



# 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

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## ARTICLE INFO

### Article history:

Received 21 March 2007

Received in revised form

25 December 2008

Accepted 28 April 2009

## ABSTRACT

Subglacial landsystems in and around Okanagan Valley, British Columbia, Canada are investigated in order 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 to 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 are incompatible with prevailing explanations of landsystem development centred mainly on deforming beds. They are best explained by meltwater erosion and deposition during ice sheet underbursts.

During the Late-Wisconsin glaciation, Okanagan Valley functioned as part of a subglacial lake spanning multiple connected valleys (few 100s 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 heat flux, 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.

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## 1. Introduction and objectives

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

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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. Extensive subglacial lakes (100s km<sup>2</sup>) and smaller water-filled cavities (km<sup>2</sup> to 10s km<sup>2</sup>) 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 remain unclear (Bell et al., 2007) due in part to 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). However, the same landforms are also attributed to the erosional and/or depositional action of turbulent meltwater sheets (Shaw, 1996; 2002). Consequently, glaciodynamic reconstructions based on glacial bedforms remain equivocal.

This paper examines landform and sediment assemblages in the Okanagan Valley-Thompson Plateau area of British Columbia, Canada (Fig. 1), an area close to the southern margin of the former Cordilleran Ice Sheet (CIS). Here, a rich record of glacial landforms and sediments, including drumlins and valleys, lends itself to regional reconstructions of glacier bed processes and glaciodynamics. Our objectives are to evaluate landscape development, regional glaciodynamics and hydrologic conditions along the southern margin of the CIS. We evaluate hypotheses of landform genesis, present an event sequence of landscape evolution, and investigate the implications of these for CIS hydrology, dynamics and geometry.

## 2. 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. 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 and 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. It is a fault-

controlled valley extending over 300 km from Shuswap Lake to the Columbia River in Washington State<sup>1</sup>, 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; we describe them separately below for organizational simplicity.

