CANQUA 2009 Pre-conference Field Trip
April 29 – May 2, 2009

Cordilleran ice sheet dynamics and decay, and postglacial landscape adjustment in southern Interior British Columbia

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Introduction

During the last glaciation the Cordilleran Ice Sheet (CIS) covered most of British Columbia. The southern margin of the CIS terminated in a series of lobes that extended into the northern Washington, Idaho and Montana (Booth 1987). In south-central British Columbia, the ice surface of the CIS was generally above 2300 m asl (Clague 1989). Ice thickness varied from ~2000 m over major valleys to a few hundred metres over major mountain ranges (Ryder *et al.* 1991).

Our understanding of the history of the Late Wisconsinan (marine oxygen isotope stage (MIS) 2) Cordilleran Ice Sheet was first summarized by Davis and Mathews (1944), and has been further refined by Fulton (1991) and Clague (1989, 2000). Davis and Mathews presented a conceptual framework of CIS evolution from its inception to glacial maximum. The work of Fulton (1965, 1967, 1969), Clague (1989, 2000), and Ryder *et al.* (1991), amongst others, has focused on the behaviour of the CIS from late glacial maximum through deglaciation.

On this trip we will examine 1) the Quaternary history recorded in the thick fills of the Fraser and Okanagan valleys; 2) evidence for MIS 2 CIS dynamics and hydrology (subglacial lakes, linked-cavities, drumlins, tunnel valleys, tills, the Chasm); 3) deglacial processes and environments (evolution and drainage of glacial lakes Thompson and Penticton); and 4) the timing and nature of postglacial landscape adjustments (postglacial incision, paraglacial fan development, and the subsequent aeolian record). Our trip through the southern Interior of BC will focus on the Fraser River valley near Clinton, the southern Fraser plateau, the Thompson plateau, and the Okanagan valley (Fig. 1).

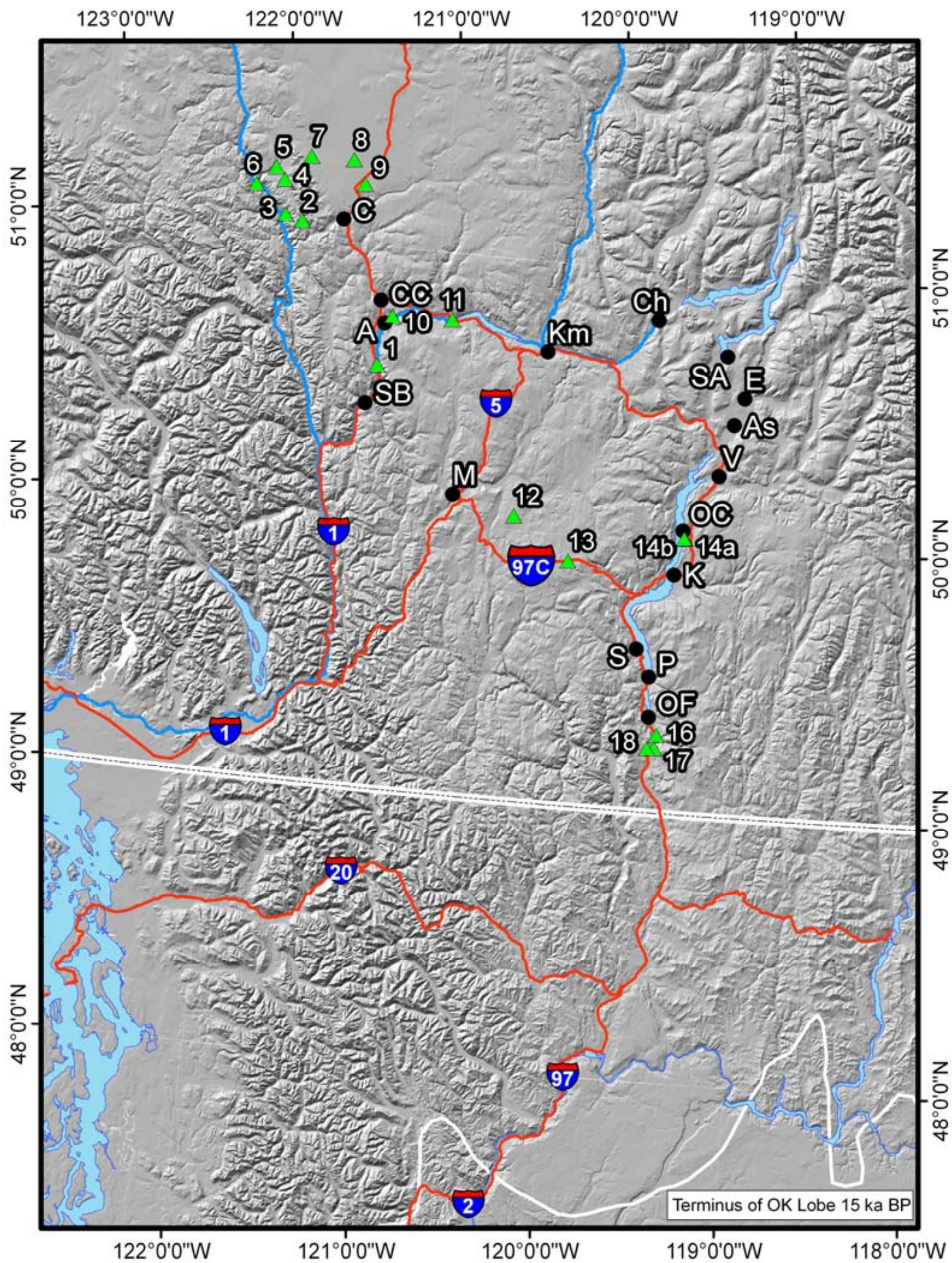

Figure 1. Stop locations (green triangles) and towns (black dots). SB = Spences Bridge, A = Ashcroft, CC = Cache Creek, C = Clinton, Km = Kamloops, Ch = Chase, SA = Salmon Arm, M = Merritt, K = Kelowna, OC = Okanagan Centre, V = Vernon, As = Armstrong, E = Enderby S = Summerland, P = Penticton, OF = Okanagan Falls. Data sources: roads, Geobase; SRTM 30 m DEM (USGS); 15 ka ice limit (Dyke *et al.*, 2003). Map projection: WGS 1984 Albers

Day 1

Stop 1: Thompson River valley (50° 33.724′ N, 121° 18.144′ W)

This is a short stop to allow field trip participants to become familiar with the valley fill and the stratigraphy and geomorphology that developed during various phases of postglacial landscape adjustment. The Thompson River valley fill near Ashcroft includes late Wisconsinan and Holocene age (MIS 1 and 2) glaciofluvial outwash, till, glacial lake sediments, fluvial sediments, and colluvial and alluvial fan sediments (Fig. 2). In places (e.g., near the mouth of the Bonaparte River) older sediments are exposed, but their age remains uncertain (Clague and Evans 2003).

Late glacial (MIS 2) glaciolacustrine sediments dominate the Thompson River valley fill north of Spences Bridge and can exceed 100 m in thickness (perhaps as thick as 140 m based on well logs). They record an ice-dammed lake, glacial Lake Thompson (Mathews 1944), that existed during the last gasps of the last Cordilleran ice sheet. Such lakes are thought to have resulted from regional backwasting and downwasting of the ice over the plateaus when the equilibrium line rose above the elevation of the ice surface (Fulton 1991; Fig. 3). Given possible ice thickness variations between the plateau and dissecting valleys, regional downwasting may have resulted in ice free plateaus when the valleys contained ice tongues that impounded glacial lakes (Fulton 1965, 1967, 1969; Ryder 1976; Clague 1989; Fig. 3). We will return to the story of glacial Lake Thompson toward the end of day 2.

Seen at this stop are glaciolacustrine sediments associated with glacial Lake Thompson into which ancestral Thompson River has incised. Relatively short periods of fluvial aggradation during postglacial incision are recognized by thick units of fluvial gravel and sand deposited on incisional terraces. As fluvial incision commenced, paraglacial fans (alluvial and/or mudflow fans) were deposited on fluvial gravel and sand. After paraglacial fan development ceased, the surfaces of the fans became capped with up to several metres of eolian sediments. These eolian units are currently vegetated and, for the most part, stable. Details on the timing of postglacial fluvial incision and paraglacial fan development will be discussed in the context of the Fraser River valley at stop 6.

Stop 2: Jesmond Road, Jesmond valley (51° 3.858′ N, 121° 49.564′ W)

This stop will introduce field trip participants to some of the till forming processes that were taking place during the last (Fraser) glaciation in mountain valleys in the region. More specifically, at this site, the till has been interpreted to have been deposited initially mainly by lodgement, but it then evolved into a deformation till as it thickened and as its pore water content increased. At that stage the till would have begun to deformed in a ductile manner beneath the glacier, and ice flow would likely have changed from compressive to extending flow in response. In other words, ice flow velocity would have increased. There is also evidence at this site to suggest a later phase of dewatering, and a return to brittle deformation. The character of the till at this stop is in contrast to that at some other sites in the valley that indicate that lodgement processes dominated throughout glaciation; sites such as this would have acted as "speed bumps" (or so-called "sticky spots") along the glacier's path (see Lian & Hicock 2000).

Also seen at this stop, directly underlying Fraser Glaciation till, are several metres of indurated gravel and sand that represents fluvial aggradation in the valley sometime before 780 ka, perhaps in response to the onset of a mid-Pleistocene glacial episode.

Stop 3: Cougar Point (51° 4.913' N, 121° 55.696' W)

This stop is situated on the east side of the Fraser River valley, some 1000 m above the valley floor. It allows for a spectacular view of the dissected and terraced Quaternary valley fill sediments that are nearly 500 m thick. At this stop an overview of the valley-fill stratigraphy will be given.

At the base of the valley fill, and resting on bedrock, are thick (up ~20 m thick) sequences of glaciofluvial-deltaic gravel and sand. These sediments are, in places, overlain with, or interbedded with, up to 25 m of matrix-supported diamicton (flow till?). These lowest units are interpreted to represent aggradation associated with the approach of glacial ice (glacial advance stage). They are overlain by up to 350 m of horizontally bedded sand and silt and minor diamicton that was deposited in glacial Lake Camelsfoot. It is capped, in places with several metres of till. Optical dating of the glaciolacustrine sediments have revealed that this glacial sequence was all deposited during the last glaciation, although valley-fill sediments interpreted to represent the remnant of an older glacial fill have been documented by others at sites ~100 km to the north. See Lian & Hicock (2001) for a detailed account of the Fraser valley fill stratigraphy in the immediate area. Set into the glacial valley fill are sedimentary units and landforms that developed after deglaciation, and these will be discussed at stop 6.

