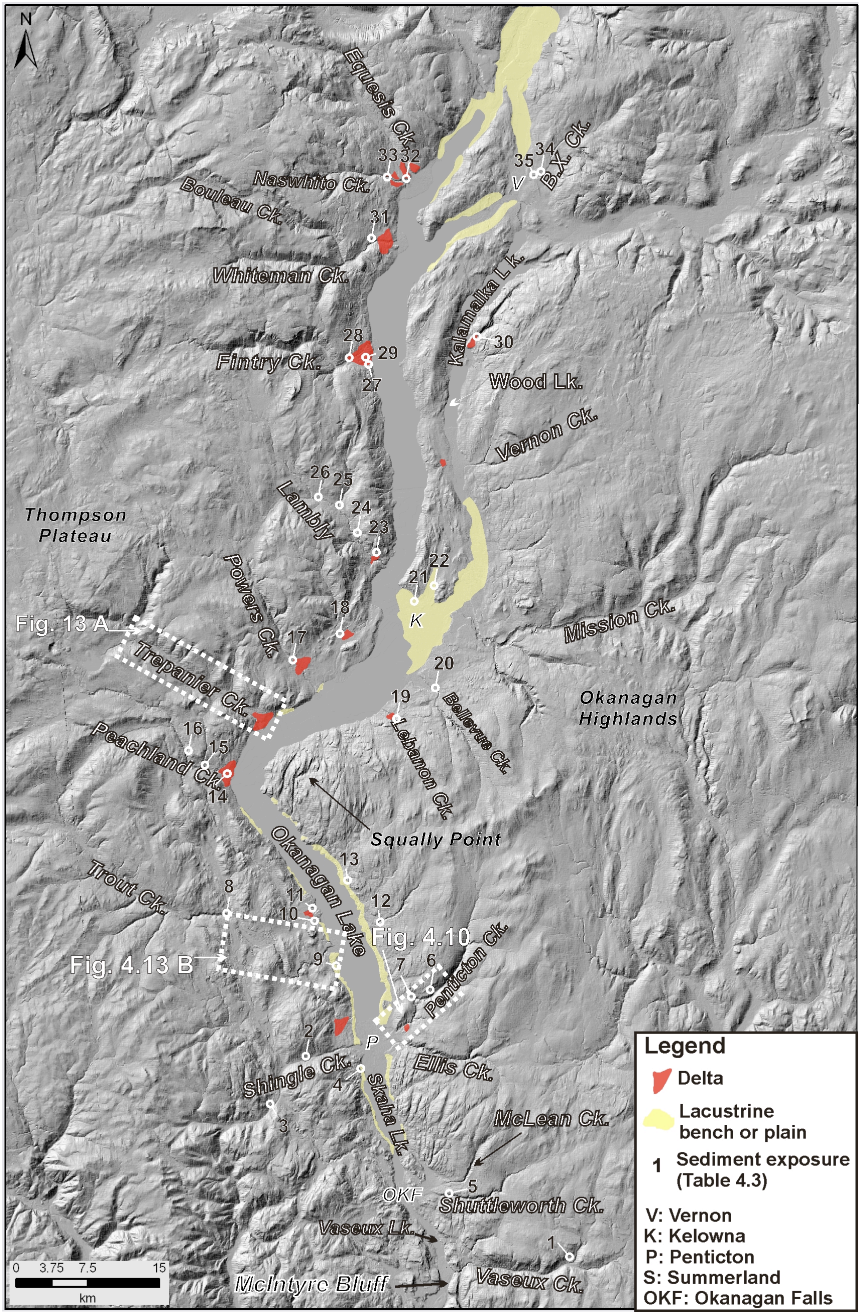


Figure 4.2. Hillshaded digital elevation model of Okanagan Valley and surrounding uplands. More than twenty one major tributary valleys drain toward Okanagan Lake from the surrounding uplands. Lacustrine sediment benches, plains, and deltas record the presence of glacial Lake Penticton. Dashed boxes locate subsequent figures. The location of 35 exposures documented in Table 4-3 are indicated on this map. Additional minor exposures are not shown on this map. Base map: BC TRIM 25 m DEM © GeoBC, by permission. Site coordinates and details given in Appendix D.

***Figure 4.3.** Landforms and sediments of the glacial Lake Penticton basin. A) Every tributary exhibits a narrow bedrock gorge in its lower reaches. Bench surfaces (arrow) occur in sediments overlying bedrock (Ellis Ck.). B) High Bench (~900 m asl) surface (arrow) overlying bedrock in a tributary valley (Shuttleworth Ck.). C) Up-valley (northeastward) view of tributary benches (arrows) resulting from valley train dissection along Penticton Ck. D) 'White Silt' terraces flanking the valley sides consisting of rhythmic beds of fine sand and silt exposed in cliffs up to 50 m high. Crouched person for scale (red circle). E) Down-valley (southward) view of McIntyre Bluff, proposed damming site for gLP (Nasmith, 1962). F) Surface (arrow) and steep front of a perched delta at the mouth of Trepanier Creek. Refer to Figure 4-2 for location of creeks and McIntyre Bluff.*

Prominent benches of fine sand, silt and occasionally clay flank the valley walls along Okanagan Lake. These benches are most prominent south of Squally Point along the southern portion of Okanagan Lake (Figs. 4.2, 4.3 D). However, more extensive lacustrine plains (with local perched benches) also occur near Kelowna and Vernon (Fig. 4.2) where they form the valley floor. Silt benches typically rise over 40-50 m above the modern valley floor. They onlap the bedrock valley walls and slope toward the valley axis. Sediments in the silt benches were deposited in glacial Lake Penticton during Late Wisconsinan deglaciation of the CIS (Flint, 1935b; Nasmith 1962; Fulton, 1967).

4.3.2. Previous research on glacial lake development in Okanagan Valley

Existing understanding of the paleogeography and paleoenvironment of gLP stems from studies conducted in different regions of the gLP basin and under varying research mandates. These include reconnaissance-scale landscape descriptions (Dawson, 1879), surficial mapping (Fulton, 1969), and lake-based (Eyles et al., 1990, 1991) and land-based (Vanderburgh and Roberts, 1996) seismic stratigraphic studies. Consequently, different and potentially conflicting paleogeographic and paleoenvironmental reconstructions (Table 4.1) arise from different geographic areas or regarding the time frame of lake evolution.

4.3.2.1. Recognition of a paleo-lake basin in Okanagan Valley

The first proposition of a paleo-lake basin within Okanagan Valley was advanced by Dawson (1879) who interpreted the valley-side flanking silt benches (Fig. 4.3 D) as Pleistocene-age lake-bottom sediments (he later reinterpreted these deposits as marine sediments (Dawson, 1896), though this interpretation was refuted by Daly (1915) based on exposures of glaciolacustrine sediments in other valleys of southern British Columbia). The glaciolacustrine origin of these sediments has since been confirmed by numerous studies (Table 4.1). Nasmith (1962) proposed the name glacial Lake Penticton (gLP) to identify this deglacial lake. The timing of deglaciation in Okanagan Valley is poorly constrained. Consequently, the duration of this lake and its timing are loosely bracketed. Glacial maximum was reached circa 16.5 k a BP (Fulton, 1991). Alley (1976) inferred gLP drainage before 8.2 k a BP (based on radiocarbon dates from peat a bog on the valley floor). More recent optical ages from loess overlying lake sediments suggest that the highstand of this lake was completed by 11 ± 1 ka BP (Lesemann et al., in review).

4.3.2.2. Paleogeography of gLP

Although gLP is generally thought to have occupied the main Okanagan Valley trough (Flint, 1935b; Nasmith, 1962; Fulton, 1969), various interpretations of its basin-wide extent have been proposed (Table 4.1). The prominent silt benches in the area between Squally Point and Okanagan Falls (Fig. 4.2) led to the suggestion that gLP developed preferentially in the southern portion of Okanagan Valley (Shaw, 1977; Shaw and Archer 1978, 1979). However, silt benches and localized silt plains occur to the north of Squally Point (Fig. 4.2), near Kelowna, and in the northern reaches of Okanagan Lake north of Vernon (Fig. 4.2). More recently, lake- and land-based seismic have revealed up to 730 m of lacustrine sediments below Okanagan Lake (Eyles et al., 1990, 1991) and extending beyond the modern lake boundaries where they make up a significant portion of the valley fill in the northern part of the gLP basin (Vanderburgh and Roberts, 1996) and along Kalamalka Valley around Kelowna (Paradis et al., 2010) (Fig. 4.1). Fulton (1969) expanded on Nasmith's (1962) work and used a limited number of tributary mouth deltas in a geographically-restricted portion of Okanagan Valley (Table 4.1) near Vernon (Fig. 4.2) to propose four stages of gLP. He postulated a maximal water plane at ~520 m asl called the Grandview Flats stage. Three other stages associated with lake lowering (490 m asl, 430 m asl and 340 m asl), were also proposed (Table 4.1). These reconstructed water planes were assumed to be representative of the entire gLP basin. Their extrapolation beyond the boundaries of the mapsheet under study (NTS mapsheet 82E sw) (Table 4.1) was not constrained by water plane indicators south of Vernon (Fig. 4.2).

4.3.2.3. Paleoenvironment of gLP

Early interpretations of the 'White Silt' benches favoured tributary streams (Flint, 1935b; Nasmith, 1962) delivering meltwater and sediment into a supraglacial lake (Flint, 1935b), or narrow ice-lateral 'ribbon lakes' developed between the valley sides and a stagnant ice tongue in the valley axis, and later merging into a proglacial basin as ice in

Table 4.1. Proposed paleogeography and paleoenvironment of glacial Lake Penticton

Author(s)	Lake type	Environmental setting	Origin of silt benches	Lake extent [1]	Supporting evidence
Flint (1935)	Supraglacial and ice lateral 'ribbon lake'	Tributary valley-supplied sediments deposited on top of stagnant ice in valley axis, and along ice margins	Collapse due to melt of valley-axis ice tongue	Okanagan Lake basin-wide extent, north of Vernon to north of Armstrong, Kalamalka Lake valley to Kelowna, Skaha Lake	Lake extent determined by distribution of silt benches. Sediment input inferred from lateral and basinward gravel to silt successions within silt benches recording proximal-distal relationships near tributaries
Nasmith (1962)	Ice lateral 'ribbon lake'	Ice lateral lake circumscribing a decaying ice tongue in valley axis. Sediments delivered via tributaries	Fill of ribbon lake basin, partial collapse from melting valley-axis ice tongue	Okanagan Lake basin-wide extent, north of Vernon to north of Armstrong, Kalamalka Valley to Kelowna	Lake extent determined by distribution of silt benches. Lake type inferred from geometry of silt benches on valley sides.Tributary sediment input inferred from presence of sediment fans
Fulton (1969)	Ice lateral and proglacial	Dead ice tongue in valley axis in front of northward-retreating decaying ice in valley[2].	Not addressed by model due to absence in study area.	Four lake stages identified in northern portion of Okanagan Lake basin, centred on Vernon and surrounding areas.	Changes in lake extent inferred from tributary mouth deltas implying some tributary sediment delivery (though not explicitely stated). Lake type and valley-axis sediment sources follow deglacial model of Kerr (1934) and Davis and Mathews (1944)
Shaw (1977)	Supraglacial and proglacial	Supraglacial delta built against and on top of decaying ice tongue in valley axis supplying meltwater and sediments	Sediment compaction and collapse due to melt of valley-axis ice tongue	Southern portion of Okanagan Lake basin (south of Squally Point), Skaha Lake, (represents temporal 'snapshot')	Lake extent reflects abundance of silt terraces in the study area and the position of hypothesized ice tongue at time of emplacement. Lake type and ice position follow model proposed by Fulton (1969)
Eyles et al. (1990, 1991)	Proglacial with supraglacial component	Initial proglacial and partly subglacial. Rapidly backwasting valley-axis glacier with calving front. Sedimentation into supraglacial lake atop/around stranded calved blocks	Sediment compaction and partial collapse due to melting of calved ice blocks	Sequential lake development into Okanagan Lake basin-wide extent, north of Vernon to north of Armstrong, Kalamalka Lake valley to Kelowna.	Lake-based seismic surveys revealing quasi-basin-wide distribution of parallel reflections interpreted as laminated glaciolacustrine fines. Environmental interpretations based on analogy to marine-based outlet glaciers calving into fjords.
Vanderburgh and Roberts (1996)	Proglacial	Retreating ice tongue in valley axis supplying meltwater and sediments. Tributary valley fan deltas contributing sediments to lake basin. Valley-side gravel sources correspond to basinward thinning and fining reflections on seismic profiles	Not addressed by model due to absence in study	Represented as northern portion of Okanagan Lake between Vernon and Enderby (reflects extent of study area).	Lake environment inferred from land-based seismic surveys and corroborated by drill cores through entire valley fill between Vernon and Enderby. Lake type and retreat style follows deglacial model of Fulton (1969).

[1] Refer to Figures 4.1, 4.2 for locations. Representations of lake extent frequently reflect the extent of the field areas and/or a short temporal 'snapshot'. [2] In this context, Fulton (1969) defines dead ice as ice too thin to deform internally. Decaying ice refers to ice disconnected from an active glacier but sufficiently thick to deform internally.

receded up-valley (Nasmith, 1962) (Table 4.1). In both these cases, the valley-side silt benches result from sediment collapse associated with melting of valley-axis ice. Remnant benches are preserved where sediments were deposited on bedrock. Nasmith (1962) also proposed that gLP occupied an isostatically-depressed basin and was dammed by ice and sediment within a valley constriction located between Okanagan Falls and McIntyre Bluff (Figs. 4.2, 4.3B). He envisaged drainage of gLP by breaching of the decaying ice and sediment dam, punctuated by stillstands, though drainage mechanics were not elaborated and it is unclear how dam breaching would lead to stillstands.

Fulton (1969) developed a regional reconstruction of the glacial lake network in southern British Columbia, emphasizing ~northward retreat of valley-occupying ice tongues, and progressive deglaciation of valleys with concomitant proglacial/ice-lateral lake development (Fulton 1969). Fulton's (1969) delta-based reconstructions imply some sediment sourcing from tributaries. However, his overall conceptual deglacial model clearly emphasizes valley-axis decay of ice tongues, presumably supplying meltwater and sediment to gLP. Fulton (1969) relied on the dam mechanism and location of Nasmith (1962) to impound three of his four proposed lake stages within a glacioisostatically-depressed basin. Final lake drainage occurred as a result of differential glacioisostatic uplift (Fulton, 1969).

Shaw (1977) and Shaw and Archer (1978) examined the sedimentology of the White Silt benches. Although Shaw (1977) speculated on the possibility of an initial 'ice lateral' lake phase to explain the maximum elevation of some lacustrine sediments, most of this sedimentologic work is couched within the paleoenvironmental context developed by Flint (1935b) and Nasmith (1962). Shaw (1977) and Shaw and Archer (1978) proposed a supraglacial origin for gLP and development of a delta built on stagnating ice and abutting a decaying ice tongue. This putative delta was located south of Squally Point, a location seemingly proposed based on the extent of the most prominent lacustrine benches around South Okanagan Lake (Fig. 4.2). The existence of this delta was rationalized based on the presence of 'winter sand' layers (Shaw, 1977; Shaw and Archer, 1978) within clayey lacustrine beds. Winter sands were attributed to delta front failures. Down-valley paleoflows in lake sediments supported the notion of a valley-axis delta as a sediment source for the White Silt benches. No sedimentologic or seismic stratigraphic evidence was (or has since been) used to support the existence of this delta. Much like Fulton's (1965) work in other glaciolacustrine basins in southern British Columbia this sedimentology work recognized that large volumes of sediment were deposited rapidly within gLP. This work also highlighted the importance of turbidity currents and grain flows as depositional agents within the lake basin, suggesting episodic (seasonal) sediment delivery events by periodic failures of the delta front (Shaw, 1977, Shaw and Archer, 1978). Inferences of abundant sediment delivery by sediment gravity flows were rationalized within the context of a decaying valley-axis ice tongue. Formation of the silt benches was ascribed to collapse and let-down by the stagnating ice tongue, and differential compaction of the silts (Shaw, 1977). Eyles et al. (1990, 1991) envisaged two phases of lake sedimentation: 1) voluminous sediment delivery near the ice margin from ice tunnels, and 2) supraglacial rhythmic deposition atop grounded calved ice blocks and partly on bedrock valley walls. Their environmental model is largely based on analogies with retreating glaciers in some coastal fjords (Eyles et al., 1990, 1991).

Vanderburgh and Roberts (1996) identified valley-side sediment wedges fining toward the valley axis. They proposed that these sediments were deposited in a proglacial lake as the ice front retreated up-valley and presumably acted as a significant (though not exclusive) meltwater source. These sediment wedges were envisaged as being derived from tributary streams and gullies. They reported no evidence consistent with ice-contact deposition, or stagnation (regional or block-scale) of ice masses within the proglacial basin (Vanderburgh and Roberts, 1996). The vast majority of gLP studies have noted the absence of dropstones within the White Silt benches.

4.4. Methods and terminology

Paleoenvironmental reconstructions presented herein rely mainly on landform identification and sedimentologic interpretations. A combination of aerial photographs and digital terrain models were employed to identify landforms and to assess their regional context. Landform genesis is determined by examining their internal composition, geomorphology and geomorphic context. Sediments within landforms are described using standard sedimentologic approaches and criteria such as sediment texture and structure, paleocurrent measurements, the nature of bed contacts, and the lateral extent and thickness of units (Table 4.2). Paleocurrents were measured from imbricate clasts (a-b plane), cross-laminations (for ripples), and cross-beds (dunes). Over 35 exposures were identified throughout Okanagan Valley, ranging in height from 2 m to over 50 m. The best exposed and most extensive exposures were logged at the decimetre scale. Regional water planes (lake extent at a particular time) are reconstructed by identifying and integrating landforms considered to be reliable water plane indicators (e.g., deltas), and others (secondary water plane indicators) that tend to underestimate water plane elevations (subaqueous fans and lake-bottom sediment benches). For paleogeographic reconstructions, landform position and elevation was surveyed using a differential Global Positioning System (dGPS) used in real time kinematic collection mode. Measurements of delta elevations and tributary benches were performed on surfaces identified in the field and from aerial photographs. Where the topset-foreset transition was exposed in deltas, elevation was also measured at this transition as this is considered to be the most accurate location of a former water plane. Measurement error is conservatively estimated between ~0.2 to 0.8 m due to discrepancies in elevations associated with outdated geodetic reference markers and the geoid model used in data processing. These errors are considered insignificant given the local relief on landforms, the scale of the study area, and the uncertainty associated with surveying delta surfaces. Results were analyzed graphically and within a Geographic Information System (GIS). Lake volume calculations were also performed using GIS software.

Paleoenvironmental and paleogeographic reconstructions in this paper rely extensively on valley-side sediment benches, subaqueous fans, and deltas. The origin of valley-side sediment benches is not always clear. Therefore, throughout this paper, the term sediment bench is used to describe valley-side sedimentary bodies. The term terrace is used genetically (e.g. elevated river floodplain) and is reserved for cases where genesis of the sediment bench is clearer. The term subaqueous fan is used *sensu lato* to refer to sediments deposited in a fan-shaped body below the lake water plane; within this context, a subaqueous fan need not have formed at the mouth of a submerged glacier tunnel (*quod vide* Rust and Romanelli 1975). GLP subaqueous fans are broadly analogous to subaqueous deltas of Nemec (1990a), the sublacustrine-fan of

Larsen and Smith (1999), and may also be analogous to some coastal 'fan deltas' that include a significant subaqueous depositional component (e.g. Prior and Bornhold, 1988, 1989). Lastly, the term delta refers to a sedimentary surface graded to a stable water plane (located at the foreset-topset bed transition).

4.5. Landform identification for paleogeographic reconstructions

4.5.1. Primary water plane indicators

4.5.1.1. Wave-cut benches and beaches

Paleo-lake reconstructions commonly rely on wave-cut benches to infer water plane elevations and reconstruct former lake extent (e.g., Johnsen and Brennand, 2004). In Okanagan Valley no wave cut benches have been reported. This may reflect unfavourable wind conditions (intensity and direction) within the gLP basin, combined with bedrock valley walls which, when compared to sediment surfaces, require more sustained wave action to develop wave-cut benches. Sediments are more frequent along the valley walls of Kalamalka Valley and North of Kelowna. Nasmith (1962) reported one beach at the northern extremity of Kalamalka Lake, though its exact location is unknown. A second beach near Oyama occurs near the base of a perched delta and may only be recognizable in the field 'under favourable light conditions' (Nasmith, 1962). Paradis (2009) mapped a series of surfaces interpreted as beach ridges between Kelowna and Oyama, and the surrounding uplands. Some of these features occur within narrow tributary valleys and are probably unrelated to gLP because they occur at elevations well above (in excess of 700 m asl) the proposed highstand of gLP near 500 m asl (Nasmith, 1962; Fulton, 1969). The origin of mapped beach ridges within the gLP basin is equivocal to some degree since no corroborating sedimentologic data are provided to justify interpretations. More significantly, these features are discontinuous and are so rare that correlations are unfeasible. Thus, paleo-water plane reconstructions from these landforms are not possible within the gLP basin.

Table 4.2. Descriptions and interpretations of lithofacies for glacial Lake Penticton basin and tributary sediment sources.

	Lithofacies name and symbol	Description	Depositional process	Depositional environment	Refs.
Clay and silty clay	Laminated clay and silty clay	Laminated clay and silty clay. Often interlaminated with silt and very fine sand in rhythmic couplets 0.03-0.1 m thick. Beds of multiple couplets can be up to 0.25 m thick. May contain soft sediment deformation structures including flames, and ball and pillow structures	Suspension settling	Subaqueous, lake-bottom distal from	1, 2,10, 25
Silt	Laminated silt	Planar to undulating laminated silt and fine sand. Normal grading sometimes visible. Parting lineations due to presence of very fine sand. Often drapes coarser beds. Soft sediment deformation structures (flames, ball and pillow) common. Beds 0.2-0.8 m thick	Suspension settling, occasional minor traction	Subaqueous, lake-bottom	2, 8, 22, 23
Sand	Normally graded sand	Normally graded coarse to fine sand, sometimes capped by silty sand. Can be massive to diffusely-graded near base. Beds 0.25-1.1 m thick	Suspension settling from waning turbidity currents. Waning stage deposits from subaerial outwash	Subaqueous, lake bottom; outwash plain (valley train)	10, 12, 13, 27
	Reverse graded sand	Reverse graded medium to coarse sand. Beds 0.2-0.8 m thick	Deposition from grain flows	Subaqueous fan; delta foreslope	1, 10
	Planar stratified sand	Planar stratified fine to coarse sand with occasional silt lenses and granules and small pebbles in clusters. May contain cm-scale soft sediment clasts and armoured clay clasts. Beds ~0.05-0.3 m thick. Bedsets up to 4 m thick.	Deposition from traction from turbidity currents. Waning stage deposits from river flows	Subaqueous fan; delta foresets; outwash plain (valley train)	4, 6, 14, 18
	Diffusely graded sand	Diffusely graded medium sand, faint stratification. Sharp (erosive) lower contacts. Occurs as laterally continuous beds (at least 3-7 m wide, 0.5-1 m thick) or within scours (3-5 m wide, up to 2 m deep). May contain silty sand rip-up clasts	Rapid suspension settling from hyperconcentrated flows and turbidity currents	Subaqueous fan	4, 5, 6, 14, 16, 18, 21
	Trough cross-stratified sand	Trough cross-stratified medium to coarse sand. Occurs as well sorted sand and less well sorted medium sand and granules with occasional pebbles. Beds up to 0.7 m thick. Often interbedded with trough cross-stratified gravel	Deposition from dune-forming flows: turbidity currents or subaerial river flows	Subaqueous fan; delta; outwash plain (valley train)	6, 20, 22
	Cross-laminated sand	Cross-laminated medium to fine sand and silty sand. Includes stoss-erosional (type A), stoss depositional (type B) and sinusoidal/draped (type S) laminations. Typically occur as cosets 0.1-5 cm thick. Type A most common, may be overlain by type B, and most often draped by type S. Type S often grades into laminated silt	Decreasing traction and increasing suspension settling during ripple deposition by	Delta foresets; subaqueous fans; lake bottom	1, 2, 3, 6, 7, 10, 19, 20

	Lithofacies name and symbol	Description	Depositional process	Depositional environment	Refs.
Gravel	Planar-stratified gravel	Horizontal to sub-horizontal planar stratified gravel beds. Clast supported rounded to sub-rounded pebble to boulder gravel. Beds normally graded but can be locally massive or rarely reversely graded. Imbricate clasts common. Coarse sand to granule matrix. Beds up to 2 m thick	Deposition from high density turbidity currents. Traction deposition of gravel sheets	Subaqueous fan; possibly outwash plain (valley train)	17, 18
	Tabular cross-stratified gravel	Steeply dipping (15-30°) beds of clast-supported pebble to boulder gravel. Individual beds typically normally graded. Imbrication commonly occurs. Beds may be openwork near base of bed and gradually more sandy upwards. Portions of beds sometimes reversely graded. Gradational transitions between normally and reversely graded portions. Beds 0.3-1.5 m thick	Deposition from traction, possible avalanching on delta foreslope	Subaqueous fan; delta	18, 19, 20
	Trough cross-stratified gravel	Clast-supported pebble to boulder gravel. Troughs are 2-5 m wide and 0.25-0.7 m deep with sharp basal contacts	Suspension and traction deposition by dune-forming underflows.Scour-fill by river flows	Subaqueous fan; outwash plain (valley train)	17, 20, 24
	Heterogeneous gravel	Poorly-sorted, clast- to matrix-supported pebble to cobble gravel with granule to silt matrix. Gravel clusters sometimes present. Matrix may be locally laminated within cluster interstices	Traction deposition with local clast support by turbulence	Subaqueous fan; outwash plain (valley train)	17, 20, 21, 24, 26
Diamicton	Stratified silty stony diamicton	Poorly sorted matrix supported silty stony diamicton including ~10% boulders, 20% cobbles, 15-20% pebbles. Frequently laminated and/or interbedded with lenses of laminated silt and sand. Clasts sub-rounded,to sub-angular, some exhibiting faint striae. Localized imbricate clusters	Cohesive debris flow, possible localized fluidal flow. May be remobilizing till	Subaerial outwash(valley train); Subaqueous fan	6, 7, 26,

References: (1) Allen (1985), (2) Ashley (1975), (3) Ashley et al. (1982), (4) Cheel (1989), (5) Cheel and Rust (1986), (6) Eyles (1987), (7) Eyles et al. (1987),(8) Fulton (1965),(9) Gorrell and Shaw (1991), (10) Johnsen and Brennand (2006), (11) Kneller (1995), (12) Lowe (1975), (13) Lowe (1982), (14) Miall (1977), (15) Miall (1983), (16) Middleton and Hampton (1976), (17) Mulder and Alexander (2001), (18) Nemec (1990a,b), (19) Nemec et al. (1999), (20) Postma (1990), (21) Russell and Arnott (2003), (22) Shaw (1977), (23) Shaw and Archer (1979), (24) Shaw and Gorrell (1991), (25) Smith and Ashley (1985), (26) Sohn et al. (1997), (27) Winsemann et al. (2007).

4.5.1.2. Deltas

Deltas record sediment delivery and aggradation to a stable water plane. Within the study area, deltas occur within every major tributary valley (Fig. 4.2). Their morphology consists of gently-sloping surfaces (sloping down tributary valley and toward Okanagan Valley) leading to a steep front (Fig. 4.3F). Deltaic bodies are frequently dissected, and display numerous inset surfaces. Deltas are identified mainly based on their morphology. In some cases, exposures allow for examination of internal architecture which consists of inclined beds of sand and imbricate gravel (foreset beds) overlain by sub-horizontal beds of sand and imbricate gravel (topsets). Due to the scarcity of exposures, paleo-water plane reconstructions rely on the elevation of the upper delta surface, instead of the topset-foreset contact. This creates a likely overestimation of the former water-plane elevation. This overestimation can be as much as 1-4 m (Gustavson et al., 1975; Thorson 1989) and does not take into account seasonal lake level fluctuations.

4.5.2. Secondary water plane indicators

Secondary water plane indicators consist of landforms that under-estimate water-plane elevation. Two types are recognized within the study area: subaqueous fans and lake-bottom sediments.

4.5.2.1. Subaqueous fans

Subaqueous fans consist of sand and gravel bodies formed below a water plane. Within the study area, subaqueous fans occur as high as 800 m asl, though most are found between 500-600 m asl along tributary valleys (Table 4.3). Subaqueous fans can have an irregular surface morphology with variable slope and individual lobate forms. In most cases however, incision and inset surfaces obscure the original form. Therefore, subaqueous fans were identified based on sedimentary facies, facies associations and landscape context (including elevation) (Table 4.3).

Table 4.3. Summary table of additional sedimentologic observations used in regional gLP reconstruction but not documented in text.

Place name	Sedimentologic observations[1]	Landform Setting[2]	Elevation (m asl)[3]	Interpretations
Vaseux Ck.	(1) Planar-bedded and possibly trough cross-bedded gravel and sand over bedrock (5 m high, 10 m long exposure)	Dissected High bench	980	Valley train and bedrock eroded by meltwater supplied from decaying tributary ice near plateau edge. Unclear relation to gLP level. May be remnant of fan extending into valley or ice-contact deposit.
Shingle Ck.	(2, 3) Planar-bedded coarse sand and cobble/boulder gravel interbedded with laminated and ripple cross-laminated fine sand and silt (12 m high, 30 m long, and 4 m high, 75 m long exposures respectively).	Dissected low benches (paired)	610, 740	Subaqueous sand and gravel deposition; incision by meltwater during ice retreat westward along Shingle Ck. basin. Unclear relation to gLP
	(4) Heterogeneous gravel interbedded with laminated silt and fine sand (3.5 m high, 40 m long exposure).	Inset fan surface at mouth of meltwater channel	410	Subaqueous fan deposits in gLP or local pond
Shuttleworth Ck.	(5) Interbedded diamicton and laminated fine sand Cobbles and large pebble clusters load and deform laminated sand (35m high, 12 m long exposure)	Dissected High bench near tributary mouth	730	Subaqueous debris flow (diamicton) and suspension deposition into lacustrine setting developed in regionally-dammed subglacial lake or ice-dammed lake in tributary valley
Penticton Ck.	(6) Planar-stratified medium to coarse sand overlying trough cross-stratified coarse sand (3m high, 10 m long exposure)	Dissected (High or gLP?) bench	645	Subaerial or subaqueous outwash deposited in regionally-dammed subglacial lake or ice-dammed lake in tributary valley (subaqueous), or valley train (subaerial)
	(7) Interbedded/interlaminated clay, silt, and ripple cross-laminated fine sand (14 m high 200 m long exposure)	Gently sloping sediment surface in local bedrock basin	775	Deposition by underflows and suspension in isolated lacustrine basin on valley sides fed by meltwater channel originating near 1100 m asl
Trout Ck.	(8) Planar-bedded coarse sand and gravel interbedded with ripple cross-laminated, and planar laminated fine sand and silt. Fines loaded by gravel clasts (7 m high, 12 m long exposure, near Faulder)	High bench remnant at tributary mouth	700	Subaqueous deposition of tributary outwash stream into local lacustrine basin or in regionally-dammed subglacial lake
	(9) Cross-laminated fine sand and silt (a, b, and s-types) beds (40 m high, 200 m long exposure, near Sun Oka Beach)	'White Silt' bench on western valley side	415	Deposition by underflows and suspension in gLP basin

Place name	Sedimentologic observations[1]	Landform Setting[2]	Elevation (m asl)[3]	Interpretations
	(10) Subhorizontal to undulating beds of planar tabular pebble and cobble gravel interbedded with planar-laminated and planar-bedded fine sand and silt (25 m high, 200 m long exposure)	Upper portion of 'White Silt' bench	420	Subaqueous fan or proximal delta bottomsets deposited in gLP
	(11) Tabular cross-stratified gravel and sand (6 m high, 30 m long exposure)	Dissected 'White Silt' bench	475	Delta foreslope of 500 m asl delta surface deposited in gLP
Naramata Ck.	(12) Sand at surface, diamict overlain by interbedded silty sand and cobble/boulder gravel at depth (subsurface data: BC water well report 82844)	Flat and gently sloping surface, sediments in lee of bedrock step/depression.	650	Sediment fan remnant flanking Naramata Ck.
	(13) Horizontal and weakly cross-stratified coarse sand and cobble gravel with occasional boulders (4 m high, 12 m long exposure)	Sediment body filling bedrock channel in structural lineament, ~flat-topped surface dipping toward valley (~260°)	520	Sediment fan (possible gLP delta based on elevation)
Peachland Ck.	(14) Surface of coarse sand and cobble/boulder gravel (30 m high and 200 m long surface)	Surface and slope of sediment fan	390-500	Delta foreslope deposited in gLP
	(15) Planar-stratified sand and gravel, downvalley paleocurrents (110-170°, n=25) (3 m high, 12 m long exposure)	Dissected High bench on northern valley side	625	Subaerial or subaqueous deposition forming tributary valley train. Possibly related to subglacial lake or ice-marginal basin
	(16) Trough cross-stratified and planar-stratifed coarse sand alternating with undulating planar-stratified gravel, downvalley paleocurrents (85-165°) (5 m high, 75 m long exposure)	Dissected High bench on northern valley side	700	Subaerial or subaqueous deposition forming tributary valley train. Possibly related to subglacial lake or ice-marginal basin

Place name	Sedimentologic observations[1]	Landform Setting[2]	Elevation (m asl)[3]	Interpretations
Powers Ck.	(17) Tabular cross-stratified gravel overlain by planar-stratified gravel (~20 m long, 4 m high exposure)	Tributary mouth. Dissected (gLP) bench	520	Delta foresets and topsets graded to gLP
McDougall Ck.	(18) Tabular cross-stratified gravel and sand overlain by planar-stratified sand (4-5 m high, ~15 m long exposure)	Tributary mouth. Flat to gently valley-ward sloping surface filling around bedrock knobs within broad receiving basin	520	Delta foresets and topsets graded to gLP
Lebanon Ck.	(19) Planar-stratified sand, locally trough cross-stratified and diffusely graded (3 m high, 12 m long exposure)	Gently down-valley sloping and hummocky surface	475	Delta bottomsets or precursor subaqueous fan downflow of gLP delta graded to 505 m asl
Bellevue Ck.	(20) Planar-stratified coarse sand and gravel, diffusely-graded medium sand with occasional cobble gravel beds. (3-6 m high, 15 m long exposure)	Gently sloping fan surface	465	Delta bottomsets or subaqueous fan downflow of gLP delta graded to 520 m asl
Mt. Knox	(21) Cross-laminated fine sand and silt (mainly B-, S-type; occasional A-type). Paleocurrent direction: 190-225°. Measured exposure: 10 m high, 30 m long. Sediments exposed intermittently in bench over 90 m distance	Valley side sediment bench	385-392	gLP ('White Silt'-type) sediment bench at the foot of Mt. Knox
Glenmore subdivision	(22) Convoluted laminated silt and fine sand, loaded and deformed by pebble gravel. Deposited against smoothed and locally striated bedrock. (3 m high, 10 m long exposure)	Valley-ward sloping sediment bench onlapping bedrock of Ellison Ridge	447	Proximal glaciolacustrine sediments in gLP re-entrant filling depression in Ellison Ridge. Sediments likely derived from ice tongue in Kalamalka Lake valley and near Ellison Ridge.