## 3. Landform suites of Thompson Plateau

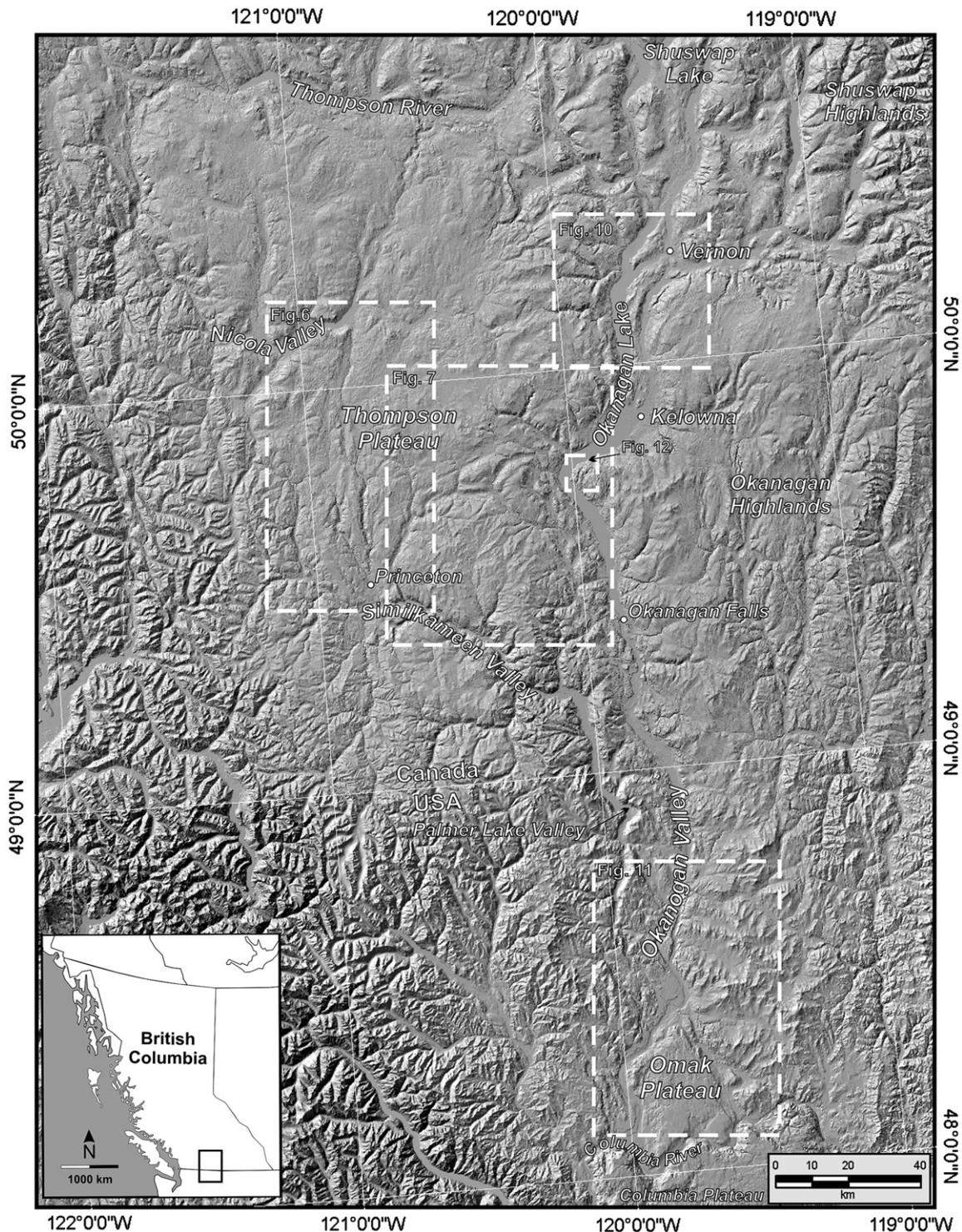
### 3.1. Drumlins

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. 1–3). 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 (Figs. 1, 2).

Topography has little effect on the orientation of the Thompson Plateau swarm: drumlins occur on the plateau surface, in some valley-bottoms, and on adverse slopes leading out of valleys. The swarm maintains its orientation across many valleys (Fig. 3). Locally, swarm orientation does reflect topography: smaller drumlin fields bifurcate around some bedrock highs. These high points are not streamlined and stand out on digital elevation models by their non-linear texture (Fig. 3). Some valleys 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). 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 Okanagan Valley between the 49th parallel and the Columbia River (Fig. 1).

Thompson Plateau drumlins contain a range of materials including bedrock, unconsolidated sediments (diamictons and glaciofluvial/glaciolacustrine sediments) and combinations of these materials (Fig. 4; Table 1) (Lesemann and Brennand, 2005). They exhibit parabolic, spindle and transverse asymmetric planforms (Fig. 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. 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. 5). At the heads of *en echelon* clusters, coalesced crescentic

<sup>1</sup> 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.