Stay in Clinton for the night.

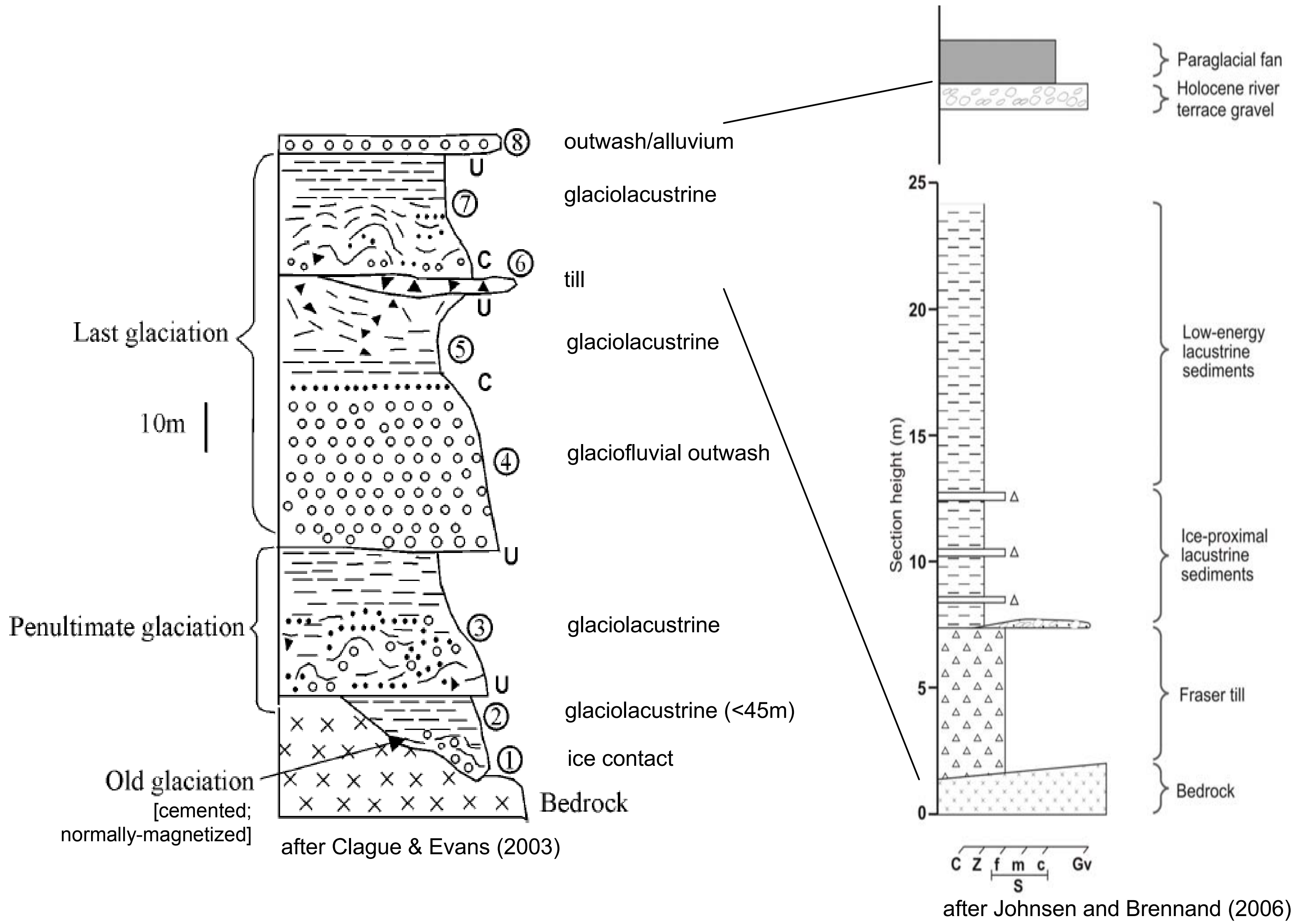

Figure 2. Generalized Quaternary sediment fill in Thompson River valley near Ashcroft

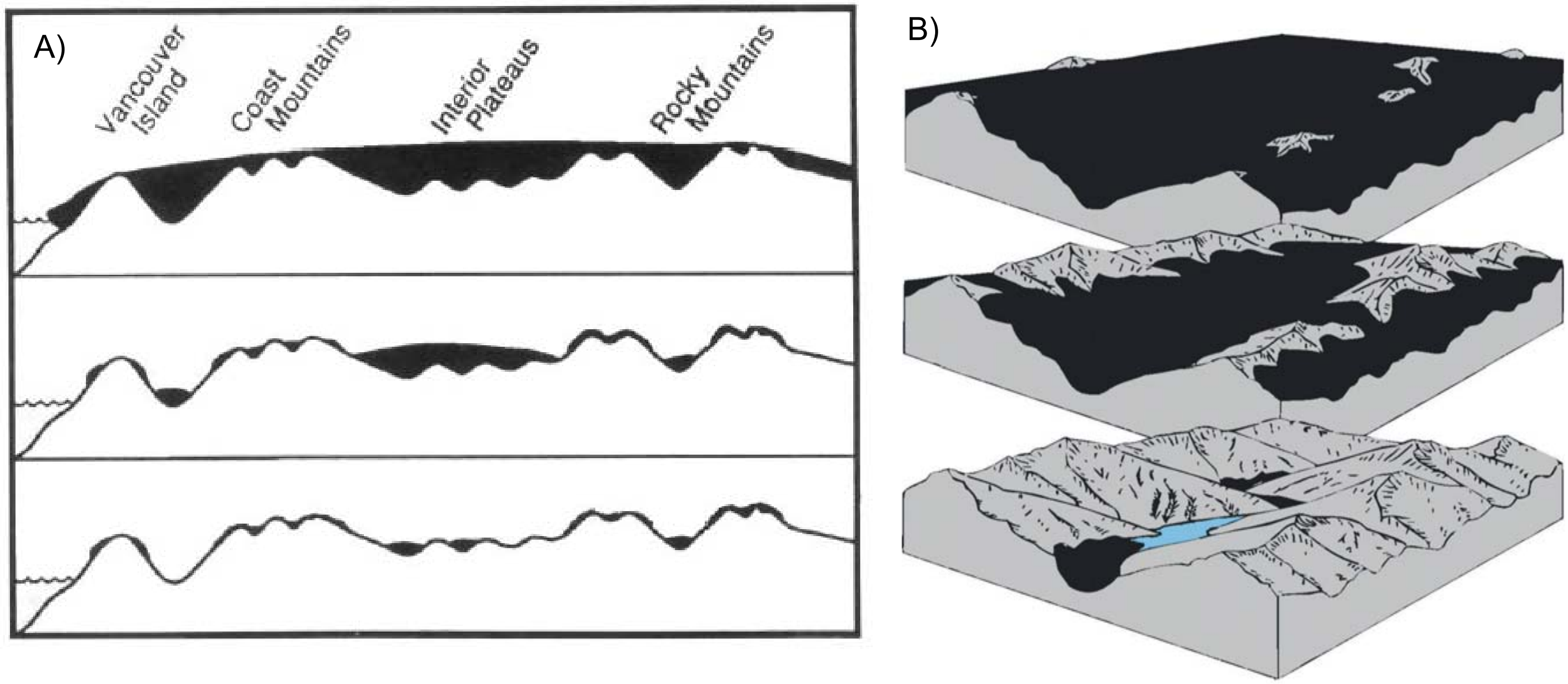

Figure 3. Conceptual models of the decay of the Cordilleran Ice Sheet. A) Following CIS climax, a rise in equilibrium line altitude to near the tops of mountains caused regional downwasting and resulted in scattered remnants of the CIS in the interior and its main valleys, and alpine glaciers at high elevation (Fulton 1991). B) In the latter stages of regional downwasting residual dead ice masses became confined to valleys and impounded glacial lakes (Clague 1989, after Fulton 1967).

Day 2

Stop 4: Jesmond Road (51° 12.608′ N, 121° 57.336′ W)

This stop is in a gravel pit and gives a view of the typical character of the glaciofluvial sediments that cap the Fraser Plateau in this region. Exposed in the gravel pit are several metres of poorly-stratified gravel and sand that rest directly on consolidated, massive, clast-supported diamicton (till). Of particular interest at this site is the appearance of large clasts of diamicton in the glaciofluvial gravel and sand, which, together with the paucity of stratification, suggests rapid deposition.

Stop 5: Big Bar Creek Road (51° 15.110′ N, 122° 0.834′ W)

From this stop one can view on the north side of the valley, in the distance, the remnants of a valley fill that represents a glacial advance in the region that occurred in the mid Pleistocene. The exposure begins about 200 m above Big Bar Creek Road (above the valley floor), and consists of a 100 m-thick sequence of indurated glacio(?)fluvial gravel and sand, which is overlain with indurated glacigenic diamicton, and then capped fluvial sand and gravel. The diamicton is interpreted to be reworked (slightly?) till, or tillite. A sand lens within the till is reversely magnetized, indicating that it represents a glacial advance that occurred more that 780 ka ago. Fluvial sand near the base of the sequence has, however, normal magnetization. This suggests that a remnant of an even older (>1.05 Ma, or more likely >2.6 Ma) aggradational sequence occurs in the valley. See Lian *et al.* (1999).