Place name	Sedimentologic observations[1]	Landform Setting[2]	Elevation (m asl)[3]	Interpretations
Lambly Ck.	(23) Sub-angular to sub-rounded boulders at land surface	Gentle down-tributary sloping surface (toward Okanagan Valley)	525	Delta surface (based on morphology) graded to gLP
	(25) Alternating beds of planar-stratified sand and gravel. Discontinuous exposures up to 10 high, 2-10 m long, occuring over a ~300 m distance.	Valley floor forming High bench near headwaters	790	Dissected valley fill in headwaters
	(26) Planar-stratified and localy diffusely-graded fine-medium sand, interbedded with planar-laminated and cross-laminated silty sand and silt. Sub-angular to sub-rounded boulders on surface (4 m high, 7 m long exposure).	River cut through High bench valley fill near headwaters	920	Valley fill, possible lacustrine basin (subglacial or ice-marginal). Possible lag to produce boulders on surface
Fintry Ck.	(27) Sand and gravel filling depressions on delta top.	Depressions on surface of delta with steep front facing Okanagan Valley	525	Sediment meltout and/or inflow of sediments into kettle holes formed in delta surface graded to gLP highstand
	(28) Interbedded planar-stratified pebble gravel, massive to diffusely-graded medium sand. Rip-up clasts of sandy silt in diffusely-graded beds (12-15 m high, 40 m long exposure)	Lower slopes/flank of delta graded to 520 m asl	405	Deposition by turbidity currents on delta bottomsets or precursor subaqueous fan deposited in gLP
	(29) m-scale boulders, cobbles and coarse sand exposed at ground surface in 75 m long excavation within lateral portion of steep-fronted tributary fan	Delta foresets (based on morphology)	485	High energy (glacio)fluvial transport and deposition within delta graded to gLP highstan
Cougar Canyon	(30) 50 m thick package of interbedded cobble-boulder gravel and muds. Gravel beds 5-10 m thick. Mud beds up to 7 m thick (BC water well log # 38850)	Delta surface (based on morphology)	500	Deltaic deposition in gLP extension into Kalamalka Lake valley, fed by decaying ice in uplands

Place name	Sedimentologic observations[1]	Landform Setting[2]	Elevation (m asl)[3]	Interpretations
Bouleau Ck.	(31) Alternating beds of planar-stratified sand and planar-bedded gravel. (5 m high, 6-10 m long exposure)	Two gLP benches on valley side.	600, 582	Dissected valley train forming lower benches merging with gLP delta graded to 525 m asl
Equesis Ck./Naswhito Ck.	(32) Cobble and boulder gravel in coarse sand matrix exposed at surface of steep-fronted fan at tributary mouth (up to 20 m long and 3-20 m high exposure)	Steeply (20°) Okanagan Valleyward dipping surface at tributary mouth	435	Foreslope of gLP delta (based on morphology) graded to 525 m asl
	(33) Alternating planar-stratified sand and planar-stratified cobble to pebble gravel. Interbedding occurs occasionally, (5 m high ,~8 m long exposure)	Valley-side remnant	580	Dissected valley floor leading to gLP delta at 525 m asl
B.X. Ck.	(34) Tabular cross-stratified pebble-cobble gravel. Planar to concave erosional surface truncates gravel cross-beds. Alternating beds of planar-bedded coarse to medium sand overly erosional surface (4-7 m high, 15 m long exposure)	Lower bench incised into flat to gently sloping surface (elevation ranging 530-480 m asl) with hummocky topography and channel scars	490	Possible foresets (cross-beds) of delta graded to gLP, though delta morphology not easily discerned on surface, or subaqueous fan
	(35) Planar-bedded coarse sand and gravel interbedded with rhythmically planar laminated silt, fine sand and rare clayey silt (4 m high, 5 m long exposure)	Gently down-tributary sloping low bench below site (34)	440	Autocyclic surface incised in and/or aggraded on subaqueous fan or delta bottomset

[1] Bracketed number refers to sites located on Fig. 4.2. [2] See text for definitions of bench types. [3] Elevation at top of exposure.

4.5.2.2. Lake-bottom sediments

Lake-bottom sediments are common within Okanagan Valley and occur predominantly as elevated valley-side benches between Okanagan Falls and Squally Point (Fig. 4.2). Discontinuous exposures of lacustrine sediments occur along the Okanagan Lake shoreline North of Squally Point (e.g., near Powers Creek, Fig. 4.3). Lacustrine benches flank the northern arms of Okanagan Lake and merge into extensive plains forming the valley floor near Vernon and further North to Armstrong and possibly beyond (Nasmith, 1962; Vanderburgh and Roberts, 1996). Lacustrine benches and plains also occur around Kelowna where they form the upper 50-75 m of the valley fill (Paradis et al., 2009, 2010). Lacustrine benches provide no information on the elevation of the water plane during sedimentation. Therefore, the extents of lacustrine benches and plains provide a lower limit on lake extent.

4.5.2.3. Re-interpreted water plane indicators

Proper identification of water plane indicators is critical to paleogeographic reconstructions. Existing paleogeographic reconstructions of gLP rely on the identification of deltas to establish four discrete lake stages associated with episodic drainage of gLP (Fulton, 1969). Re-examination of some deltas used to infer paleo-water planes raises doubt about their origin and their significance as a paleo-water plane indicator at this location. One such feature occurs within the B.X. Creek drainage basin north of Vernon (Fig. 4.2) where the creek has built a sediment fan into Okanagan Valley. The work of Fulton (1969) builds on the ice retreat model of Nasmith (1962). Nasmith (1962) mapped alluvial fans, not a delta, at B.X. Creek and recognized that a number of inset alluvial fans had developed due to incision. The B.X. Creek 430 m asl 'delta' (site 35, Table 4.3) is reinterpreted as a largely erosional terrace resulting from incision into the B.X. Creek fan (an autocyclic surface, Muto and Steel, 2004). This reinterpretation is rationalized from the following arguments: 1) There is uncertainty about the origin of the surface described by Fulton (1969). This stems from a lack of sedimentologic information. Fulton (1969, 1975) identified a sediment surface along B.X. Creek which he interpreted as a delta, and used to establish the B.X. Stage of gLP at ~430 m asl (1400 ft asl). Fulton (1975) reported down-fan dipping beds of stratified sand and gravel grading to lacustrine silt within flat-topped triangular sediment fans developed at many tributary mouths. However, these descriptions are not from the B.X. Creek site. They are general statements about the sedimentology of similar landforms found throughout southern British Columbia and do not reflect detailed sedimentologic descriptions of sediments at B.X. Creek. 2) The purported delta has no corresponding surface in neighbouring tributaries. It is an isolated surface that cannot be regionally correlated. 3) Incision of the fan surface has produced numerous benches. These benches are common features of most fans in Okanagan Valley and are 'autocyclic surfaces' (Muto and Steele, 2004) (§ 4.7.4) implying that all flat-topped surfaces may not necessarily be deltas. Based on the reasons stated above, no conclusive evidence of water plane stabilization at the proposed elevation of the putative B.X. stage is identified. It is further suggested that many 'delta terraces' used by Fulton (1969, 1975) to infer waterplane elevations are in fact autocyclic surfaces (§ 4.7.4) and cannot be used to infer water plane stabilization.

4.6. Paleogeographic reconstruction of the gLP basin

4.6.1. Identification of stable water plane(s)

Valley side sediment benches in tributary valleys and at tributary mouths are used to reconstruct stable water planes in the gLP basin. Four groups of benches are recognized:

1. The most prominent, termed *gLP Benches*, are benches grading to perched deltas (§ 4.6.1) developed at tributary mouths between Okanagan Falls and Vernon. These are graded to an elevation envelope between 500-525 m asl (Figs. 4.2, 4.4). No deltas occur south of Okanagan Falls in the valley segment leading to McIntyre Bluff (Fig. 4.2). No latitudinal trends in delta elevation occur within Okanagan Valley (Fig. 4.2; Table 4.3).

2. Within tributary valleys, down-valley sloping benches occur discontinuously on the valley sides. Frequent interbedding of sand, gravel and laminated silt are noted in these surfaces up to elevations of 650 m asl (Table 4.3) (§ 4.7.2.2). In exceptional cases, some tributaries exhibit remnant sediment benches containing interbedded silt and sand at elevations reaching 800 m asl (Fig. 4.4). These surfaces appear

to grade to elevations above the gLP envelope (500-525 m asl) (Fig. 4.4) and may be associated with a water plane ponded at a higher elevation than the gLP envelope (§ 4.6.4). These surfaces are termed *High Benches*.

3. *Low Benches* are sediment surfaces occurring below the *gLP Benches* and graded to a level below the 500-525 m asl envelope, and above the elevation of Okanagan Lake (342 m asl).

4. *Recent Benches* are river terraces grading to fan deltas in modern Okanagan Lake (342 m asl).

Although many benches seem to result from incision into pre-existing sediments, some benches likely result in part from aggradation. The aggradational/erosional designation is not known with certainty for all benches because most do not exhibit accessible sediment exposures in the bench risers. The sedimentology of these benches is discussed in § 4.7.2. The bench classification presented at this stage is therefore based on geomorphology (rather than sedimentology) and relies on the absolute elevation of benches in relation to the regional gLP elevation envelope.

Within Okanagan Valley, *gLP Benches* grading to deltas at tributary mouths record regional water plane stability and delta aggradation/progradation in gLP at an elevation of 500-525 m asl. *High Benches* grading to a water plane above the gLP envelope imply the presence of a lacustrine basin as high as 800 m asl. This could be associated with ice-marginal lakes developed against a valley ice tongue. Alternatively, it could be associated with a subglacial lake in Okanagan Valley (Lesemann and Brennand, 2009; Chapter 3). A full examination of these possibilities is beyond the scope of this chapter and will not be examined further. Similarly, *Recent Benches* grade to Okanagan Lake and are unrelated to gLP. Therefore, discussion in this chapter will focus mainly on *gLP Benches* and *Low Benches.* *High Benches* are discussed when their elevation makes it unclear whether they graded to gLP or not (i.e., they may be *gLP Benches*).

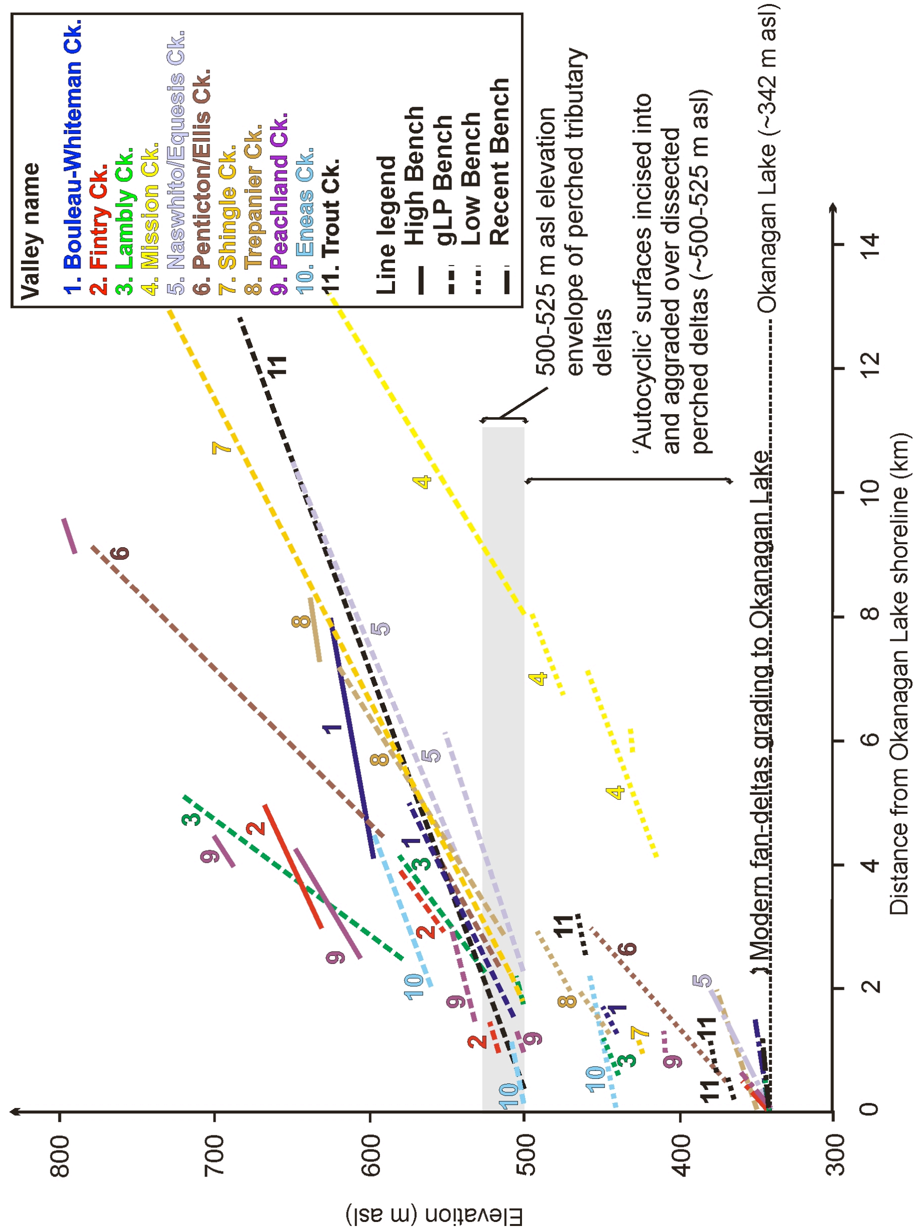


Figure 4.4. Generalized down-valley longitudinal profiles of the highest High Benches, gLP Benches, Low Benches, and Recent Benches in main tributary valleys of Okanagan Valley. Upper benches are graded to a water plane above the highstand of gLP. gLP benches are graded to perched deltas within an elevation envelope of 500-525 m asl representing the highest gLP water plane. Low benches are 'autocyclic' surfaces resulting from incision and local aggradation during a period of base level drop associated with drainage of gLP. Recent benches are terraced fan deltas graded to the elevation of Okanagan Lake. Details on the methods used to construct this figure are given in Appendix E.

4.6.2. *Glacioisostatic rebound reconstructions*

Nasmith (1962) reported the presence of two distinct valley-side 'beaches' (likely wave-cut benches) (~ 13 km apart) occurring on the hillsides above Kalamalka Lake. Based on correlation of these beaches, he proposed that a northerly glacioisostatic rebound gradient of ~ 0.6-0.7 m/km (3.5 ft/mile) was present in Okanagan Valley and persisted through development and drainage of gLP. However, limited descriptions of these "beaches" and uncertainty regarding their location prevented re-examination and surveying for this study. The origin of these landforms is therefore unclear. More significantly, features reported by Nasmith (1962) occur as single, not paired, 'beaches' and this raises the possibility that they are not contemporaneous features. Thus, it is unclear if these landforms can reliably be used for water plane tilt reconstructions.

Based on general understanding of post-glacial crustal response, crustal depression and glacioisostatic rebound is expected in Okanagan Valley. Studies of glacioisostatic rebound in the Merritt basin (Fulton and Walcott, 1975) and in Thompson Valley (Johnsen and Brennand, 2004) both report NW-directed rebound (i.e. SE surface tilt). If the glacioisostatic rebound parametres of Johnsen and Brennand (2004) are applied in Okanagan Valley, then ~ 95-125 m of rebound would be expected along the valley segment where perched deltas occur (following a northerly valley-parallel azimuth). These values are well in excess of the elevation range of deltas (25 m). Therefore, based on the basin-wide correlation of perched deltas associated with bedrock sills at Macintyre Bluff (§ 4.6.4), and the absence of significant (i.e. greater than 25 m) latitudinal trends in delta elevation, limited evidence of glacioisostatic rebound can be discerned in Okanagan Valley. Two factors may account for the apparent absence of glacioisostatic rebound in Okanagan Valley.

First, crustal and mantle conditions in British Columbia offer appropriate conditions for rapid glacioisostatic rebound. Geophysical surveys and coastal glacioisostatic rebound models suggest a warm, low viscosity mantle (Clague and James, 2002), and a relatively thin lithosphere (~33 km thick) (Clowes et al., 1995). These conditions favour rapid crustal response to loading and unloading and were invoked by Johnsen and Brennand (2004) to explain very high glacioisostatic tilts of shorelines from glacial Lake Thompson and glacial Lake Deadman in southern British

Columbia. These responsive lithosphere/mantle conditions in British Columbia could provide the appropriate conditions for rapid crustal response occurring prior to or within the time span needed for delta development in gLP. Mathews et al. (1970) postulated that, for the Canadian Cordillera, the Maxwell relaxation time (time for ~63% of rebound to occur) during deglaciation could occur within a few hundred years. Recent modeling of crustal rebound in coastal British Columbia has postulated low mantle viscosities on the order of 3-5 x 10^{18} Pa s (Clague and James 2002). In studies of glacioisostatically tilted shorelines from pluvial lakes of the Western United States, mantle viscosities similar to those calculated by Clague and James (2002) yielded Maxwell relaxation times on the order of 300 years (Adams et al., 1999), in accord with the early estimates of Mathews et al. (1970). Based on sedimentologic similarities between gLP sediments and glacial lake sediment in Thompson Valley (Fulton, 1965; Shaw, 1977) gLP may have only existed for a few decades or centuries at most (§ 4.6.2). Therefore it is conceivable that rebound took place over a short time period and that a large portion of this rebound took place prior to or within the timespan of delta development within the gLP basin.

Nonetheless, Maxwell relaxation time only account for ~63% of the potential glacioisostatic rebound. The remaining ~37% should, in theory, have a discernible rebound signature in the landscape. A second and complimentary possibility is that ice thickness was relatively thin and of uniform thickness along the axis of Okanagan Valley. Uniform ice thickness could result in nearly uniform rebound (i.e. without a detectable gradient). A similar argument was used by Sawicki and Smith (1992) to explain the apparent absence of glacioisostatic rebound in glacial Lake Invermere (Rocky Mountain Trench of British Columbia). Limited ice thickness over Okanagan Valley could reflect development of a subglacial lake (Lesemann and Brennand, 2009; Chapter 3). Based on the uncertainties associated with reconstructions of glacioisostatic rebound in Okanagan Valley, no corrections for glacioisostatic rebound are applied to water plane reconstructions in this chapter.

4.6.3. Lake areal extent and volume

During its stillstand (500-525 m asl), gLP covered a minimum area of ~1220-1350 km^2 (Fig. 4.5 A). GLP was dammed at McIntyre Bluff (Nasmith, 1962) (§ 4.6.4). Its northern extent is unclear. The valley fill north of Vernon contains lacustrine sediments

(Vanderburgh and Roberts, 1996) and this is treated as a minimum extent of gLP. It is possible that gLP extended further north and may have been contiguous with lakes in the Thompson Valley (Fulton, 1969), though this is unverified. Based on the elevation and extent of White Silt benches, mean lake depth varied between ~90 m to over 120 m, and lake volume was ~103 km^3 (over four times the 25 km^3 volume of Okanagan Lake). Calculated lake volume should be treated as a 'first order' estimate due to various sources of uncertainty associated with its calculation. One uncertainty stems from the assumption that a horizontal lake floor extended from the edge of the White Silt benches. The valley-ward extent of these benches is unknown and the likelihood of sediment compaction makes it difficult to assess the extent and volume of sediments that may have existed beyond the edge of the White Silt benches. A second source of uncertainty stems from the fact that the volume calculation assumes an absence of icewithin the lake basin. Based on the presence of eskers at 550-580m asl near Fintry (Paradis et al., 2010) and ice-contact deltas in Kalamalka Valley (500 m asl), it is possible that varying amounts of ice were present along the valley axis, particularly in the valley segment north of Fintry (Fig. 4.2, § 4.9). The timing and extent of valley ice cannot be readily assessed due to an absence of chronologic and landform data allowing for precise ice-marginal reconstructions. However, regional hydrologic connectivity recorded by the reconstructed water plane, and seismic data (Eyles et al., 1990, 1991; Vanderburgh and Roberts, 1996) suggest that extensive ice masses did not stagnate in the valley bottom (§ 4.9.2). The lake volume estimate is therefore based on an absence of ice within the lake basin. Lastly the estimate uses the land surface DEM (Geobase ®) to estimate lake bottom beyond the white silt benches. DEM uncertainty is ±10 m elevation at 90% confidence for interpolated points (British Columbia Government 1992). The modern DEM also leads to an overestimation of lake volume because it introduces post-lake drainage Holocene incision volumes beyond the white silt benches.

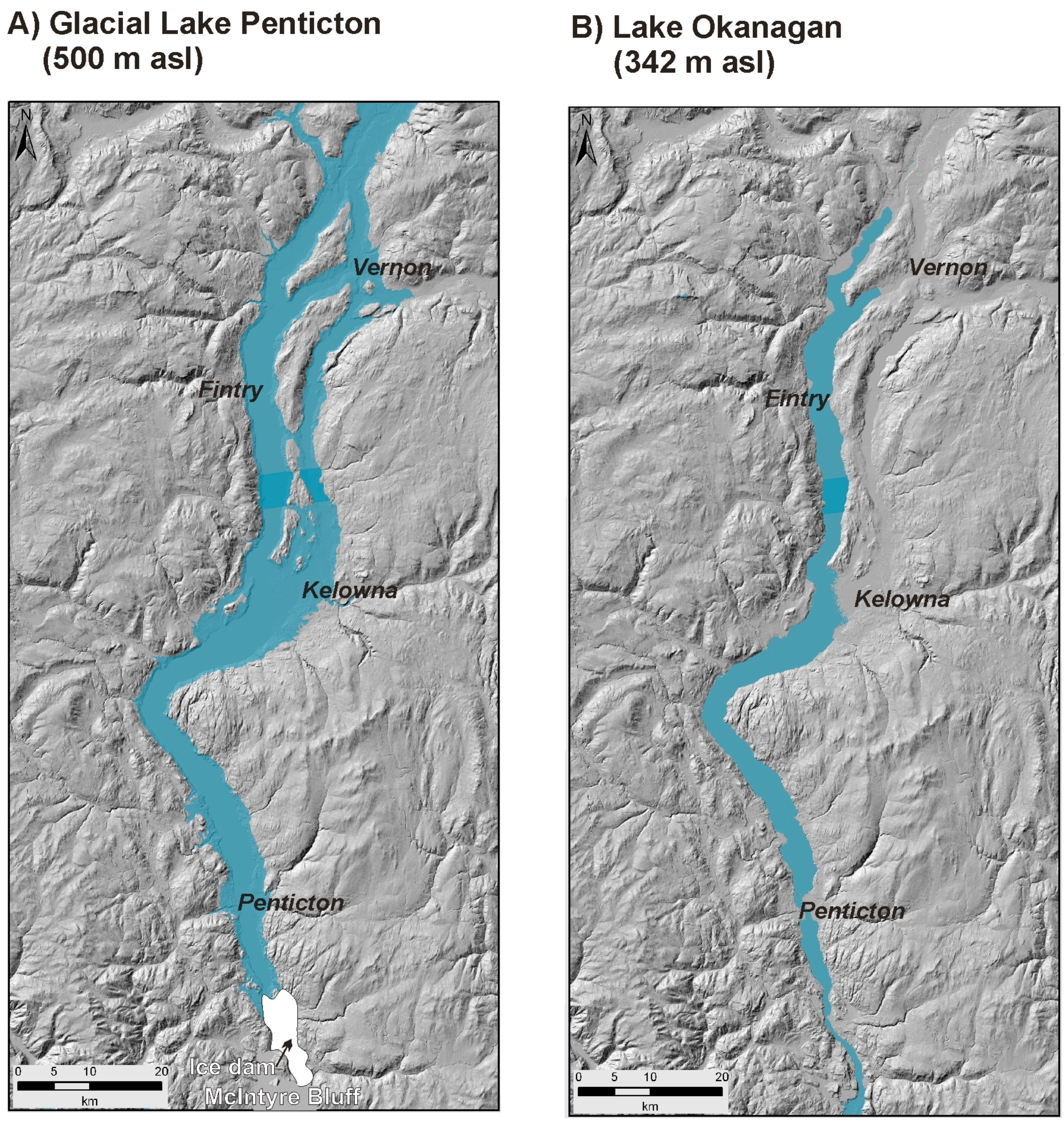


Figure 4.5: A) Extent of the glacial Lake Penticton highstand based on regional delta correlation assuming no isostatic tilt (see text for details). Ice within and around gLP basin not shown given uncertainty of location and timing of ice margins. B) Current extent of Okanagan Lake and modern valley floor. P: Penticton, K: Kelowna, V: Vernon. Data: BC Government TRIM I dataset, 25 m resolution (© GeoBC, by permission).

4.6.4. Lake dam characteristics

As proposed by Nasmith (1962) the gLP dam likely consisted of trapped ice blocks buttressed by sediments within a 5 km-long valley segment between Okanagan Falls and McIntyre Bluff (Figs. 4.2, 4.5, 4.6, 4.7). McIntyre Bluff marks the narrowest portion of the valley where two vertical rock faces are separated by only 1.5 km (Fig. 4.7 A). Bluff height reaches 680 m asl on the eastern valley side (Fig. 4.7 A).

Sediments near the valley floor within the valley constriction consist of perched benches of gravel and discontinuous beds of laminated silt and fine sand. These deposits have steep frontal slopes and irregular hummocky morphology characterized by numerous closed depressions. Modern tributary valleys have built alluvial fans atop these benches (Figs. 4.6 A, B). On the western valley side a meltwater channel along Marron River has built a prominent sediment bench near 454 m asl (Fig. 4.6 A). A group of undulating ridges and depressions with ~50-90 m spacing between crests occurs near the southern portion of this bench (Figs. 4.6 C, D)

Sediments in the perched benches near the valley floor are interpreted as erosional remnants of sediments buttressing stagnant ice in the valley constriction during damming of gLP. Hummocky topography and closed depressions are probably kettle holes and reflect the presence of buried ice within the sediments. Discontinuous laminated silts are interpreted as lacustrine sediments associated with small ponds within the dam site. Sediment benches between 425-460 m asl (including the height of overlying alluvial fans) occur 40-100 m below the elevation of the inferred highstand of gLP (500-525 m asl) indicating that they could not have directly contributed to lake damming.

One possibility is that ice blocks acted as the main damming agents and sediment benches only buttressed ice blocks. An additional and complimentary possibility is that these sediment benches have been partially eroded during lake drainage. Undulating ridges and troughs on the Marron River sediment bench are similar to forms interpreted as lake drainage bedforms in glacial lake drainageways elsewhere (Clague and Rampton, 1982; Johnsen and Brennand, 2004). Therefore, the Marron River sediment bench ridges are interpreted as bedforms associated with drainage of gLP and may be indicative of erosion of buttressing sediment benches. Alluvial fans overlying dissected benches developed following gLP drainage. The steep and sharp edge of the Orofino Creek fan in the lee of McIntyre Bluff and extension of the Atsi Klak spillway toward channels dissecting the Wolfcub Creek fan (Fig. 4.6 A) suggest that they may have been dissected during gLP drainage.

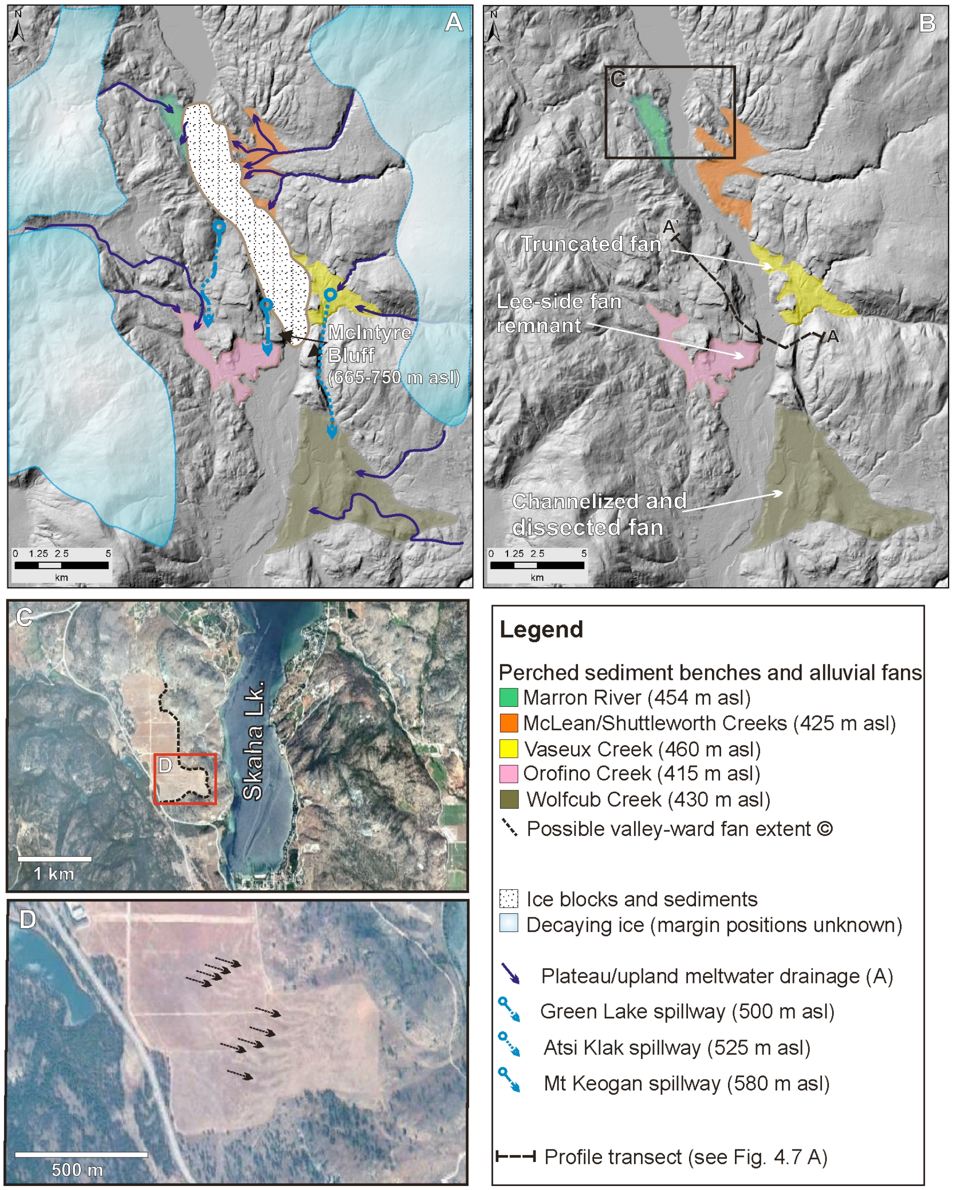


Figure 4.6. Damming and drainage evolution of glacial Lake Penticton. A) Ice blocks and sediments buttressed by sediments dam gLP between Okanagan Falls and McIntyre Bluff. Three spillways (Atsi Klak, Green Lake and Mt. Keogan) lead to increasing water plane stabilization as lake level drops toward 500 m asl. Decaying ice weakens the dam and leads to gLP drainage. B) Modern distribution of valley floor sediments, perched sediment benches resulting from erosion, and tributary valley alluvial fans. Data: BC Government TRIM I dataset, 25 m resolution (© GeoBC, by permission).C) Marron River perched sediment bench (edge indicated by dashed line) (Image: Google Earth). D) Close-up view of possible bedforms produced during lake drainage (crests marked by arrows) (Image: © Google Earth).

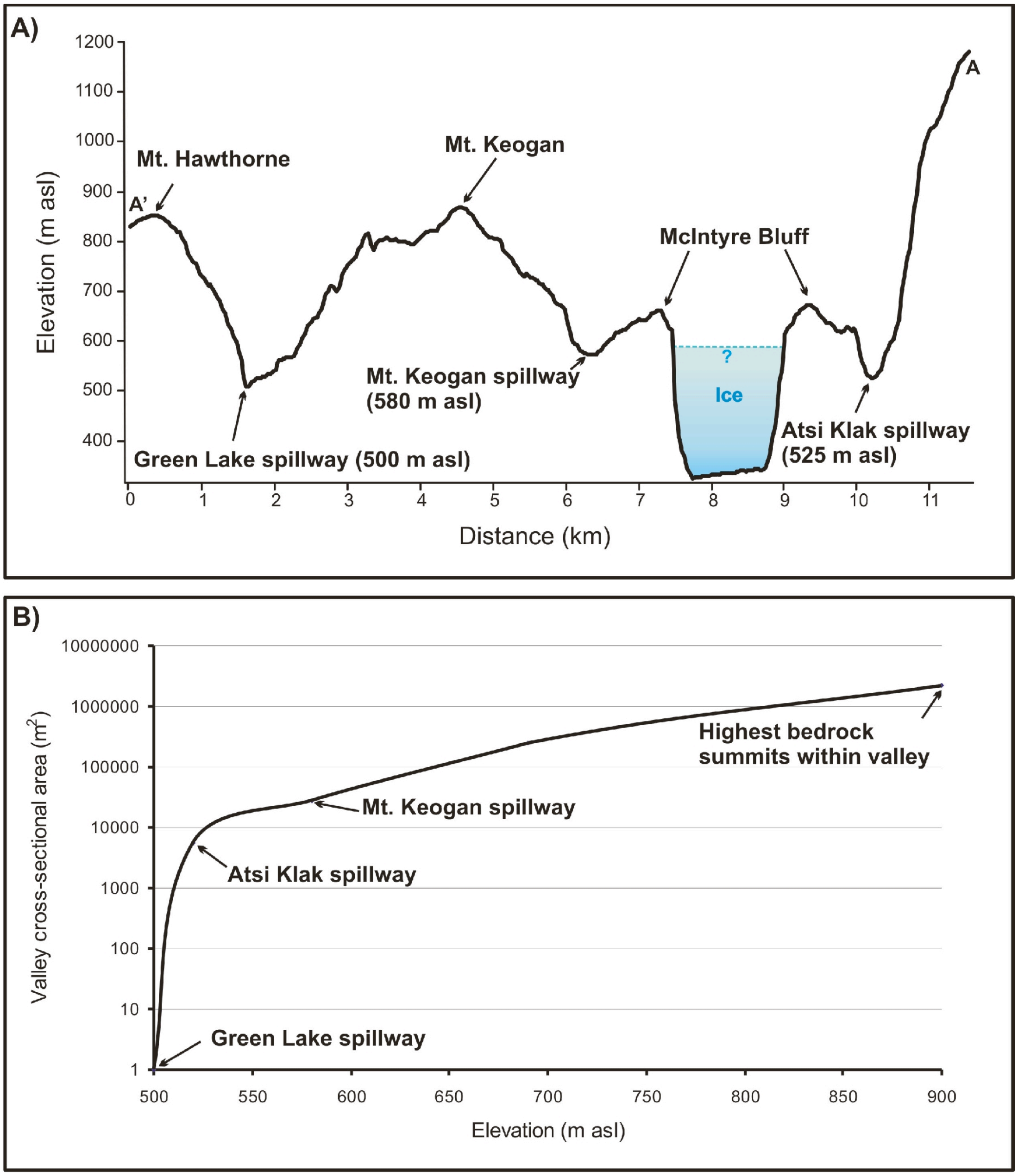


Figure 4.7. ***A) Topographic profile along transect crossing McIntyre Bluff and surrounding valley sides (refer to Figure 4-6 B for transect A-A' location). Three spillways provide outlets for glacial Lake Penticton toward a stable water plane at ~500 m asl. Note: representation of ice dam thickness in valley axis (cf. Nasmith (1962)) is schematic. Exact thickness is unknown but assumed to reach at least 500 m asl. B) Semi-log plot of changes in valley cross-sectional area with elevation. A dropping water plane leads to rapid lake confinement between valley walls and water plane stabilization as cross-sectional area decreases toward the lowest spillway elevation.***

4.6.5. Controls on lake level

Three bedrock channels dissect the valley near McIntyre Bluff and offer possible drainageways for gLP. Two prominent channels (Mt Keogan and Atsi Klak channels with floor elevations of 580 and 525 m asl respectively, (Figs 4.6, 4.7) flank the bedrock bluff along the valley axis. A third channel traverses the bedrock walls to the west of McIntyre Bluff (Green Lake channel, 500 m asl) (Figs 4.6, 4.7). These three bedrock channels operated sequentially and controlled lake level (Figs. 4.6, 4.7) as long as ice blocks and sediments blocked drainage along the valley axis. As lake level dropped, the gLP water plane became increasingly confined within Okanagan Valley. Valley cross-sectional area decreases rapidly as elevation decreases (Fig. 4.7 B). This suggests that water plane stabilization increased as the water plane approached elevations between 500-525 m asl and drainageways became progressively abandoned as lake level dropped. Regional development of perched deltas within this elevation envelope indicates water plane stability. At this time, any potential lake level rise due to meltwater input to the gLP basin could be evacuated through the Green Lake spillway.