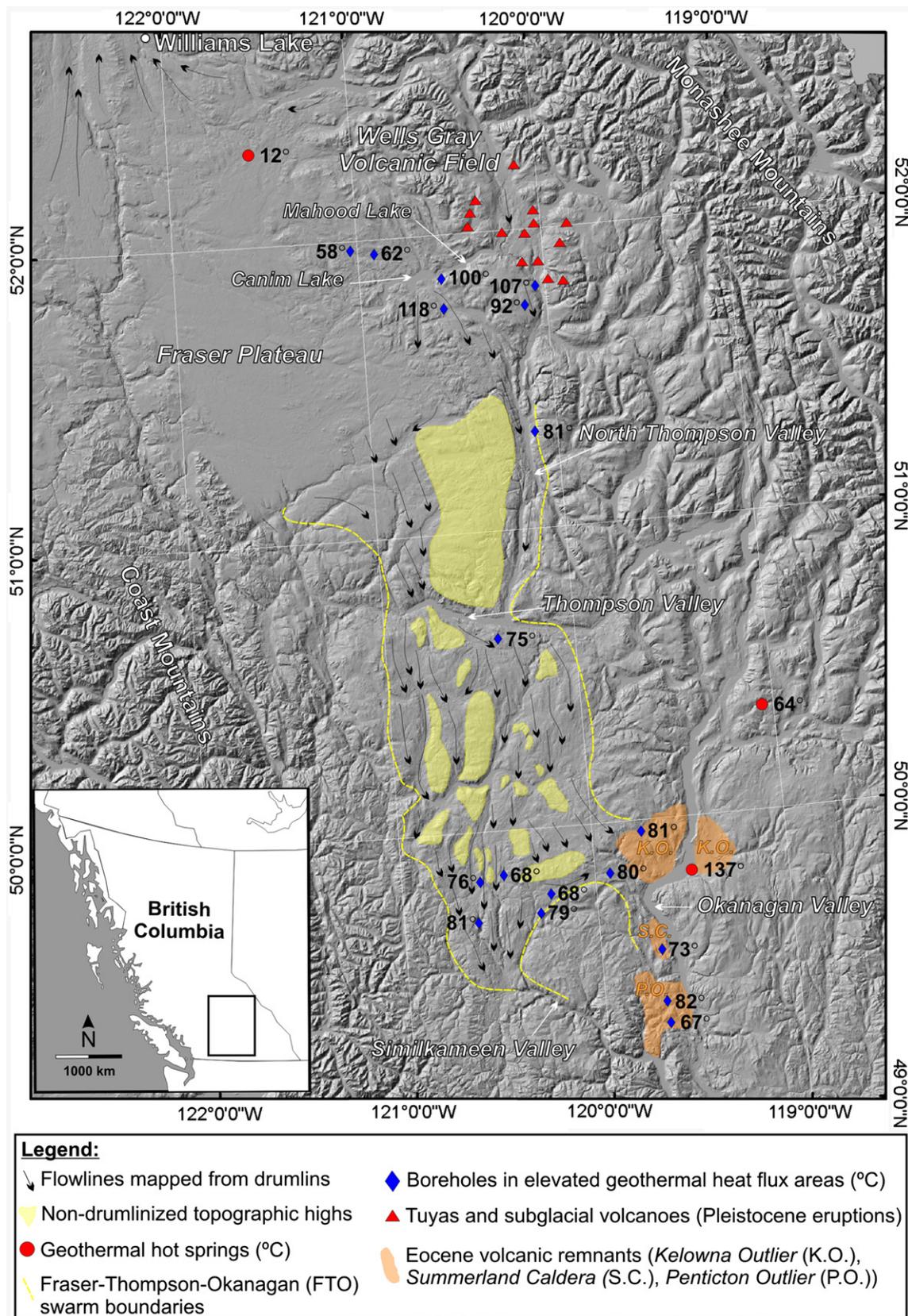
**Fig. 1.** Hillshaded digital elevation model of the field area highlighting major physiographic elements of the British Columbia southern Interior Plateau and northern Washington State discussed in the text. Dashed rectangles show locations of figures. Inset situates map within British Columbia and Washington State. Elevation data for Canadian regions: BC Government TRIM 1 dataset, 25 m resolution. Data for American regions: USGS 10 m DEM reprojected to BC Albers, NAD 83.

depressions in places form sinuous channel-like troughs that may be water-filled (Fig. 5A).