Stop 6: Big Bar Creek lookout (51° 10.953′ N, 122° 7.157′ W)

This stop gives an excellent view of the landforms associated with postglacial adjustment in this region of the Fraser River valley. Several prominent terraces can be seen on the west side of the valley, set into glacial Lake Camelsfoot sediments. The terrace stratigraphy here is virtually identical to that which was seen along the Thompson River valley at stop 1, although here many of the larger paraglacial fans have been incised by tributary steams. This longitudinal incision of the fans, and of the underlying terrace sediments, provides exposures where the sedimentology and stratigraphy can be studied, and samples for optical dating can be collected. Optical dating of fluvial sediments on river terraces indicate that ancestral Fraser River was flowing about 170 m above its present position at about 11 ka (cal yrs), and had incised to within a few metres of its present position sometime between 6 and 4 ka. Optical dating has also indicated that the initial phase of paraglacial sedimentation had essentially ended by this time. This information is largely unpublished, but the results of an initial study can be found in Lian & Huntley (1999)

Stop 7: Fraser Plateau, Big Bar Lake Road

This "stop" includes various stops along the Big Bar Lake road on the Fraser Plateau, north of the Marble Range, to view the landscape surface. Of particular interest are many boulders of local lithology (flow basalt), many of which do not appear to have been glaciated.

Stop 8 (only if time permits): Esker on the Fraser Plateau (51° 18.508′ N, 121° 33.534′ W)

This stop gives a close up view of one of the eskers on the Fraser Plateau. This esker is of particular interest as it represents part of a fluvial system that leads into The Chasm (stop 9).

Stop 9: The Chasm (50° 33.724′ N, 121° 18.144′ W)

This stop is at Chasm Provincial Park where a remarkable bedrock canyon (a retreating knickpoint) exists that is thought to have formed as a result of glacial meltwater activity. It is currently unknown whether it formed entirely at the end of the last glaciation, or whether it is the product of fluvial activity associated with several glaciations. Also observed at this stop is the remarkable flood basalt/burnt paleosol stratigraphy of the upper 40 meters of the Fraser Plateau (see Farrell *et al.* 2007, 2008).

Stops 10 and 11: Glacial Lake Thompson

High cliffs (up to > 100 m) of late Wisconsinan lake sediments, deltas and wave-cut benches occur within the Thompson valley from the outlet of Kamloops Lake at Savona to a few kilometres south of Spences Bridge (Fig. 4). These sediments record the presence of a late-glacial ice-dammed lake within the Thompson basin. Previous research has documented two glacial lakes in the Thompson basin: glacial Lake Thompson and glacial Lake Deadman, each with two stages (Fulton 1969; Ryder 1981). These lakes were dammed by ice to the west or south (e.g., near Spences Bridge), had different eastern outlets (within the Shuswap basin and at Kamloops), yet similar paleogeography within the Thompson basin (Fulton 1969, Ryder 1981); they can reasonably be viewed as different stages of the same lake (glacial Lake Thompson; Mathews 1944; Johnsen and Brennand 2006). We will make two stops to observe landforms and sediments associated with this lake and to discuss its paleogeographic evolution and associated paleoenvironment.

Stop 10: Paleo-Bonaparte subaqueous fan north of Ashcroft (50° 44.877′ N, 121° 14.510′)

This stop is a ~100 m high valley-side exposure that provides a view of the character of sedimentation in glacial Lake Thompson and informs our understanding of lake paleoenvironment during CIS decay.

At the base of the exposure sediments of the paleo-Bonaparte subaqueous fan are visible (1, Fig. 5). This fan was build by sediment delivered by the paleo-Bonaparte outwash stream as evidenced by local topography, the southeast dip of the fan surface and paleoflow measurements (Figs 5, 6). The fan is overlain by fine-grained lake-bottom sediments (mainly laminated silt, minor sand), river terrace gravel and paraglacial fan sediments (Fig. 5). Numerous normal faults in the glaciolacustrine sediments and river gravels record the melting of a large buried ice block (~60 m tall and 150 m wide). Paraglacial fan sediments containing tephra (likely Mazama tephra, J. Clague personal communication 2000; 7627 ± 150 cal yr BP, Zdanowicz *et al.* 1999) filled the kettle hole; they are not faulted.

The subaqueous fan contains coarse-grained lithofacies characteristic of high-energy sedimentation dominated by hyperpycnal flows (Johnsen and Brennand 2006). Undulatory stratified gravel and diffusely graded sand suggest very high rates of deposition from suspension. Planar stratified sand, trough cross-bedded gravel and sand, and imbricate gravel beds suggest deposition from traction transport. Diamicton beds record subaqueous debris flows. Beds of laminated silt punctuate fan sediments and record quiet-water suspension sedimentation during pauses in inflow. Gravel and sand beds overlying laminated silt beds often contain large laminated silt rip-ups and exhibit sharp, irregular or loaded lower contacts, evidence of the erosive power of, and rapid sedimentation from, hyperpycnal flows. Clastic dikes injected through diffusely graded sand beds record rapid loading and dewatering of deeper sediments. These lithofacies and lithofacies associations are similar to those attributed to jökulhlaup deposition (e.g., Russell and Arnott 2003). The Chasm lies at the head of Bonaparte valley, lending credence to the possibility of an extreme discharge event related to the formation of the paleo-Bonaparte subaqueous fan. That a subaqueous fan rather than a delta developed at the mouth of the paleo-Bonaparte outwash stream is likely due to the presence of a deep basin (insufficient sediment to fill the accommodation space), the short lifespan of the lake (insufficient time to fill the accommodation space), and the character of the outwash valley (Johnsen and Brennand 2006). The high sediment load required to generate hyperpycnal flows could have been generated and maintained in a steep valley-confined stream draining plateau-remnant ice and bordered by erodible, unvegetated slopes.

Glacial Lake Thompson was an ice-contact lake: it was dammed by ice near Spences Bridge, and stratigraphically, glaciolacustrine sediments overly Late Wisconsinan till, may be interbedded with diamicton at their base (lowest 5 m) or are separated from the till by discontinuous lenses (< 3 m thick) of fluvially-sorted gravel, suggesting that as soon as the ice melted from a location within Thompson valley, a lake developed in which sediments accumulated. This lake may have had a short-lived supraglacial phase, in places, as evidenced by numerous kettle-holes and kettle fills (Fig. 5). However, the majority of its shoreline abutted valley walls and was not in contact with ice. Indeed, the location of deltas and subaqueous fans at tributary mouths, and paleoflow direction measurements and trends in lithofacies away from tributary mouths suggest that tributaries were the dominant source of sediment and water for glacial Lake Thompson, not valley-occupying ice (Fig. 6). Taken together, lithofacies, lithofacies associations and landform-sediment relationships record high rates of sedimentation and a dynamic and often high-energy lake environment. These conditions, in combination with a fining upward sequence in lake bottom sediments, suggests that glacial Lake Thompson was coeval with decaying ice on the plateaus (Mathews 1944; Johnsen and Brennand 2006). This pattern of ice decay is perhaps counter to the prevailing view that rapid downwasting of Cordilleran ice over high relief terrain resulted in ice free plateaus when ice-dammed lakes occupied the valleys (e.g., Fulton 1967; Clague 1989), and leads us to question whether ice over all Interior valleys was necessarily thicker than plateau ice at the onset of deglaciation: antecedent ice streams and (or) subglacial lakes in deep valleys draining the CIS may have resulted in thinner ice than previously thought in some valleys.

Stop 11: Deadman delta and wave-cut benches of glacial Lake Thompson (50° 45.359' N, 120° 53.359' W)

This stop is on Deadman delta on the north flank of Thompson valley, and (if the light is right) we can see a pair of wave-cut benches on the south flank of the valley (6a, 6b, Fig. 4; Johnsen and Brennand 2004). Deadman delta is the largest Gilbert-type delta in the region (13 km^2; Fig. 6). The variable dip directions and angles of its foresets record lobate delta progradation. The size of the delta, paleoflow measurements away from delta (Fig. 6), record of hyperpycnal flows in delta foresets and lake bottom sediments, and underfitness of the Deadman River suggest that the paleo-Deadman outwash stream played a dominant role in delivering water and sediment to glacial Lake Thompson. This role was likely facilitated by the presence of remnant Cordilleran ice in the headwaters of Deadman outwash stream (Johnsen and Brennand 2006).

The elevation of deltas and wave-cut benches in Thompson valley allow paleogeographic reconstruction of lake extent and evolution (Johnsen and Brennand 2004). Reconstructions confirm the southern extent of glacial Lake Thompson, and likely final position of its ice dam, near Skoonka Creek (~7 km south of Spences Bridge), where glacial lake sediments terminate abruptly. Two lake stages have been regionally identified and define a narrow (> 6.5 – 1.5 km), deep (> 140 – 50 m) lake (Johnsen and Brennand 2004). At high stage, glacial Lake Thompson contained ~84 km^3 of water and drained east along ice margins and topographic divides into the Shuswap basin and thence to the Okanagan basin (Fulton 1969; Fig. 7). As ice backwasted and downwasted, and lake level lowered the eastern outlet became a spillway incised into the South Thompson silts east of Kamloops (Fulton 1969; Fig. 7); spillway incision totaled ~65 m prior to lake drainage. Most of the lake level lowering between the high and lowest stages was likely progressive (or punctuated) rather than catastrophic because (1) wave-cut benches and inset delta surfaces are found between the two stages (Fig. 4), and (2) base level was mainly controlled by readily eroded valley fill (lake bottom silt; Johnsen and Brennand 2004). At its lowest stage, glacial Lake Thompson contained ~24 km^3 of water.

Deadman delta exhibits nested delta surfaces: three dominant delta surfaces at 540, 480 and 450 m asl (topset-foreset contacts), and several intermediate surfaces (Fig. 8). Nested delta surfaces can result from steady or episodic lake-level lowering (Muto and Steel 2004). The eroded lower slopes of Deadman delta exhibit several kettle holes (Fig. 8), some with concentric retrogressive slump failure ridges suggesting collapse following the meltout of buried ice blocks.

Glacial Lake Thompson drained catastrophically when the ice dam around Spences Bridge failed. The resulting jökulhlaup discharged ~20 km^3 of water (~4 km^3 was retained east of Deadman delta and formed the high stage of Kamloops Lake; Johnsen and Brennand 2004) southward along the Thompson and Fraser valleys. Drainage likely occurred sometime before 9 740-10 210 BP (GSC-193, Dyck *et al.* 1965; organic silt from the bottom of Otter Lake spillway of glacial Lake Shuswap near Armstrong and the minimum age for all glacial lakes in the southern Interior, Fulton 1969), and marine flood deposits (exotic muds) dated between 11 940 and 10 190 BP in the Strait of Georgia (Conway *et al.* 2001) and Saanich Inlet (Blais-Stevens *et al.* 2001) may have been derived from the glacial Lake Thompson (and (or) Glacial lake Fraser) jökulhlaup (Johnsen and Brennand 2004).