4.6.6. gLP drainage and lake evolution

Reconstruction of a single stillstand for gLP (500-525 m asl) suggests that gLP drained during a single event without intervening water plane stabilization. An unconformity at ~215 m asl truncating lacustrine sediments below (§ 4.8.4) may have resulted from erosion during drainage, suggesting that the total lake volume (103 km^3) likely drained. This requires steady or catastrophic drainage through a passage without bedrock sills that could have lead to temporary water plane stabilization. Glacial and non-glacial lake drainage may occur through various processes including dam overspill (O'Connor, 1993), dam flotation (Walder and Costa, 1996), or tunnel enlargement (Tweed and Russell, 1999). It is difficult to determine exact dam failure mechanisms for gLP. However, dam floatation can probably be ruled out given the likelihood that sediments were mixed with, and possibly buttressed ice blocks within the dam site (Fig. 4.6 A). Dam overspill or hydrofracture could initiate dam failure. Drainageway enlargement by erosion would likely accompany dam failure. The bedrock walls of

McIntyre Bluff contain crescentic erosional marks (sichelwannen, Kor et al., 1991) formed by turbulent water flow, and add support to inferences of high velocity flow through McIntyre Bluff.

A single gLP drainage conflicts with Fulton's (1969) lake level reconstructions that suggest episodic drainage with intervening stability of the gLP water plane. Fulton (1969) proposed four lake stages (Table 4.1) occurring between 1700 and 1160 ft asl (520-360 m asl). These stages were postulated using 'delta terraces' (Fulton, 1969) at tributary mouths near Vernon. Although Fulton (1969) states that some lake stages are constrained by landforms at many tributary mouths, the majority of lake stages are in fact constrained by a single 'delta terrace' and no formal regional correlation of deltas has been performed. Furthermore, recognition of 'delta terraces' was based mainly on morphology with very limited sedimentologic support. For these reasons, evidence for the proposed lake stages is critically re-examined here. The two highest lake stages proposed by Fulton (1969) are the *Long Lake* (1600-1700 ft asl – 500-520 m asl), and *Grandview Flats* (1600 ft asl - 500 m asl) stages. Given the elevation overlap between these two lake stages, it is questionable whether postulating two distinct lake stages is in fact warranted. Identification of two lake stages reflects Fulton's (1969) paleogeographic reconstruction, with the Long Lake stage developing initially in the northern extent of Kalamalka Valley and later expanding into Okanagan Valley to form the Grandview Flats stage as ice retreated from Kalamalka Valley. It is notable that no field evidence was used to support the pattern of ice retreat (and its relative timing). Instead, ice retreat patterns are inferred by correlating glaciolacustrine deltas, which lends a degree of circularity to arguments used to infer lake stages from distinct deltas. Based on the fact that elevations of both lake stages correspond to the 500-525 m asl envelope recognized throughout the gLP basin (Fig. 4.4), it is suggested that the Long Lake and Grandview Flats stages in fact record a single metastable water plane marking the standstill of gLP. Although varying amounts of ice may have been present in Kalamalka Valley and Okanagan Valley (Paradis et al., 2010), regional delta correlations imply hydrologic connectivity between Kalamalka Valley and Okanagan Valley, indicating that local ice masses exerted no significant control on the extent and elevation of this water plane. Atsi Klak and Green Lake spillways controlled the elevation of this water plane (Figs 4.6, 4.7).

Identification of the third postulated lake stage, the *B.X.* stage (1400 ft asl -- ~438 m asl), relies on a single terrace surface north of Vernon interpreted as a delta (Fulton, 1969, 1975). It is shown (§ 4.5.2.3) that this surface is likely an erosional bench. This raises doubts about the existence of a lake stage at this elevation. Fulton (1969) argued that the B.X. stage was the best constrained gLP stage based on the presence of delta terraces and the occurrence of shorelines (based on the work of Nasmith, 1962). However, Nasmith's (1962) interpretation of shorelines is problematic (§ 4.5.2.3) and closer re-examination and topographic surveying of surfaces below 500 m asl reveal no conclusive evidence of regional delta development at the elevation of the postulated B.X. stage (~438 m asl, Fig. 4.4). The last proposed lake stage (*O'Keefe* 1160 ft. asl – 360 m asl) relies on differential glacioisostatic rebound (cf. Nasmith, 1962) of the northern portion of the gLP basin (relative to its outlet in the south). However, the reliability of Nasmith's (1962) glacioisostatic reconstruction has been questioned based on the field evidence used to infer rebound (see § 4.5.2.3). No deltas record this putative water plane.

Based upon re-examination of the evidence for proposed multiple lake stages (Fulton, 1969), the field evidence only supports the existence of a single regional water plane within an elevation envelope of 500-525 m asl, equivalent to the Long Lake and Grandview Flats stages of Fulton (1969) identified locally around Vernon. Little conclusive evidence supports the notion of punctuated drainage required to form the last two lake stages of Fulton (1969). These conclusions are consistent with the proposed origin of the White Silt benches and the drainage dynamics inferred within the rest of the gLP basin to the south (§ 4.8.4).

Development of autocyclic surfaces (Low Benches below 520-525 m asl, Fig. 4.4) (§ 4.7.4) within tributaries suggests fluvial adjustment of tributary streams following a single drainage event. The depth of incision and sediment scouring during drainage is unclear. However, a submerged lacustrine terrace ~25 m below the level of Okanagan Lake (St John, 1973) suggests erosion below modern lake level. Furthermore, the presence of an unconformity in the lacustrine valley fill near ~ 215 m asl may mark a lower limit of erosion in some portions of the basin. Following gLP drainage, tributary valleys delivered sediment and built alluvial fans and fan-deltas that eventually dammed the valley to form the modern basins of Okanagan Lake and Skaha Lake (e.g., coalescence of alluvial fans from Penticton, Ellis and Shingle Creek, Fig. 4.2). Therefore, a period of lake level increase occurred after gLP drainage (though the timing is unknown) and culminated in the stability of the water plane defining Okanagan Lake (~342 m asl).

4.7. gLP paleoenvironment

4.7.1. Lake bottom sediments and paleocurrent measurements

Lake bottom benches slope toward the valley axis at 6-8° (but locally as much as 15°). The steepest slopes occur near the valley walls where sediments onlap the steeply-dipping bedrock surface. As a result, relief can vary as much as 50 m along a transect between the valley wall and the bench edge. An elevation of 400-410 m asl, taken at the bench edge, is commonly used as a general value for bench elevation south of Squally Point (Flint, 1935b; Nasmith, 1962; Shaw, 1977) (Fig. 4.2). This elevation is also consistent with benches in the northern arms of Okanagan Lake and the valley floor around Vernon (Fig. 4.2). Benches near Kelowna occur below this elevation (375-390 m asl) (Fig. 4.2).

An overall upward-thinning and fining pattern characterizes sediments in the White Silt benches (Shaw, 1977; Shaw and Archer, 1978). This is generally attributed to decreasing meltwater and sediment supply (Fulton, 1965; Shaw and Archer, 1978, 1979). Basal beds in the White Silt benches are m-thick and frequently contain convolutions resulting from syndepositional slope failures (Shaw, 1977; Shaw and Archer, 1978, 1979). Rhythmic depositional patterns in lacustrine benches led Fulton to suggest that many of these beds were varves. Based on varve counts, Fulton (1965) estimated that glacial lakes in southern British Columbia were short-lived (~200 yrs). However, Shaw and Archer (1978) showed that episodic coarse-grained sediment inputs were present within the White Silt bench sediments; these complicate recognition of varves. Consequently, the duration of sedimentation recorded in the White Silt benches remains unclear. Based on the thickness of beds and lithofacies characteristics within the White Silt benches, it is generally agreed that abundant sediment volumes were delivered rapidly to the gLP basin (Shaw, 1977; Shaw and Archer, 1978, 1979; Eyles et al., 1990, 1991).

Glaciolacustrine lithofacies south of Squally Point have been discussed in detail by Shaw (1977) and Shaw and Archer (1978, 1979). Lithofacies range in grain size from clay to cobble gravel. However, silt and fine sand dominate lithofacies within the White Silt benches. Lithofacies record a range of depositional processes and depositional energy. Fine sand lithofacies include type A, B, and S ripples (Ashley, 1975; Ashley et al., 1982) resulting from deposition by underflows (Kent, 1978; Shaw, 1977; Shaw and Archer, 1978). Some coarse sand lithofacies (‘winter sands’, Shaw and Archer, 1978) were associated with episodic slope failures on deltas. Laminated silt records mainly suspension deposition under decreasing energy conditions. Clayey and silty clay lithofacies also record quiescent conditions. However, these lithofacies are rare and only occur discontinuously in the upper 4 m of the White Silt benches. Coarser gravely diamictic lithofacies are rare and occur near tributary valley mouths (Flint, 1935b; Shaw and Archer, 1979). These typically consist of discontinuous beds (m-scale in lateral extent and cm-scale in thickness) and lenses. They have been interpreted as flow tills (Shaw, 1977) but may also result from subaqueous debris flows (Eyles et al., 1991; Johnsen and Brennand, 2006).

Paleocurrent measurements provide critical information regarding sediment sources for gLP. Both tributary and valley axis sediment sources have been proposed (Flint, 1935b; Fulton, 1969; Shaw, 1977; Table 4.1). Paleocurrent data allow for testing of these hypotheses. Measurements from the White Silt benches between Squally Point and Okanagan Falls reveal complex paleocurrent distributions (Fig. 4.8). As was noted in previous studies (Kent, 1978; Shaw, 1977) distributions are frequently bimodal to multimodal, often exhibiting both downvalley (southward) and valley-ward (~eastward/westward) paleocurrents (Fig. 4.8). To date, paleocurrent data were largely considered within the context of a valley-axis ice tongue (Shaw, 1977) with relatively limited consideration for the location of measurement sites in relation to tributary valleys. For example, Kent (1978) and Shaw (1977) documented paleocurrents from locations that are consistently downflow and/or downvalley of major tributaries, yet these locations were not considered in their interpretations of paleoflows. Here, paleocurrent data are re-examined, paying closer attention to their geomorphic context.

4.7.1.1. Paleocurrents from a single, valley-axis source

Valley-axis parallel paleocurrents likely support notions of a dominant valley-axis sediment source (e.g. Fulton, 1969; Shaw, 1977). However, the location of measurement sites on the valley sides raises questions about the validity of these sites for testing the location of sediment sources in gLP. The biggest obstacle to evaluating the possibility of a single valley-axis sediment source is the uncertainty regarding the valley-ward extent of White Silt benches. If White Silt benches were deposited against a valley ice-tongue, then they likely contain a fairly complete record of sedimentation in an 'ice lateral' lake (assuming limited erosion of the top of the sequence). However, most proposed typologies for gLP suggest a proglacial and/or supraglacial lake setting (Table 4.1, § 4.3.2). Therefore, based on these lake types, it is reasonable to expect that lake sediments extended beyond the modern edge of the White Silt benches.

These lake types further imply that some volume of lake sediments was either displaced by supraglacial let-down, or is missing due to erosion of proglacial sediments. Consequently, a portion of the gLP fill is likely missing/displaced and it is unclear whether lake sediments were continuous across the valley or if they formed discontinuous surfaces sloping toward the valley axis. Therefore, White Silt benches preserve mostly sediments from the gLP basin margins and this may potentially limit reconstructions from paleocurrent data. A critical question stemming from the paleocurrent data is how a single valley-axis sediment source could produce downvalley paleocurrents along the gLP basin margins, given the propensity of turbidity currents to follow topography within a depositional basin (Allen, 1970). It is conceivable that, with sufficientl head (such as that associated with glacier hydraulic head), some turbidity currents could overcome slight rises in the lake bottom and deposit sediments near the valley sides. Repeated deposition could form small turbidite channels and modify the gLP lake floor paleoslope, further focusing flows near the valley sides. The gLP paleoslope is difficult to assess from the surface of the White Silt benches given the possibility of erosion, and syn- and post-depositional sediment compaction (Shaw, 1977). It is reasonable to assume that paleoslopes probably dipped toward the valley axis based on the fact that lacustrine sediments directly onlap bedrock dipping steeply toward the valley axis. Convolutions within syndepositionally-deformed beds dipping

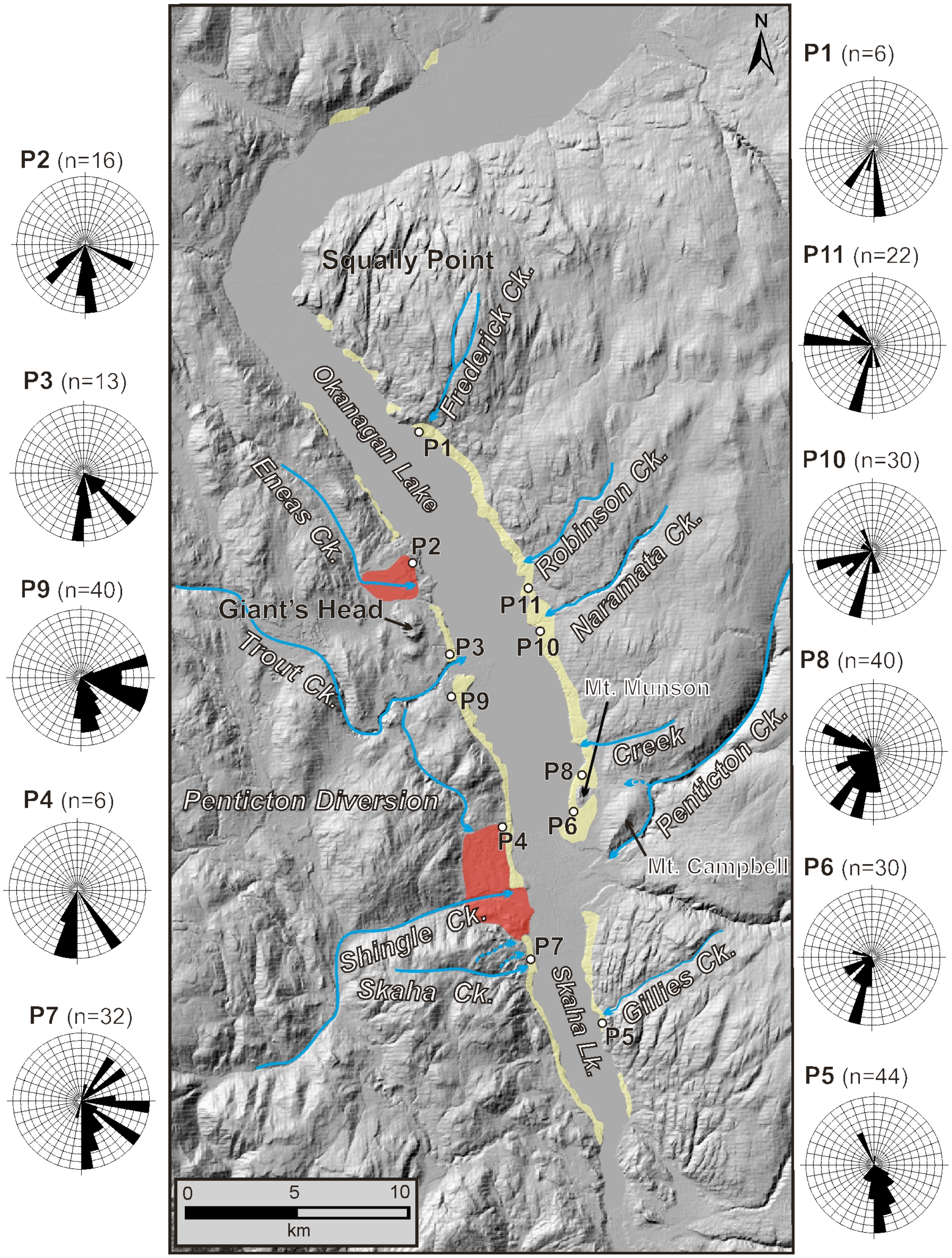


Figure 4.8. Paleocurrent measurements from ripples in rhythmically-bedded sediments in the White Silt benches (distribution mapped in yellow). Prominent perched deltas mapped in orange. Sites P2, P4 from Kent (1978). Site P1, P3, P5 from Shaw (1977). Sector divisions: 10°. Sites P1-P5 converted from distribution percentage and redrawn. also suggest bed failures on surfaces dipping toward or oblique to the valley axis (cf. Shaw, 1977). Data used to produce this figure are presented in Appendix F.

toward the valley axis further supports the inference of a lake paleoslope dipping toward the axis of Okanagan Valley.

Sites P1, P2, P3, P4, and P5 exhibit some of the strongest downvalley paleocurrents (Fig. 4.8) and possibly record the presence of remnant ice in Okanagan Valley. The least equivocal evidence for a single meltwater source along Okanagan Valley axis may be at site P1. This site occurs in proximity to Frederick Creek which, at first glance, could have been a possible sediment delivery point. However, paleocurrent measurements measured up-valley of the tributary junction record dominantly downvalley axis-parallel flows. This pattern is difficult to reconcile with a tributary sediment source and instead suggests either an Okanagan Valley-axis source or, alternatively, decaying ice on the valley sides near Squally Point and southward meltwater drainage along prominent bedrock channels (Fig. 4.8).

Sites P2, P3, and P4 possibly record dual sediment sources. For example, near site P2, Eneas Creek built a delta within the regional 500-525 m asl envelope (Fig. 4.4). Tributary sediment inputs from Eneas Creek and shifting distributaries on the Eneas Creek delta could account for southward and southeastward paleocurrents at this site. However, southwestward paleocurrents are difficult to rationalize with this explanation. These paleocurrents likely require an additional sediment input point north of the Eneas Creek delta. This could consist of decaying ice on or near the bedrock ridge separating Eneas Creek valley from Okanagan Valley, consistent with the evidence of remnant ice within the Prairie Valley area and the Trout Creek valley train (§ 4.7.2.7), Alternatively, paleocurrents could result from sedimentation (or sediment redistribution) by valley axis parallel flows, sourced from Okanagan Valley-axis ice to the north, or from flows originating at Squally Point and entering tributary embayments near Eneas Creek. Paleocurrents at site P3 exhibit dominant south and southeastward directions. Site P3 is located close to the confluence of Trout Creek and Okanagan Valley. However, in contrast to Eneas Creek, Trout Creek did not build a delta within the regional elevation envelope due to the likely presence of remnant ice which diverted meltwater though the Penticton Diversion (Nasmith, 1962; § 4.7.27). Therefore, paleocurrents at site P3 likely result from sediment delivery from a source near Giant’s Head (ice blocks) and/or a decaying ice mass in Okanagan Valley.

Valley-axis parallel paleocurrents at Site P5 could suggest an Okanagan Valley meltwater source. However, examining the local topography at this site suggests that a tributary sediment source is also possible. At site P5, southeastward paleocurrents strongly suggest Okanagan Valley-parallel flows. Shaw (1977) interpreted steeply-dipping gravel beds as delta foresets sourced from ice in Okanagan Valley. However, the measurement site is located downvalley of the creek input, in a bedrock embayment at the mouth of a prominent bedrock channel extending from Gillies Creek. The orientation of this channel is nearly parallel to Okanagan Valley. Shifting of sediment input points and topographic funnelling of turbidity currents could account for the southward paleocurrents

Two additional possibilities, applicable to all sites discussed above, could account for downvalley paleoflows along the basin margins originating from a valley-axis source. First, turbidity currents, originating from melting valley ice, could have been generated within ‘ribbon lakes’ developed between the lateral margins of an ice tongue and the valley walls (cf. Nasmith, 1962; Table 4.1). This possibility needs to be contextualized within a discussion of lake type and will be explored further in § 4.8. The second possibility is that ice-contact deltas developed between the bedrock valley walls and a valley-axis ice tongue in the valley. These deltas would build parallel to, and possibly toward, the valley walls. Although these types of ice-contact deltas developed in Kalamalka Valley (Paradis et al., 2009), this option is largely discounted here because existing deltas are most clearly linked to tributary valleys.

4.7.1.2. Variable paleocurrents from tributary valley sources

If tributary sediment meltwater sources are favoured over a valley-axis sediment source, they must be capable of accounting for the paleocurrent distributions (Fig. 4.8). Paleocurrents reflect the initial input angle as a stream delivers sediments. Some variability may be expected as a result of slight shifts in the sediment input point and/or routing of turbidity currents on the lake floor. Sites P6 and P8 occur near the edge of an extensive lake bottom bench (Fig. 4.8) and are both distal to a perched subaqueous fan north of Mt. Campbell (§ 4.7.2.5). Site P8 is downvalley of an unnamed creek entering Okanagan Valley. Southwestward paleocurrents at site P6 likely reflect topographic funnelling of turbidity currents between Mt. Campbell and Mt. Munson. Paleocurrents at site P8 have a stronger westerly component than those at site P6 and this may reflect preferential routing of turbidity currents north of Mt. Munson due to shifts in the sediment input points or variations in lake-bottom topography.

Sites P7, P10, and P11 reflect sediment inputs from tributaries with near orthogonal junction angles with Okanagan Valley, which may have allowed for a greater range of sediment dispersal within the basin. Paleocurrents at site P7 likely reflect sediment inputs from meltwater flow along the main channel of Skaha Creek or along meltwater channels adjacent to Skaha Creek (dashed lines, Fig. 4.8). Southeasterly paleocurrents at this site either result from shifting meltwater sources (i.e. from meltwater channels or Skaha Creek) or they could record sedimentation distal to the extensive fan built by Shingle Creek. Paleocurrents at site P10 are dominantly oriented toward the valley axis as a result of turbidity currents presumably following the basin paleoslope. Paleocurrents at site P11 are bimodal and may reflect two sediment input points. Westerly and northwesterly paleocurrents might be associated with northward shifting flows from Naramata Creek. More southwesterly paleocurrents could reflect inputs from Robinson Creek. Kent (1978) documented an exposure ~500 m north of site P11 and immediately south of Robinson Creek. He noted a dominance of sandy sediments and an absence of cross-laminated silts, and consequently reported no paleocurrent data from this exposure. These coarser sandy sediments might be proximal deposits to those documented at site P11, accounting for the southwesterly paleocurrents at this site. Lastly, paleocurrents at site P9 are dominantly southeastward. Though this site is close to the modern course of Trout Creek, ice blockage near the Penticton Diversion precluded sediment delivery via Trout Creek (§ 4.7.2.7). Consequently, turbidity currents at site P9 could be associated with decaying ice blocking the course of Trout Creek.

Therefore, despite bimodal to multi-modal paleocurrent distributions that include significant down-valley components, the regional pattern of paleocurrents appears to record both sediment delivery from tributary valleys and sediment delivery or redistribution by valley axis-parallel flows. Some downvalley paleocurrents used to support the notion of a valley-axis delta (Shaw 1977) may reflect routing of tributary-sourced material along bedrock channels, and along the lake floor paleoslope, though this latter option is equivocal. In other cases (e.g. P2, P3, P5) paleocurrents may partly record the presence of Okanagan Valley-axis ice or local ice blocks. The likelihood of combined valley-side and localized Okanagan Valley-axis ice sources is further illustrated by the difficulty of bed correlations in the White Silt benches (Shaw, 1977; Kent, 1978). In these studies, cross-valley bed correlations were unsuccessful, even for some of the thickest beds. This result is consistent with multiple sediment input points (tributaries) and possibly localized Okanagan Valley-axis sediment sources, rather than large-scale depositional events capable of producing cross-valley units as implied by the correlation attemps (Kent, 1978; Shaw, 1977) The presence of extensive Okanagan Valley-axis ice is discussed and largely discounted in § 4.9.2 where this topic is further addressed.

4.7.2. Tributary valley geomorphology

Tributary valleys consist of narrow and steep-sided bedrock gorges, commonly eroded across, or oblique to, the bedrock structural grain. Drainage basin pattern generally consists of low order arborescent valleys near the plateau edge, leading to a deeply incised bedrock gorge forming a single channel traversing the Okanagan Valley sides (Figs 4.2, 4.3 A, B, C). Waterfalls frequently occur within the bedrock gorges. The slope of tributary valleys decreases toward Okanagan Lake where the gorge widens and may sometimes coalesce with a neighbouring tributary valley (e.g., Penticton and Ellis Creeks, Fig. 4.2).

Within Okanagan Valley, sediments mainly occurs within eroded bedrock channels. Greater volumes of sediment occur within tributary valleys. These sediments form dissected valley trains exhibiting discontinuous valley-side benches (Figs 4.3 A, B, C). These benches onlap the bedrock valley walls and form discontinuous treads that can be traced and sometimes correlated down-valley. In other cases, discontinuous surfaces cannot be traced or correlated with certainty. In some tributary valleys, benches occur as paired treads across the valley (i.e. treads occur at the same elevation). Their down-valley extent is typically discontinuous, though some treads can extend for as much as 1-10 km length, though most extend for 100s m. Some down-valley treads can be correlated based on their slope and, especially, their position with respect to neighbouring benches. The lateral extent of benches varies from a few tens of metres to several hundred metres, depending on the local tributary geometry. Along a tributary valley, bench treads generally have a flat-topped morphology with occasional channels.

Areas of hummocky topography and clustered closed depressions occur locally. As tributaries widen down valley, some bench treads grade to sediment fans or deltas near the tributary valley mouth (Fig. 4.3F). The location (or elevation) of this transition varies between tributaries. Section 4.7.2.1 examines the sedimentology of tributary valley fills and sediment benches.

4.7.2.1. Sedimentation in a deep and narrow receiving basin: a case study of Penticton Creek valley fill

The sedimentology of valley-side benches along Penticton Creek is examined, starting near the plateau edge and progressing toward the lowest fan sediments near the modern valley floor (Figs. 4.9; 4.10). Overall, gravel, and especially sand, dominate the sediments of the valley fill (Fig. 4.9; Table 4.2). Laminated and massive silt and fine sand become more abundant in lower tributary reaches near the floor of Okanagan Valley and near the elevation of the White Silt benches (~430 m asl).

4.7.2.2. Sedimentology of *High Benches* within the confined bedrock valley: proximal subaqueous fan sediments

4.7.2.2.1. Observations

Over 15 m of sediments are exposed within a *High Bench* (~800 m asl, Site PI, Fig. 4.9 A, Fig. 4.10) approximately 2 m below the bench tread. Sediments at the base of the exposure consist of weakly stratified sandy-silt diamicton (Fig. 4.11 A) containing discontinuous silty laminations and lenses of interlaminated silt and fine sand. The diamicton contains ~30-50% clast content (by area) (pebble gravel and coarser) and most gravel clasts are rounded to sub-rounded and occur as part of clusters or asisolated clasts deforming silty laminations. Striae are present on some pebbles. The lower contact of this unit was not visible. The upper 30 cm of this unit consist of planar stratified medium sand interbedded with silt and diamicton. A gravel bed, with a sharp lower contact, overlies the diamicton. The uppermost 12 m of the exposure consist of planar-stratified and heterogeneous gravel interbedded with planar-stratified and cross-stratified fine and medium sand. Silt is interlaminated with some cross-laminated sand (type A and type S cross-laminations, Table 4.2). Paleocurrent measurements from imbricate gravel clasts and cross-laminations are consistently down valley (200-300°)

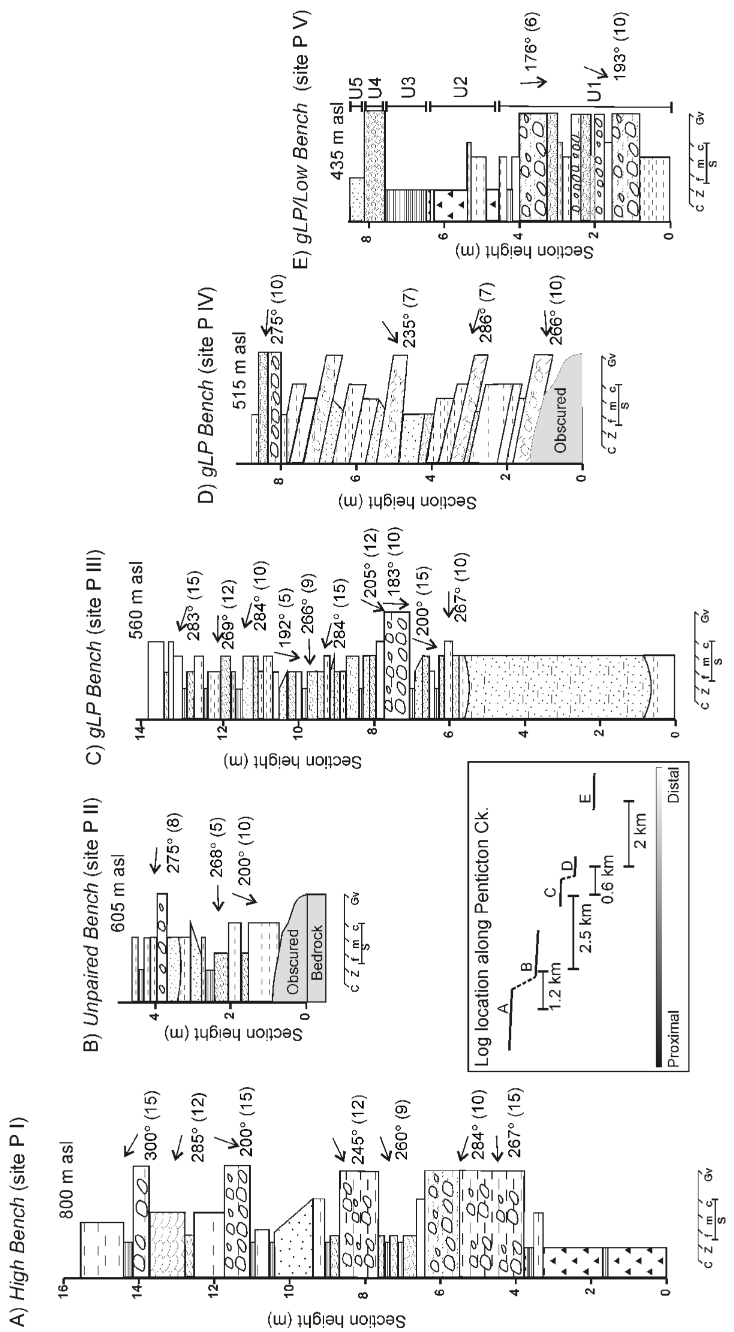


Figure 4.9. Sediment logs illustrating down-valley facies changes within Penticton Ck. valley fill. Valley fill consists mainly of sand and gravel. A, B) Diamicton, gravel and sand dominate the High Benches. C) Sand-dominated gLP Bench interpreted as a subaqueous fan. D) Dominance of coarse sand and gravel in gLP Bench interpreted as a delta. E) Unpaired bench at tributary mouth recording subaqueous deposition and development of subaerial braided river. Refer to Table 4.2 for lithofacies legend. Refer to Figure 4.10 A for site locations.

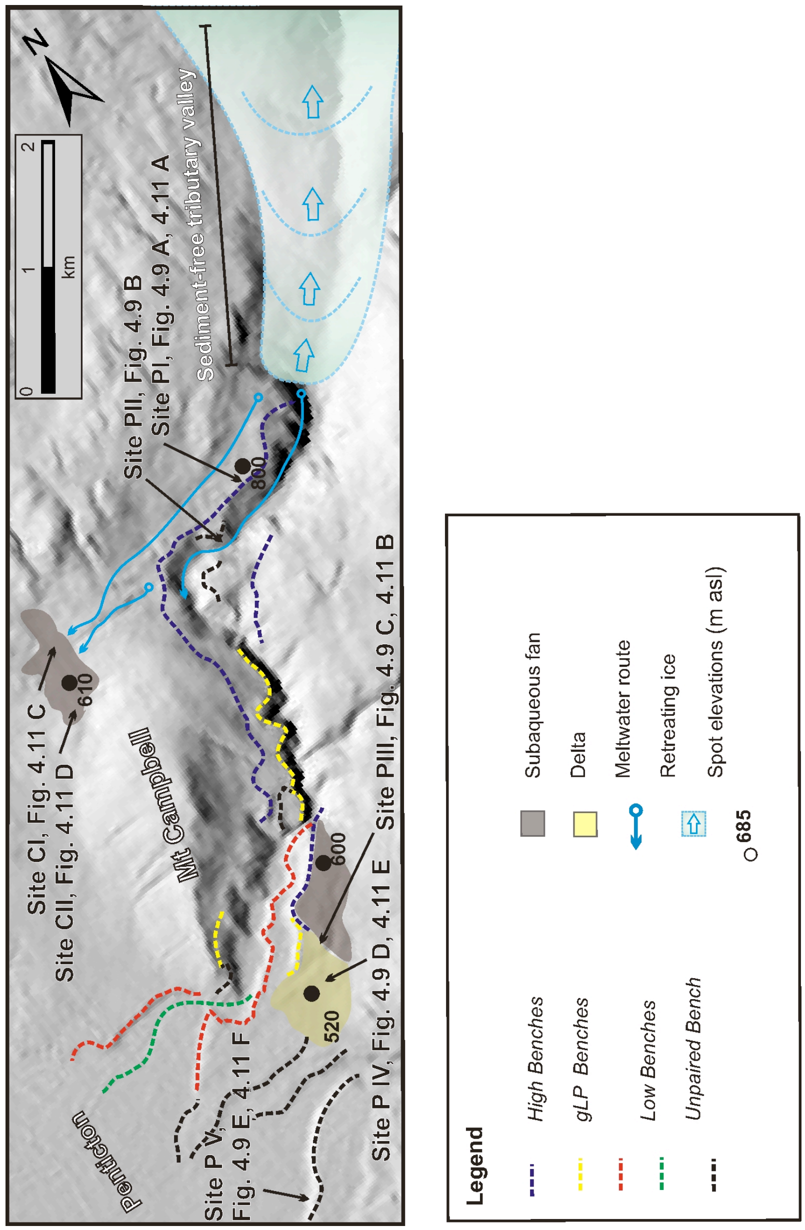


***Figure 4.10.** Landforms within the sediment fill along Penticton Creek valley. This valley exhibits subaqueous fans and deltas . Incision has dissected fan surfaces, producing benches that can be paired (coloured dashed lines) or unpaired (dashed black lines). Two subaqueous fans occur along Penticton Creek (north and south of Mt. Campbell) deposited within a deep receiving basin. A delta has developed on top of subaqueous fan sediments. Note: Position of the ice margin only serves to illustrate an upland/plateau meltwater and sediment source, it is not constrained by geomorphic or sedimentary records. Refer to Fig. 4.2 for location of map. Base map: BC Trim 25 m DEM, GeoBC, use by permission.*

(Fig. 4.9 A). More pronounced northwesterly paleoflows occur in the upper 1.8 m (Fig. 4.9 A). Sediments in neighbouring *High Bench* (Site PII, Fig. 4.9 B, Fig. 4.10) above 610 m asl exhibits similar facies, textures, and paleoflows as those exposed at Site PI.