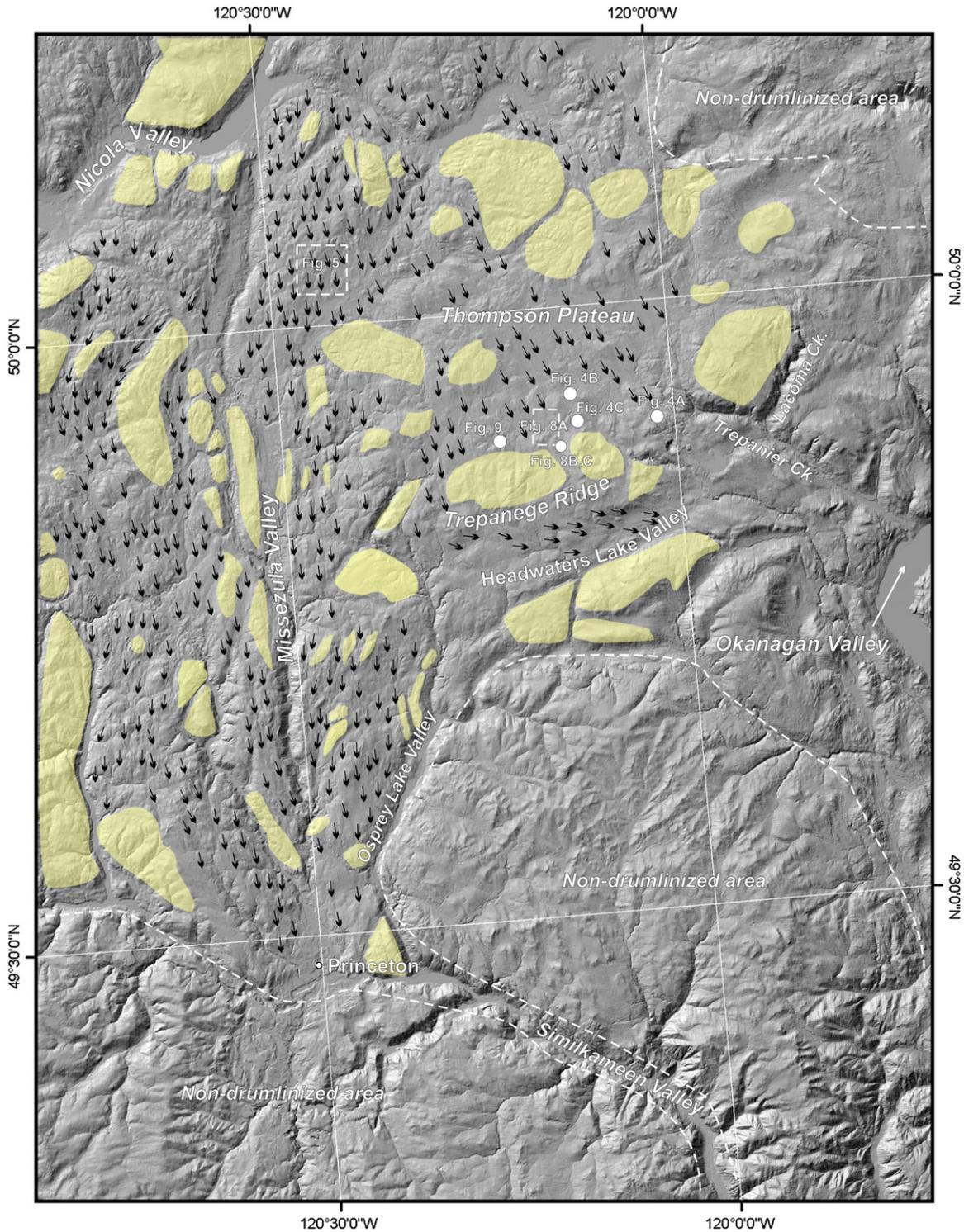
### 3.2. Valleys

Valleys crossing Thompson Plateau are eroded into bedrock and sediment. Two scales of valleys exist. The largest are as long as

100 km with variable depths reaching a maximum of 150 m (Miszezula Valley) (Figs. 3, 6; Table 2). They are parallel to the regional drumlin swarm. Some traverse the entire plateau and connect transverse bounding valleys such as the Nicola and Similkameen valleys. Drumlins are present on the valley walls and shoulders. Modern rivers have incised into a discontinuous sediment fill along most of their length.