The lower slopes of Deadman delta record evidence of catastrophic lake drainage in the form of lake drainage bedforms (Mathews (1944) noted a "peculiar undulating … surface"). These bedforms are two-dimensional and asymmetrical (steeper stoss slopes) with wavelengths of 100-230 m and heights of 3-7 m (some are cross-cut by the kettle-slump ridges) (Fig. 8); they are perhaps best classified as antidunes (Carling *et al.* in press). Delta foreset truncation in ground penetrating radar profiles (Fig. 8), deep scours in the troughs of some antidunes, near-surface backset beds in shallow (1.5 m deep) sediment trenches, and ridge parallel surface boulders suggest that bedforms in the upflow part of the field are the product of phases of erosion and deposition associated with spatially and temporally varying water depth and velocity induced by the topography of the bedform and lake drainage progress. Downflow the bedforms increase in size and are composed of gravel backset beds. The increase in flow depth and hence accommodation space may explain the apparent downflow transition to fully depositional antidunes. Flutes ornament these downflow antidunes (Fig. 8), indicating late-stage erosion by longitudinal vortices. Steep stoss slopes were likely oversteepened through erosion by roller vortices developing at these upflow-facing steps. Paleohydraulic reconstructions using antidune wavelength, suggest that the minimum water depth during antidune formation was ~16-36 m (consistent with reconstructed water depths), and the flow velocity was ~13-19 m s^{-1} (Carling *et al.* in press).

Following glacial Lake Thompson drainage, Kamloops Lake developed to the east of Deadman delta (Fig. 7), and the Thompson River flowed southward and incised through ~150 m of valley fill to bedrock, deposited gravel and developed numerous terraces (Johnsen and Brennand 2004). Incision to within a few metres of present river level was achieved by the mid-Holocene (Ryder 1981; 7 510-7 670 BP, Hallett *et al.* 1997); ~14 km^3 of valley fill has been eroded during and since lake drainage. The former shorelines of glacial Lake Thompson have been tilted (up to 1.8 m km^{-1} to the north-northwest) by differential glacio-isostatic uplift (Fig. 4; Johnsen and Brennand 2004), lending support to the notion of an ice divide centred on the Fraser Plateau north-northwest of the Thompson valley (e.g., Ryder *et al.* 1991).

Stay in Merritt for the night.

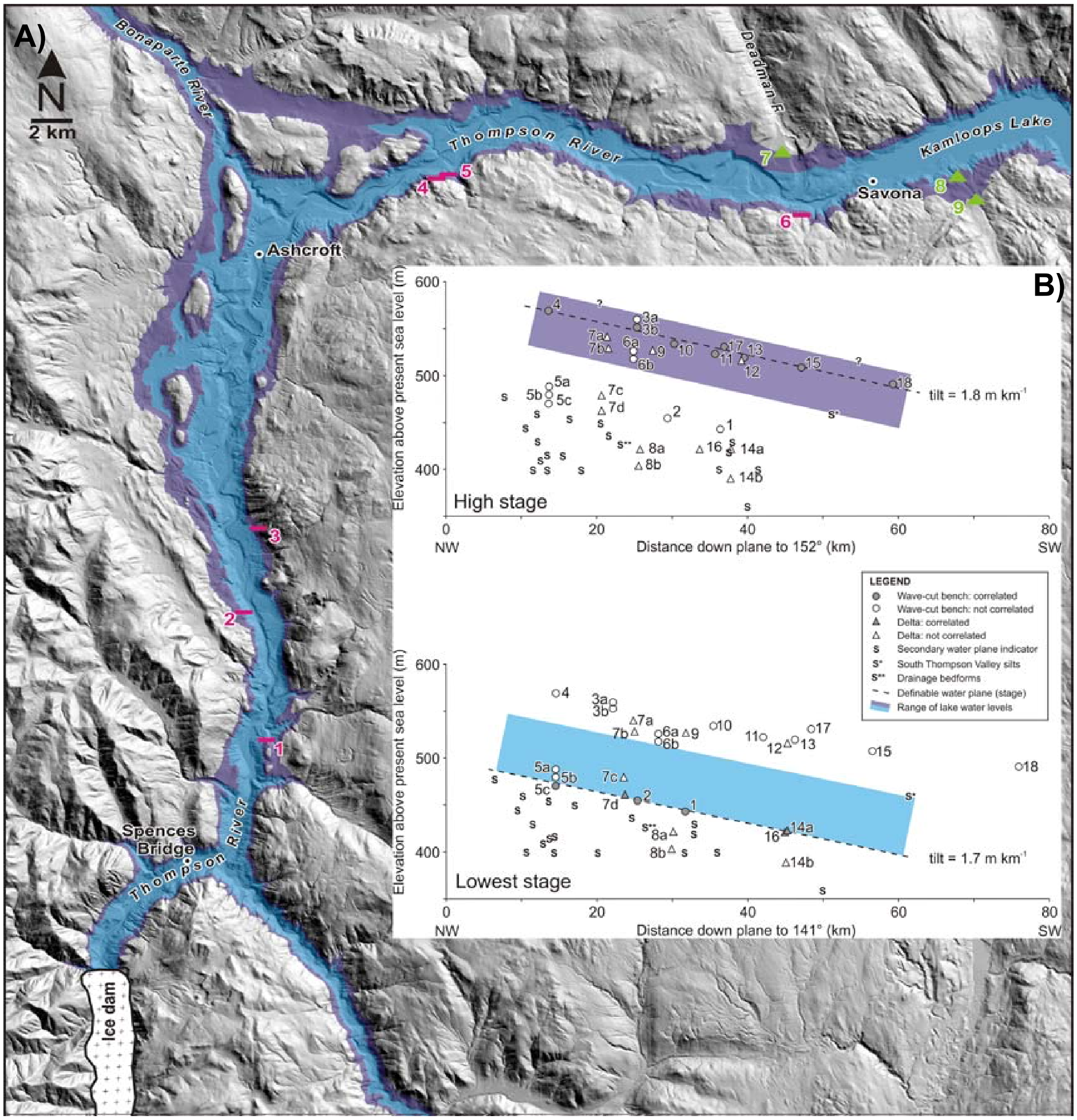

Figure 4. Stages of glacial Lake Thompson. A) Map of a high and the lowest stage of glacial Lake Thompson in the western part of the Thompson basin. Location of ice dam, deltas (green triangles) and wave-cut benches (pink bars). B) Best-fit (first order trend surface) water planes for a high and the lowest stage of glacial lake Thompson. Shaded bars indicate the range of water levels with outlets in the Shuswap basin (high stage) and through the south Thompson silts (lowest stage). Modified from Johnsen (2004) and Johnsen and Brennand (2004).

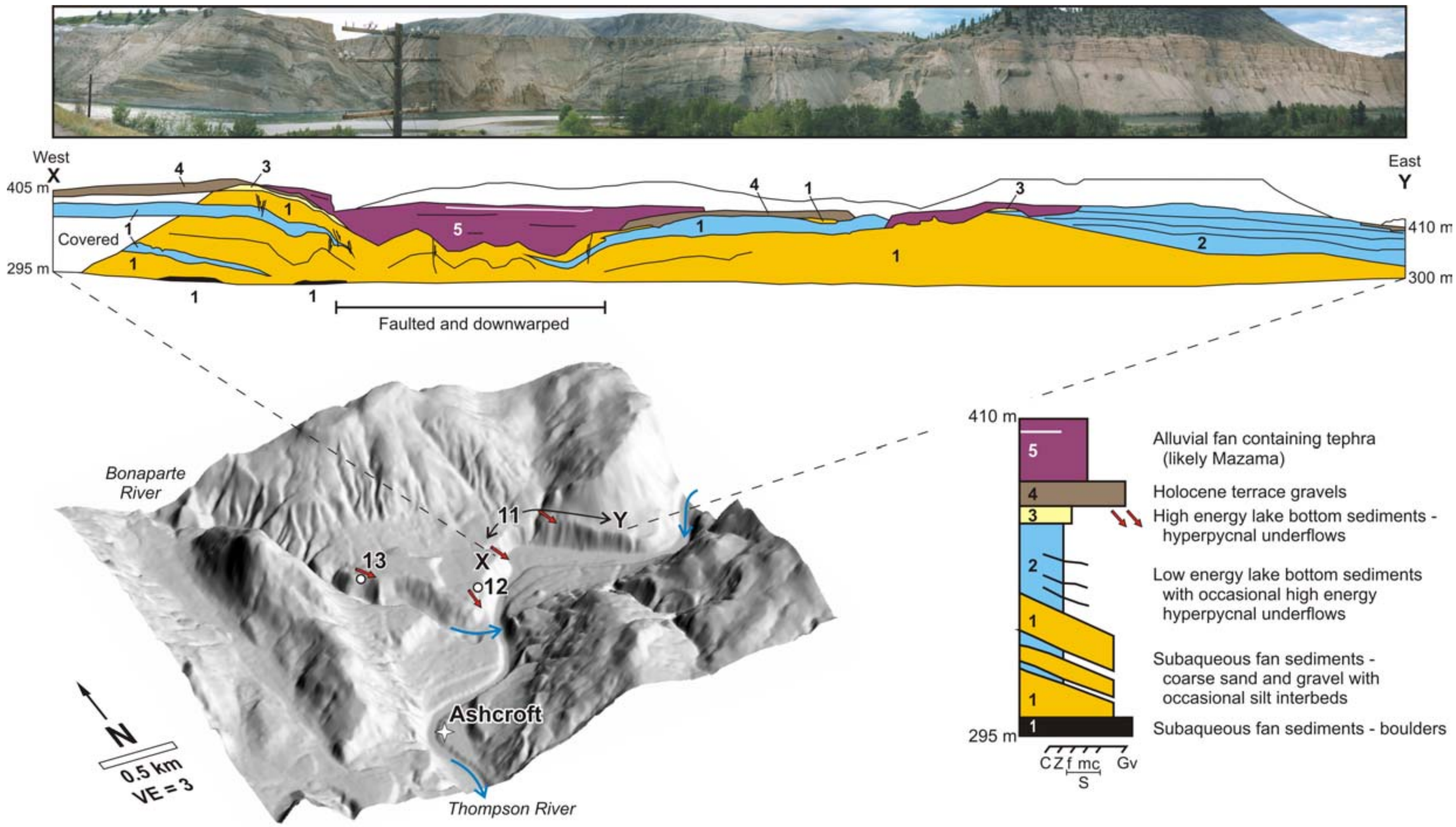

Figure 5. Paleo-Bonaparte subaqueous fan. A) Photographic panorama and interpreted sedimentary sequence of section 11 (2 km long, 110 m high). Colluviated portions of sections inferred. B) Perspective hillshaded digital terrain model showing geomorphic context of sections (white dots and black arrows), paleoflow directions (red arrows) and present river direction (blue arrows) (DEM data, British Columbia Government 1996). C) Generalized sedimentary sequence of section 11. Modified from Johnsen and Brennand (2006)