4.7.2.2.2. Interpretation

In both the *High Bench* and the *Unpaired Bench* the stratified appearance of diamictons and the inclusion of silty laminations and lenses deformed by outsized clasts suggest that these deposits record debris flows (Table 4.2). The stratified appearance likely reflects the presence of shear planes associated with multiple sediment pulses within a debris flow event (Mohrig et al., 1998). Gravel clusters suggest increased turbulence and traction transport, consistent with flow transformation as a result of water incorporation within a subaqueous debris flow (cf. Mohrig et al., 1998; Elverhoi et al., 2007; Breien et al., 2007). Similarly, dilution of a subaqueous debris flow and associated sediment lofting (Sohn, 2000; Felix and Peakall, 2006; Breien et al., 2007) may explain the presence of laminated lenses which record suspension settling of silt between debris flow events. Preservation of these lenses implies limited erosion by successive events, possibly as a result of hydroplaning (cf. Mohrig et al., 1998; Elverhoi et al., 2000; Breien et al., 2007). The presence of striae on some clasts may indicate a till source for these debris flows (Dreimanis, 1982). However, striae can also be produced by clast-clast contact within debris flows (Atkins, 2004) making a glacial origin for these deposits equivocal. The rounded to sub-rounded character of clasts, and the absence of other types of glacigenic wear features (e.g., facets, keels, plucked ends, etc.) suggest either a non-till origin or limited glacial transport of these clasts. Heterogeneous gravel with imbricate clusters and laminated matrix suggest traction transport and localized clast support by turbulence. Planar-stratified gravel and sand are reported from subaerial fan systems where they record deposition by bedload sheets (Nemec, 1990; Zielinski and Van Loon, 2000). They may also result from deposition by turbidity currents in subaqueous settings (Table 4.2) (Sohn et al., 1997; Postma, 2000). The occurrence of cross-laminated sand grading to laminated silt suggest deposition of sand by turbidity currents where suspension deposition of silt follows the passage of a current.

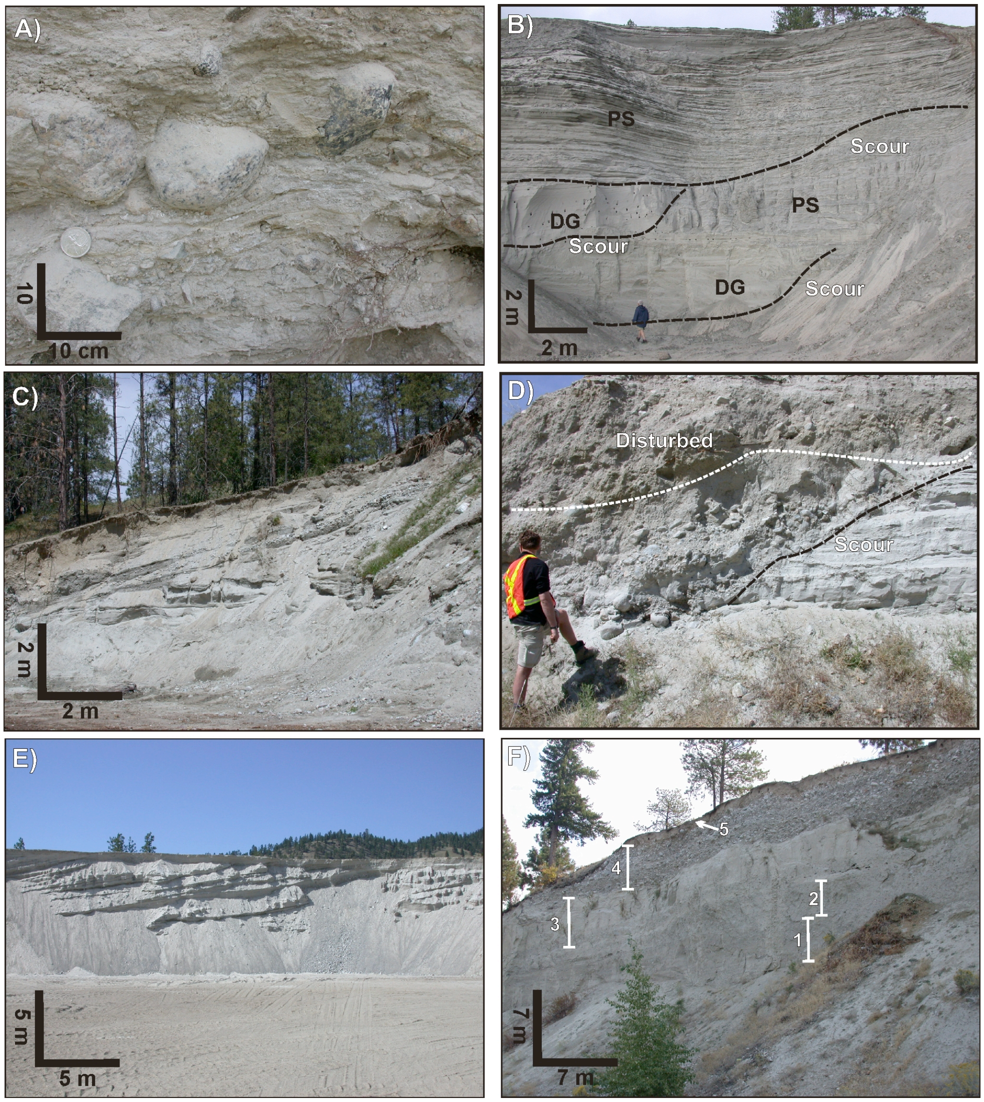


Figure 4.11. Sediments and landforms associated with Penticton Ck. valley train. A) Stratified diamicton including laminated silty sand deformed by gravel clasts (coin diameter: 2.5 cm) near the base of the High bench. B) Planar stratified (PS) and diffusely graded (DG) medium to fine sand in subaqueous fan (~600 m asl). C) Cross-stratified sand and gravel in a subaqueous fan north of Mt. Campbell near 610 m asl. D) Heterogeneous gravel filling scour in planar bedded and diffusely-graded fine sand. Distal exposure in same subaqueous fan as exposed in C. E) Cross-stratified sand and gravel forming delta foresets inset below subaqueous fan of photo B, F) Cross-section through distal valley train sediments exhibiting planar bedded and trough cross-bedded sand and gravel (1) passing into a diamicton with a sheared and graded lower contact (2), overlain by laminated and massive fine sand and silt (3). Imbricate and trough cross-stratified gravel (4) truncate the underlying fine sand and silt (3). Eolian sediments cap the entire sediment package (5). Refer to Figure 4.10 A for photograph locations.

Similarly, laminated silt capping some gravel beds results from suspension deposition after gravel emplacement. Interbedding of laminated silt throughout the exposure further indicates subaqueous deposition of the entire sediment thickness. Therefore, sediments within the *High Bench* and the *Unpaired Bench* (site PI and PII, Fig 4.10) are interpreted to form part of a subaqueous fan deposited within the confines of the bedrock tributary valley.

4.7.2.3. Sedimentology of mid-valley, *gLP benches*: subaqueous fan and delta sediments

4.7.2.3.1. Observations

Extensive paired and unpaired *gLP Benches* are preserved between 600-500 m asl, south of Mt. Campbell, where Penticton Creek valley widens (Fig. 4.10). ~14 m of sediments are exposed within a *gLP bench* at site PIII (Fig. 4.9 C) near 600 m asl (Fig. 4.10). These sediments consist mainly of medium to coarse sand. Two m-scale scours filled with 4 m of diffusely-graded sand and up to 8 m of planar-stratified sand, and cross-laminated sand (type A, and occasionally type B, Table 4.2) grade to laminated silt and silty clay (Fig. 4.11 B). Contacts frequently exhibit mm- to cm-scale diapiric and flame structures. A 1 m thick planar-stratified cobble gravel bed occurs near 7.5 m height (Fig. 4.9 C). Paleoflow measurements from cross-laminations are dominantly down valley, following the flanks of Mt. Campbell (Figs. 4.9 C, 4.10).

Sand and gravel deposits occur at approximately the same elevation to the North of Mt. Campbell (sites CI, CII, Fig. 4.10). These deposits occur in a small bedrock basin formed by structural lineaments, immediately downflow from a bedrock step. A continuous bench surface, extending along Penticton Creek from ~800-630 m asl (Fig. 4.10), ends at the bedrock step. Basin sediments have accumulated below this step and are exposed in two areas. An uppermost group of exposures (CI) occurs against the flanks of the bedrock basin. Exposed sediments occur westward (i.e. toward Okanagan Valley) of this point. A second group of exposures (CII) occurs south-westward and ~35 m below the CI exposures (Fig. 4.10). Sediments at site CI consist of cm- to m-scale inclined beds of planar-stratified sand, planar-stratified pebble gravel, and weakly stratified heterogeneous gravel. These beds dip away from the basin bedrock walls (~195° azimuth). Dip angles vary from 15° near the exposure base and increase toward the top of the exposure where they reach ~ 20-25° (Fig. 4.11 C). No other deposits overlie these dipping beds as they are conformable to the land surface sloping away from the edge of the bedrock basin. Sediments at site CII consist of sub-horizontal to gently dipping (~3-5°) beds of fine to medium sand, interbedded and occasionally interlaminated with silty sand and silty clay. Fine sand beds are commonly planar-stratified to massive, but may also be cross-laminated and diffusely graded. Fine sand beds are capped by silty clay.

Deformation structures such as flames and ball and pillow structures occur in these beds and repeated scouring of these bedsets produces broad undulations. Normal faults occur along scour margins. Centimetre-scale fine sand beds capped by silty clay fill the scours. In other cases, scours are filled by heterogeneous gravel interbedded with silty fine sand, and cobble gravel (Fig. 4.11 D).

Approximately 45 m below site P III cross-sets of weakly planar-stratified, reverse-graded medium sand, planar-stratified gravel, and occasional tabular cross-stratified gravel beds are exposed in a gravel pit (515 m asl; Site PIV, Figs. 4.9 D, 4.10, 4.11 E). Cross-sets dip at ~15-20° and are overlain by near horizontal planar-stratified sand and gravel, and trough cross-stratified sand and gravel beds. Imbricate clasts within planar-stratified gravel beds record westerly to southwesterly flows (Fig. 4.9 D).

4.7.2.3.2. Interpretations

Sediments at sites PIII, CI and CII record deposition within subaqueous fans located along Penticton Creek valley, and the valley north of Mt. Campbell. These may be downflow extensions of the subaqueous fan(s) inferred from sites PI and PII or separate subaqueous fans formed later than those at PI and PII.

Local variations in basin geometry lead to distinct sedimentary architecture. In the exposure located along Penticton Creek (site PIII), planar-stratified and cross-laminated sand records deposition by turbidity currents (Table 4.2). Transitions from type-A to type-B cross-laminations reflect the increasing importance of deposition from suspension. Vertical gradation to laminated silt and silty clay further records the latest stages of suspension fallout following the passage of turbidity currents. Frequent loaded contacts result from rapid deposition onto water-saturated sediments, and bolster the inference of subaqueous deposition. Diffusely-graded sand filling scours is attributed to rapid deposition from hyperconcentrated flows due to flow expansion in a scour zone associated with a hydraulic jump (Gorrell and Shaw, 1991; Russell and Arnott, 2003), or more sustained deposition from hyperconcentrated flows (Kneller and Branney, 2006). Diffusely-graded facies are common in subaqueous fans of glaciated basins (Rust and Romanelli, 1975; Russell and Arnott, 2003; Winsemann et al, 2007). Overall, these subaqueous fan deposits record frequent sediment delivery events punctuated by quiescent periods allowing for suspension settling of fines. Sediment delivery may reflect either individual flow events or pulses within a longer-lived sustained flow (e.g., hyperpycnal flow, Mulder and Alexander, 2001). The abundance of medium and fine sand within the fan likely reflects a sandy sediment source or a depositional setting distal to coarser sediments. Consequently, coarse gravel beds (Figs. 4.10, 4.11) record episodic high magnitude events in response to higher discharge or possibly a more proximal depositional setting.

Within the bedrock basin north of Mt. Campbell, sandy and gravely cross-sets at site CI could be interpreted as delta foresets. However, the absence of sediments overlying these cross-sets (possible topsets) makes a deltaic interpretation equivocal. Consequently, the cross-sets at site CI may record initial infilling of the bedrock basin and sediment avalanching as channels deliver sediments to the basin rim where the slope increases beyond the break. Sediments at CII record subaqueous deposition distal to site CI. Interpretations of site CII are similar to those at site PIII. Diffusely graded sands with flame and loading structures record subaqueous deposition punctuated by quiescent periods allowing for suspension settling of finer material. Frequent scours may result from erosion by sediment flows originating at site CI. Gravel-filled scours reflect episodic high energy events and/or proximal depositional conditions.

At site PIV, the overall architecture and bed dip angles are consistent with those associated with deltas (possibly Gilbert-type deltas) (Nemec, 1990a and references therein). In this context, cross-sets record progradation of delta foresets and the overlying planar-stratified and local trough cross-stratified beds record delta topsets. Lithofacies suggest sediment avalanching down the delta foreslope as turbidity currents (possible hyperpycnal flows) and grain flows (reverse-graded beds). Clast imbrication in the topsets suggests more fluidal flows consistent with flows in river channels or distributaries. The delta surface is at 515 m asl and falls within the gLP water plane envelope.

4.7.2.4. Sedimentology of tributary-mouth *gLP Benches*: distal subaqueous fan and alluvial fan sediments

4.7.2.4.1. Observations

Tributary-mouth *gLP Benches* occur at the confluence of Ellis and Penticton creeks (Site PV, Fig. 4.10). An ~10 m high exposure in the riser of one of these benches reveals five units (Figs. 4.9 E, 4.11 F) that are laterally continuous for over 150 m. Sand and gravel dominate the exposed material in the lowermost 6 m (unit 1) (Fig. 4.9 E). These sediments consist of planar-stratified, and trough cross-stratified gravel and sand with a few heterogeneous gravel beds and open framework pebble and granule beds. An armoured mudball and other silty rip-up clasts occur in some sandy beds. Unit 2 is composed of diamicton (Fig. 4.9 E). Its lower contact with unit 1 varies from gradational to sharp. The gradational transition occurs over ~70 cm thickness and consists of interbeds of laminated fine sand and silt, planar-bedded sand and silty diamicton. Silty and clayey stringers occur where the contact is sharp. Clasts within the diamicton are rounded to sub-rounded and occasionally sub-angular, and rarely exhibit striae. Unit 3 is composed of thinly-bedded and laminated silt and silty sand interbedded with diamicton of unit 2 (Fig. 4.9 E). Unit 4 occurs above a sharp and undulating contact with unit 3 (Fig. 4.11 F). It consists mainly of trough cross-stratified sand and gravel (Fig. 4.9 E). Unit 5 consists of 30-40 cm of massive fine sand infilling scours in the top of unit 4 (Figs. 4.9 E, 4.11 F). Braided channel patterns are visible on the bench surface.

4.7.2.4.2. Interpretations

Gravel in unit 1 reflects rapid deposition by a combination of hyperconcentrated flows and high-density turbidity currents (Table 4.2) in a subaqueous setting or, alternatively, subaerial deposition on an outwash plain. The occurrence of an armoured mudball within sandy beds is indicative of erosive flows entraining lacustrine sediments (Little, 1982; Krzyszton, 1984 in Etienne et al., 2006). The diamicton of unit 2 is interpreted as a debris flow. Silty and clayey stringers toward the lower contact of unit 2 are may be associated with shear planes due to debris flow movement (Hampton, 1975; Lachniet et al. 1999). The paucity of striae on diamicton clasts suggests that the material probably does not originate from till. However, this possibility cannot be completely ruled out since some rare striae occur on some clasts. The diamicton material may, therefore, be associated with channel bank failures or mobilization of glacigenic sediments within the tributary or at an ice front in the tributary. Laminated silt and silty sand of unit 3 records mainly suspension deposition of fines. The elevation of unit 3 is within the elevation range of gLP and the White Silt benches, suggesting that this unit may be associated with gLP. Interbedding of silts (unit 3) and diamicton (unit 2) also suggests a subaqueous depositional setting for unit 2. By corollary, this strengthens the subaqueous interpretation for unit 1 given some gradational contacts between units 1 and 2. Sharp contacts associated with shear planes at the base of unit 2 may still be produced in a subaqueous setting (Mohrig et al., 1998). The sharp lower contact of unit 4 is attributed to erosion of the underlying silt bed (unit 3). Trough cross-stratified sand and gravel (unit 4) results from scouring and filling within fluvial channels or by turbidity currents in a lacustrine basin (Postma, 1990; Mulder and Alexander, 2001). The high degree of sorting of the gravel, and general absence of silt in the sand and gravel would favour a subaerial depositional setting such as an alluvial fan or a braided river. The latter interpretation is consistent with observations of scours within this unit and braided channel pattern on the bench surface. Unit 4 is therefore interpreted as an alluvial fan surface or, alternatively, a braided river system, Lastly, the texture, massive appearance, and stratigraphic position of unit 5 suggest that this material is eolian sediment. Eolian sediment frequently occurs near the top of sediment benches in southern British Columbia (Church and Ryder, 1972).

4.7.2.5. Summary of Penticton Creek valley fill: an example of sedimentation in a deep and narrow receiving basin

Documented sites along Penticton Creek record mainly subaqueous sediment deposition in a deep and narrow receiving basin. Basin geometry affects subaqueous fan development. In this setting, slope angle of the receiving basin and density of incoming flows exert important controls on sediment deposition: a steep basin slope beyond the sediment input point creates a voluminous basin. In combination with dense sediment flows, subaqueous fans develop as sediments cannot aggrade to the water plane (during the time span of sediment delivery and/or water plane stability) to form deltas (Fig. 4.12 A). However, One exception occurs at site PIV where a delta records subaerial deposition and the presence of a stable water plane (Fig. 4.12 A). It is hypothesized that this delta is inset within the subaqueous fan sediments filling most of the tributary valley. Sequential incision and downvalley sediment shunting fills the receiving basin and allows for delta aggradation.

The elevation of some exposures in Penticton Creek valley requires the presence of a subaqueous setting at an elevation exceeding the height of spillways at McIntyre Bluff (§ 4.6.5). Consequently, the highest exposures recording subaqueous deposition may not be associated with gLP. The implication is that a higher water plane occurred within Penticton Creek valley prior to gLP development. Two possible explanations for an elevated water plane include ice-marginal lakes developed within a tributary (cf. Paradis et al., 2010) or subglacial lake development within Okanagan Valley (cf. Lesemann and Brennand, 2009, Chapter 3).

Due to the availability of exposures in Penticton Creek valley, observations and interpretations of the valley fill are most complete here. A similar depositional setting occurs along Mission Creek and Shingle Creek (Fig. 4.2). Consequently, interpretations of Penticton Creek valley fill is considered to be generally applicable to the sediment fill of Mission Creek and Shingle Creek.

4.7.2.6. Sedimentation in a shallow receiving basin: a case study of Trepanier Creek valley fill

4.7.2.6.1. Observations

Trepanier Creek is characterized by a continuous valley train extending for over 15 km between Okanagan Valley and the steep bedrock slopes leading to the regional plateau. A prominent sediment fan, characterized by a subhorizontal surface marked by channel scars, irregular depressions, and a steep-fronted (~30°) down-valley dipping arcuate morphology (Fig. 4.3 F) marks the end of the valley train where Trepanier Creek enters Okanagan Valley (Figs. 4.13 A). A single *High Bench* surface occurs on the northern valley side at 625 m asl. Limited exposures reveal diffusely graded coarse sand interbedded with boulder gravel. A discontinuous *gLP Bench* occurs above the valley train near the tributary mouth. *Low benches* occur within the fan (Fig. 4.13A).

4.7.2.6.2. Interpretations

Despite limited exposures, *High Bench* sediments are tentatively interpreted as subaqueous fan deposits, largely due to the presence of diffusely graded sands (Russell and Arnott, 2003). The sediment fan at the tributary mouth below the *High Bench* is interpreted as a delta based on its morphology: its near-horizontal upper surface and the presence of faint channel scars interpreted as distributaries. Irregular depressions are interpreted as kettle holes associated with ice block melting. Coarse gravel exposed at the surface and the possibility of kettles within the delta suggest energetic (glacio)fluvial sediment transport. Kettle holes record melting of buried ice blocks. These blocks may have originated as bergs within gLP or, alternatively were transported during jokulhlaups (Fay, 2002). The limited valley-ward extent of the delta front may reflect the presence of a pronounced slope break as Trepanier Creek enters Okanagan Valley, leading to sediment avalanching into the deeper basin of Okanagan Valley proper. The abrupt termination of some deltas has been used to argue for an ice-contact origin for these features (Nasmith, 1962, Paradis et al., 2010). This is not necessarily required however. Basin geometry limits delta progradation and accounts for the morphology and valley-ward extent of Trepanier Creek delta. This depositional setting (Fig. 4.12 B) is recognized in other valleys such as Fintry Creek, Naswhito/Equesis Creek, Bouleau/Whiteman Creek, and Lambly Creek (Fig. 4.2).

4.7.2.7. Multibasinal sedimentation along tributary valleys: a case study of Trout Creek basin

A multibasinal depositional setting is characterized by sedimentation in one or more extensive basin occurring between the tributary mouth and the main depocentre in Okanagan Valley (Fig. 4.12 C). It is exemplified by the Trout Creek valley (Fig. 4.13 B).

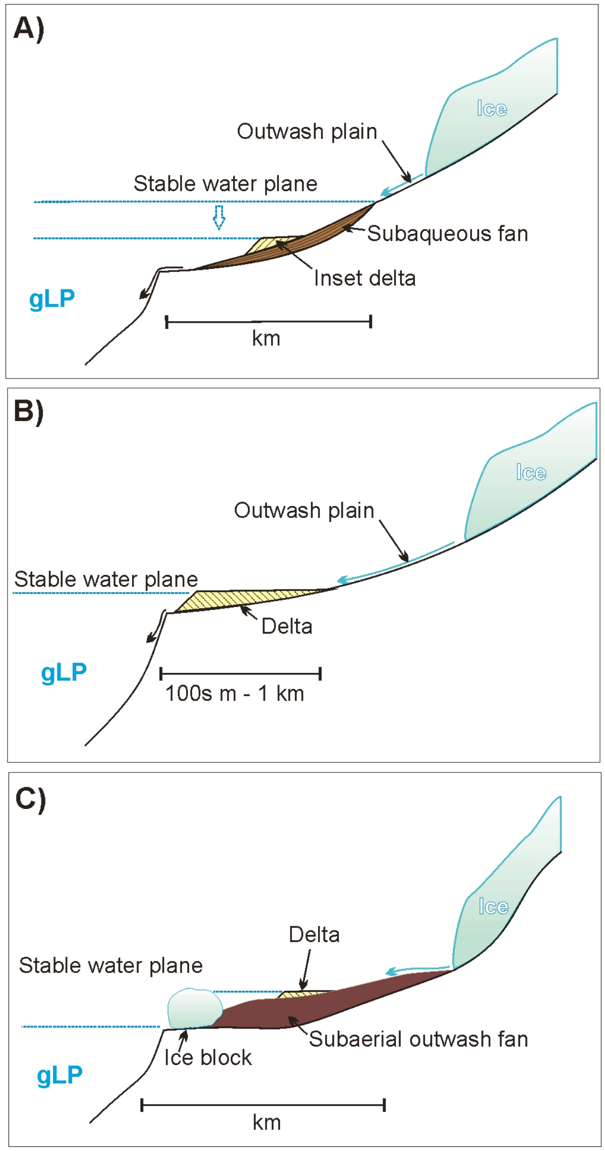


Figure 4.12. Three geomorphic settings for development of sediment sequences within the glacial Lake Penticton basin. A) Deep and narrow receiving basin: a steep bedrock surface and deep receiving basin promotes subaqueous fan development. An inset delta graded to a lower water plane develops following erosion of the subaqueous fan. B) Shallow receiving basin: a delta connected to an outwash plain develops in the lake basin. C) Multi-basin deposition: a long receiving basin between the tributary mouth and the Okanagan Valley basin leads to deposition of a valley train which may extend to the lake basin but this is not a necessary condition. Ice blocks may lead to local ice-dammed lakes, delta development, and river diversion. Note: Ice margin position is schematic and used to illustrate the meltwater/sediment source.

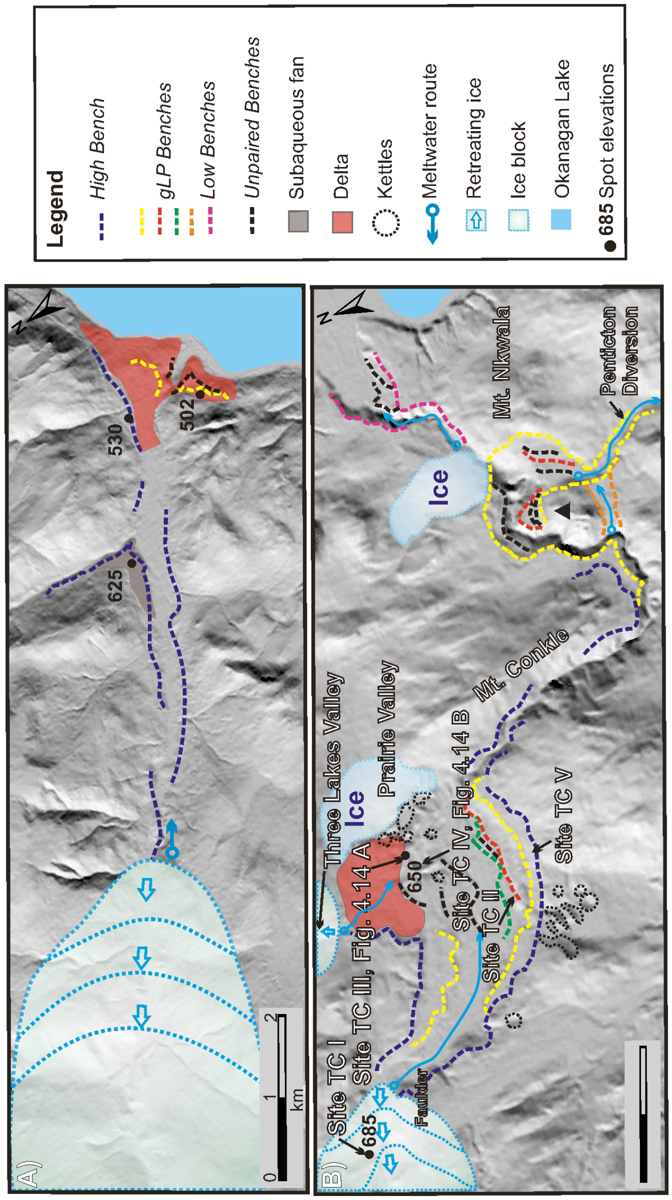


Figure 4.13. Tributary valley trains, subaqueous fans, deltas, and main benches along Trepanier Creek (A) and Trout Creek basin (B). A) Trepanier Creek is characterized by a continuous valley train leading to a delta at the tributary mouth. A High Bench suggests dissection of a surface partly characterized by subaqueous fans. Terraced delta surface results from incision during lake drainage. B) The Trout creek basin is characterized by an extensive valley train with a possible inset delta. Complex basin topography and possible remnant ice blocks have led to drainage diversions through bedrock channels (Penticton Diversion). Note: Position of the ice margin is schematic and only serves to illustrate an upland/plateau meltwater and sediment source, and is not constrained by geomorphic or sedimentary records. Base maps: BC TRIM 25 m DEM © GeoBC, by permission.

4.7.2.7.1. Observations

Trout Creek drains the regional uplands leading to the edge of Thompson Plateau (~1200-1400 m asl) and exits into a ~10 km long bedrock basin (here named Trout Creek basin) before reaching Okanagan Valley (Fig. 4.13 B). Smaller bedrock and sediment floored tributary valleys (Darke Creek, Eneas Creek, Fig. 4.2, and Three Lakes Valley, Fig. 4.13 B) also drain into this basin from intermediate uplands (~700-800 m asl) leading to Thompson Plateau. Trout Creek enters the bedrock basin from a deeply-incised ~250-300 m wide valley. Basin width increases rapidly down valley and reaches 1.5 km near Faulder (Fig. 4.13 B) and over 4 km at the edge of the White Silt benches above Okanagan Lake. An extensive valley train fills the Trout Creek basin and extends, uninterrupted for ~5km from the break in slope of the regional uplands (685 m asl near Faulder) toward Prairie Valley (Fig. 4.13 B). The upper valley train contains *gLP Benches* and is characterized by an irregular surface topography which includes metre-scale semi-circular closed depressions and more linear decametre-scale depressions. The valley train ends abruptly in a steep-fronted zone of hummocky topography characterized by closed depressions (620-650 m asl, Fig. 4.13 B). However, the southern portion of the valley train continues southeastward against the western flanks of Mt. Conkle and toward the Penticton Diversion (a bedrock channel at ~560 m asl draining toward Penticton) (Nasmith, 1962) (Fig. 4.13 B). Valley train incision has produced multiple paired and unpaired *gLP Benches and* over 20 unpaired and/or dissected *Low Benches*. Channel scars occur frequently on the more extensive bench surfaces. Only the most prominent *gLP* and *Low Bench* surfaces are mapped in Figure 4.13 B. The *High Bench* is nearly continuous along the entire length of the valley train, save for a gap south of Mt. Conkle. In contrast, Low Benches disappear as the valley train follows the western flanks of Mt. Conkle. It reappears where the valley widens south and east of this mountain. There, the *High Bench* and some *gLP Benches* curve around a bedrock summit (triangle in Fig. 4.13 B) and merge into the bedrock channel of the Penticton Diversion (Fig. 4.13 B). Subsequent *Low Benches* by-pass the Penticton Diversion (560 m asl) and slope toward Okanagan Lake. Areas to the east of Prairie Valley (~530 m asl) consist of a gently sloping valley floor that extends almost continuously to the edge of the White Silt Benches (~415 m asl). Bedrock outcrops and prominent bedrock knobs (e.g. Giant's Head, Fig. 4.8) extend above the valley floor.

Well sorted sand and gravel (sub-rounded to rounded), interbedded with fine sand and silt dominate the sedimentology of the valley train. Sediments exposed at the head of the valley train near Faulder (Site TCI, Fig. 4.13 B) consist of gently dipping beds of planar-stratified coarse sand, and pebble to cobble gravel. Beds dip at ~ 2-5° and are generally conformable with the surface slope of the *High Benches*. Rhythmically-bedded and laminated fine sand and silt are interbedded with coarser sand and gravel. Convolutions and loading structures occur frequently within the finer beds (site 8, Table 4.3, Fig. 4.2). Exposures along bench risers (Site TCII, Fig. 4.13 B) reveal subhorizontal beds of planar-stratified sand and gravel, and trough cross-stratified sand. Cross-laminated fine sand and silt, and planar-laminated silt and clayey silt are interbedded with the coarser sand beds. Based on limited exposures and some water well logs, these fine-grained beds appear discontinuous and occur throughout the valley train thickness.

Near Prairie Valley, exposures in an unpaired *Low Bench* below the *High Bench* (650m asl) reveal at least 15 m thick dipping beds of planar-stratified coarse sand, and stratified and imbricate gravel, overlain by discontinuous subhorizontal beds of planar-stratified sand and planar-stratified pebble and cobble gravel (Site TCIII, Figs. 4.13 B, 4.14 A). Sandy waveforms are exposed at Site TCIV, ~ 200 m south and ~20 m higher than the dipping beds at Site TCIII (Fig. 4.13 B). Stratigraphic continuity between these two sites cannot be verified however. At site TCIV, bed dip directions and paleocurrent measurements from imbricate clasts in planar-bedded gravel record dominantly southward flows.

Exposures in discontinuous *High Benches* (possibly *gLP Benches*) within Trout Creek basin (site TCV, located ~600 m south and 30 m below site TCIII, Fig. 4.14 B) reveal 3.5-4 m wavelength formsets exhibiting 0.6 m high planar cross-stratified and trough cross-stratified medium to coarse sand (Fig. 4.14 B). Paleocurrent measurements from cross-stratification record southeastward to southward flow toward the narrow valley segment flanking Mt. Conkle. (Fig. 4.13 B). Water well logs reveal that fine-grained sediments, a few metres thick, form the valley floor east of Prairie Valley. However sand and gravel dominate north of Prairie Valley in the area between Eneas Ck (Fig. 4.8) and the edge of the White Silt benches (corresponding to the area of Summerland (Fig. 3.2) at an elevation of 500 m asl). Exposures near the edge of the

White Silt benches contain m-scale sand and gravel cross-sets dipping toward Okanagan Valley (Kvill, 1976).

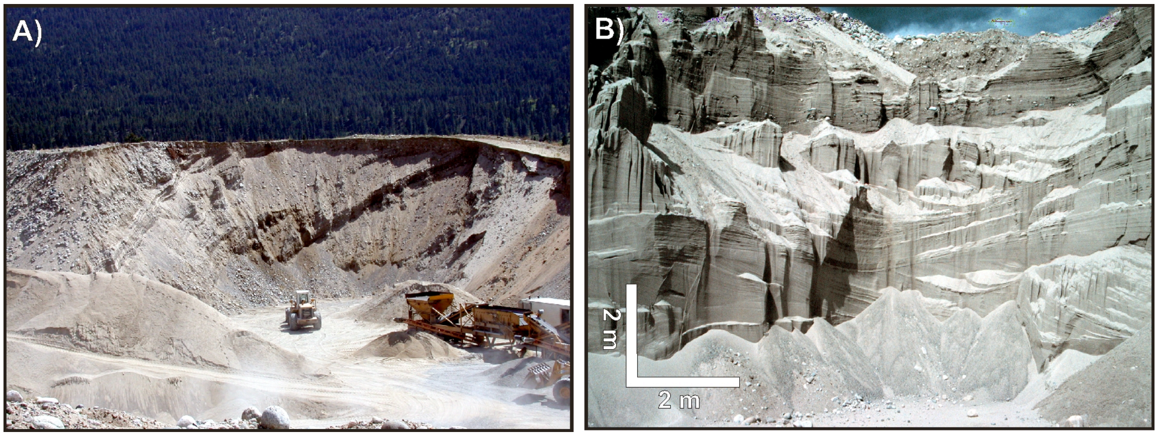


Figure 4.14. Sediments within the Trout Creek valley train. A) Sand and gravel foresets in a delta inset into the valley train (Site TC III, 650 m asl). B) Sandy waveforms within the Trout Creek valley train (Site TC IV, 620 m asl).

4.7.2.7.2. Interpretations

Benches within the Trout Creek basin valley train result from incision. Evidence for sediment aggradation on *Low Bench* surfaces is difficult to assess given limited exposures in bench risers. However, sediment removal along the axis of the valley train strongly suggests dominance of erosional processes, though localized aggradation may have occurred on some surfaces. The irregular hummocky surface and the presence of closed depressions within the *High Bench* of the valley train are interpreted as kettle holes resulting from melting of buried ice blocks as suggested by Nasmith (1962) and Kvill (1976). The abrupt termination of the valley train near Prairie Valley in an area of steep-sided hummocky topography suggests ice contact deposition, requiring stagnant ice in Prairie Valley (Nasmith, 1962). As well, presence of ice in the lower reaches of Trout Creek valley is inferred in order to deflect drainage toward the Penticton Diversion (Nasmith, 1962). An ice dam (Fig. 4.13 B) would need to be at least as wide as the gap where Trout Creek presently flows (~500 m wide). *gLP Benches* along the Penticton Diversion record incision of sediments within this channel. Flow into the Penticton Diversion ceased as incision lowered the outwash stream to ~540 m asl. Eastward flow toward Okanagan Valley resumed as the ice dam melted and/or further incision re-established drainage within the bedrock walls of the Trout Creek canyon. It is unclear exactly when this eastward rerouting took place. However, sedimentation by the stream led to development of modern Trout Creek delta graded to Okanagan Lake. No other delta developed along this stream within the regionally correlated delta envelope. This then suggests that a significant portion of the incision of the Trout Creek valley fill post-dates drainage of gLP and is therefore the result of late deglacial/early Holocene fluvial activity. These surfaces are therefore *Low Benches* and some may be *Recent Benches*.