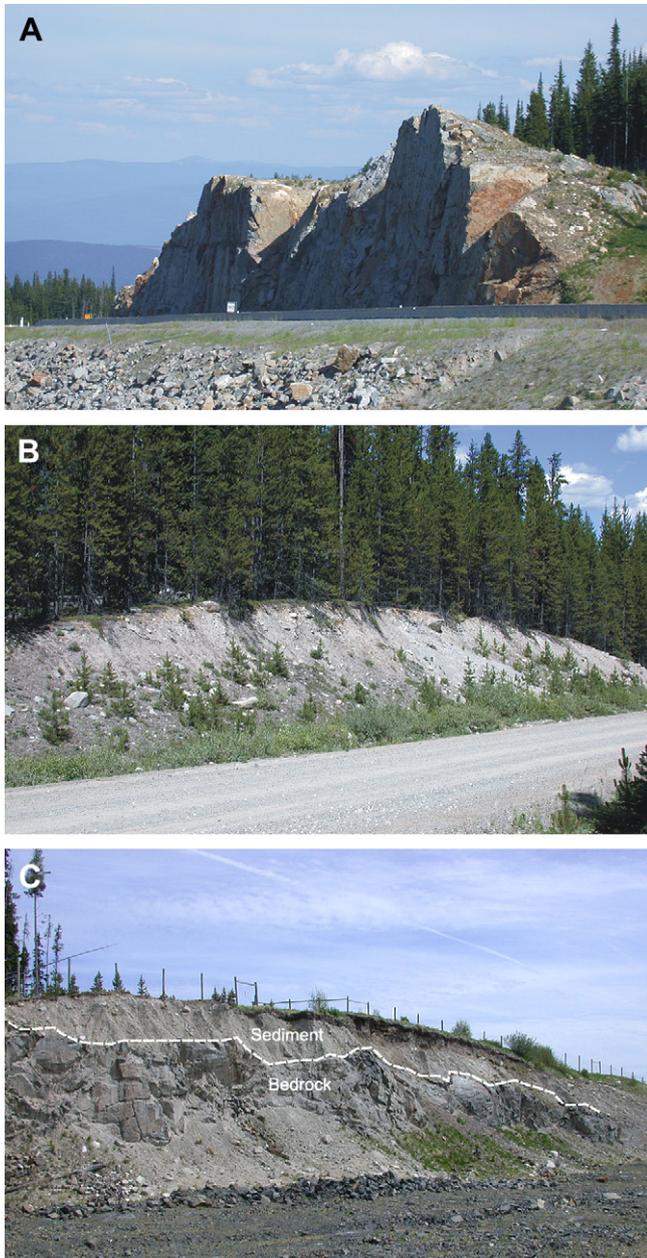
**Fig. 2.** Generalized flowline map of the Interior Plateau within British Columbia (inset map). Flowlines are mapped from drumlins in the Fraser-Thompson-Okanagan (FTO) swarm, beginning on Fraser Plateau and extending southward across Thompson Plateau toward Okanagan Valley and Similkameen Valley. 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. BC Albers projection, NAD 83. Geothermal flux data: Fairbank and Faulkner, 1992; Gratsby and Hutcheon, 2001.



**Fig. 3.** Hillshaded digital elevation model of Thompson Plateau. Drumlins are pervasive on Thompson Plateau and maintain consistent orientations across valleys. Refer to Fig. 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. BC Albers projection, NAD 83.

Smaller valleys occur in groups and traverse non-drumlinized bedrock highs such as Trepanege Ridge (Fig. 7). Their length varies between 10 and 15 km and their depth can be as much as 50 m (Table 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. 8A). In all cases, valley incision is greatest along sediment reaches and shallower along bedrock reaches (along their elevated mid-section). Some bedrock valleys form unbranched networks (Garnet and Osprey Lake valley networks, Fig. 7).



**Fig. 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 subglacial cavities (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). See Fig. 3 for locations.