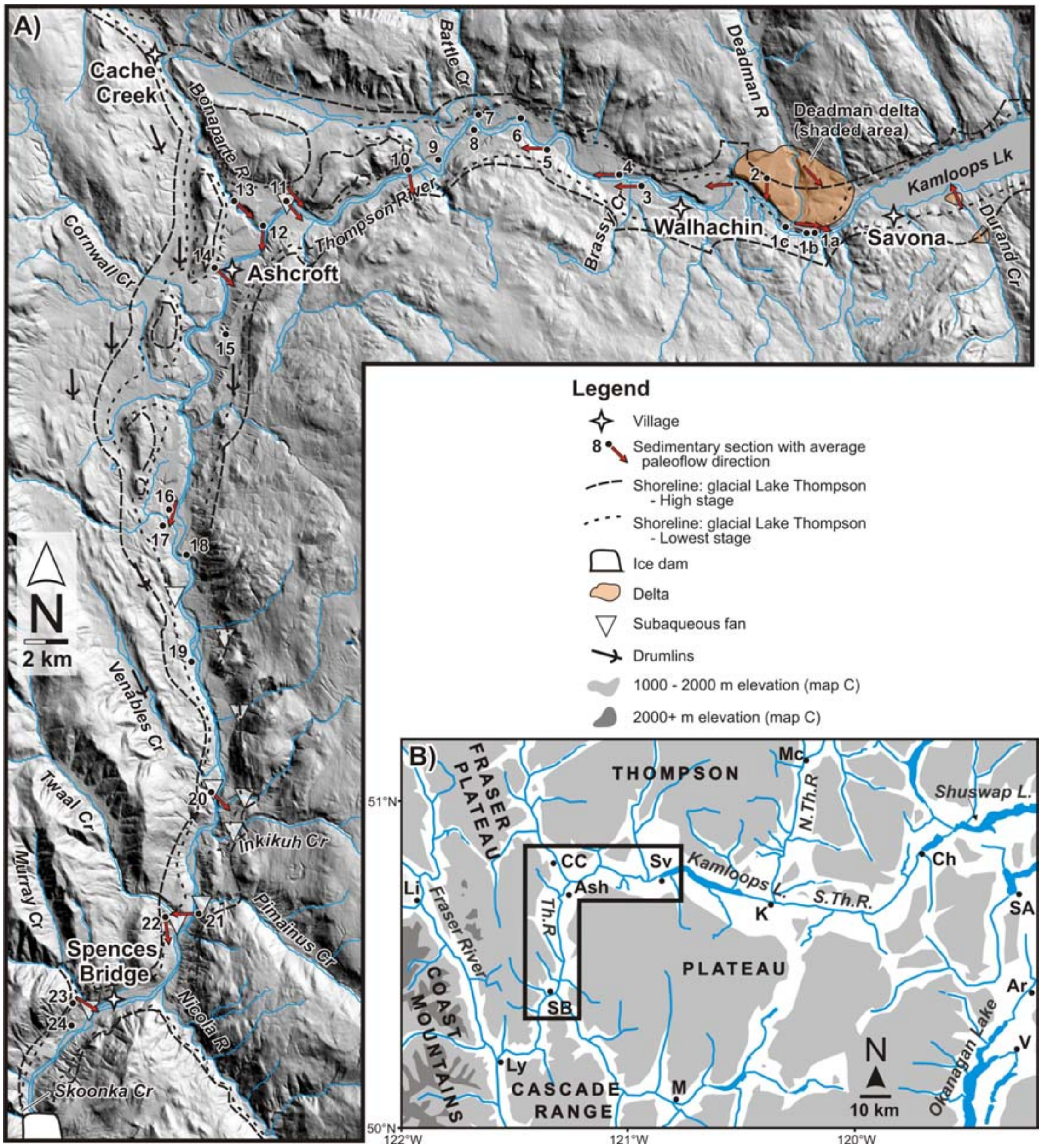

Figure 6. A) Hillshaded digital terrain model of the Thompson basin showing locations of sedimentary sections, deltas, subaqueous fans, measured paleoflow directions, paleoshorelines and the ice dam associated with glacial Lake Thompson (DEM data, British Columbia Government). B) Regional physiographic context and place names. Map abbreviations: Li = Lillooet, Ly = Lytton, SB = Spences Bridge, M = Merritt, Ash = Ashcroft, CC = Cache Creek, Sv = Savona, K = Kamloops, Mc = McClure, Ch = Chase, SA = Salmon Arm, Ar = Armstrong, and V = Vernon. *Th.R* = Thompson River, *N.Th.R* = North Thompson River, and *S.Th.R* = South Thompson River. Modified from Johnsen and Brennand (2006).

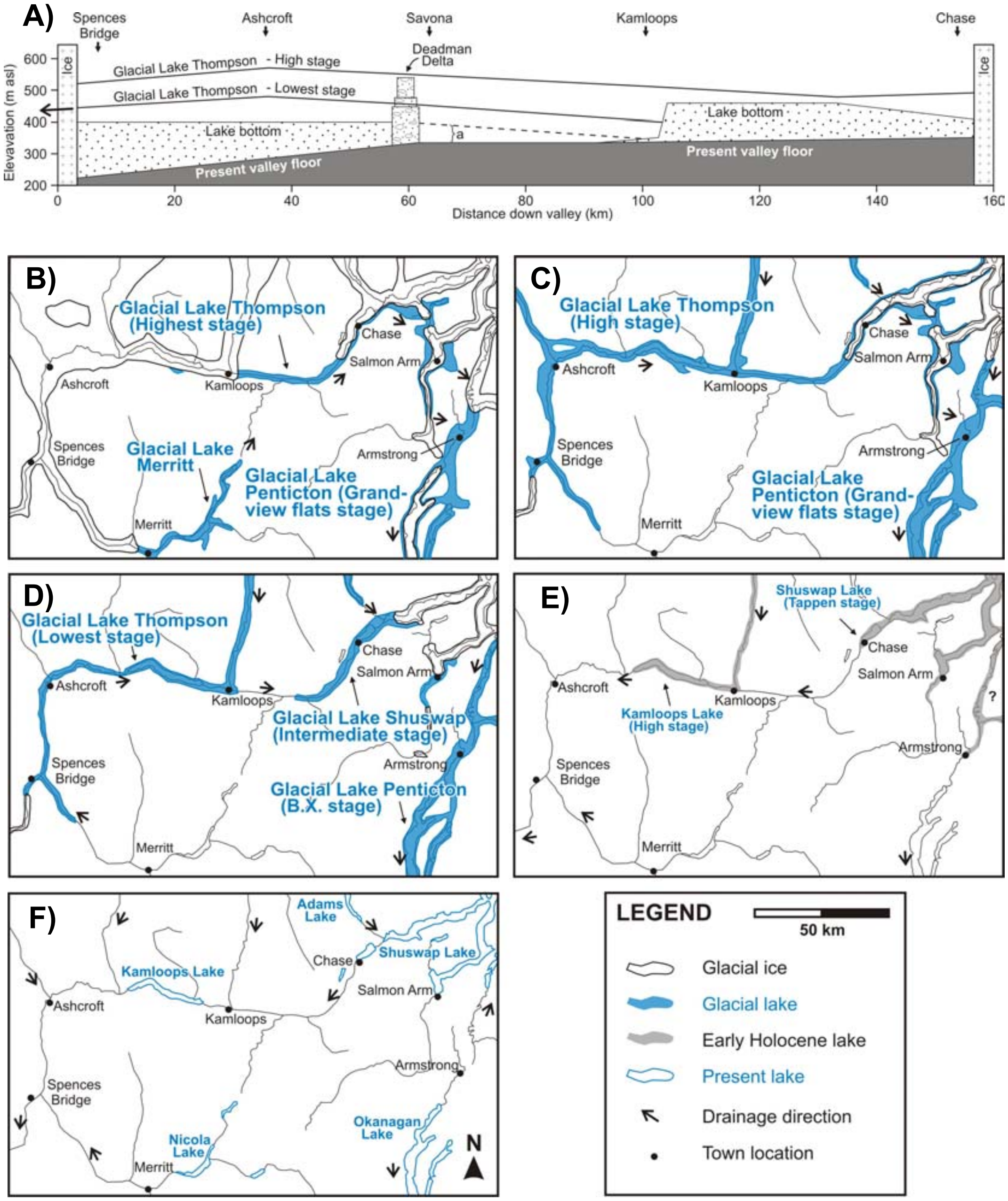

Figure 7. Late glacial lake evolution in the southern Interior of British Columbia (modified from Johnsen and Brennand 2004). A) Cartoon of down-valley projections of water planes and inferred lake bottom elevations from Spences Bridge to Chase along Thompson and South Thompson valleys. Changes to the slope of the water plane result from changes in valley orientation. B-F) Map view of lake evolution. Three stages of glacial Lake Thompson are identified. Two stages of Kamloops lake shown. Glacial lakes and stage names from outside the Thompson basin after Fulton (1969). Earlier stages of glacial Lake Merritt and glacial Lake Penticton not shown. F) shows modern drainage.