The sorted and sub-rounded to rounded character of sediments in the Trout Creek basin valley train record mainly fluvial transport. The inference of buried ice blocks within the valley train surface suggests a dominance of glaciofluvial processes for valley train deposition. Planar-stratified and trough cross-stratified sand and gravel within benches record (glacio)fluvial deposition including bedform migration (dunes) (Miall, 1979), and possibly local aggradation on bench surfaces. Rhythmically-bedded fine sand and silt record quiescent subaqueous depositional conditions. The apparently limited lateral extent of these beds indicates that sediments may have been deposited within ponds and small basins on the valley train surface. Dipping beds at site TCIII may record delta foresets. However, this interpretation is equivocal because the presence of topset beds cannot be verified given gravel pit activity, limited access, and the quality of exposures. If these sediments do in fact record deltaic sedimentation (Fig. 4.14 B), then they record a stable water plane near ~650 m asl associated with a local ice-dammed lake. An alternative interpretation is that the dipping beds result from large-scale collapse due to melting of buried ice blocks in the *High Bench* and possibly associated with the hummocky topography near Prairie Valley. However, an absence of faults within the dipping beds speak against this possibility. Metre-scale waveforms (Fig. 4.14B) including diffusely-graded sand record deposition by hyperconcentrated flows (Gorrell and Shaw, 1991). These forms likely result from much higher energy events than other sediments in the Trout Creek valley train and are potentially associated with episodic high magnitude events such as jokulhlaups (cf. Duller et al. 2010). Kvill (1976) interpreted the sediments in Prairie Valley as a local lacustrine plain. Extensive sand and gravel deposits at Summerland leading to gravely foresets near the lacustrine benches record development of a delta emanating from Eneas Creek (Fig. 4.8) (Kvill, 1976) and graded to the regional delta envelope.

In conclusion, Trout Creek valley fill is characterized by sedimentation in multiple sub-basins within an area between the regional uplands and Okanagan Valley (Fig. 4.12 C). These sub-basins result from a combination of topographic obstacles and local ice dams that deflected meltwater drainage and led to temporary ice-dammed lake development. Bedrock knobs create complex basin topography which likely played a role in trapping and retaining dead ice within this valley. Multibasinal sedimentation (Fig. 4.12 C) is also recorded along Powers Creek valley (Fig. 4.2).

4.7.3. Genesis of tributary valleys

The low channel orders of many bedrock channels connecting Okanagan Valley and surrounding plateaus record focused erosion along these valleys. Given a strong regional structural grain generally oriented parallel to oblique to Okanagan Valley, there is a high likelihood of structural control on channel patterns. However, this cannot be the only pattern control as some channels are eroded across the bedrock structural grain. Erosional processes are therefore responsible for a significant portion of the channel pattern. On the plateaus, channels of similar pattern (but larger dimensions) have been interpreted as remnants of Eocene fluvial systems (Tribe, 2005). Similarly, fluvial gorges in glaciated montane settings have been shown to persist through glacial episodes and resist strong modification during glaciation (Montgomery and Korup, 2010). Some bedrock channels within Okanagan Valley and on the Thompson plateau operated as tunnel valleys (subglacial meltwater conduits) (Lesemann and Brennand, 2009). Such channels are probably polygenetic, experiencing repeated occupation by ice and meltwater during glaciation and streams during interglacials. The v-shaped channel cross-sections further support inferences of dominant (glacio)fluvial erosion. Had glacial erosion dominated, more u-shaped valley cross-sections might be expected (Harbor, 1992, Montgomery, 2002). Many variables control valley cross-sectional shapes. These include bedrock type, geologic structure and the nature and intensity/duration of erosional agents. However, in areas with similar geologic characteristics, the effects of different erosional agents can be assessed. For example, Riedel et al., (2007) documented, in the nearby North Cascades, clear juxtaposition of u-shaped and v-shaped valleys. u-shaped valleys were associated with glacier erosion. V-shaped valleys were associated with meltwater drainage (Riedel et al., 2007). Based on valley cross-sectional shape, and fills containing an abundance of glaciofluvial sediments terminating in deltas and fans, tributary valleys of Okanagan Valley are interpreted as major meltwater throughways between plateaus and Okanagan Valley during deglaciation of the most recent CIS (and possibly earlier glaciations).

4.7.4. Development of tributary valley trains and benches

Tributary valley benches result mainly from incision through valley train sediments. Early studies of the gLP basin recognized the presence of terraces near tributary mouths (Flint, 1935b, Nasmith, 1962). These terraces were attributed to episodic changes in local base level. For example, Nasmith (1962) identified four main terrace surfaces in Trout Creek and associated each surface with a distinct water plane in gLP. Taylor (1961) identified seven terraces and equated their development to regional climatic changes leading to base level fluctuations during deglaciation. However, these studies likely oversimplify the association between terraces and base level fluctuations (cf. Muto and Steel, 2004). For example, Nasmith (1962) only considered the most prominent terraces in Trout Creek and did not recognize the significance of smaller benches occurring between his four prominent benches. Smaller benches also record erosion associated with base level drop. The focus placed on the larger surfaces exaggerates their relative importance at the expense of smaller ones. The implication is that larger benches record processes more clearly than smaller ones. This fails to recognize that bench size is also a result of erosional patterns as channels migrate, and that varying discharges will affect the erosional efficiency.

Rudimentary bench correlations based on absolute elevation and elevation relative to regionally-correlated deltas illustrate that narrow bench segments may be correlative with more extensive benches within the same tributary valley (yellow and red dashed lines, Fig. 4.13 B). More importantly, not all benches (especially the *gLP Benches* and *Low Benches*) terminate in a reliable water plane indicator such as a delta. For example, at least two (and possibly more) prominent *gLP Benches* within the Trout Creek valley converge toward the Penticton Diversion (Fig. 4.13 B). However, only a single perched delta has developed at the distal end of the Penticton Diversion (Fig. 4.8), rather than multiple deltas as would be expected if each bench resulted from episodic base level drop and water plane stabilization. Therefore, the case of the

Penticton Diversion illustrates that multiple benches may be associated with a single water plane indicator. This pattern is repeated in most tributary valleys where a stable water plane indicator is recorded by a delta in the 500-525 m gLP water plane envelope (Fig. 4), yet no other delta has developed during subsequent incision (save for the fan deltas graded to the modern lake level, and deltas developed in local basins such as in Trout Creek basin). *High Benches*, *Low Benches* and *Recent Benches* do occur above and below the gLP delta elevation envelope but these surfaces are mainly eroded into existing valley train/delta/lake bottom sediments. Therefore, although most of the sediments within the *High Benches* are derived mainly from the surrounding plateaus, subsequent incision into this surface leads to sequential down-valley shunting of sediments as meltwater incised older (higher) deposits and transfers sediments down-valley. This process may be partly responsible for longitudinal sediment sorting and may account for the dominance of fine sand and silt in the White Silt benches. Local aggradation may occur on some bench surfaces, though the extent of this is difficult to assess given a paucity of exposures.

Given that tributary valley fill does not record episodic lake level drops and stability, the presence of nested benches is best explained by 'autoincision' (Muto and Steel, 2004). This mechanism was identified in laboratory experiments and builds on the concepts of 'complex response' of fluvial systems (Schumm, 1974) by recognizing that an autogenic incision mechanism controlled mainly by channel slope and the sediment/water ratio can lead to formation of multiple benches (terraces) during a single period of base level fall (Muto and Steel, 2004). The morphologic results consist of paired and unpaired benches whose width and down-valley extent reflects channel migration during incision. Additionally, fluvial response (incision) may lag base level drop if sediment supply is sufficiently high to prevent autocyclic incision (Leeder and Stewart, 1996; van Heijst and Postma, 2001; Muto and Steel, 2004). One important implication of this process is that in the gLP basin, the fluvial response of tributaries to regional base level fluctuations will be specific to each tributary: timing of incision and resulting bench patterns will be unique to each tributary due to different channel slopes, and rates of sediment and water supply. Bench correlations within a tributary, and especially between tributaries, are therefore difficult and, more importantly, it is questionable whether they provide meaningful insights into regional base-level changes. Deltas remain the most reliable water plane indicators as they represent water plane stability over a time span where the influence of autogenic processes is dampened (Muto and Steel, 2004).

4.8. Lake type

The fundamental question surrounding the type of lake that occupied Okanagan Valley has yet to be fully resolved, despite various propositions (Table 4.1). In this section, we test existing lake type hypotheses (ice-lateral 'ribbon lake', supraglacial lake, proglacial lake) in light of geomorphic, sedimentologic and published seismic data. The crux of any lake type interpretation centres on the distribution of White Silt benches and lacustrine plains, and the process(es) responsible for their distribution. Specifically, the main question is whether the limited valley-ward extent of the 'White Silt' benches results from limited deposition along the axis of Okanagan Valley (Flint, 1935b, Nasmith, 1962), supraglacial deposition and collapse of lake sediments along the axis of Okanagan Valley (Nasmith, 1962; Shaw, 1977; Eyles et al., 1991), or erosion of lacustrine sediments along the valley axis (Kvill, 1976).

4.8.1. Ice lateral 'ribbon lakes'

The narrow and elongate silt benches along the valley sides in Okanagan Valley led to the proposition that gLP developed as ice-lateral lakes along the margins of a valley-occupying ice tongue. Flint (1935b) argued for this lake environment based on the presence of listric faults within the White Silt bench sediments, which he interpreted as evidence of removal of lateral support during melting of the valley ice tongue. Nasmith (1962) argued for a similar lake environment based on the valley-ward extent of silt benches and their abrupt termination in silt cliffs. His proposed model also seems conditioned by interpretations of regional ice stagnation and meltwater drainage along the margins of decaying ice bodies (cf. Davis and Mathews, 1944). Nasmith (1962) used the presence of numerous (interpreted) kame terraces in tributary valleys to support this model of regional stagnation, though this argument is somewhat circular. Given the trough-like morphology of Okanagan Valley, and assuming a decaying ice tongue in the valley, these lakes would likely expand valley-ward in step with surface lowering of the valley ice tongue. Similarly, as ice decayed, 'ribbon lake' basins could become hydrologically connected during lake expansion in front of, and transgression over, a stagnant ice tongue as suggested by Flint (1935b) and Nasmith (1962). Indeed, regional delta correlations recording a single regional water plane (rather than unconnected ice lateral lakes), support a regional hydrologically connected lake. Thus, 'ribbon lakes' could have developed locally, but this setting would likely be transitional to a more extensive proglacial or supraglacial lake (Flint; 1935b). This type of transition is implicit in Fulton's (1969) summary diagrams of lake development though the environmental changes are not explicitly recognized. Thus; the pattern of White Silt benches can be reconciled with three types of lake.

Paleocurrent measurements from the White Silt benches (Fig. 4.8) may help differentiate between these lake types. Paleocurrents mostly record evidence of valley-ward deposition by turbidity currents originating from tributary valleys. Evidence of turbidity currents originating from valley-axis ice as would be expected in a ribbon lake confined by ice in the valley axis; is more limited and speaks against extensive 'ribbon lake' development. For this reason; further discussions focus on the two other types of lakes that have been proposed for gLP (supraglacial and proglacial).

4.8.2. Supraglacial lake

Flint (1935b) first proposed that gLP developed as a supraglacial lake and this lake type has largely been favoured since (Table 4.1) (Fulton 1969; Shaw 1977; Eyles et al., 1990. 1991). Two main elements underpin propositions of supraglacial lake development along the Okanagan Valley axis: 1) observations of listric faulting in the White Silt benches (for the same reasons described in § 4.8.1) (Flint; 1935b; Shaw and Archer; 1978); and 2) a regional deglacial model suggesting extensive valley-bottom stagnant ice (e.g.; Davis and Mathews; 1944; Fulton, 1991). However neither of these elements is unequivocal because whereas it is true that listric faults record removal of lateral support, their occurrence says nothing about the mechanism by which lateral support is removed; removal of lateral support can be equally well explained by let-down of supraglacial sediments or erosion along the valley axis. Furthermore, rationalizing the presence of a valley-axis deglacial lake (supraglacial or otherwise) from the regional deglacial model is somewhat circular since it was, along with the great depth of valleys, the presence of extensive valley-bottom lacustrine benches that led to the postulate that decaying ice tongues remained in valley-bottoms to supply meltwater to deglacial lakes. Shaw (1977) rationalized supraglacial lake development based on the likely preservation of ice in the valley bottom due to the great depth of the lake basin and its irregular topography which were unlikely to result from fluvial incision of the lacustrine fill following deglaciation. The absence of river terrace gravels atop the White Silt benches, as seen in other valleys of southern British Columbia (Ryder et al., 1991; Johnsen and Brennand, 2006), supports this interpretation.

Eyles et al., (1991) argued for a 'hybrid' proglacial-supraglacial lake. They envisaged gLP to form atop stranded ice blocks ("supraglacial") in front of an actively retreating ice margin (proglacial). White Silt benches result from letdown of sediments deposited atop the stranded ice blocks. Eyles et al., (1990, 1991) rely mainly on seismic facies and process analogies from coastal fjord environments to support their lake evolution model. However, interpretations of seismic records from Okanagan Lake (Eyles et al., 1990, 1991) are difficult to assess due to contradictory interpretations, inconsistencies between the appearance and descriptions of seismic profiles, and inconsistent subdivisions of the valley fill. For example, Eyles et al., (1991) describe their Unit IIa as massive throughout and resulting from rapid deposition and deformation. This deformation is correlated with deformed beds at the base of the White Silt benches and is used to argue for continuity of Unit IIa to the top of the benches. However, many seismic profiles (figs 4, 5 in St John, 1973, fig. 10 in Eyles et al., 1991) show a ~50-75 m thick package of parallel reflections grading from the chaotic character of the lower portions of this unit, and sometimes passing into their unit III. The gradational change of Unit IIa from chaotic to parallel suggests a change in the style of sedimentation of Unit IIa and fewer instances of deformation near the top of this unit. Yet, Eyles et al (1990, 1991) argue that deformation dominates the upper surface of Unit IIa where foundering and melting of ice blocks atop the lake floor produced the hummocky appearance of this surface. Inconsistent divisions of the valley fill also complicate interpretations. On some profiles, Unit III encompasses the entire package of parallel reflections near the top of the valley fill (fig. 9 in Eyles et al., 1991). In other cases, Unit III only encompasses a small portion of these parallel reflections (figs 10, 12 lower in Eyles et al., 1991), leaving a ~50-60 m thick package of unaccounted parallel reflections between the top of the chaotic facies of Unit IIa and the base of Unit III. More significantly, Eyles et al., (1990,

1991) do not mention the fact that Unit III is defined above an unconformity truncating the upper portion of Unit IIa. In addition, conflicting interpretations of Unit III, first associated with Holocene sedimentation in Okanagan Lake (Eyles et al., 1990) and later correlated with the White Silt benches (Eyles et al., 1991) raises questions about interpreted environmental changes recorded within the upper portions of the valley fill below Okanagan Lake.

Relationships between seismic facies and the distribution of White Silt benches raise further doubts about the environmental reconstruction of Eyles et al., (1991). The extent of grounded ice blocks on the lake floor (top of Unit IIa) is not explicitly stated by Eyles et al., (1991) though their figures (e.g., figs 16 B, 16 C in Eyles et al., 1991) suggest quasi valley-wide ice extent (approaching the extent of ice implicated in other supraglacial lake scenarios). If the top of Unit IIa is in fact correlative to the White Silt benches, it is difficult to reconcile the presence of continuous parallel reflections with large-scale ice block melt and sediment let-down. Even if concessions are made for the fact that figure 16 of Eyles et al. (1991) is conceptual and ice extent may have been more restricted, allowing for local preservation of parallel reflections near the top of Unit IIa, the sequence of seismic facies remains difficult to explain within the depositional model of Eyles et al. (1991). The difficulty arises from the fact that some seismic lines (e.g. line 15-16, fig. 9, Eyles et al., 1991), exhibiting parallel reflections, cross Okanagan Lake in areas where lacustrine benches occur on the valley sides. Based on the model of Eyles et al. (1991), lacustrine benches should be associated with ice blocks in the valley and with collapsed sediments (chaotic seismic facies) near the top of Unit IIa. Therefore the pattern of White Silt benches does not result predominantly from letdown of sediments deposited atop ice blocks.

4.8.3. Proglacial lake

In the northern reaches of the gLP basin, Vanderburgh and Roberts (1996) argued for proglacial lake development based on the absence of White Silt benches and on the longitudinal and valley-wide continuity of seismic reflections in the valley fill which showed no evidence of deformation that would be expected in a supraglacial lake setting.

In the southern reaches of the gLP basin, chaotic seismic facies dominate the valley fill and are inferred to record rapid deposition of sediments from subglacial tunnels (Unit IIa, Eyles et al., 1991). Eyles et al. (1991) argued that sediment deformation further contributed to the acoustically-transparent character of this unit. However, Vanderburgh and Roberts (1996), using drill core and seismic data, showed that acoustically-transparent seismic units (their unit IIa) can in fact contain poorly developed stratification and do not necessarily record deformation processes. Near the top of Unit IIa (Eyles et al., 1991), the gradational transition between the chaotic reflections and increasingly parallel and laterally continuous reflections are consistent with proglacial sedimentation as argued by Vanderburgh and Roberts (1996). This gradational change suggests that gLP developed as a proglacial lake following rapid deposition of the majority of the chaotic valley fill. An absence of unconformities near the top of Unit IIa (Eyles et al., 1991) suggests a gradual transition from subglacial sediment delivery to a proglacial basin (Eyles et al., 1991). Based on the paleocurrent data (§ 4.7.1), sediment was delivered mainly to the gLP basin via tributary valleys and possibly from the glacier front in the valley axis, though its exact position in the valley is unclear. Parallel seismic reflections near the top of Unit IIa (Eyles et al., 1991) suggest no extensive deformation of these sediments by buried ice blocks. However, an unconformity between Unit IIa and Unit III (Eyles et al., 1991) records erosion of a portion of Unit IIa and leaves open the possibility that some evidence of buried ice blocks has been removed. However, it can be stated that if buried ice was present within the lake sediments, it was likely of sufficiently minor proportions to leave no discernible evidence on seismic profiles of Eyles et al., (1990, 1991) (within the resolution limits of the data). Extensive ice masses were therefore not buried in the lacustrine fill.

4.8.4. Drainage of proglacial gLP and formation of the White Silt benches

If the evidence suggests that gLP developed as a proglacial lake, this interpretation must account for the distribution of White Silt benches on the valley sides. An event sequence for the evolution of gLP is proposed to specifically address the outstanding questions of gLP drainage – an issue ignored by previous lake evolution models – and the formation of White Silt benches.

Kvill (1976) suggested that erosion of lake-bottom sediments along the valley axis could potentially account for the distribution of the White Silt benches. His rationale was based mainly on the perceived implausibility of preserving a narrow and very thick ice tongue along the valley axis (Kvill, 1976). The present-day distribution of lacustrine benches may largely reflect erosion along the valley axis during drainage. Reservoirs (natural or man-made) impounded behind dams have significant erosive power during drainage events (Major et al., 2008). However, direct observations of the drainage of a lake similar in size and depth to gLP do not exist and analogues must be used to approximate the effects and processes occurring during drainage. The most appropriate analogues are documented erosion and fluvial adjustments following dam removal along rivers (Stewart, 2006; Major et al., 2008). Most of these reservoirs have a considerable sediment wedge forming a subaqueous basin floor that extends up valley along the reservoir length. One key difference may be that, in a man-made reservoir, this sediment wedge eventually grades to the alluvial plain of the feeder river upflow of the dam. This situation may not have been present in gLP where multiple sediment input points were present along the basin margins.

During drainage events, erosion occurs by knickpoint retreat. Two types of knickpoints (*stepped* or *rotating*) can develop as a function of the bed material and the critical shear stress for sediment entrainment (Holland and Pickup, 1976; Stewart, 2006). Specifically, the relative erosion rates at the top and base of the knickpoint face will determine which type of knickpoint develops (Holland and Pickup, 1976; Stein and Julien, 1993; Stewart, 2006). Rotating knickpoints (Stein and Julien, 1993) develop when erosion at the top of the face is greater than at the base. This results in decrease of the angle of the knickpoint face over time, and a concomitant decrease in erosional efficiency. In contrast, if erosion at the top of the knickpoint is slower than at its base, a stepped knickpoint will develop (Stein and Julien, 1993) and erosion will generally proceed more quickly than with rotating knickpoints (Stewart, 2006) through development of headward-retreating erosional notches.

Within a km-wide basin such as the gLP basin, characterized by a stratified and heterogeneous valley fill, stepped knickpoints could develop where coarse gravely beds or cohesive fine grained beds overlie sandy beds. In addition, multiple knickpoints may be active at a given time. The erosional efficiency of stepped knickpoints is potentially very high. In one of the rare studies in a natural setting, initial knickpoint retreat following dam breaching was on the order of 100s m per hour (Major et al., 2008), downcutting the former reservoir sediments over 10 m to produce remnant valley side benches in the former reservoir fill. It was noted that much of the widening occurred through slope failures as the knickpoint retreated and sediments were removed along the knickpoint path (Major et al., 2008). The scale differences between this study and the dimensions of the gLP basin complicate comparisons. Nonetheless, this study gives a 'first-order' impression of the potential for sediment erosion during gLP drainage, and suggests that this is a viable mechanism for removal of a significant portion of the lacustrine fill during drainage of gLP.

A number of factors favour efficient knickpoint migration along the axis of the gLP basin during drainage. Firstly, knickpoint migration speed increases with an increase in drop height (i.e. knickpoint relief) (Stein and Julien, 1993; Robinson and Hanson, 1996), so long as a stepped knickpoint can be maintained. Maintenance of a stepped knickpoint depends on the efficiency of sediment removal and the persistence of a plunge pool at the base of the knickpoint (Holland and Pickup, 1976). Given the fine texture of the lacustrine sediments, both of these requirements could be easily met as eroded sediment would be rapidly evacuated within the flow. Knickpoint geometry can be estimated from lake bathymetry and seismic profiles. The elevation of the water plane at the onset of drainage is taken as 500 m asl (elevation of Green Lake spillway, Figs 4.6,4.7). The lake bottom elevation is recorded by the top of the White Silt benches ~ 400 m asl (assuming limited erosion of this surface). The depth of erosion can be estimated from the height of the White Silt benches (~60 m) and the water depth (average 150 m depth, ~ 210 m asl), and sediment thickness to the top of seismic unit IIa of Eyles et al. (1991) (Max. ~15 m). Erosional depth is therefore ~ 225 m and corresponds to the base of the erosional surface produced by up-valley migration of the knickpoint. It is also representative of the maximum thickness of eroded sediments along the valley axis. This is likely an overestimation since beds of lacustrine sediments slope toward the valley axis where the total sediment thickness may have been less than 210 m. Furthermore, the retreating knickpoint(s) need not span the entire valley width between the White Silt benches. Knickpoints can occupy only a fraction of the valley width, especially during the early stages of drainage when knickpoints incise rapidly into the sediments and later widen (Straub, 2007). A significant portion of erosion and widening of the eroding lake floor takes place through slope failures and sediment flushing as a result of knickpoint migration during drainage.

A more critical question relates to the initial drop height at the point of dam failure, and persistence of this height along the valley in order to maintain erosional efficiency. An isolated bedrock outcrop (~348 m asl) occurs near the valley axis at Okanagan Falls (Fig. 4.2) in the northern extent of the postulated dam site. The presence of a bedrock sill within the gLP drainageway would limit knickpoint retreat during drainage. However, this outcrop is localized and water well logs immediately to the east of the valley axis at Okanagan Falls reveal that bedrock occurs ~275 m below the modern valley floor (~ 65 m asl) (well #24129, BC Water Resource Atlas), providing ample depth for knickpoint initiation and migration. In addition, the great bedrock depth is maintained south of McIntyre Bluff (Toews, 2007), and provides an efficient throughway for eroded sediments.

Seismic surveys reveal evidence of erosion through gLP sediments. For example, lacustrine terraces in Skaha Lake and south Okanagan Lake were identified by St John (1973). These terraces occur ~25 m below the modern lake plane but their origin is not clear. It is conceivable that they result from erosion of the lake bed during drainage. A prominent unconformity occurs near the top of seismic Unit IIa (Eyles et al., 1991) and truncates parallel reflections near the top of the valley fill. This unconformity may mark the base of knickpoint erosion during gLP drainage. Finally, normal faults within the White Silt benches are associated with removal of lateral support during lake bed erosion, rather than with melting and collapse of valley-axis ice (cf. Flint, 1935b).

4.8.5. Distribution of lacustrine sediments in Okanagan Valley

Incision of the lake floor during drainage of gLP explains the distribution of the White Silt benches south of Squally Point. However, lacustrine benches are largely absent between Peachland Creek and Fintry Creek (with the exception of the Kelowna basin and Kalamalka Valley, and some isolated benches near Powers Creek) (Fig. 4.2). Does their absence result from complete erosion or from non-deposition along certain parts of Okanagan Valley? The possibility of limited lake sediment deposition is inherent in depositional models of Shaw (1977) and Shaw and Archer (1978). However, lake-based seismic surveys reveal that parallel reflections near the top of seismic Unit IIa occur throughout Okanagan Valley and indicate that glaciolacustrine sedimentation did not preferentially occur south of Squally Point. Regionally-correlated tributary mouth deltas record basin-wide sediment inputs and are consistent with the inference of lacustrine sedimentation throughout the gLP basin.

Valley depth to bedrock and the density of tributary valleys along segments of Okanagan Valley offer an explanation for the distribution of lake sediments. A first-order explanation could relate to bedrock depth and accommodation space (valley width remains fairly constant so it is not considered in this discussion of accommodation space) within portions of Okanagan Valley: lacustrine benches would occur along the shallowest bedrock floor where decreased accommodation space leads to sediment aggradation to the greatest elevation. The greatest overdeepenings occur between Peachland Creek and Fintry Creek (~640-450 *below* sea level, Eyles et al., 1990) (Fig. 4.2). Shallower bedrock depths (200 m below sea level, Eyles et al., 1990) occur near Trepanier Ck (Fig. 4.2) but no lacustrine benches occur in this area. In contrast, bedrock depths of ~400 m below sea level (Eyles et al., 1990) are maintained along the valley segment where the White Silt benches occur. Therefore, bedrock depth alone cannot be the sole controlling variable on the occurrence of lacustrine benches.

Sediment supply may control the distribution of silt benches. Six major tributary valleys (Penticton Creek, Ellis Creek, Shingle Creek, Penticton Diversion, Eneas Creek Naramata Creek) and a number of smaller bedrock channels occur within the ~20 km-long valley segment containing the White Silt benches (Figs 4.2, 4.8). Sediment delivery from these tributaries and possibly backwasting ice in the valley axis to the north was sufficient to allow aggradation of the White Silt benches (despite significant bedrock depth). Lacustrine benches also occur on the valley sides and the valley floor near Vernon where large tributary valleys converge (Equesis Creek, Salmon River, Fig. 4.2). Similarly, four major tributary valleys (Lebanon Creek, Bellevue Creek, KLO Creek, Mission Creek, Fig. 4.2) and decaying ice in Kalamalka Valley (Paradis et al., 2010) delivered sediment into the Kelowna basin where bedrock depth is 250-275 m above sea level (i.e. at least ~400 m shallower than the shallowest portions along the axis Okanagan Valley) (Pugin and Pullan, 2009 in Paradis et al., 2010). In contrast, no other valley segment contains the same density of tributary valleys as that near the White Silt benches. Therefore, sediment delivery from tributary valleys likely controlled basin filling and thus the production of benches during lake drainage. This is well illustrated by the absence of benches in the valley segment between Peachland Creek and Powers Creek (Fig. 4.2) where no major tributary valleys drain the eastern valley side (Squally Point) and only Trepannier Creek drains from the western plateau to Okanagan Valley. Similarly, lacustrine benches are absent on the valley sides in the 40 km-long valley segment between Lambly Creek and Equesis Creek (Fig. 4.2) because only three tributary valleys delivered sediment into Okanagan Valley from the west. Ellison Ridge blocked drainage from Kalamalka Valley and its uplands, limiting additional sediment input. Lastly, steep valley side slopes may have further affected valley-side accumulation of lacustrine sediments. Valley side slopes in the area where White Silt benches occur are gentler than further north where benches are absent. Steep valley-side slopes may have favoured mass movements and, along with limited sediment supply, may have prevented significant accumulation of silt on the valley sides.

4.9. Water sources for sediment transfer and ice distribution within gLP basin

4.9.1. Ice in the uplands and on the plateaus: extent and deglacial evolution

Tributary valley trains, deltas and subaqueous fans, and gLP paleocurrent data record water and sediment transfer from the uplands and plateaus to the valley bottom via tributary streams. Regionally correlated deltas indicate hydrologic connectivity between tributaries. Tributary sediment transfer was coeval with gLP. Systematic basin wide thinning and fining patterns in lake bottom rhythmites (Fulton, 1965; Shaw, 1977; Shaw and Archer, 1979) suggest a climate-driven deglacial scenario where sediment transfer decreases over time. Consequently, it is argued that meltwater transferred sediment along the tributaries, which requires dwindling ice in the uplands and plateaus around Okanagan Valley while gLP occupied the valley bottom. However, it is also conceivable that the thinning and fining patterns partly reflect the presence of a valley-axis ice front. Although valley-axis ice does not negate the possibility of tributary inputs, it could in places dominate over tributary sources. The prevailing deglacial model (Davis and Matthews, 1944; Fulton, 1991) suggests that plateaus were largely ice-free as gLP developed and substantial meltwater production would not be expected from upland/plateau areas. Two possible scenarios arise from this situation: 1) ice was in fact present on the plateaus while gLP occupied the valley, requiring a reassessment of the regional deglacial model, or 2) alternative sources of water need to be considered.

A possible alternative water source involves abundant rainfall or snowmelt during a moist climatic period near the end of deglaciation. At this time, hillsides would be free of vegetation, and sediments could be easily mobilized (cf. Church and Ryder, 1972). Paleoclimatic and sedimentologic factors argue against this possibility however. Paleoclimatic records suggest a cool climate at the end of the last glacial period followed by a transition to a warming but dry climate between 10-12.3 k cal a BP (Heinrichs et al., 1997; Walker and Pellatt, 2008) during which time gLP drained (drainage complete by 11 ± 1ka BP, Lesemann et al., in review). A cooler and moister climate followed until ~4 k cal a BP. A warming deglacial climate would be conducive to rapid ice sheet decay and abundant meltwater production. Indeed, this process was invoked by Fulton (1991) who postulated that a rapid rise in glacier equilibrium line altitude was responsible for decay of the most recent CIS. The thinning and fining pattern in lake sediments is consistent with this scenario. Lastly, sediment volumes near the top of the valley fill also favour meltwater over precipitation sources. Based on paleoclimatic data (Heinrichs et al., 1997; Walker and Pellatt, 2008) and the timing of lake drainage, up to ~ 75-90 m of sediments (top of Unit IIa, Eyles et al., 1991) were deposited under warming but dry conditions. In contrast, only up to ~30 m (Unit III. Eyles et al., 1991) were deposited under a moister climate since deglaciation. Close to three times more sediments were deposited in Okanagan Valley under drier conditions where precipitation was presumably suppressed (though the possibility of flash floods under a dry climate regime may make these arguments equivocal). Although sediment availability does affect rates of sediment transfer, the abundance of sediments stored in tributaries suggests that this may not have been a limiting factor for sediment transfer. For these reasons, variations in meltwater drainage best account for the valley fill characteristics and volume. Decaying ice on plateaus remains the most viable explanation for sediment transfer along tributaries. Some episodic rainfall events might occur during deglaciation and be responsible for some high-magnitude sediment transport events (e.g. Denner et al., 1999; Lawson et al., 1998), though they would likely be of subordinate importance over the time scale of deglaciation.

4.9.2. Ice in Okanagan Valley: extent and abundance

Preservation of valley ice is predicted in the prevailing deglacial model (Kerr, 1934; Davis and Mathews, 1944; Fulton, 1991). Therefore, a more pertinent paleoenvironmental question focuses on the extent of ice within the gLP basin. Existing paleoenvironmental reconstructions (Nasmith, 1962; Fulton, 1969; Shaw, 1977; Shaw and Archer, 1978) have mostly assumed the presence of extensive stagnant ice tongues in the valley bottom during deglaciation. Nasmith (1962) interprets many tributary valley benches as kame terraces based on this assumption and the presence of kettle holes within these benches. Kettle holes also occur in some perched tributary deltas (e.g., Fintry Creek, Fig. 4.2) and in the extensive sediment fan downflow of the Penticton Diversion (Fig. 4.8). Although these kettle holes record deposition over/around ice blocks, their discontinuous occurrence and scale are insufficient to infer regional stagnation of valley-scale remnant ice tongues. More extensive stagnant ice may have been preserved around the Trout Creek basin (Fig. 4.13B). Again, however, this is a relatively small area and stagnation in this location may have been favoured by the geometry of this basin and the presence of multiple bedrock outcrops that could trap ice within the basin.

Interpretations of gLP as a proglacial lake (§ 4.3.2.3) imply the presence of an ice front within Okanagan Valley. However, the timing of ice frontal positions is unknown and potentially highly variable depending on interpretations of valley landforms and sediments. For example, valley-parallel paleocurrents in the White Silt benches south of Eneas Creek delta (P2, Fig. 4.8, § 4.7.1) might require the presence of Okanagan Valley ice in this southern portion of the gLP basin. However, a large proportion of paleocurrent data (sites P5 and P2 excluded) can be rationalized within a model of tributary sediment delivery (§ 4.7.1.2). Evidence for Okanagan valley-axis ice is therefore equivocal and not regionally consistent.

Paradis et al., (2010) inferred the presence of a valley glacier in Okanagan Valley based on mapping of ice-contact outwash terraces (including some tributary mouth deltas) near Peachland Creek and Trepanier Creek (Fig. 4.2). These kame terraces also occur as narrow subhorizontal surfaces flanking the valley sides and sometimes merge with the perched delta surfaces at tributary mouths. Ice contact interpretations rely on two lines of evidence: the steepness of terrace (delta) foreslopes, suggestive of lateral ice support, and the presence of kettle holes within deltas. However, terrace foreslopes are not oversteepened and are near or below angle of repose for gravely material within the deltas. In addition, the valley-ward extent of these deltas is partly controlled by the basin geometry (§ 4.7.2) that explains the abrupt termination of some delta fronts. As well, merging of these delta surfaces with so-called kame terraces give a false impression of their steepness and lateral continuity. This is mainly because sediments within kame terraces fill linear structurally-controlled bedrock channels aligned with the valley walls. In places, this fill is only a sediment veneer overlying bedrock The apparent steepness of their foreslopes is therefore strongly influenced by the underlying bedrock. Channel fills can be linked to bedrock meltwater channels draining from the surrounding uplands. Thus, the meltwater and sediment source for these terraces is from decaying ice in the uplands and valley sides, not from ice marginal drainage along Okanagan Valley. The second line of evidence relying on kettle holes has been discussed above and is equally applicable in this situation. Some tributary valley benches (Fig. 4.10) occurring above the highstand of gLP may be associated with ice-contact deposition or local lakes dammed by ice in Okanagan Valley. However if these surfaces do record ice-contact deposition, their occurrence ~50 m above the inferred stillstand of gLP does not require extensive stagnant ice in these portions of Okanagan Valley.

Lastly, Nasmith (1962) mapped moraine ridges (interpreted as likely lateral moraines) atop sedimentary surfaces located immediately below the bedrock valley walls between Mission Creek and Bellevue Creek (Fig. 4.2). Although these features were used to infer the presence of active ice within the Kelowna area, re-examination of these landforms reveals that these interpretations are tenuous and reflect a preconceived notion of ice marginal retreat. For example, Nasmith's (1962) interpretations are based entirely on landform identification, without supporting sedimentologic evidence. Nasmith (1962) hypothesized that some of these ridges might contain stratified drift if they were deposited within an ice-contact basin. However, no evidence supports this hypothesis nor is there evidence that moraine ridges are composed of diamicton or colluvial material, as might be expected for lateral moraines. The origin of these landforms is therefore unclear. Interpreting these landforms as moraines implies that they are constructional features. However, based on air photo interpretation and cursory field observations, the ridges are in fact defined by channels incised within the sedimentary body. The 'moraine' ridge morphology may therefore be an erosional remnant rather than a depositional feature. Consequently, ice marginal positions cannot be confidently inferred from the characteristics of the East Kelowna sediment bench.