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. 6, 7; Table 2).

### 3.3. 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. Erosion has exploited joints and other bedrock weaknesses. 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 (Lesemann and Brennand, 2005). Above 1200 m asl, sediment cover is absent and only bedrock is exposed. These higher surfaces show incipient drumlinization. For example, bedrock valleys controlled by geologic structure on Trepanege Ridge leave upstanding bedrock knobs as interfluvial showing poorly developed streamlining (Fig. 7). Other interfluvial exhibit patches of deeply weathered bedrock (Fig. 9). These patches have limited areal extent (100s m<sup>2</sup>). The estimated depth of weathering from exposures is at least 7 m and may be greater (Fig. 9). Surfaces of deeply weathered outcrops exhibit 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. 9).

### 3.4. Landform associations

There is a close spatial association between drumlins and valleys on Thompson Plateau. We explore the example of Trepanege Ridge (Fig. 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-drumlined bedrock occur (above 1200 m asl.). Valleys begin in sediment among drumlins (Fig. 8A) on the north side of Trepanege Ridge and continue through the transition in drumlin types. Above the drumlin termination (on the northern ridge flank), bedrock valleys cut across the ridge (Fig. 5). Ridge elevation decreases gradually to the west where drumlins curve around the western ridge flank toward Headwaters Lake Valley (Fig. 7). There, they form a small swarm within the confines of the southern flanks of Trepanege Ridge and Mt. Kathleen Ridge (Fig. 6). Glaciofluvial sediments (from 4 to >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. 7). This small swarm terminates abruptly against the north slope of Mt. Kathleen Ridge. Steep-sided and narrow bedrock valleys cut across this ridge.

## 4. Landform suites of Okanagan Valley

### 4.1. Drumlins

Drumlins occur within the confines of Okanagan Valley. They are mainly composed of bedrock and exhibit variable planforms (Table 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. 10), and three fields in the Okanagan Valley between the Canada–USA Border and the Withrow Moraine (Fig. 11).

The Swan Lake drumlin field (Fig. 10A,B; Table 1) occurs on a prominent bedrock ridge rising above the modern valley floor (Fig. 10A). These bedrock drumlins exhibit stoss-side crescentic depressions. A discontinuous, decimetre-thick veneer of glaciolacustrine silt onlaps these drumlins. A more prominent bedrock interfluvial separating Okanagan Lake from Kalamalka Lake and Wood

**Table 1**  
Summary characteristics of drumlins on Thompson Plateau and in Okanagan Valley.

Location		Single drumlin morphometrics						Morphology	Composition
Drumlin field name	Coordinates <sup>a</sup>	Area (km <sup>2</sup> )	Length (m)	Width (m)	Height (m)	Elongation ratio	Planform; Longitudinal profile <sup>b</sup>		
<b>Okanagan Valley (Canada)</b>	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; asymmetrical	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	Whaleback flanked by channels; undulating crestline; symmetrical and asymmetrical	Bedrock
<b>Okanagan Valley (USA)</b>	Tonasket	48°44' N 119°31' W	30	380–700	180–350	10–40	1.8:1–2.2:1	Classical morphology; symmetrical and asymmetrical	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; asymmetrical	Bedrock, gravel, sand, diamicton
	Columbia Plateau <sup>c</sup>	47°53' N 119°30' W	550	100–2000	100–500	14–30	3:1–8:1	Classical asymmetrical drumlins, parabolic, spindle; asymmetrical	Diamicton, stratified sand and gravel, bedrock knobs at heads <sup>c</sup>
<b>Thompson Plateau</b>	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; symmetrical and asymmetrical	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; asymmetrical	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; asymmetrical	Bedrock, diamicton, glaciofluvial sediments

<sup>a</sup> Coordinates are approximate and refer to a point close to the centre of a designated drumlin group.

<sup>b</sup> Morphological terminology after Shaw and Kvill (1984).

<sup>c</sup> Some drumlin composition data from Kovanen and Slaymaker (2004).