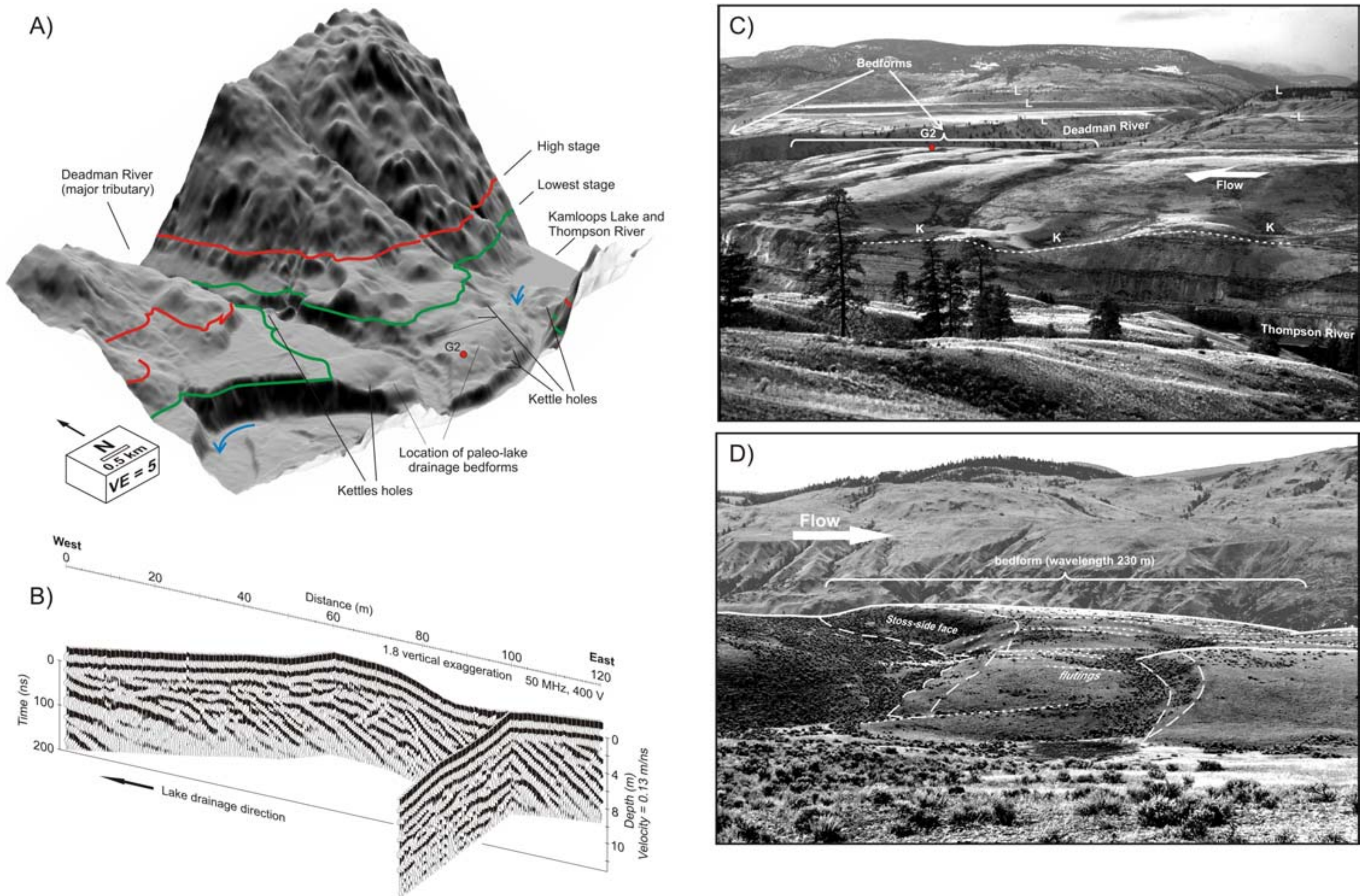

Figure 8: Deadman delta. A) Perspective (5x vertical exaggeration) hillshaded digital terrain model (DEM data, British Columbia Government 1996) showing projected lake stages, inset delta surfaces, kettle holes, lake drainage bedform fields and location of ground penetrating radar (GPR) survey (G2). Thompson River flow direction, blue arrows. B) GPR profile, G2, streamwise across a lake drainage bedform in the upflow (eastern) part of the field. Antidune bedforms C) in the upflow (eastern) portion of the field and D) in the downflow (western) portion of the field. L, nested delta surfaces, K, kettle or scour holes. Note the steep stoss slopes and superimposed flutings in D). Modified from Johnsen and Brennand 2004, and Carling *et al.* in press.

Day 3

Stop 12: Douglas Lake drumlins (50° 3.385′ N, 120° 25.908′ W)

Drumlins composed of bedrock and/or sediment are common within the footprint of the CIS. They have been attributed to enhanced ice abrasion (Evans 1996), till accretion (Goldstein 1994), meltwater erosion (Shaw 1996), and a combination of subglacial deformation and meltwater erosion (Mate 2000). They have been associated with ice flow directions, ice streams (Evans 1996), ice retreat (Goldstein 1994), and flood (underburst) tracks (Shaw 1996). Given the lack of consensus on drumlin genesis, perhaps equifinality and (or) polygensis in CIS drumlin production are possibilities, yet detailed form-process studies on which to evaluate these propositions are lacking, in particular knowledge of the internal architecture of drumlins is rarely known, yet is critical to evaluating process models.

The Douglas Lake drumlins occur within a N-S oriented, drumlinized corridor (the Fraser-Thompson-Okanagan corridor), upward of 65 km wide, that extends ~250 km from Wells Gray Volcanic field on the eastern margin of the Fraser Plateau, to Okanagan and Similkameen valleys, and arguably some 500 km to the terminus of the Okanagan Lobe (although certainly discontinuous in this extension). Non-drumlinized patches occur within this corridor. At this stop we will discuss drumlin shape (Fig. 9; Lesemann and Brennand in press) and composition and their implications for drumlin genesis. We will present unpublished electrical resistivity surveys of drumlins collected by Tracy Brennand, Jerome Lesemann and Darren Sjogren.

Stop 13: Trepanege Ridge drumlins and tunnel valleys (49° 54.415′ N, 120° 5.953′ W)

At this road cut we can view the internal composition of the tail of a spindle-shaped drumlin on the flanks of Trepanege Ridge. This drumlin is composed of bedrock and diamicton. The diamicton is fissile, with a loamy texture (~10-15% clast content), contains silt stringers and exhibits fractures dipping to the southeast. Clasts exhibit striae, plucked ends and keels and have a spread unimodal fabric that parallels Trepanege Ridge and is oblique to the drumlin axis. We will discuss the implications of drumlin composition and sedimentology for drumlin genesis both at this site and more regionally within the Trepanege Ridge (and basin) drumlin swarm. This information is largely unpublished and has mainly been assembled by Jerome Lesemann.

N-S oriented valleys up to ~10 km in length cross the highway around stop 13 (Fig. 10; Lesemann and Brennand in press). These valleys follow a path up and over Trepanege Ridge; they exhibit convex longitudinal profiles. They are sediment-walled and deeper (to ~100 m deep) upflow and downflow from Trepanege Ridge, but bedrock-walled and shallower (25 m) over the ridge crest. Sorted and heterogenous gravel deposits partially fill the valleys, and eskers are present on some valley floors. These landforms are tunnel valleys that record channelized subglacial meltwater flow.

Stop 14: Okanagan Centre section (50° 1.386′ N, 119° 26.277′ W)

The Okanagan Lake basin is approximately 120 km long and up to 3-5 km wide. The modern valley floor is ~330-340 m asl, a relief drop of ~900 m from Thompson Plateau, but the bedrock floor of the valley is up to -650 m asl in places (MacAulay and Hobson 1972; Eyles *et al.* 1991), this overdeepening mainly attributed to glacial erosion (Fulton 1972, Eyles *et al.* 1991, Vanderberg and Roberts 1996). Valley fill is up to ~800 m thick in places (Eyles *et al.* 1991). Interpretations of paleoenvironmental change contained within this fill are contentious and equivocal.

Pioneering work by Fulton (1972) and Fulton and Smith (1978), led to the identification of at least two glacial and two interglacial/interstadial intervals (dating back to MIS 5), in the thick sedimentary sequences contained within the deeply incised valleys of the southern Interior. The Okanagan Centre section is a ~100 m thick exposure of sediments on the shore of Okanagan Lake and is the type section for MIS 4 sediments (Okanagan Centre Drift) in the southern Interior (Fig. 11). The assignment of Okanagan Centre Drift to MIS 4 relies on geoclimatic markers such as tills (glacial interval) and paleosols (interglacial/interstadial intervals) for which little detailed sedimentology is reported, and on stratigraphic correlation to other regional stratotypes, some of which have been dated with radiocarbon ages (detrital and *in situ*) or tephrachronology (Fulton and Smith 1978).

Using lithostratigraphic correlation of boreholes (two of which contained organic material (likely detrital) dating back to the last interglacial), seismic interpretations, and the Okanagan Centre section as a stratigraphic framework for the northern end of the Okanagan basin, Fulton (1972) suggested that the Okanagan valley fill records a complex depositional history spanning multiple glacial-interglacial cycles, although he acknowledged that Fraser till was generally absent along the valley axis. In contrast, more recent lake-based (Eyles *et al.* 1991) and land-based (Vanderburgh and Roberts 1996) seismic surveys have revealed a relatively simple valley fill architecture, possibly limited to MIS 2 (Late Wisconsinan (Fraser) glaciation) (Vanderburgh and Roberts 1996). Both studies highlight the presence of thick subglacial meltwater gravels at the base of the sequence (although Eyles *et al.* (1991) acknowledge that these gravels may be significantly older than MIS 2) and even thicker deglacial lake sediments forming the bulk of the fill. These potentially conflicting interpretations, and the reliance upon the Okanagan Centre section as a stratigraphic model, provide impetus for refining our understanding of the timing and nature of environmental changes preserved in the Okanagan Centre section.

A detailed sedimentological investigation of the lithostratigraphic units comprising the Okanagan Centre sequence reveals successive deposition of subaqueous and subaerial outwash possibly during glacier advance, deposition of subglacial till during glaciation, followed by glaciolacustrine sedimentation during glacier retreat, and postglacial fan and loess deposition (Lesemann *et al.* unpublished). In this interpretation, diamicton units record a single glacial advance and debris flow events that were deposited in a paraglacial fan, not separate glacial advances. Tephrachronology (glass shard chemistry and glass encased magnetite chemistry) has been inconclusive to date, but optical age estimates support these interpretations: most sediments were deposited between 3 000 and 25 000 yrs ago (MIS 2). This chronology is consistent with seismic interpretations (Eyles *et al.* 1991; Vanderburgh and Roberts 1996) and sedimentological interpretations (Vanderburgh and Roberts 1996) supporting an MIS 2 age for the majority of Okanagan valley fill.