Landform evidence for valley ice is most abundant on the valley sides in the northern portion of Okanagan Valley (near Fintry) and along Kalamalka Valley. An esker near the Fintry delta (Fig. 4.2) (Paradis et al., 2010) suggests the presence of ice near 560-580 m asl. It is oriented ~parallel to slightly oblique to Okanagan Valley. However, it's limited length (few 100s m) precludes any regional reconstruction of ice distribution. An absence of moraines further limits reconstructions of glacier geometry in this area. It is therefore unclear if this esker developed from a decaying valley ice tongue or if it was associated with more extensive glacier complexes extending toward the uplands. Additional eskers, small discontinuous moraine ridges (100s m length), and some meltwater channels on the eastern valley sides of Kalamalka Valley (Nasmith, 1962; Paradis et al., 2010) further record the presence of ice around this portion of the gLP basin.

Lake-based seismic surveys potentially offer the most conclusive evidence for an ice frontal position in Okanagan Valley near Fintry (Fig. 4.2). These data reveal a distinctive interdigitate transition between seismic Units IIa and IIb (St John, 1973; Eyles et al., 1991). Unit IIa dominates the valley fill south of this transition. However, unlike many coastal fjords affected by glacier margin fluctuations, neither the lower portion of Unit IIa nor its gradational transition to increasingly parallel reflections associated with sedimentation in gLP shows evidence of glacier grounding and marginal fluctuations (within the limitations of the data resolution). Glacier margin fluctuations produce distinctive wedges of deformed sediments, thrust sheets, interdigitate patterns, and subaqueous fans at grounding lines (Lonne and Syvitski, 1997; Fiore et al., 2011). None of these features appear on seismic profiles. Therefore, if deposition of the bulk of Unit

IIa was accompanied by backwasting of the valley glacier as proposed by Eyles et al., (1991), then retreat must have occurred without much interruption or episodic stillstands. Calving could favour rapid glacier retreat (Eyles et al., 1991). However, the lateral continuity and undisturbed character of parallel reflections at the top of Unit IIa argue against the possibility of calving and *en masse* stranding of ice blocks on the lake floor (cf. Eyles et al., 1991) (§ 4.9.2). Similarly, a paucity of dropstones in the White Silt benches argues against iceberg armadas in gLP. However, the White Silt benches afford a spatially-limited view of the former lake basin fill and dropstones could have been present within the portion of the lake basin that was eroded, and in the gLP sediments that now underlie Okanagan Lake. Kettle holes in some tributary deltas and in the delta downflow of the Penticton Diversion might indicate that some icebergs were present in gLP, though these kettle holes could also result from remnant ice blocks trapped by topography. Evidence for calving in gLP is therefore equivocal. If calving did occur, the paucity of dropstones in the White Silt benches suggests that it was a minor component of ice retreat and/or bergs carried little sediment. This last option could result from extensive scouring of basal sediments during subglacial deposition of Unit IIa (Eyles et al., 1991; Lesemann and Brennand, 2009; Chapter 3).

Consequently, landforms and seismic data suggest that, during existence of gLP, there was a limited volume of ice in Okanagan Valley (at least over the area where Unit IIa occurs). What volume of ice did occur was probably not grounded on the lake floor. If ice blocks were buried within the sediments of seismic Unit IIa, they were likely of limited extent since their melting has not led to significant disturbance of parallel reflections at the top of Unit IIa. Although Eyles et al. (1991) do not state the resolution of their seismic data, seismic profiles resolve ~7-10 parallel reflections within a sediment thickness of ~100-120 m near the top of Unit IIa. Within these data, no significant evidence of bed disturbance can be seen and this suggests that if buried ice blocks were present, any deformation associated with melting probably affects less than 12-15 m of sediments.

It is only near Fintry that evidence for grounded ice in the valley is more conclusive. The pronounced differences in the seismic character of Units IIa and IIb indicate different depositional styles (St John 1973; Eyles et al., 1991). The dominance of parallel reflections in Unit IIb to the north of the interdigitate transition zone is similar to the seismic character of proglacial sedimentation behind grounding line fans in coastal fjords (Lonne and Syvitski, 1997). Unit IIb could therefore have developed as ice retreated northward from its grounding line and sediments filled this local proglacial sub-basin. The fining upward trends in a portion (Seismic Units II upper, IIIa, and IIIb, Vanderburgh and Roberts, 1996) of the valley fill north of Okanagan Lake reported by Vanderburgh and Roberts (1996) could represent increasingly distal depositional conditions within this sub-basin. Grounded ice near Fintry is also consistent with observations of eskers, meltwater channels, and moraines in this portion of the valley and with the inference of a glacier front at a similar latitude in Kalamalka Valley where ice-contact deltas developed against the glacier margin in a proglacial lake (Paradis, 2009). Overall, these data suggest limited extent and influence of grounded ice in the gLP basin. Where grounded ice did occur, its retreat was coeval with gLP expansion in order to allow for development of a regional water plane controlling tributary delta elevations throughout the gLP basin.

4.9.3. Seasonality of sediment delivery recorded in lake sediments and lake duration

A seasonal signal of sediment delivery could be expected within the lacustrine sediments. However, evidence of seasonal sedimentation is not clear within most lacustrine benches. Although Fulton (1969) and Shaw (1977) identified thinning and fining patterns, and recognize rhythmic sedimentation patterns in glaciolacustrine sediment benches, the seasonality of these rhythmites has not been conclusively established. Fulton (1969) assumed that rhythmic couplets recorded seasonal sediment input and inferred lake duration of a few hundred years by counting sediment couplets (treating the rhythmic couplets as varves). However, Shaw (1977) and Shaw and Archer (1978) recognized that thin beds of silt-clay couplets were punctuated by sand in the capping clay. Although clay occurs in the upper metres of lacustrine benches, it is generally absent from the bulk of gLP sediments. One exception occurs near Summerland (Fig. 4.2) where clay beds and laminations occur throughout a 32 m high section (cf. Shaw 1977). The selective occurrence of clay within lacustrine benches indicates that this material was present in the lake basin. The absence of clay in most exposures therefore does not reflect a supply limit. Instead, it probably results from non-deposition of fines due to prevailing conditions within the lake basin. Johnsen and Brennand (2006) noted similar selective clay deposition in glacial Lake Thompson and attributed this to a range of possible factors including impeded suspension settling due to strong stratification of the water column by sediments, an energetic lacustrine environment due to multiple sediment input points, flushing of suspended clay during lake drainage, and deposition of the clay fraction within turbidites. Evidence of seasonal deposition is therefore difficult to infer from the beds of the White Silt benches.

In contrast, seasonality may be more easily recognized where clay occurs in the upper few metres of the lacustrine benches. Shaw and Archer (1978) recognized a peculiar depositional pattern where sand beds and partings were frequently intercalated within the 'winter' clay layer (Shaw 1977, Shaw and Archer, 1978). These authors interpreted these 'winter sands' as turbidity currents triggered by delta front collapse during winter. This sedimentation model assumes the presence of a supraglacial delta in the valley axis and although evidence for a valley-axis delta is equivocal (§ 4.7.1), tributary deltas could certainly act as sediment sources for 'winter sands'. However, are 'winter sands' and their bracketing clayey beds necessarily indicative of quiescent depositional conditions during winter?

Varve-like sediments can also be produced by a range of processes such as suspension deposition from mud-rich interflows, the distal components of low density turbidity currents (Pickering et al., 1986) or sustained turbidity currents (Lambert and Hsu, 1979). These mechanisms transfer sediments at or near the lake bed and result in shorter settling heights than sediment inputs by overflows. More importantly, these processes have no inherent seasonal control so that multiple beds could be produced during a single season (Etienne et al., 2006). More recently, the importance of grain flocculation on the settling velocity of clay particles has been recognized in glacial lakes (Hodder and Gilbert, 2007; Hodder, 2009). The settling velocity of macroflocs approaches that of larger clastic particles. Consequently, macroflocs settle clay particles to the lake floor at a higher rate than individual clay grains (Hodder, 2009) and clay accumulation may occur over short time spans and without complete shut-down of the hydrologic system delivering sediments. Lastly, laboratory experiments on settling of clays and silty clay mixtures indicate that these sediments can, in a few hours, consolidate to a degree where they may resist erosion (Hawley, 1981). If winter sand deposits are examined within the context provided by these data, clay beds underlying the winter sands might not necessarily require 'winter' conditions for their deposition.

These arguments cast doubt on the lines of evidence used to argue for seasonal lacustrine sedimentation. A corrolary interpretation is that clay beds record the end of underflow depositional events (cf. Johnsen and Brennand, 2006). In some cases, dm- to m-thick beds likely record high energy events delivering high sediment volumes to the lake basin, especially during the early stages of lake development (Fulton, 1969; Shaw, 1977). As well, thinning and fining of beds toward the top of the White Silt benches indicates that sediment volume decreased over time, presumably in response to decreasing meltwater volumes and/or sediment exhaustion, though this latter possibility seems less likely given the abundance of sediments stored in tributaries. Given the inferred meltwater sources for sediment transfer to the gLP basin, seasonal controls on sedimentation would be expected, in step with the seasonal meltwater regime of the decaying CIS. However, additional factors could obscure any seasonal signal present within the sedimentary record. For example, glacier melt responds to diurnal temperature cycles, and more episodic events such as rainstorms (Marren, 2005). In addition, multiple tributaries, contemporaneously delivering sediment to the same basin as underflows and overflows create a complex spatio-temporal depositional environment where seasonal signals might be dampened and confounded with short-lived events. Difficulties in ‘varve’ correlations in the gLP basin (Kent, 1978) were in part attributed to these multiple sediment input points. Given this complexity, it is questionable whether any meaningful inferences can be made regarding the duration of gLP from rhythmites.

4.10. Conclusions

- Late Wisconsinan deglaciation of Okanagan Valley led to development of glacial Lake Penticton. A number of conclusions on the paleogeography and paleoenvironment of this lake can be drawn:
- Glacial Lake Penticton (gLP) developed as a proglacial lake impounded by an ice and sediment dam occupying a ~5 km long valley constriction (McIntyre Bluff).
- Regionally-correlated perched deltas developed at the mouth of tributary valleys. In association with bedrock spillways near the dam site these deltas define the only stable water plane for gLP within an elevation envelope of 500-

525m asl. Earlier propositions of multiple lake stages are rejected based on re-examination of field evidence.

- Despite evidence for glacioisostatic rebound in coastal British Columbia and in other glacial lake basins of interior British Columbia, no clear evidence of glacioisostatic rebound can be recognized in the gLP basin. This is attributed to either rapid glacioisostatic rebound controlled by low viscosity mantle conditions occurring prior to development of landforms used for rebound reconstructions, and/or to potentially uniform rebound over the gLP basin due to thin valley ice of relatively uniform thickness
- At its maximum extent, gLP covered an area of ~1220 km^2 and contained a water volume of ~103 km^3.
- An absence of deltas below the 500-525 m asl (save for modern deltas graded to Okanagan Lake level) suggest a single, possibly catastrophic, drainage event likely by hydrofracture and tunnel enlargement. A possible unconformity within the upper portion of the valley fill suggests that the lake drained fully and eroded the lake floor. The resulting outburst of water was routed along Okanagan Valley and likely drained into the Columbia River in Washington State (Chapter 3).
- Erosion accompanied lake drainage and the present distribution of White Silt benches reflects lake bed erosion by knickpoint retreat, and post-erosion slope failures of lacustrine benches.
- A substantial portion of the gLP sediments were derived from tributary valleys. Sediments were transferred by meltwater originating from dwindling ice masses in the uplands and plateaus surrounding the gLP basin.
- Limited evidence for valley-axis ice is present in the gLP basin. The only exception occurs in the northern portion of the basin where ice was likely grounded at some stage during evolution of gLP.
- Coeval plateau ice and limited valley ice raise questions about the applicability of the prevailing CIS growth and decay model which suggests that plateaus were ice-free while large ice masses stagnated in valley-bottoms.

5. Conclusions

A number of research objectives were outlined in order to refine understanding of subglacial processes and glacier dynamics operating during CIS build-up and spanning glacial maximum, and deglacial processes and patterns associated with CIS decay. Research findings are summarized below.

5.1. Chapter 2 objectives and findings

Within Chapter 2 describing the development and operation of linked cavities at the ice-bed interface on Thompson Plateau, research objectives were to:

1. Reconstruct paleo-CIS bed environments and (sub)glacial processes within an area of streamlined bedforms

Findings indicate that, within an area of streamlined bedforms (drumlins) on the Thompson Plateau, sediments within drumlins, in areas between drumlins, and in non-drumlinized zones record a complex suite of subglacial processes where fluctuations in porewater pressure play a primary role in substrate genesis. Despite the occurrence of drumlins on the Thompson Plateau, pervasive substrate deformation was not a dominant and process. However, localized deformation is recorded in diamictons. Three lithofacies record a broad spectrum of subglacial erosional and depositional processes including evidence of lodgement, deformation, and suspension settling of fines within proglacial and/or subglacial cavities. The three lithofacies consists of 1) *consolidated diamictons* recording subglacial processes such as abrasion and plucking associated with clast ploughing and lodgement, and the variable influence of sediment deformation; 2) *poorly-consolidated diamictons* resulting from local deformation under elevated porewater pressure conditions. These diamictons have a mud-poor matrix and this is attributed in part to lithologic controls and possibly to washing and elutriation of the matrix by porflow and pipeflow. Stratification within these sediments is associated with remobilization by debris flow processes; and 3) *sorted and stratified sediments* deposited in proglacial and/or subglacial water-filled cavities. The possibility that some of these ponds instead form during deglaciation cannot be completely discounted.

2. Explore the implications of inferred bed conditions on our understanding of paleo-CIS dynamics

The sedimentology of the Thompson Plateau substrate suggests high subglacial water pressures within bed materials, potentially leading to the operation and evolution of a linked-cavity subglacial drainage system. These subglacial water-filled cavities were up to 10's of metres (and possibly larger) in diameter and dm's to m's deep. In terms of glaciodynamics, subglacial water-filled linked cavities in the sedimentary basin of the southern Thompson Plateau likely decreased bed friction and would have promoted sliding (e.g., Lliboutry, 1983). Such effects may have facilitated enhanced flow toward Trepanege Ridge where ice would have thickened. Although linked cavity systems have been implicated as a bed condition favouring rapid ice flow (e.g., Kamb, 1987), it is unlikely that prolonged rapid ice flow occurred over this portion of Thompson Plateau. This is likely due to the presence of Trepanege Ridge which impeded ice flow by blocking ice discharge south of the regional basin.

5.2. Chapter 3 objectives and findings

Within Chapter 3, consisting of a regional reconstruction of subglacial hydrology and glaciodynamic behaviour along the southern margin of the Cordilleran Ice Sheet in British Columbia, Canada and northern Washington State, USA, stated objectives were to:

1. Evaluate landscape development, regional glaciodynamics and hydrologic conditions along the southern margin of the CIS

Landsystems of the Thompson Plateau–Okanagan Valley region area record evidence of erosion and deposition by subglacial meltwater underburst(s). Proposed underburst(s) occur in an area previously identified as a potential corridor for rapid ice flow. However, this prior proposition of ice streaming relied on the presence of drumlins, and physiographic characteristics such as overdeepened bedrock troughs that resemble some modern ice stream settings. Landforms include drumlins of various shapes composed of sediment and bedrock, and tunnel valleys eroded in bedrock and sediment.

2. Evaluate hypotheses of landform genesis and present an event sequence of landscape evolution

Prevailing genetic explanations for some landforms on Thompson Plateau and in Okanagan Valley (mainly drumlins) focus on sediment deformation, but are unable to account for the composition of drumlins, their morphology and spatial arrangement, the morphology of tunnel valleys and their composition and fill, landform associations and sediment flux conservation. A proposed unifying landscape model involves erosion and/or deposition by subglacial meltwater underbursts. These processes account for drumlin genesis, the spatial patterns and association of drumlins with bedrock valleys, and thick sediment fills in valleys of interior British Columbia.

3. Investigate the implications of landform genesis and landscape evolution for CIS hydrology, glaciodynamics and geometry

The meltwater underburst model of landscape genesis put forth in this chapter raises questions about the subglacial hydrology of the CIS, particularly subglacial storage and release of meltwater: large meltwater reservoirs are required to store and release water. A subglacial lake in Okanagan Valley offers a potential reservoir site and mechanism for meltwater storage. Geodynamic conditions (thermal springs and subglacial volcanic eruptions) are favourable conditions for subglacial lake development and are possible underburst triggers. In addition, pronounced regional sags in the ice surface, resulting from regional geodynamic events could have become the sites for supraglacial lakes, the drainage of which could have facilitated subglacial lake growth and/or triggered subglacial lake drainage. Although the scale of these inferred events far exceeds any documented processes in modern ice sheets, the hydrologic conditions and processes are consistent with our understanding of modern subglacial hydrological conditions in Iceland, Greenland and Antarctica.

In terms of glaciodynamic behaviour, the Thompson Plateau and Okanagan Valley landsystems may (indirectly) record rapid ice flow: (i) short-lived localized accelerations, akin to surges, occurring during underbursts, and (ii) longer-lived accelerations during subglacial lake build-up. In both cases, glacier acceleration results from loss of resistance to flow, not sediment or ice deformation.

Interpretations of the Thompson Plateau and Okanagan Valley landsystems also have repercussions for understanding of CIS geometry and ice volumes. Subglacial lake development and transient rapid flow can redistribute ice mass within an ice sheet. In the CIS, preferential ice flow over valleys (and the presence of lakes there) would thin valley ice, influencing patterns and processes of deglaciation. Thin valley ice and comparatively thicker plateau ice could lead to nearly coeval deglaciation of valleys and plateaus, or even deglaciation of valleys before plateaus. This deglacial pattern is a potential reversal of the prevailing model of CIS decay in southern British Columbia. It suggests that endogenous processes may partly control CIS decay, in addition to exogenous variables like climate forcing.

5.3. Chapter 4 objectives and findings

Within Chapter 4, examining the paleogeographic and paleoenvironmental evolution of glacial Lake Penticton (gLP) during decay of the Cordilleran Ice Sheet in British Columbia, Canada, stated objectives were to:

1. Document gLP paleogeography by focusing on the geographic extent of gLP in Okanagan Valley through regional water plane reconstruction

A single regionally-correlated water plane is identified in the gLP basin at an elevation of 500-525 m asl. This water plane is constrained by perched deltas developed at tributary valley mouths and by bedrock spillways near the dam site. Earlier propositions of multiple lake stages are rejected based on re-examination of field evidence.

2. Assess the evidence for glacioisostatic rebound in Okanagan Valley and examine the glaciodynamic conditions responsible for glacioisostatic rebound patterns

Despite evidence for glacioisostatic rebound in coastal British Columbia and in other glacial lake basins of interior British Columbia, no clear evidence of glacioisostatic rebound can be recognized in the gLP basin. This is partly attributed to rapid glacioisostatic rebound controlled by low viscosity mantle conditions occurring prior to development of landforms used for rebound reconstructions, and to potentially uniform rebound over the gLP basin due to thin valley ice of uniform thickness.

3. Examine the the evolution of gLP, focusing on lake extent and volume, controls on lake level, and drainage of gLP.

At its maximal extent, gLP covered an area of ~1220 km^2 and contained ~103 km^3 of water. A single (catastrophic) drainage event dropped lake level as much as ~210 m (to an elevation of 320 m asl). Drainage led to erosion of a portion of the valley fill. Near Penticton, tributary alluvial fans prograded across Okanagan Valley and dammed the valley to form the modern basins of Okanagan Lake and Skaha Lake.

4. Reconstruct paleoenvironmental conditions in and around the gLP basin by examining water and sediment delivery pathways to the gLP in order to better constrain lake type, the distribution of ice in and around the gLP basin, and the style of lake drainage and its effects within the lake basin.

A substantial portion of the gLP sediments were derived from tributary valleys. Sediments were transferred by meltwater originating from dwindling ice masses in the uplands and plateaus surrounding the gLP basin. Limited evidence for valley-axis ice is present in the gLP basin. The only exception occurs in the northern portion of the basin where ice was likely grounded. Coeval plateau ice and limited valley ice raise questions about the applicability of the prevailing CIS growth and decay model which suggests that plateaus were ice-free while large ice masses stagnated in valley-bottoms.

Glacial Lake Penticton (gLP) developed as a proglacial lake (possibly ice-contact in the northern portion of the basin) impounded by an ice and sediment dam occupying a ~5 km long valley constriction (McIntyre Bluff). An absence of deltas below the regionally-correlated water plane suggests a single, possibly catastrophic, drainage event likely by dam burst and tunnel enlargement. The resulting outburst of water was routed along Okanagan Valley and likely drained into the Columbia River in Washington State. Erosion accompanied lake drainage and the present distribution of White Silt benches reflects lake bed erosion by knickpoint retreat, and post-erosion slope failures of lacustrine benches.

5.4. Future avenues of research

As part of this research a number of potential future avenues of research were identified:

- **Seismic surveys on Okanagan Lake**. By far, the most complete and potentially most informative sedimentary archive of environmental changes, glacier processes and deglacial processes and patterns lies in the valley fill of Okanagan Valley. Existing seismic surveys have barely begun to lift the veil on this archive (Eyles et al., 1990, 1991, Vanderburgh and Roberts, 1996). Unfortunately, the resolution of existing data makes it unlikely that more meaningful information will be extracted from these datasets. Consequently, new lake-based seismic surveys could be performed in order to better resolve individual bed characteristic, the characteristics of the valley fill near the bedrock surface, and the nature and extent of unconformities. Lastly, shallow targeted seismic surveys could attempt to resolve the along-valley changes in seismic facies, particularly those in the upper portion of the valley fill in order to elucidate the potential links with the White Silt benches.
- **Coring/drilling of the valley fill**. If seismic surveys offer a regional view of basin architecture, interpretations of this architecture are tenuous without the corroborating evidence of exposures and/or drill cores. Drill cores used to support land-based seismic surveys have shed light on valley fill characteristics (Vanderburgh and Roberts, 1996). Similar corroborating datasets should be used to truth the lake-based seismic data. Unfortunately, this is a monumental task given the great water depth of Okanagan Lake, and the immense thickness of the valley fill in places. There is a paucity of equipment capable of drilling through such a thickness of sediments. Equipment capable of drilling to these depths is typically used by sea-going research vessels. Only one ‘portable’ drilling system (DOSECC) is currently capable of drilling at these depths within inland lakes.
- **Refine the chronology of deglaciation on plateaus and in Okanagan Valley**. Currently, newer dates on the drainage of gLP (Lesemann et al., in

review) suggest that plateaus and valleys deglaciated nearly at the same time (10.5-11 k cal a BP). Refined chronologic controls on deglaciation would help to test the validity of the proposed CIS growth and decay model.

- **Examine the sedimentology of tributary valley fills above the elevation of the gLP highstand**. Development of a subglacial lake in Okanagan Valley was hypothesized in this thesis. This hypothesis could be tested by examining the characteristics of tributary valley sediments in their upper reaches and near the plateau edge.
- **Geochemical fingerprinting of underburst/outburst sediments**. If a subglacial underburst is in fact responsible for regional erosion in Okanagan Valley and its surrounding plateaus, then eroded sediments were likely flushed beyond the CIS margin and through the Columbia River system. Similarly, drainage of gLP may also have removed some of the valley fill. These eroded sediments may have travelled toward (and possibly into) the Columbia River. Geochemical fingerprinting of sediments along the Columbia River could potentially shed light on the volumes of sediments removed from Okanagan Valley, and the depositional style of these sediments beyond the ice margin (cf. Zuffa et al., 2000, Prytulak et al., 2006).

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Appendices

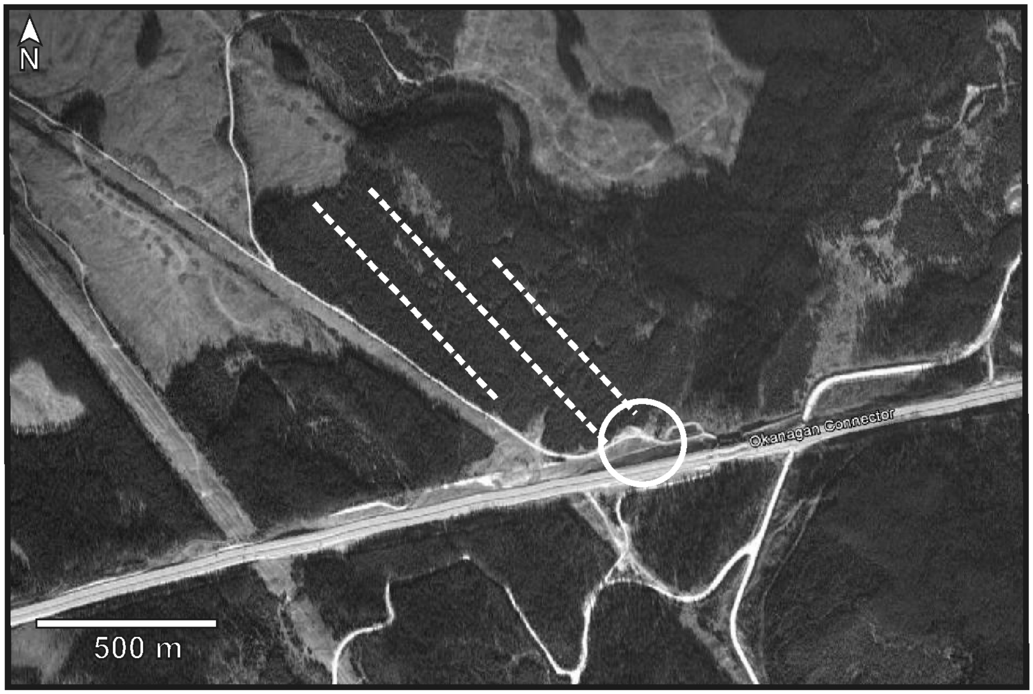
Appendix A: Site locations and geomorphic setting of sites on Thompson Plateau

A.1: Table of site locations from Thompson Plateau

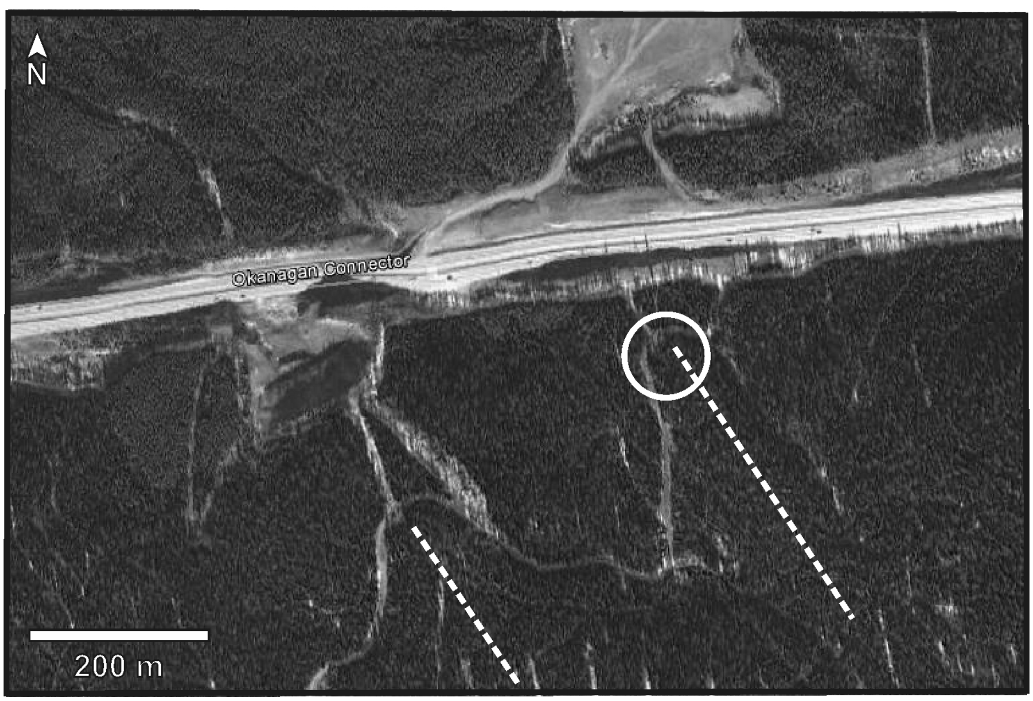
Site No.	Sediment type	Landform context	Latitude	Longitude	Appendix
1	Consolidated diamicton	Drumlinized	49.906711	-120.100646	A.2.1
2	Consolidated diamicton	Drumlinized	49.910809	-120.049912	A.2.2
3	Consolidated diamicton	Weakly-drumlinized	49.909564	-120.077457	A.2.3
4	Consolidated diamicton	Weakly-drumlinized	49.879873	-120.19433	A.2.4
5	Consolidated diamicton	Drumlinized	49.922974	-120.371958	A.2.5
6	Consolidated diamicton	Weakly-drumlinized	49.897987	-120.164227	A.2.6
7	Poorly-consolidated diamicton	Non-drumlinized	49.930121	-120.110966	A.2.7
8	Poorly-consolidated diamicton	Drumlinized	49.973659	-120.168891	A.2.8
9	Poorly-consolidated diamicton	Drumlinized	49.940480	-120.067578	A.2.9
10	Poorly-consolidated diamicton	Weakly-drumlinized	49.947941	-120.175613	A.2.10
11	Poorly-consolidated diamicton	Non-drumlinized	49.960590	-120.027759	A.2.11
12	Poorly-consolidated diamicton	Drumlinized	49.990884	-120.024071	A.2.12
13	Sorted and statified sediments	Drumlinized	49.947359	-120.175172	A.2.13
14	Sorted and statified sediments	Non-drumlinized	49.948915	-120.17303	A.2.14
15	Sorted and statified sediments	Drumlinized	49.967059	-120.131707	A.2.15
16	Sorted and statified sediments	Non-drumlinized	49.946065	-120.313415	A.2.16
17	Sorted and statified sediments	Drumlinized	49.987743	-120.095851	A.2.17
18	Sorted and statified sediments	Non-drumlinized	49.962620	-120.006709	A.2.18

A.2: Geomorphic setting of field sites on Thompson Plateau

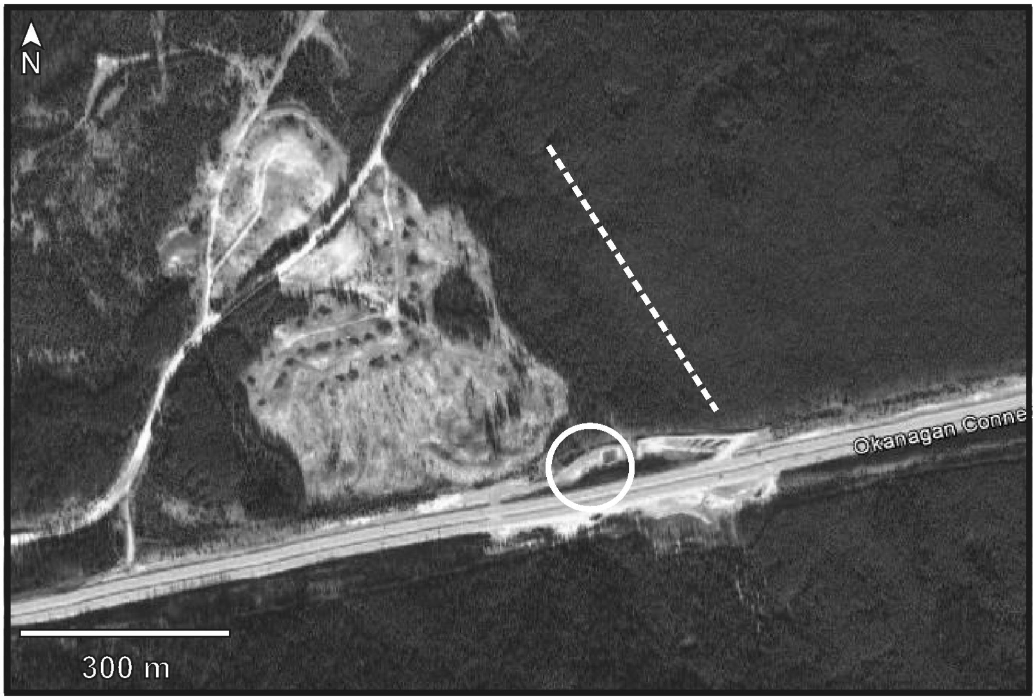
A.2.1: Site 1-Consolidated diamicton, drumlin tail.



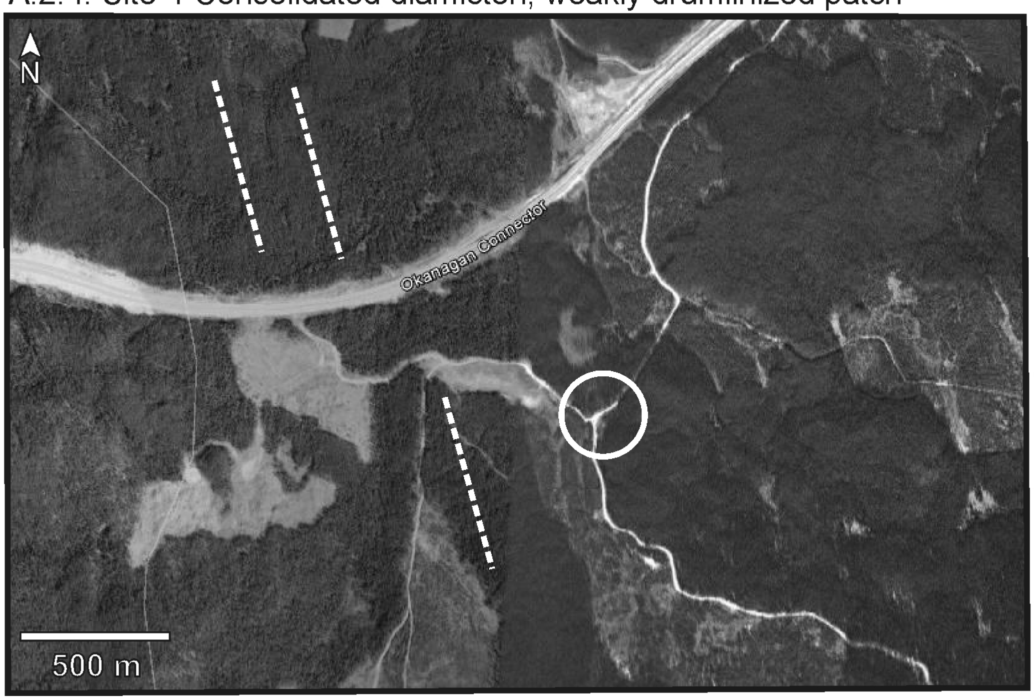
A.2.2: Site 2-Consolidated diamicton, drumlin nose



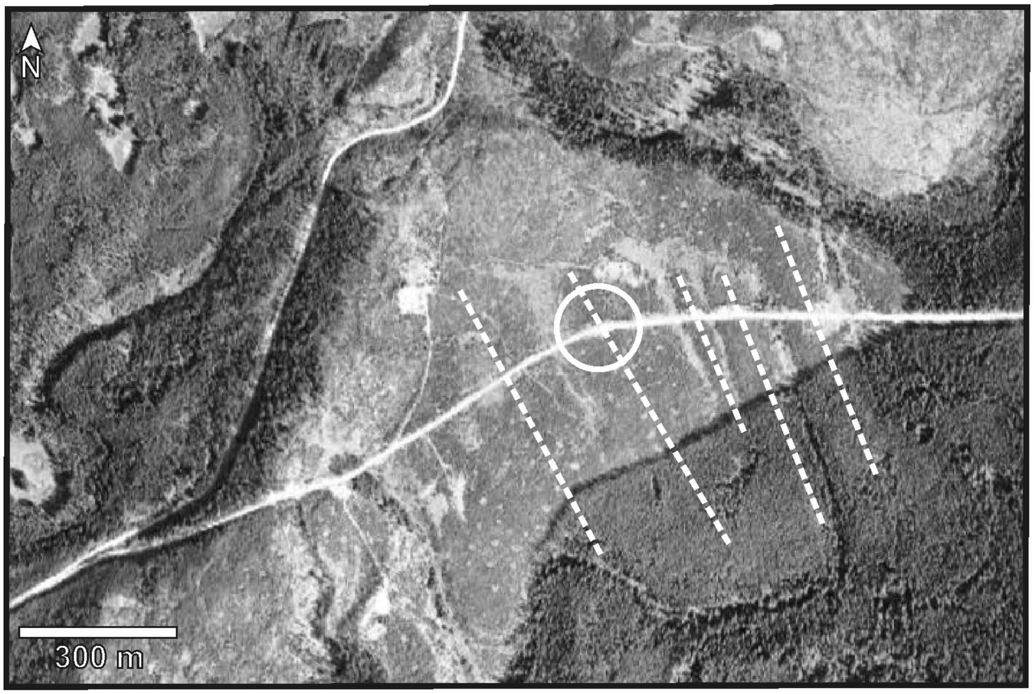
A.2.3: Site 3-Consolidated diamicton, western drumlin flank in weakly-drumlinized area



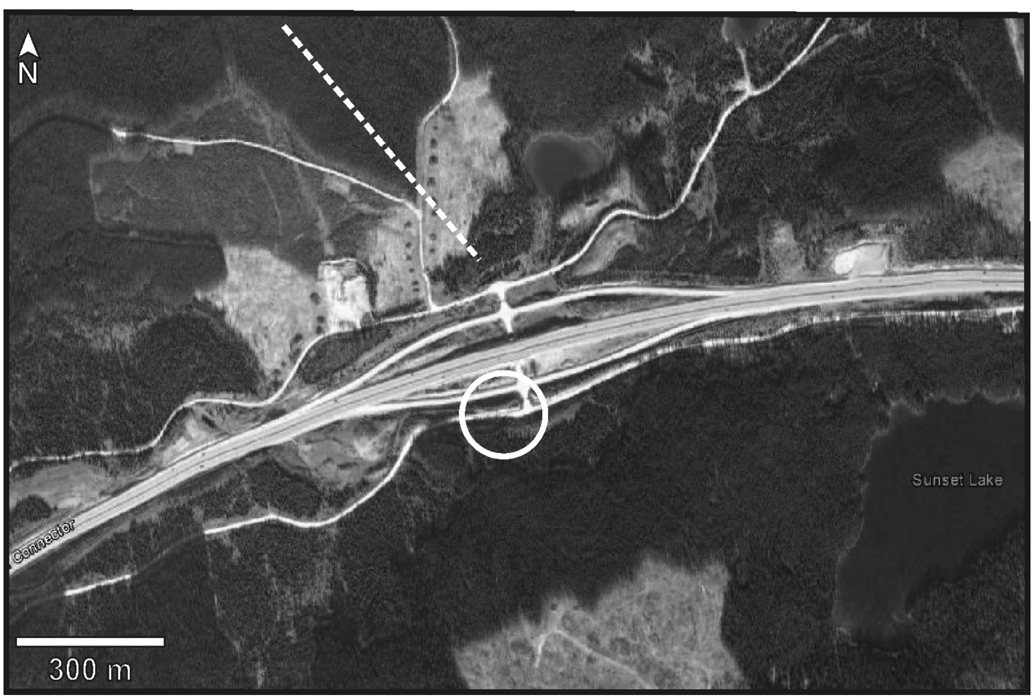
A.2.4: Site 4-Consolidated diamicton, weakly drumlinized patch



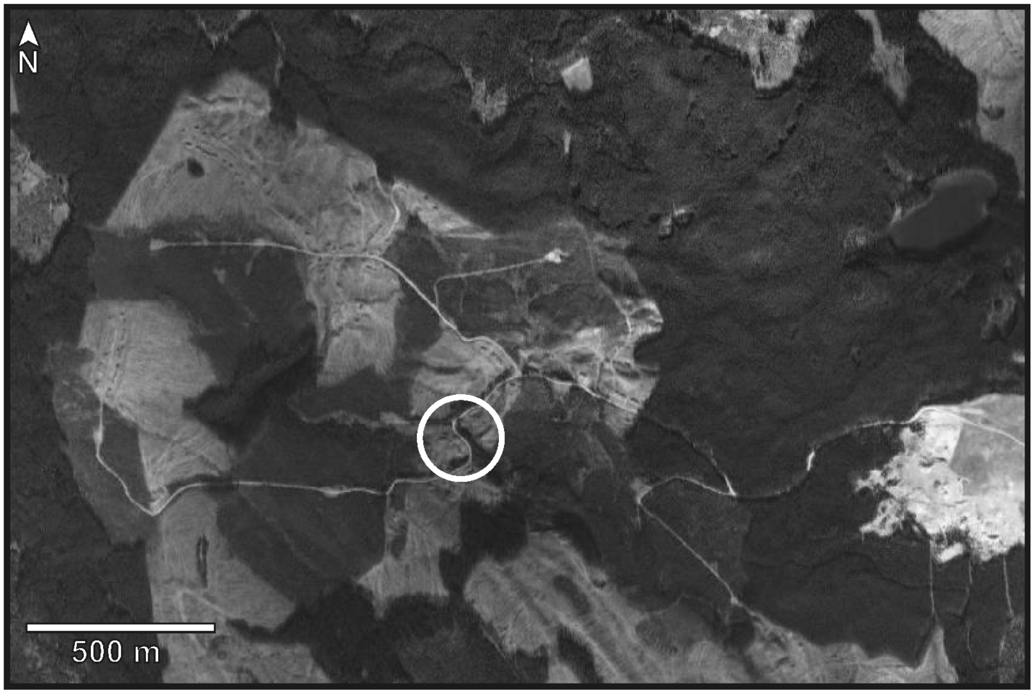
A.2.5:Site 5-Consolidated diamicton, drumlin flank near nose



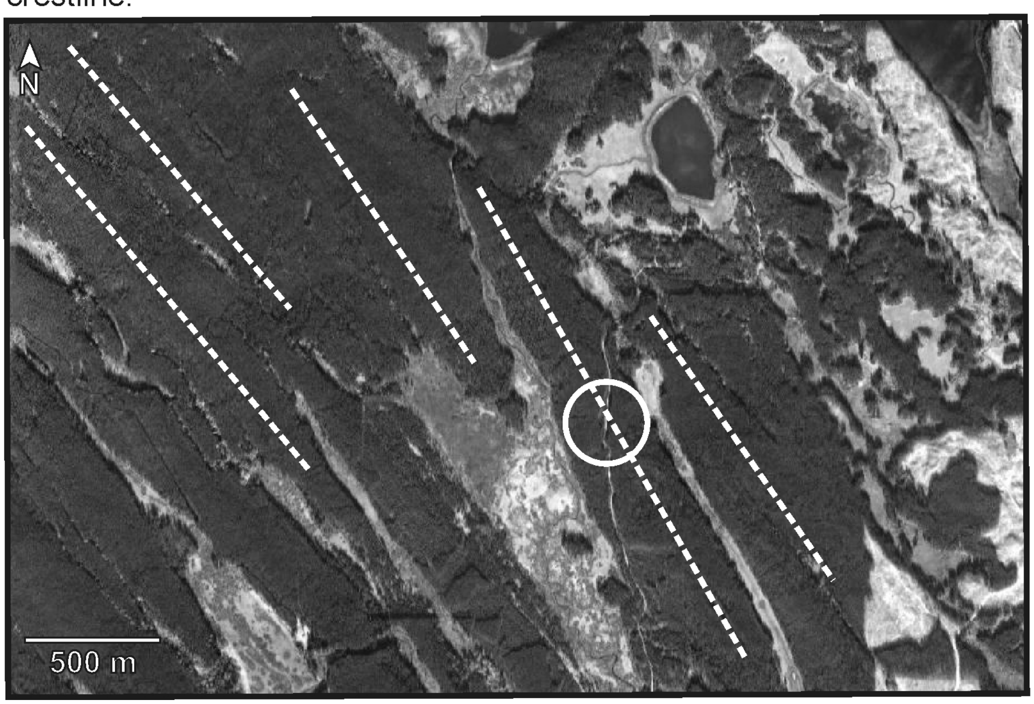
A.2.6: Site 6-Consolidated diamicton, weakly-drumlinized.



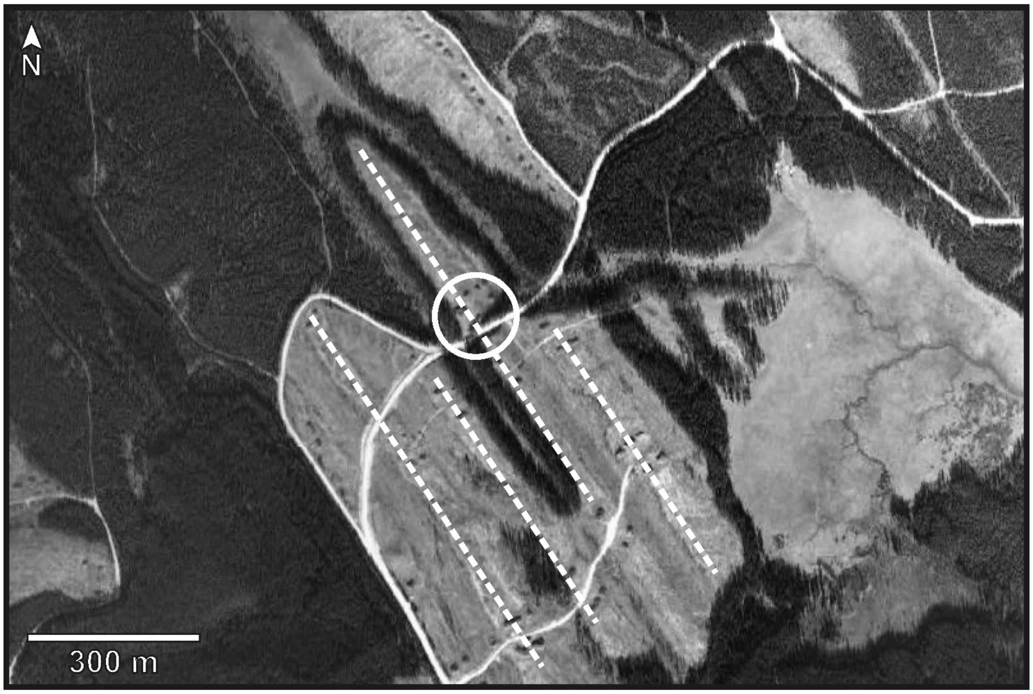
A.2.7: Site 7- Poorly-consolidated diamicton, non-drumlinized.



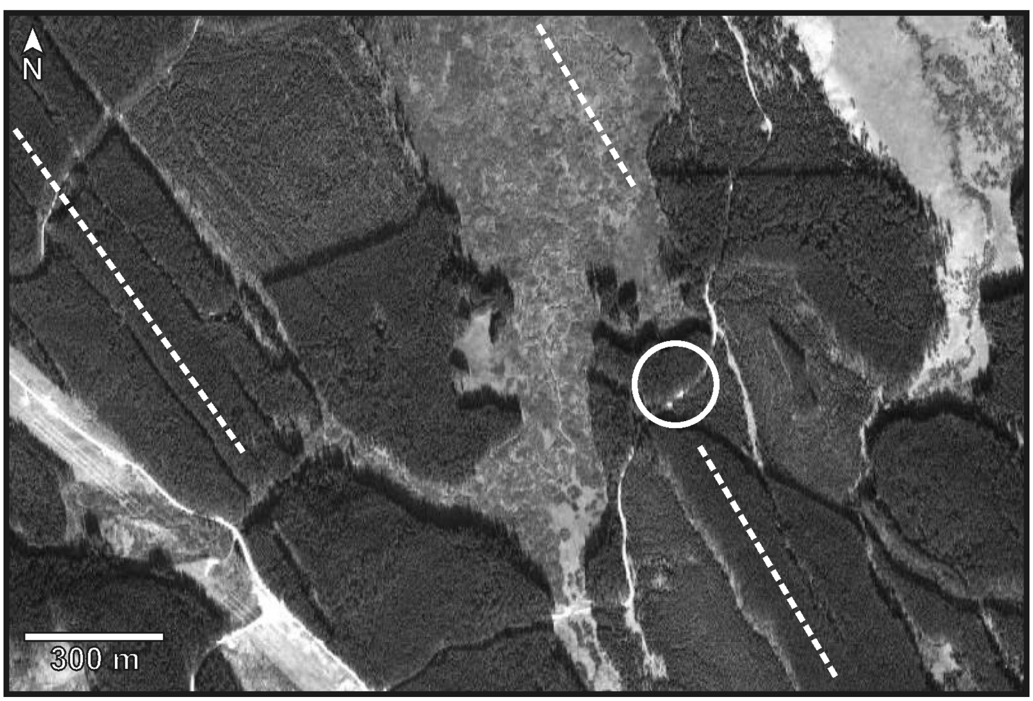
A.2.8: Site 8-Poorly-consolidated diamicton, midway along drumlin crestline.



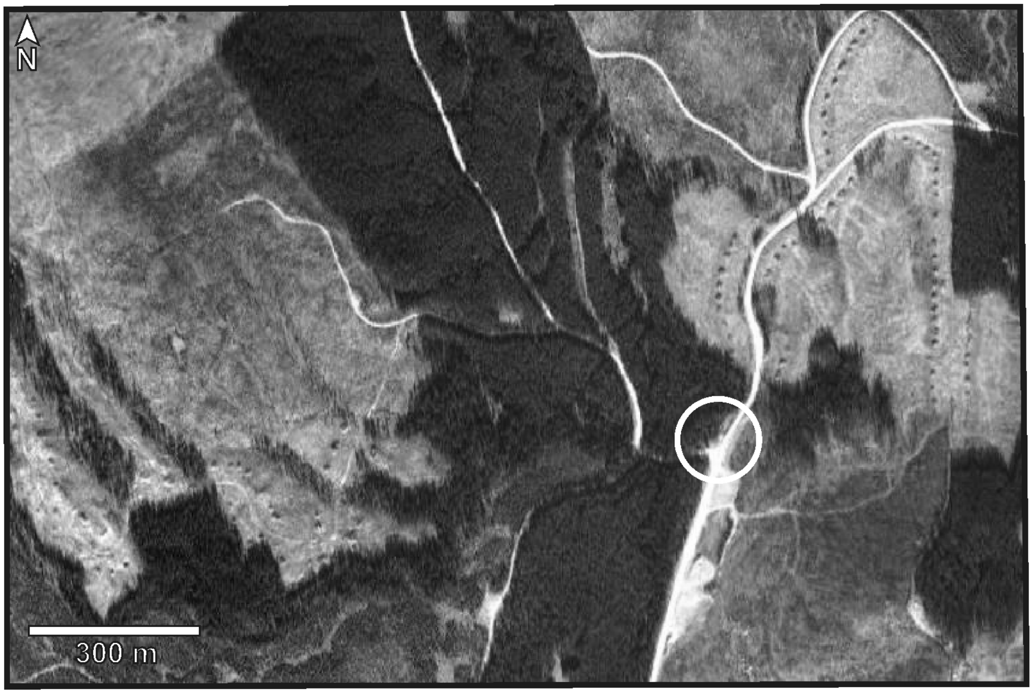
A.2.9: Site 9-Poorly-consolidated diamicton, midway along drumlin crestline.



A.2.10: Site 10-Poorly-consolidated diamicton, weakly-drumlinized area.

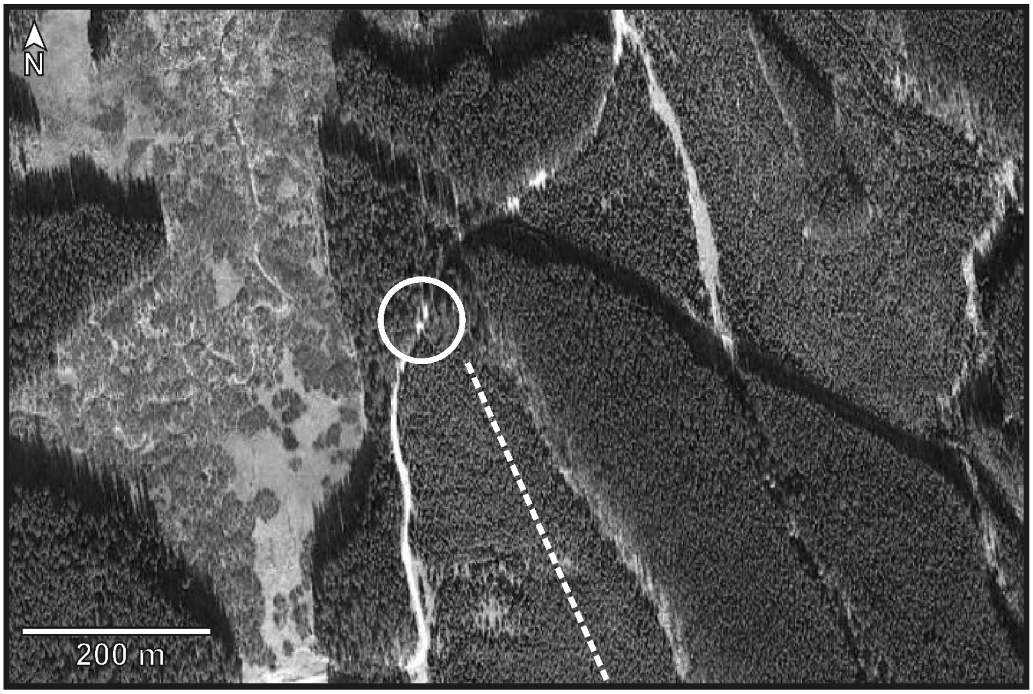


A.2.11: Site 11-Poorly-consolidated diamicton, non-drumlinized, marginal to weakly-drumlinized area.

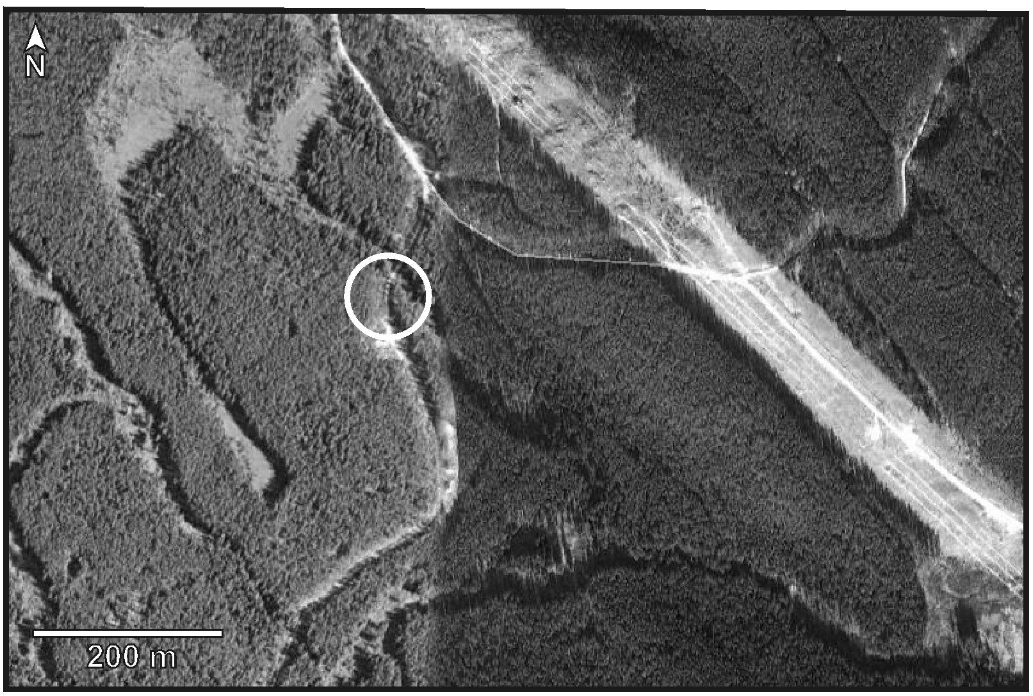


A.2.12: Site 12- Poorly-consolidated diamicton, flute superimposed on crestline of drumlinized patch

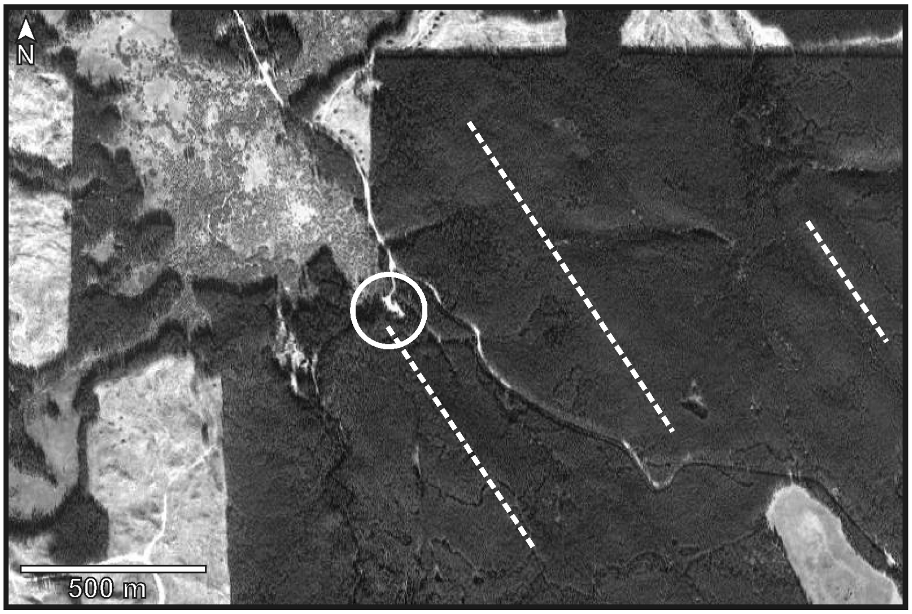
A.2.13: Site 13-Sorted and stratified sediments, near drumlin nose



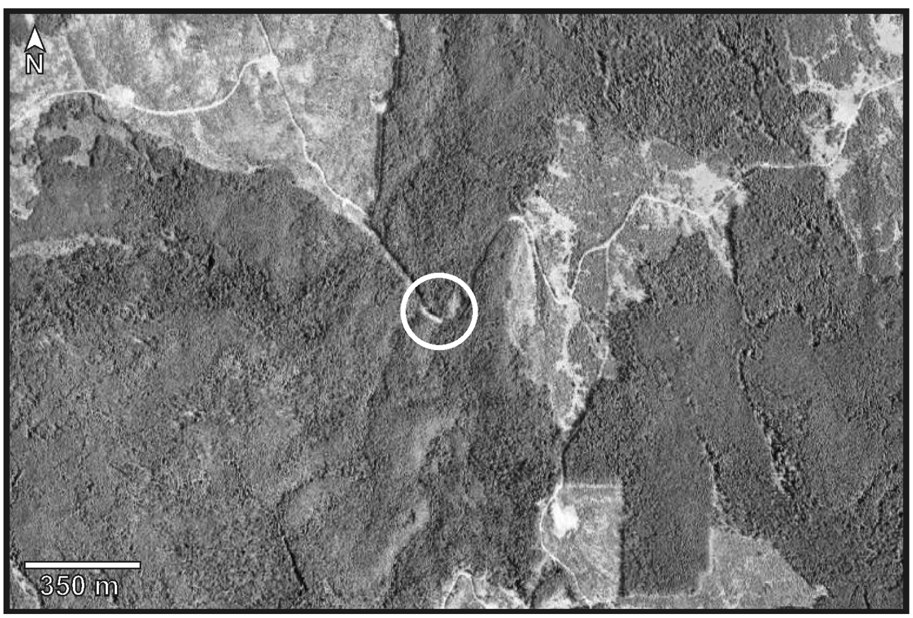
A.2.14: Site 14-Sorted and stratified sediments, in non-drumlinized zone.



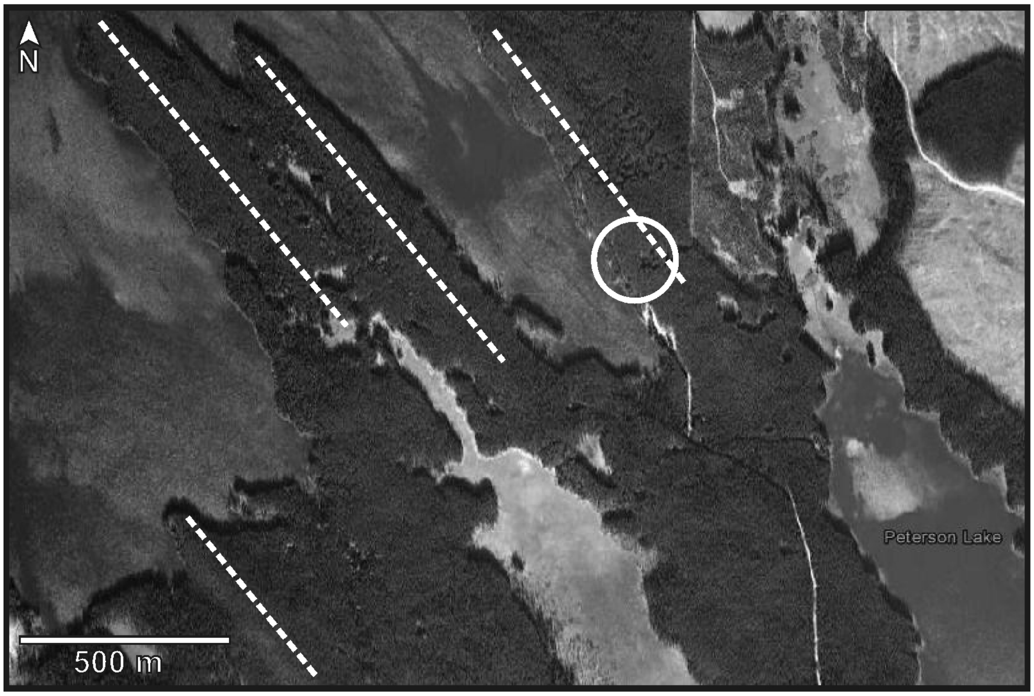
A.2.15: Site 15-Sorted and stratified sediments, near drumlin nose.



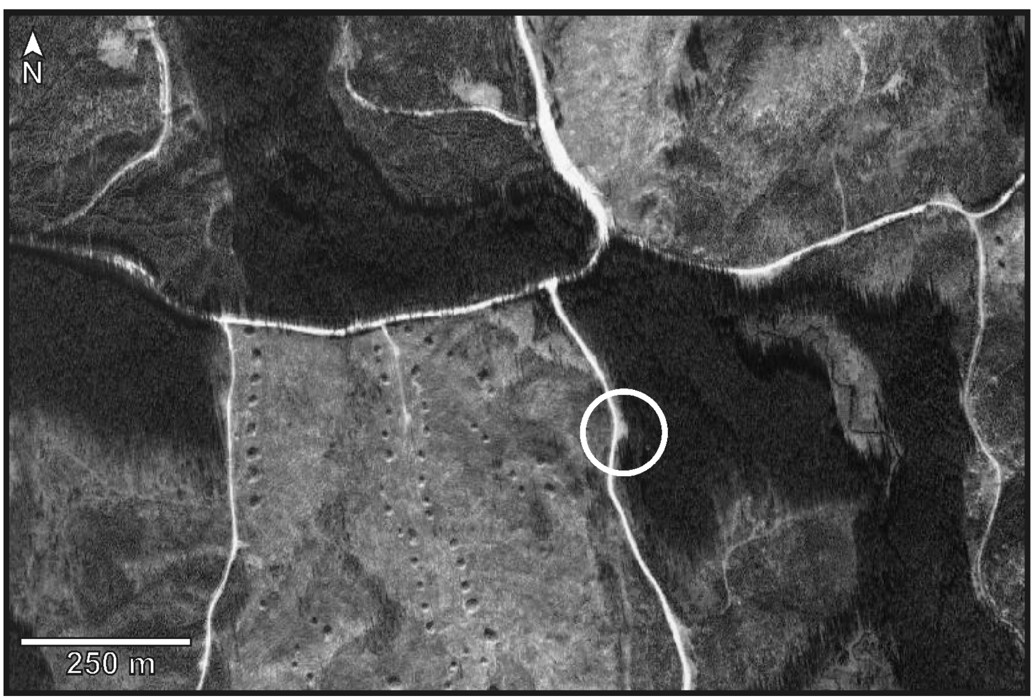
A.2.16: Site 16-Sorted and stratified sediments, in non-drumlinized area.



A.2.17: Site 17-Sorted and stratified sediments, in drumlin flank near tail.



A.2.18: Site 18-Sorted and stratified sediments, in non-drumlinized area.



Appendix B: Grain size data, Thompson Plateau

	Grain Size (%)						
Site Number	Clay (<2 Ξm)	Silt (2-62 Ξm)	V. fine sand (62-125 Ξm)	Fine sand (125-250 Ξm)	Medium sand (250-500 Ξm)	Coarse sand (500 Ξm-1 mm)	V. Coarse Sand (1-2 mm)
1	33.6	26.4	11.3	9.7	11.4	6.7	0.9
2	34.2	16.3	8.7	10.4	9.8	9.2	11.4
3	39.3	19.7	15.3	11.7	7.8	5.4	0.8
4	18.2	15.1	9.5	12.6	19.6	17.1	7.9
5a	40.9	32.8	8.4	8.3	6.4	1.7	1.5
5b	34.7	34.3	10.3	8.6	7.9	3.1	1.1
6a	29.3	28.9	12.7	10.1	10.6	6.5	1.9
6b	34.5	22.1	11.2	8.4	10.4	8.7	4.7
7a	18.3	10.8	7.4	16.4	23.2	13.2	10.7
7b	16.9	14.2	10.8	15.3	22.5	12.6	7.7
8a	20.4	20	5.9	18.4	28.2	4.8	2.3
8b	10.7	29.1	10.1	17.2	23.4	7.7	1.8
8c	7.3	30	12.6	21.2	19.5	7.9	1.5
9	24.4	11.3	11.4	14.5	22.9	12.6	2.9
10	8.7	10	13	23.6	25.1	11.8	7.8
11a	14.4	19.8	17.5	20.2	18.6	6.6	2.9
11b	10.5	22.4	12	13.4	21.8	14.4	5.5
12	17.3	30.3	9.4	16.7	13.8	8.8	3.7
13a	13.8	35.4	10.6	18.1	11.7	6.6	3.8
13b	15.3	39.9	11.4	18.6	8.9	4.1	1.8
14a	25.6	68.3	2.7	3.4	0	0	0
14b	30	64.5	3.1	2.3	0.1	0	0
14c	65.1	24.8	3.7	4.5	1.9	0	0
14d	62.5	30	2.6	4.1	0.8	0	0
14e	65.3	28.3	4.2	1.6	0.6	0	0
14f	69.8	24.4	3.9	1.2	0.7	0	0
14g	76.8	22.9	0.3	0	0	0	0
14h	68.7	26.5	3.8	0.8	0.2	0	0
15a	26.1	70.3	3.5	0	0.1	0	0
15b	15	46.7	16.4	13.5	6.6	1.8	0
15c	19.9	41.9	14.9	17.5	4.2	1.4	0.2

Refer to Figs 2.3 and 2.7 for site numbers and locations.

Appendix C: Till fabric, clast wear, and structural data from Thomspon Plateau.

All measurements corrected for declination (19° E in 2002) from true north.