Stay in Penticton for the night.

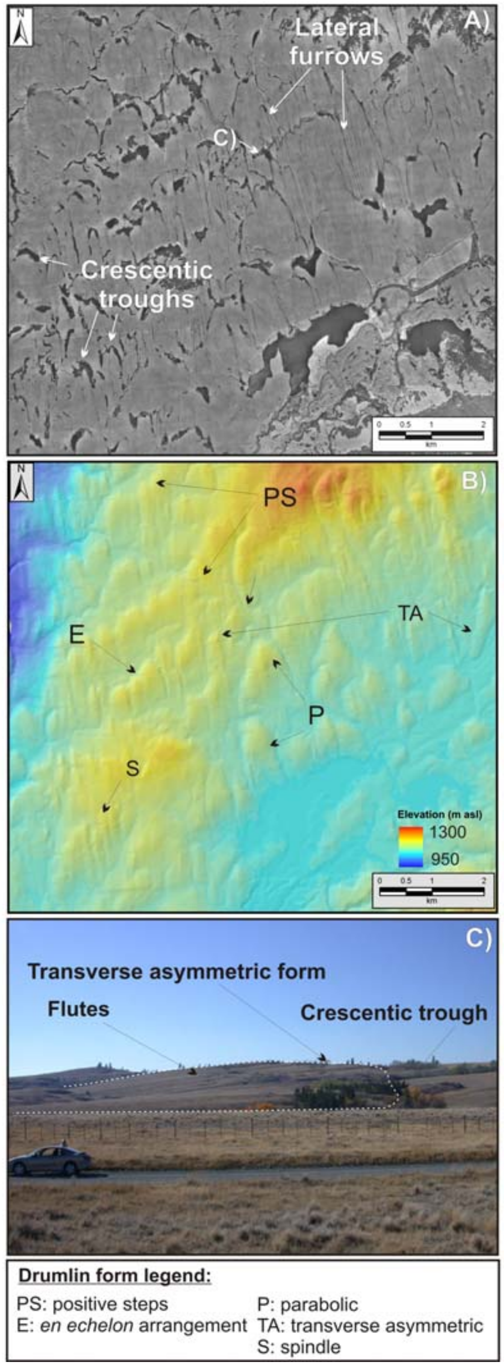

Figure 9. Morphology of drumlins on Thompson Plateau. A) Drumlins exhibit stoss-side crescentic troughs (occupied by lakes) and lateral furrows. B) Drumlins develop preferentially over positive steps and have *en echelon* arrangement. Drumlin shapes include paraboloc forms, transverse asymmetrical forms, and spindle forms. C) Field photograph of bedrock flutes superimposed on larger transverse asymmetric forms (dotted line) with a prominent crescentic trough at its head. Height of transverse asymmetric form is 6 m. From Lesemann and Brennand (in press).

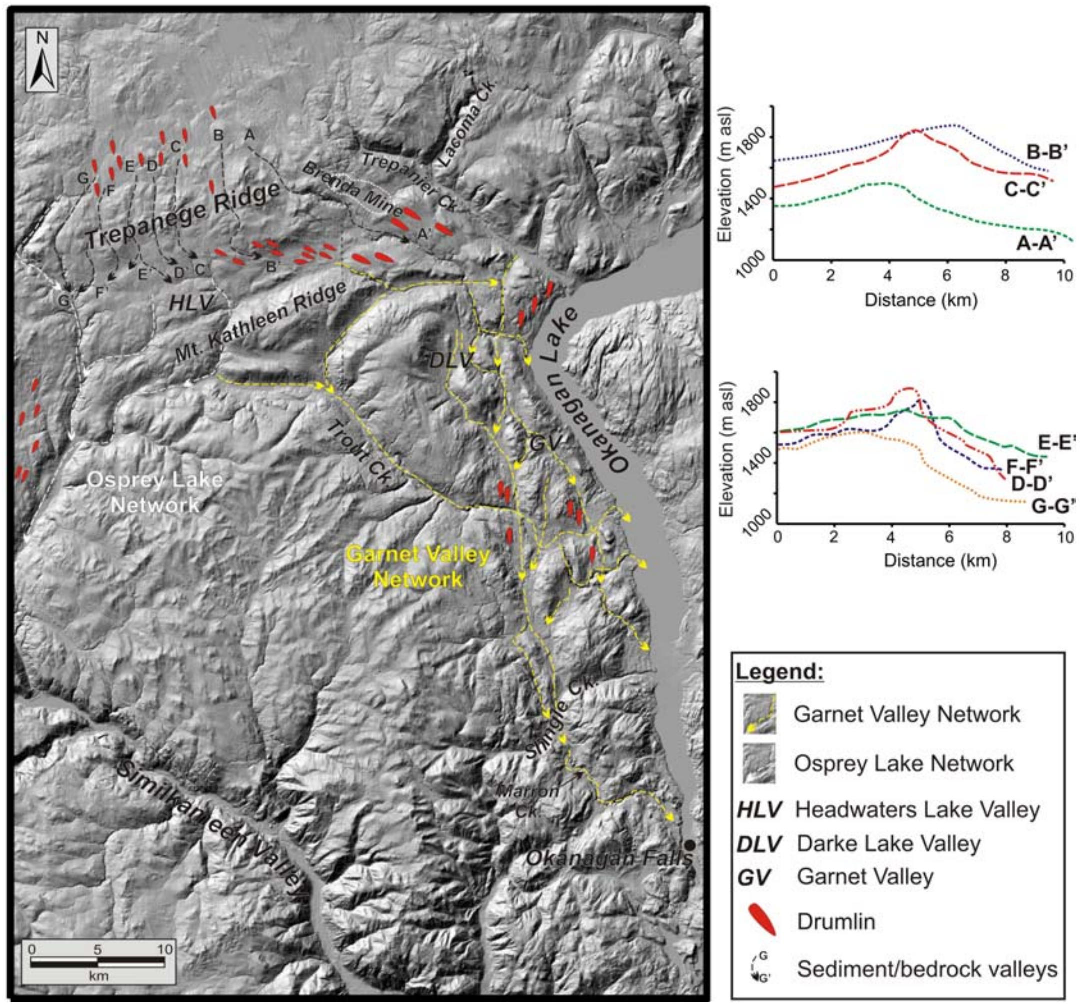

Figure 10: Bedrock valleys along the west side of Okanagan lake. Valleys (A-G) traversing Trepanege Ridge exhibit convex longitudinal profiles. Valley walls are composed of sediment in their headward and downflow reaches and bedrock along elevated middle sections across Trepanege Ridge. Valllesy initiate among drumlins upflow of Trepanege Ridge. From Lesemann and Brennand (in press)

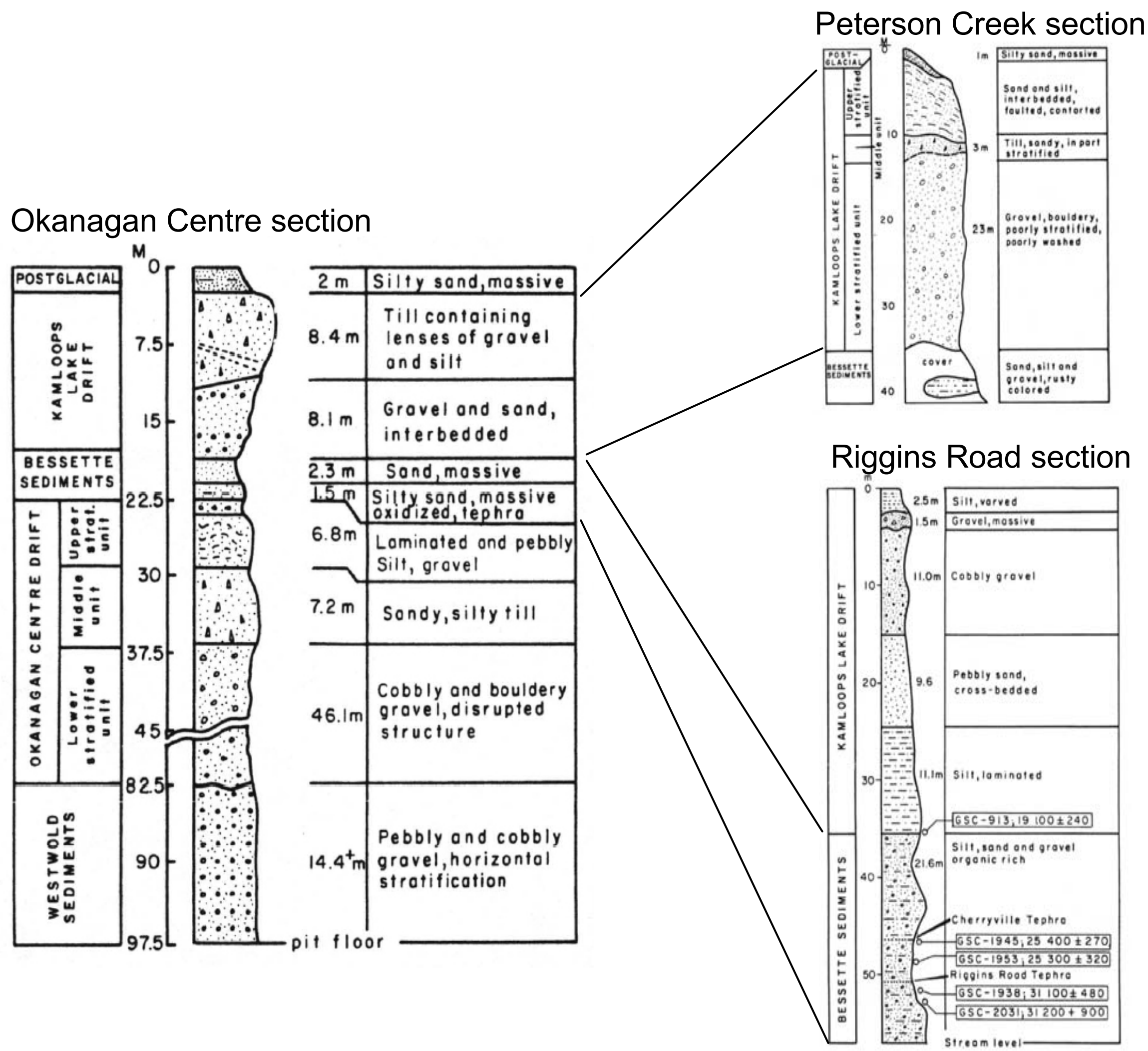

Figure 11: Lithostratigraphic correlation of units in the Okanagan Centre section with those elsewhere in the southern Interior (after Fulton and Smith 1978).