Fabric Site 1

Azimuth (°)	Plunge (°)	Comments
64	9	
253	18	
238	14	
215	14	Striae on clast top, aligned with a-axis (190°-10°)
248	7	
79	15	
133	4	Possible keel on clast bottom, no striae visible (igneous clast)
203	10	Striae on clast top, oblique to a-axis (4°-184°)
126	11	
239	12	Striae on clast top, aligned with a axis (55°-235°)
47	54	Striae on clast top, aligned with a axis (48°-228°)
98	7	
138	5	
229	27	Striae on clast top, aligned with a axis (50°-230°)
249	16	Plucked end (248°)
48	23	
54	32	
112	10	
56	20	
92	4	Plucked end (91°)
145	33	
92	2	
148	5	Striae on clast top, oblique to a-axis (157°-337°), faint keel (158°-338°)
104	6	Plucked end (98°)
283	6	
357	2	Striae on clast side, aligned with b-axis (60°-240°)
72	11	Striae on clast top, aligned with a axis (74°-254°)
48	25	
202	23	
228	4	Striae on clast top, aligned with a axis (50°-230°)

Fractures (Azimuth, Dip): (93,45); (90,53); (143,46); (142,52)

Fabric Site 2

Azimuth (°)	Plunge (°)	Comments
159	7	
35	7	
359	10	
224	10	
174	15	Striae aligned with b-axis (89°-269°), plucked end (173°)
199	9	
358	8	Striae on clast top, aligned with a-axis (358°-178°), plucked end (179°)
353	3	
43	10	Striae on clast top, aligned with a-axis (43°-223°)
355	5	
162	6	
139	8	Plucked end (139°)
209	4	
41	3	
169	13	
186	7	Plucked end (186°)
63	17	Striae on clast top, slightly oblique to a-axis (60°-240°)
137	6	
114	11	
44	4	Striae on clast top, aligned with a-axis (45°-225°)
133	27	
294	17	
109	17	Striae on clast top, slightly oblique to a-axis (105°-285°), plucked end (111°)
138	4	Striae aligned with b-axis (52°-232°), curved striae on edges and side
360	14	
209	23	
219	3	
127	6	
41	14	
167	12	

Fabric Site 3

Azimuth (°)	Plunge (°)	Comments
200	44	
305	9	
324	4	
180	9	
252	4	Striae on clast top, aligned with a-axis (70°-250°)
88	8	
230	4	Plucked end (230°)
186	12	Keel (6°-186°)
200	4	
332	8	Striae on clast top, aligned with a-axis (152°-332°)
109	54	
194	8	Plucked end (192°)
32	3	
38	4	
217	7	
215	4	Plucked end (215°)
315	40	
29	13	
155	7	Striae on clast top, aligned with a-axis (155°-335°)
130	6	Striae on clast top, aligned with a-axis (130°-310°)
19	3	
325	39	
17	3	
342	7	Keel (167°-347°)
191	11	
230	24	Striae on clast top, aligned with a-axis (50°-230°), faint plucked end (15°), keel in non-ploughing position, striae around clast edges
219	11	
335	20	
81	17	
316	4	

Fabric Site 4

Azimuth (°)	Plunge (°)	Comments
62	24	Keel (70°)
108	27	
281	25	
116	27	Striae on clast top, aligned with a-axis (116°-296°)
170	17	
132	23	
134	20	
163	9	
175	21	Striae on clast top, aligned with a-axis (179°-359°)
100	5	
202	24	Striae on clast top, aligned with a-axis (204°-24°), keel in non-ploughing position
96	6	
120	19	
		Striae on clast top, aligned with a-axis (121°-301°), curved striae on edges
90	47	
144	34	
168	26	
136	25	
120	11	
96	27	
137	36	
118	9	
320	8	
194	9	Striae on clast top, aligned with a-axis (1°-181°), keel (186°), plucked end (195°)
82	35	Keel (80°)
96	36	Striae on clast top, aligned with a-axis (96°-276°)
155	14	
172	27	
138	21	
110	41	
106	21	

Fractures (Azimuth, Dip): (128,57); (127,54); (96,57); (101,51); (122, 57)

Fabric Site 5

Azimuth (°)	Plunge (°)	Comments
346	16	
171	6	
301	14	
155	9	
349	11	Keel (350°)
317	42	Striae on clast top, aligned with a-axis (137°-317°)
343	11	
59	17	
54	4	
240	12	
184	22	Striae on clast top, aligned with a-axis (4°-184°)
332	23	
66	3	
108	34	Plucked end (90°)
51	9	
158	9	
126	7	
287	4	Keel (290°)
308	27	
319	6	Striae on clast top, aligned with a-axis (139°-319°)
144	3	
166	2	
149	4	
324	4	Striae on clast top, aligned with a-axis (144°-324°), plucked end (324°)
165	19	
350	32	Striae on clast top, aligned with a-axis (1°-180°)
28	8	Striae on clast top, aligned with a-axis (28°-208°), plucked end (208°), keel (207°)
20	12	
36	29	Striae on clast top, aligned with a-axis (36°-216°), plucked end (216°)
32	31	

Shears (Azimuth, Dip): (127,36); (146,45); (178,43); (173,34); (172, 35)

Fabric Site 6

Azimuth (°)	Plunge (°)	Comments
12	24	
13	3	
334	11	Keel (154°-334°)
61	33	
47	28	
12	10	
188	9	
88	49	Striae on clast top, aligned with a-axis (88°-268°)
382	6	
181	19	
154	34	
138	24	
188	19	
358	32	
13	29	
357	41	Keel (175°-355°), plucked end (357°)
31	14	Keel (32°-212°), plucked end (212°)
58	20	
318	27	
324	20	
1	7	Striae on clast top, aligned with a-axis (180°-360°)
179	13	
322	21	
151	7	
154	24	
346	7	
63	47	Striae on clast top, aligned with a-axis (64°-244°)
20	39	Striae on clast top, aligned with a-axis (20°-200°), plucked end (20°)
348	17	
68	24	Striae on clast top, aligned with a-axis (66°-246°), keel (67°-247°)

Fractures (Azimuth, Dip): (29,6); (42,11); (78,42); (83,38)

Fabric Site 8

Azimuth (°)	Plunge (°)	Comments
19	9	
185	6	
150	0	
46	24	Striae on clast top, aligned with a-axis (46°-226°)
210	36	
116	4	
274	3	
52	6	
164	24	
236	4	
193	7	
166	3	Striae on clast top, aligned with a-axis (166°-346°)
227	27	
36	15	Plucked end (38°)
184	34	
63	47	
206	4	
71	18	Plucked end (73°)
273	32	Striae on clast top, aligned with a-axis (94°-274°), plucked end (263°)
346	16	Striae on clast top, aligned with a-axis (160°-340°), plucked end (346°)
69	6	
341	18	
4	7	Striae on clast top, aligned with a-axis (4°-184°)
21	37	
283	14	
95	19	
226	9	
144	3	
186	3	
194	27	

Fabric Site 9

Azimuth (°)	Plunge (°)	Comments
28	31	
10	24	
350	16	
325	20	
289	22	Striae on clast top, aligned with a-axis (109°-289°), plucked end (294°)
325	24	
345	9	
325	23	
319	9	
307	14	
25	25	Striae on clast top, aligned with a-axis (25°-205°), plucked end (294°)
229	6	Striae on clast top, aligned with a-axis (47°-227°)
322	21	
142	16	
345	36	
260	16	Striae on clast top, aligned with b-axis (37°-217°)
356	22	Striae on clast top, aligned with b-axis (48°-228°), Keel (176°-356°), plucked end (354°)
294	4	
92	11	
170	28	
208	5	Striae on clast top, aligned with a-axis (28°-208°)
185	28	
109	42	
349	4	
105	10	Striae on clast top, aligned with a-axis (105°-285°)
105	6	Striae on clast top, aligned with a-axis (107°-287°)
202	20	Striae on clast top, aligned with a-axis (22°-202°)
338	14	
311	36	
353	12	

Fabric Site 10

Azimuth (°)	Plunge (°)	Comments
149	11	
139	14	
56	16	
99	26	Striae on clast top, aligned with a-axis (99°-279°), plucked end(97°)
333	17	Striae on clast top, aligned with a-axis (150°-330°), keel (167°-347°)
161	14	keel (161°-341°)
151	12	Striae on clast top, aligned with a-axis (152°-232°)
14	17	Striae on clast top, aligned with a-axis (15°-195°)
162	13	Striae on clast top, aligned with a-axis (162°-342°)
51	9	Striae on clast top, aligned with a-axis (52°-132°)
122	48	
295	7	Striae on clast top, aligned with a-axis (115°-295°), plucked end (291°)
46	20	
135	21	Striae on clast top, aligned with a-axis (135°-315°)
67	18	
90	16	Striae on clast top, aligned with a-axis (90°-270°), plucked end (91°)
180	4	
308	11	Striae on clast top, aligned with a-axis (128°-308°)
43	17	
332	11	
124	24	
50	23	Striae on clast top, aligned with a-axis (50°-230°)
275	7	Striae on clast top, aligned with a-axis (97°-277°), plucked end (277°)
15	18	
73	21	
227	12	Striae on clast top, aligned with a-axis (°46-226°), plucked end (227°)
199	10	
6	9	Striae on clast top, aligned with a-axis (6°-186°)
185	1	Striae on clast top, aligned with a-axis (5°-185°), plucked end (185°)
25	12	

Fabric Site 11

Azimuth (°)	Plunge (°)	Comments
106	19	Striae on clast top, aligned with a-axis (106°-296°), plucked end (105°)
104	13	Striae on clast top, aligned with a-axis (104°-284°)
135	28	
3	13	
159	9	
132	18	
183	33	
343	19	
200	37	
189	18	Plucked end (189°)
138	21	Keel (138°-318°), plucked end (138°)
304	13	Keel (127°-307°), plucked end (307°)
303	14	
232	12	Striae on clast top, aligned with a-axis (52°-232°)
206	24	
171	8	
359	20	
127	25	
163	22	
20	29	
232	12	Striae on clast top, aligned with a-axis (52°-232°)
198	16	
51	24	
94	37	Striae on clast top, aligned with a-axis (94°-274°)
353	6	
101	10	Striae on clast top, aligned with a-axis (101°-291°), plucked end (101°)
26	3	
131	9	
144	23	
207	7	

Fabric Site 12

Azimuth (°)	Plunge (°)	Comments
284	2	
124	6	
324	2	
146	3	
333	14	Striae on clast top, aligned with a-axis (150°-330°), plucked end (330°)
156	14	
294	5	
286	14	
300	48	
326	47	
155	22	
204	15	
137	19	
192	3	
114	56	
113	51	
295	46	
275	5	Striae on clast top, aligned with a-axis (88°-268°), plucked end (268°)
83	12	Striae on clast top, aligned with a-axis (83°-263°), plucked end (83°)
16	12	Striae on clast top, aligned with a-axis (16°-196°), plucked end (16°)
138	32	
351	10	
129	19	Striae on clast top, aligned with a-axis (129°-309°), keel (132°-312°)
124	31	
142	13	
229	20	
20	14	
271	23	Striae on clast top, aligned with a-axis (90°-270°)
96	6	
245	28	Striae on clast top, aligned with a-axis (65°-245°), plucked end (245°)

Fractures (Azimuth, Dip): (168,50); (177,47)

Appendix D: Coordinates and access information for sites documented in Chapter 4.

	Coordinates		
Site number	Latitude (°)	Longitude (°)	Notes
1	49.248425	-119.379233	Private access
2	49.471770	-119.687689	Access off Green Mtn. Rd., North side
3	49.432488	-119.743865	Access off Apex Mtn. Rd., North side
4	49.447397	-119.617974	Access from Old Penticton Oliver Highway
5	49.331080	-119.493656	Access from N. Shuttleworth Creek Rd
6	49.518144	-119.512837	Private access
7	49.523427	-119.544488	Access from Greyback Mtn. Rd.
8	49.613020	-119.783066	Junction Trout Creek Rd. and Faulder Rd.
9	49.561138	-119.638829	South of Sun Oka Beach, access off Highway 97
10	49.599919	-119.657265	Access south of Faircrest Drive or Butler Rd.
11	49.615938	-119.665615	Gully accessed from Kean St., Summerland
12	49.593398	-119.563540	BC waterwell report 82844
13	49.635827	-119.612462	Abandonned gravel pit, west of Naramata Rd.
14	49.746105	-119.764486	Access along Bulyea Ave., Peachland
15	49.755693	-119.794513	North of Princeton Ave., Peachland
16	49.758535	-119.808489	North of Princeton Ave., Peachland
17	49.839271	-119.647717	Access west of Popp Rd., Glenrosa/Westbank
18	49.867744	-119.586101	Private gravel pit, north of Stevens Rd., Westbank
19	49.818013	-119.471585	Housing development, south of Mission Ridge Rd.
20	49.833033	-119.449226	Private gravel pit, east of Bedford Rd., South Kelowna
21	49.904871	-119.487967	Access of Knox Mtn. Rd., Kelowna
22	49.900264	-119.441765	Housing development, Denali Dr., Glenmore
23	49.932468	-119.531791	Private access from Bear Creek Rd.
24	49.961204	-119.549606	Access from Lambly Creek Rd.
25	49.990813	-119.583091	Access from Lambly Creek Rd.
26	49.999942	-119.636785	Forestry Rd. branching WSW from Bear Main Rd.
27	50.135706	-119.510041	Access from Westside Rd.
28	50.101830	-119.507589	Abandonned gravel pit, west of Westside Rd.
29	50.117528	-119.513466	Resort development, access along westside Rd.
30	50.127201	-119.358005	Private, no access. BC water well report 38850
31	50.236961	-119.484550	Access at Bouleau Lake Rd. and Whitemans Creek Rd.
32	50.288117	-119.429638	Access along Six Mile Creek Rd.
33	50.296023	-119.475296	Access along Siwash Creek
34	50.279393	-119.230567	Access along B.X. Rd.
35	50.278772	-119.250271	Access from 16th St., Vernon

	Coordinates		
Site number	Latitude (°)	Longitude (°)	Notes
P I	49.520025	-119.513121	Access from Greyback Mtn Rd.
P II	49.518718	-119.528146	Private land, access from Greyback Mtn Rd.
P III	49.490186	-119.545715	Private gravel pit, access from Lawrence Ave, Penticton
P IV	49.488941	-119.549547	Private gravel pit, access from Lawrence Ave, Penticton
P V	49.472746	-119.570495	Access from Dartmouth Dr., Penticton
C I	49.524554	-119.545308	Access from Spiller Rd.
C II	49.521771	-119.547769	Access from Reservoir Rd.
TC I	49.612627	-119.783021	Trout Creek Rd at Faulder
TC II	49.587621	-119.747904	Access from Trans Canada Trail
TC III	49.595943	-119.732436	Private gravel pit, access on Princeton-Summerland Rd.
TC IV	49.590747	-119.730606	Private gravel pit, access on Bathville Rd.
TC V	49.583824	-119.741988	Access from Bathville Rd., Summerland

Refer to Figure 4.2.for regional perspective of sites in this table.

Appendix E: Procedure used to identify tributary sediment benches and to generate Figure 4.4.

Figure 4.4 (Chapter 4) is a generalized representation of tributary valley sediment benches. The procedure used to develop this figure from data within tributary valleys of Okanagan Valley is explained in more detail in the following section.

Bench identification:

Benches were identified using DEMs and aerial photographs. These results were verified in the field where additional benches were also identified.

Bench surveying:

Bench elevations were measured by using a combination of surveying and existing topographic data, depending on bench accessibility. Accessible benches were surveyed using a differential GPS (dGPS). Elevations of a few inaccessible benches were obtained from 1:50 000 topographic maps. Benches were surveyed at/near the bench edge. Bench surfaces generally slope toward the tributary valley axis at varying inclinations. In some cases, especially for the smallest benches (< 200 m width) it is difficult to assess the true bench slope. This is due to the presence of a ~tributary valley-parallel slope on the bench surface and a tributary valley-normal slope across the bench surface. This cross-valley slope component is created in part by the build-up of colluvial aprons (from the valley sides) atop the bench surface and by dissection and erosion of the bench surface and treads. The elevation of bench treads was not projected toward the axis of the tributary valley. Consequently, bench profiles are 2-dimensional representations of elevation changes along the bench surface.

Identification of bench groups:

Data plots allow for identification of different bench groups based on their elevation, position within a tributary valley, and their potential intersection with the gLP water plane. Examples from Lambly Creek, Penticton Creek and Naswhito Creek are used to illustrate the procedure.

Lambly Creek valley (Figs. E.1, E.2):

i) The highest *High Bench* (red pentagons) is identified by linking multiple bench segments recording the highest bench surface within Lambly Creek valley (Figs. E.1, E.2).

ii) An additional *High Bench* (yellow circles) is identified using aerial photographs and orthophotos. Topographic maps provide elevation data. Surface correlation is tentative due to the uncertainty in the elevation of linked bench segments. Dashed lines represent this uncertainty. This surface is not represented in Figure 4.4.

iii) A *gLP Bench* (green circles) is identified below the *High Benches* and consists of a projected surface terminating within the gLP elevation envelope (500-525 m asl). Linked bench segments define this surface. However, some linked segments are far apart due to erosion within the tributary valley and there is some uncertainty associated with correlating distant bench segments using only elevation. Therefore, correlation is informed by geomorphology and the benches are correlated based on their position within the tributary valley and their relative position between groups of benches. A sloping bedrock surface on the valley sides and accumulation of colluvium on bench treads can create m-scale elevation differences on bench surfaces. Therefore, some surveyed points are ignored during bench correlations. Two examples of such points are illustrated on Figure E.2.

iv) A second potential *gLP Bench* is identified near the tributary mouth (blue circles). This forms a nearly-horizontal surface and may be part of a delta graded to gLP. For illustrative purposes, this surface is differentiated from the other *gLP Bench* (green circles). However, these surfaces may be associated.This surface is not represented in Figure 4.4.

v) A number of *Lower Benches* (variously coloured squares) are identified below the gLP elevation envelope. These occur as small bench segments (10s m to a few 100s metre length). Their extent is strongly influenced by the presence of a deeply-incised bedrock gorge in the lower reaches of Lambly Creek valley. Only one Low Bench is shown in Figure 4.4.

vi) A *Recent Bench* (light green triangles) occurs as part of the modern Lambly Creek delta prograding into Okanagan Lake (~342 m asl).

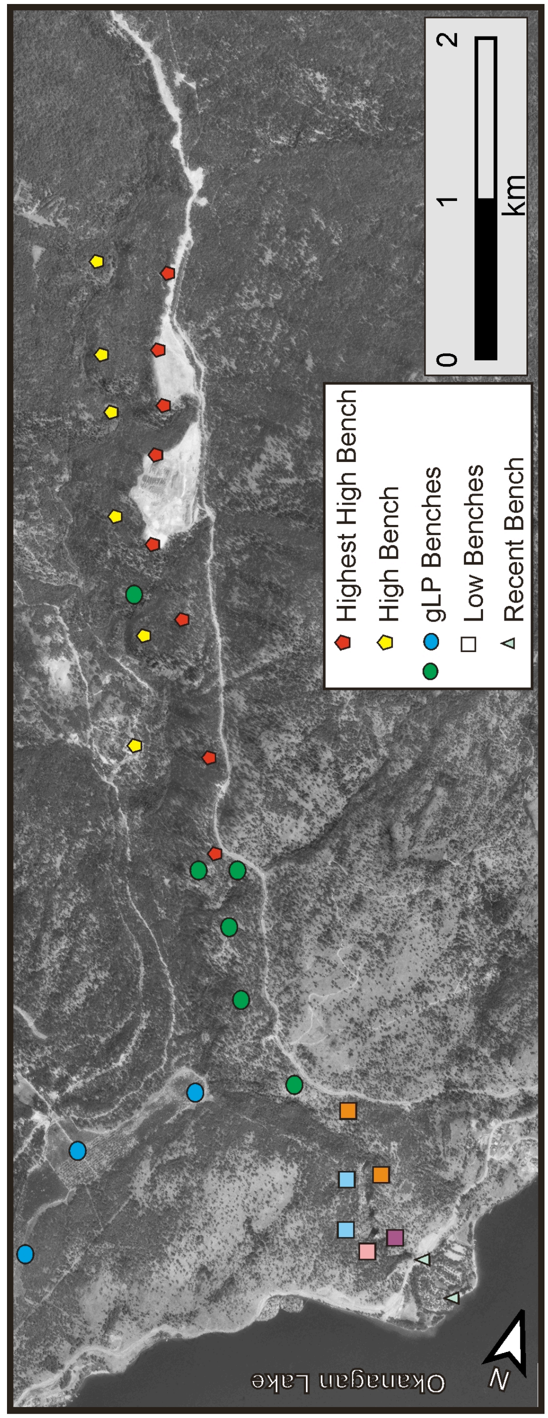


Figure E1: Tributary valley bench correlations in Lambly Creek (Data source: BC Orthophoto 82eNW draped over TRIM 1, 25 m resolution DEM. © GeoBC, by permission.

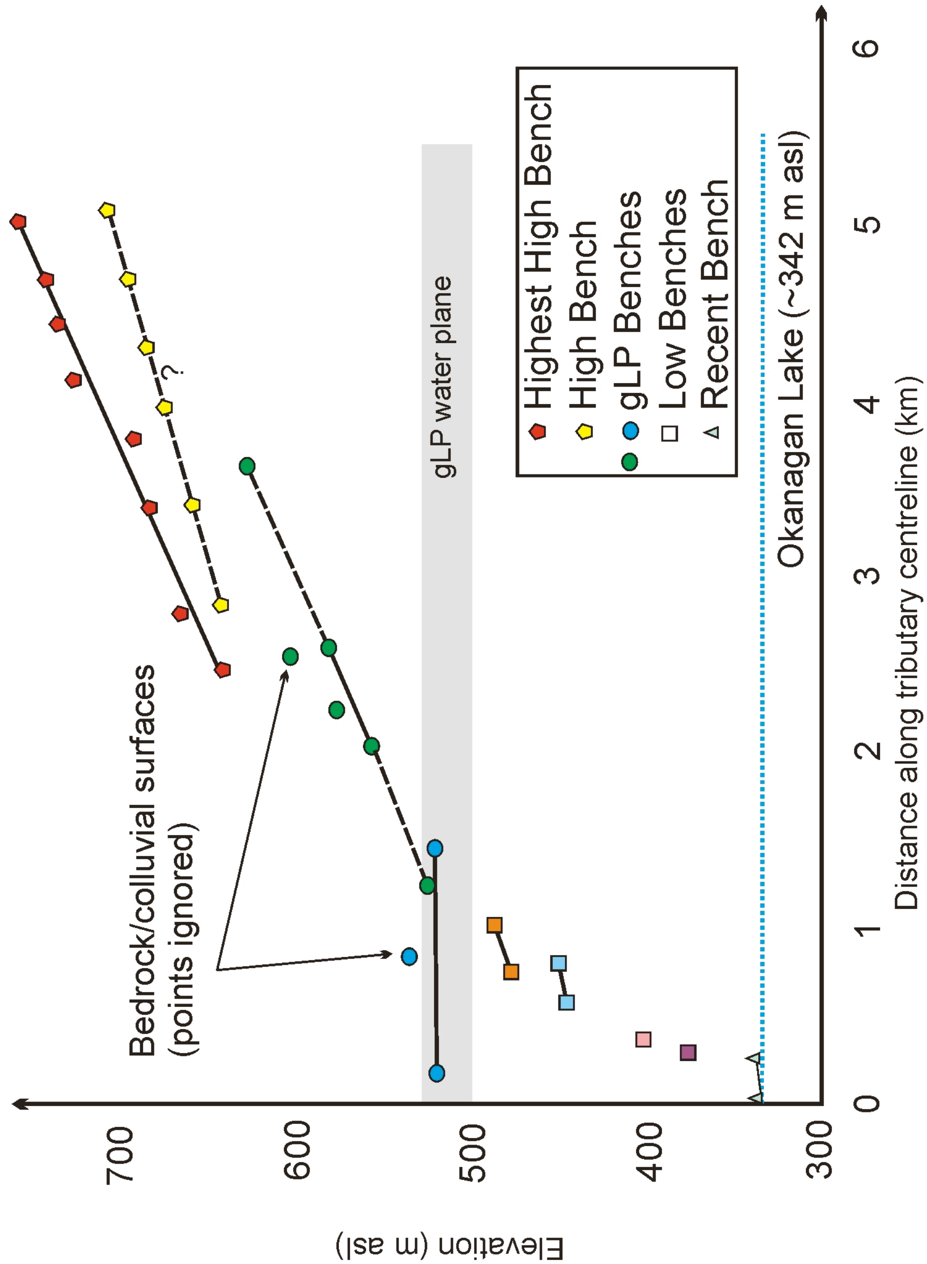


Figure E.2: Bench slopes in Lambly Creek valley rmapped using a combination of field surveys with differential GPS, aerial photographs, DEMs, and topographic maps.

Penticton Creek valley (Figs. E.3, E.4):

i) Short segments of a *High Bench* (red pentagons) are identified in Penticton Creek valley. This interpretation is tentative given a general absence of other correlative segments.

ii) Two *gLP Benches*, intersecting the gLP water plane (if projected) are identified below the *High Bench* (green and blue circles). The lowest of these two benches (green circles) terminates in a delta within the gLP elevation envelope. The up-valley extent is uncertain because the surface cannot be traced continuously along the tributary. Correlation is based on the elevation of this surface and its relative position in the tributary valley. These two gLP benches may be related to the metastable nature of the inferred gLP stillstand. It is proposed that bedrock spillways controlled stabilization of the gLP water plane between 500-525 m asl. These tributary benches may reflect these lake level fluctuations.

iii) Intervening bench segments occur between the *High* Bench and the *gLP benches.* These segments are identified as *gLP Benches* but they may be *Low Benches*. The limited bench extent precludes a more conclusive interpretation.

iv) Two *Low Benches* (orange and blue squares) form extensive surfaces grading to the Okanagan Valley floor in Penticton.

v) A *Recent bench* (light green triangles) occurs as part of the modern Okanagan Valley floor in Penticton.

Naswhito Creek valley (Figs. E.5, E.6):

i) *High Benches* were not identified in Nashwito Creek valley. However, two *gLP Benches* were identified (green, blue circles). The highest of these benches (green circles) forms the highest sediment surface extending up-tributary valley. This bench extends toward the tributary mouth and terminates in a lobate, steep-fronted surface interpreted as a delta.

ii) Two *Low Benches* (orange and light blue squares) occur within the gLP elevation envelope. They form short bench segments and are likely autocyclic surfaces incised into the delta surface.

iii) Additional *Low Benches* (pink, magenta, and dark blue squares) occur below the gLP envelope and result from autocyclic incision.

iv) A *Recent Bench* (light green triangles) occurs as part of the modern delta graded to Okanagan Lake.

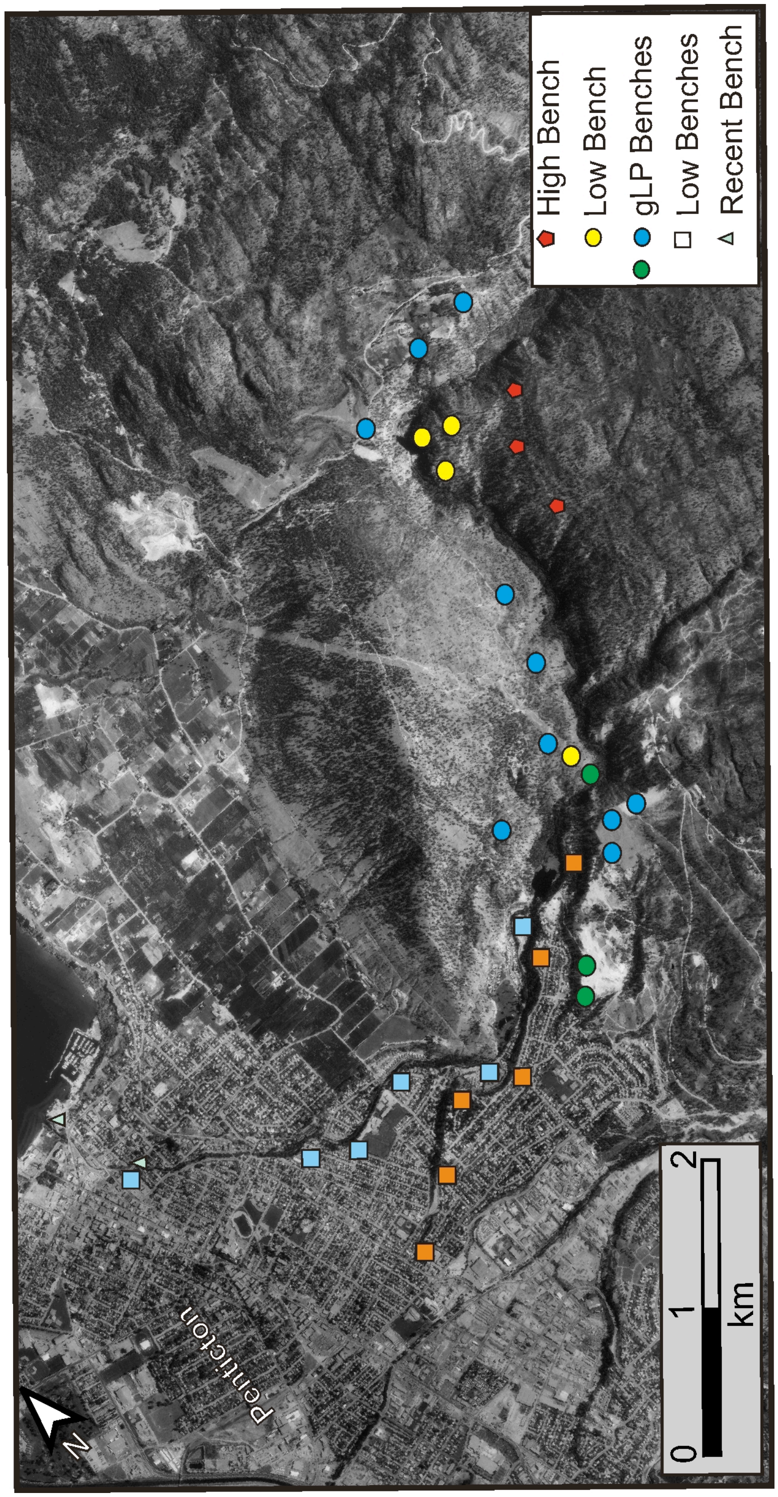


Figure E3: Tributary valley bench correlations in Penticton Creek (Data source: BC Orthophoto 82eNW, © GeoBC, by permission.

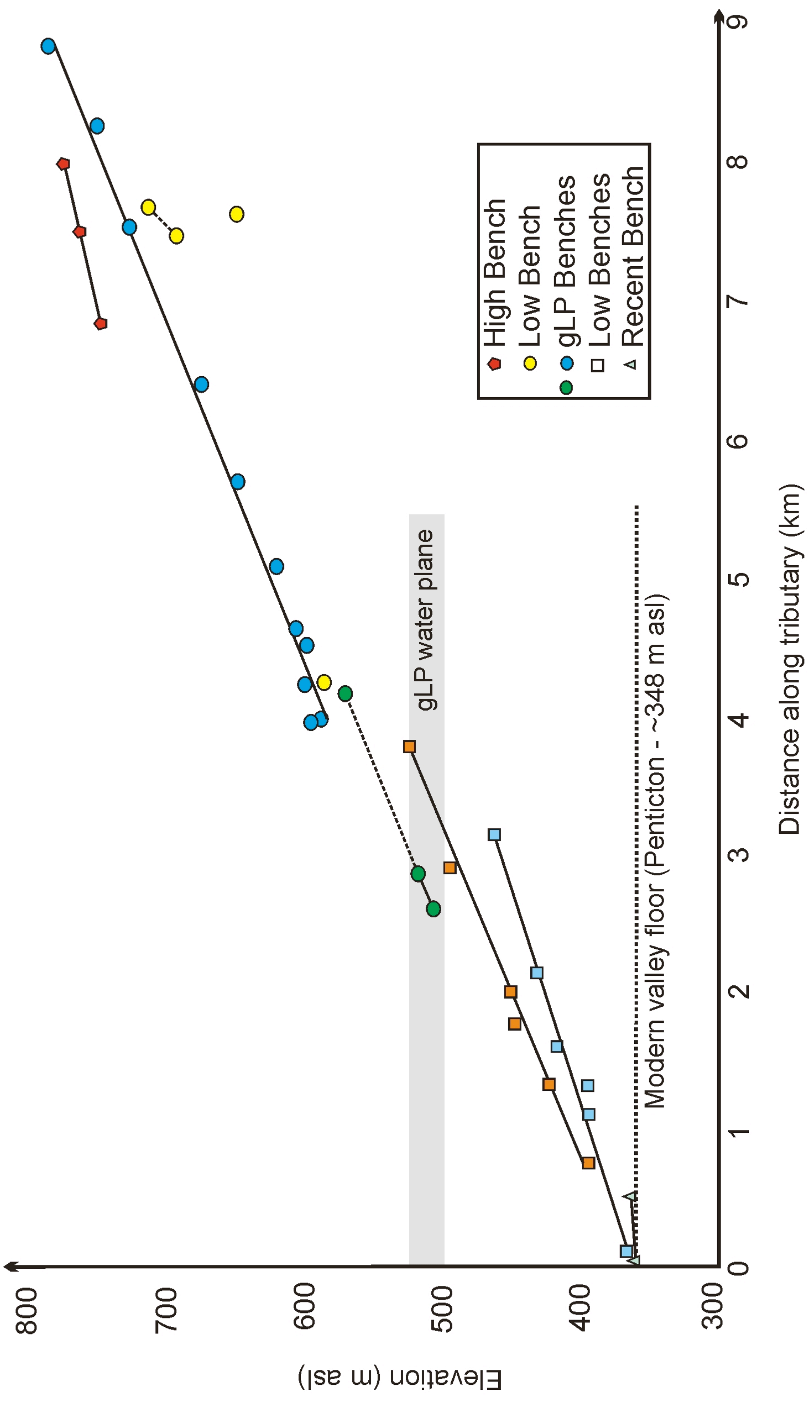


Figure: E.4: Bench slopes in Penticton Creek valley mapped using a combination of field surveys with differential GPS, aerial photographs, DEMs, and topographic maps.

Figure E.5: Tributary valley bench correlations in Naswhito Creek valley (Data source: BC Orthophoto 82eNW, © GeoBC, by permission.

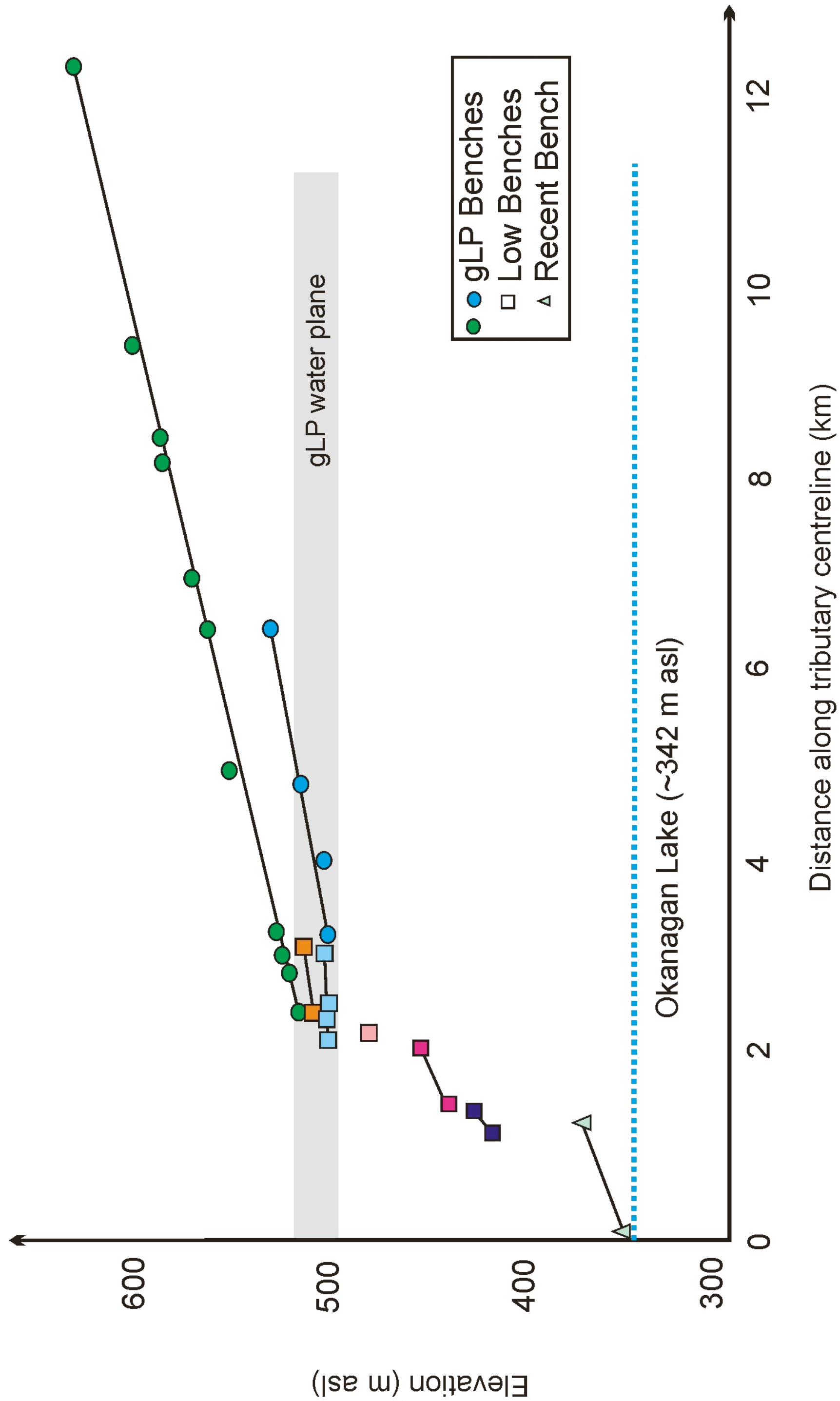


Figure: E.6: Bench slopes in Naswhito Creek valley mapped using a combination of field surveys with differential GPS, aerial photographs, DEMs, and topographic maps.

Appendix F: Location of paleocurrent measurement sites in the gLP basin.

Site ID	Location	Coordinates (°Lat./°Long.)	Elevation (m asl)	Description
P6	South Mt. Munson	49.511751 -119.580382	380	Gulley in sediment bench, west of Kettle Valley Rail Trail
P7	Skaha Creek channels	49.448116 -119.618385	400	Gulley south of abandonned gravel pit off old Penticton-Oliver Highway. Penticton Indian Band land
P8	North Mt. Munson	49.522786 -119.575562	377	Slump scar and gulley, west of Kettle Valley Rail Trail
P9	Sun Oka	49.561304 -119.642502	390	Gulley west of highway 97
P10	Naramata Creek	49.580407 -119.589838	398	Gulley east of Sammet Rd. (private road access)
P11	Robinson Creek	49.609728 -119.59792	400	Robinson Creek confluence with Okanagan Valley, terraced gulley, northeast of Mill Rd.

Appendix G: Paleocurrent data measured in the White Silt benches of the gLP basin.

Measurement Sites					
P6	P7	P8	P9	P10	P11
185	100	241	165	250	271
177	35	223	112	244	187
200	20	218	95	186	195
231	175	176	74	193	165
200	192	180	175	200	300
236	158	300	87	300	315
241	123	200	103	284	275
194	141	215	77	260	192
187	111	193	115	200	198
200	128	284	108	193	220
213	167	205	164	177	279
168	172	197	155	194	288
300	164	216	149	177	317
284	177	285	102	260	330
199	169	177	131	255	273
281	48	186	98	222	194
206	57	289	103	256	181
217	130	293	188	165	192
226	52	302	154	170	283
245	100	201	184	225	312
261	128	194	178	236	335
259	94	189	163	340	290
220	37	217	141	257	
194	83	184	120	310	
188	56	263	117	335	
205	100	255	96	328	
193	39	249	79	228	
210	133	235	105	234	
238	81	229	114	241	
241	147	257	68	192	
	158	248	75		
	173	208	83		
		212	177		
		198	159		
		234	164		
		305	183		
		312	179		
		294	192		
		334	75		
		296	84		

Note: All measurements corrected for magnetic declination (19° E in 2002) from True North