Day 4

Glacial Lake Penticton

During deglaciation the Okanagan Lake basin was filled by glacial Lake Penticton (Flint 1935, Nasmith 1962, Fulton 1969). Nasmith (1962) mapped meltwater channels, kettled outwash sediments, kame terraces, deltas and lake floor sediments (silt benches) along the Okanagan Valley (and into tributary mouths) between Osoyoos Lake and Enderby. He concluded that glacial Lake Pentiction was a partially supraglacial (early stage), partially ice-marginal (lateral to an active tongue of ice occupying the Okanagan Valley trough) and partially proglacial lake (distally) that was dammed by ice and sediment in the vicinity of McIntyre Bluff, and expanded northward as the ice tongue in Okanagan Valley downwasted and backwasted northward; these ideas echo those of Flint (1935). A gradually falling lake level with only a few periods of stability was explained by the nature of the dam and outlets. Sediment and water were supplied both from the valley-confined ice mass and from deglaciated tributaries. Fulton (1969) defined four stages of glacial Lake Penticton (Long Lake ~520 m asl, Grandview Flats ~490 m asl, B.X. ~430 m asl, and O'Keefe ~350 m asl) based on the identification and elevation of deltas and delta terraces in the North Okanagan basin (Fig. 12). He depicted glacial Lake Penticton evolving from an ice marginal (lateral to a stagnating ice mass along the centre of Okanagan Valley) to a proglacial, ice-contact lake. In an attempt to explain the presence of winter sands within the lake varves, Shaw and Archer (1979) suggested that sediment was mainly supplied from an ice lobe occupying the valley; they placed a supraglacial delta in the vicinity of Squally Point. The existence of the silt benches overlying bedrock south of Squally Point has been attributed to (i) the presence of a stagnating mass of ice along the valley axis (non-deposition, Fulton 1969), (ii) supraglacial sedimentation over a stagnating ice lobe along the valley axis followed by lowering of lake bottom sediments during ice melt (Shaw 1977), or (iii) supra-ice deposition over a field of grounded icebergs followed by lowering of lake bottom sediments during ice melt (Eyles *et al.* 1991). Interpretations from land-based and lake-based seismic profiles and drilling do not appear to support the presence of a grounded ice lobe along the axis of Okanagan Valley: no till is reported (e.g., Vanderburgh and Roberts 1996), and Eyles *et al.* (1991) postulate that the chaotic character of their lower lacustrine unit (seismic facies IIA, Eyles *et al.* 1991) may be due, in part, to the melting of buried ice blocks or the foundering of ice blocks produced at a calving margin.

On day 4 we will visit several stops to view evidence for glacial Lake Penticton and discuss lake paleogeography (lake level reconstructions), paleoenvironment (lake type and sediment delivery) and drainage; the possibility of an antecedent subglacial lake and its effects on the pattern of ice sheet decay will also be discussed. The new information that will be presented is largely unpublished and has been assembled by Jerome Lesemann.

Stop 15: <u>Giant's Head Park: lake paleogeography and paleoenvironment</u> (49° 35.150' N, 119° 40.107' W)

From this vantage point we can see the extensive white silt benches (~ m asl) between Squally Point and Penticton, the coalescing fans that separate Okanagan and Skaha Lakes at Penticton, and a series of perched and sloped sediment surfaces (nested and terrace-like in some instances, but often individual benches) grading to different water levels of glacial Lake Penticton. The gravel pit in Prairie Valley exposes a kettled delta surface (faulted foresets). We will discuss new paleogeographic reconstructions of the lake, sediment delivery to the lake, and the possible existence of an earlier subglacial lake in the valley.

Stop 16: McIntyre Bluff: damming and drainage mechanisms (49° 17.296′ N, 119° 31.022′ W)

This stop is on a dissected alluvial fan (Vaseux Creek fan) that was part of the ice-sediment complex that dammed glacial Lake Penticton at the narrow bedrock gorge (~900 m wide), bordered by McIntyre Bluff, and extending for ~12 km south of Okanagan Falls. Kettle lakes in the distal portion of the fan suggest that ice was buried within the fan sediments. The final drainage of glacial Lake Penticton was responsible for fan dissection. The elevation of McIntyre bluff is 660 m on the east and 740 m on the west. So long as the gorge is filled with ice or sediment, McIntyre Bluff is sufficiently high to retain the highest reconstructed levels of glacial Lake Penticton. South of this fan (east side of the bluff) we can see the Atsi Klak bedrock spillway (530 m asl), drainage through which was partially responsible for dissection of the Wolfcub Creek fan south of McIntrye Bluff (visible from stop 18).

Stop 17 (if time permits): McIntyre Bluff: evidence for lake drainage in the bedrock gorge (49° 14.367′ N, 119° 31.457′ W)

Drainage of glacial Lake Penticton through the bedrock gorge produced s-forms along the walls of McIntyre bluff. S-forms are visible through the trees on the east side of the bluff.

Stop 18: McIntyre Bluff: evidence for lake drainage downflow of the dam (49° 14.255′ N, 119° 33.883′ W)

This stop is on a sandy fan south and west of McIntyre Bluff that was built by streams issuing from decaying ice on the bedrock slopes above the valley and possibly also from Green Lake Spillway. It is preserved in the lee of the bluff, but its southern extent was truncated during drainage of glacial Lake Penticton. Wolfcub Creek fan on the east side of the valley was also dissected at this time, and an expansion bar formed as flow exited the bedrock gorge.

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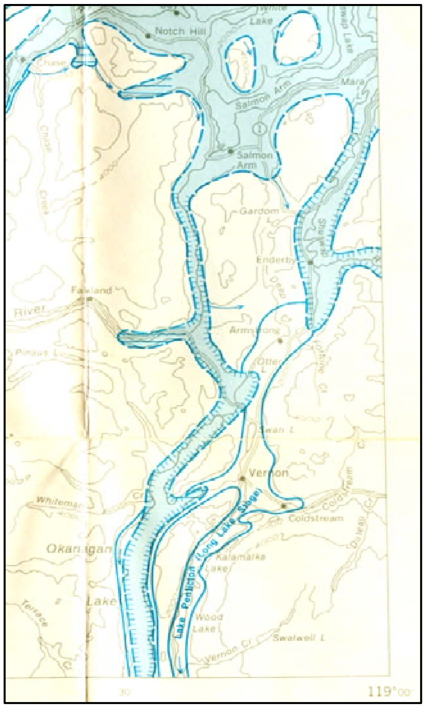
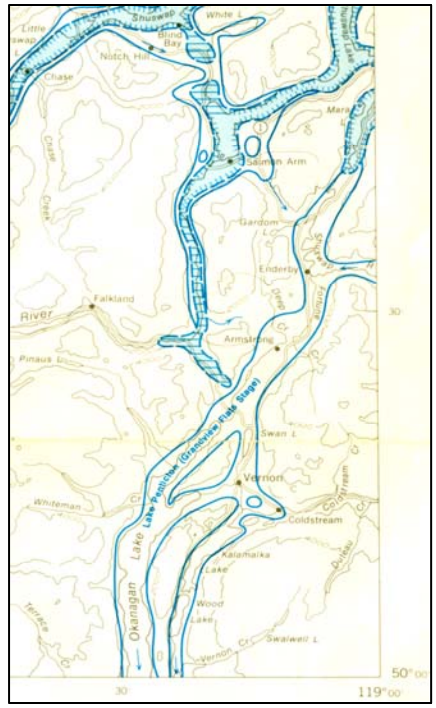
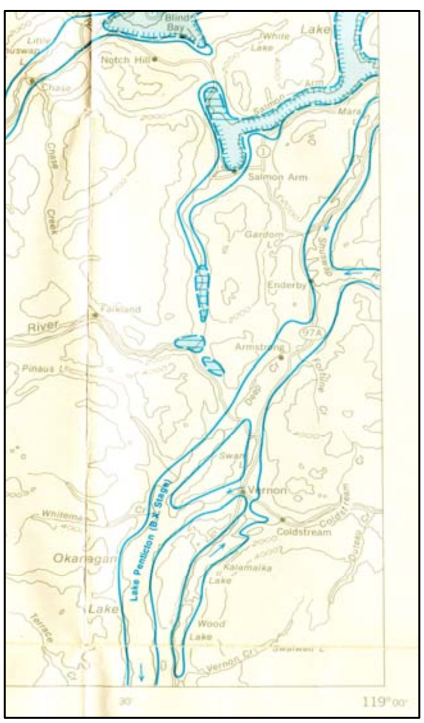
Long Lake stage (1700’, 520 masl)



Grandview Flats stage (1600’, 490 masl)



B.X. stage (1400’, 430 masl)



O’Keefe stage (1160’, 350 masl)

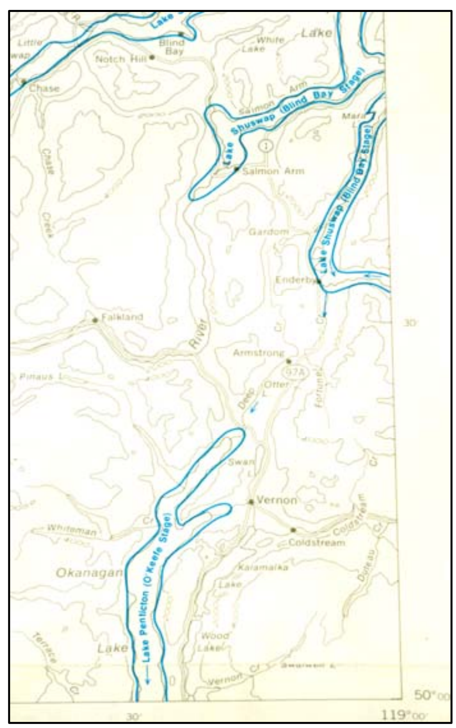


Figure 12. Stages of glacial Lake Penticton (Fulton 1969)

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