Lake preserves the Ellison Ridge drumlin field, southeast of Vernon (Fig. 10C; Table 1). These drumlins are composed entirely of bedrock and occur on the crest and western flank of the interfluvium. They exhibit rhomboidal planforms with a space-filling arrangement controlled by bedrock joints (Fig. 10C). 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. 10D; Table 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 floor have reoccupied and dissected these troughs (Fig. 10D). Shorelines from an extensive deglacial lake also overprint the forms.

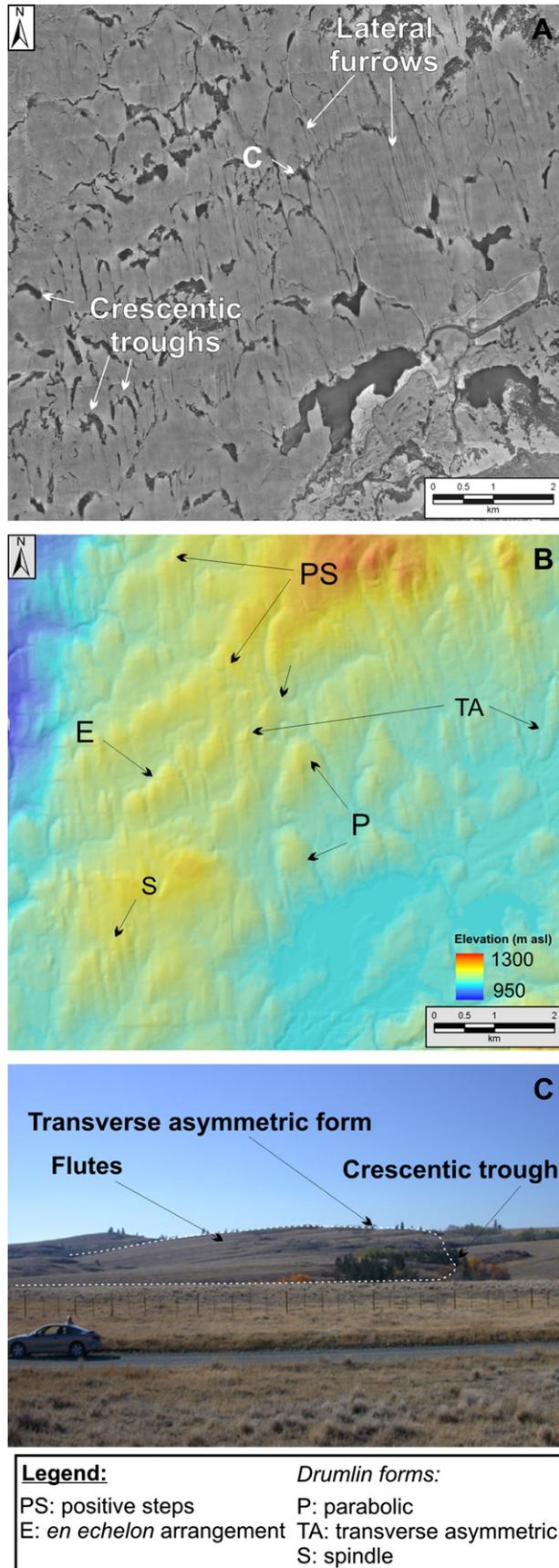
In Okanagan Valley, bedrock drumlins occur near Tonasket, WA, in small fields on interfluvies between valleys (bedrock coulees) (Fig. 11). More extensive fields occur on Omak Plateau (Fig. 11) and on the Columbia Plateau (Fig. 1) where they are composed of sorted and stratified sediments (based on water-well logs) and diamictons

(Kovanen and Slaymaker, 2004) respectively. On Omak Plateau, individual drumlins are continuous across sediment-bedrock outcrops. This drumlinized surface also exhibits a 'scabbed' appearance (Fig. 11) characterized by patches of exposed bedrock.

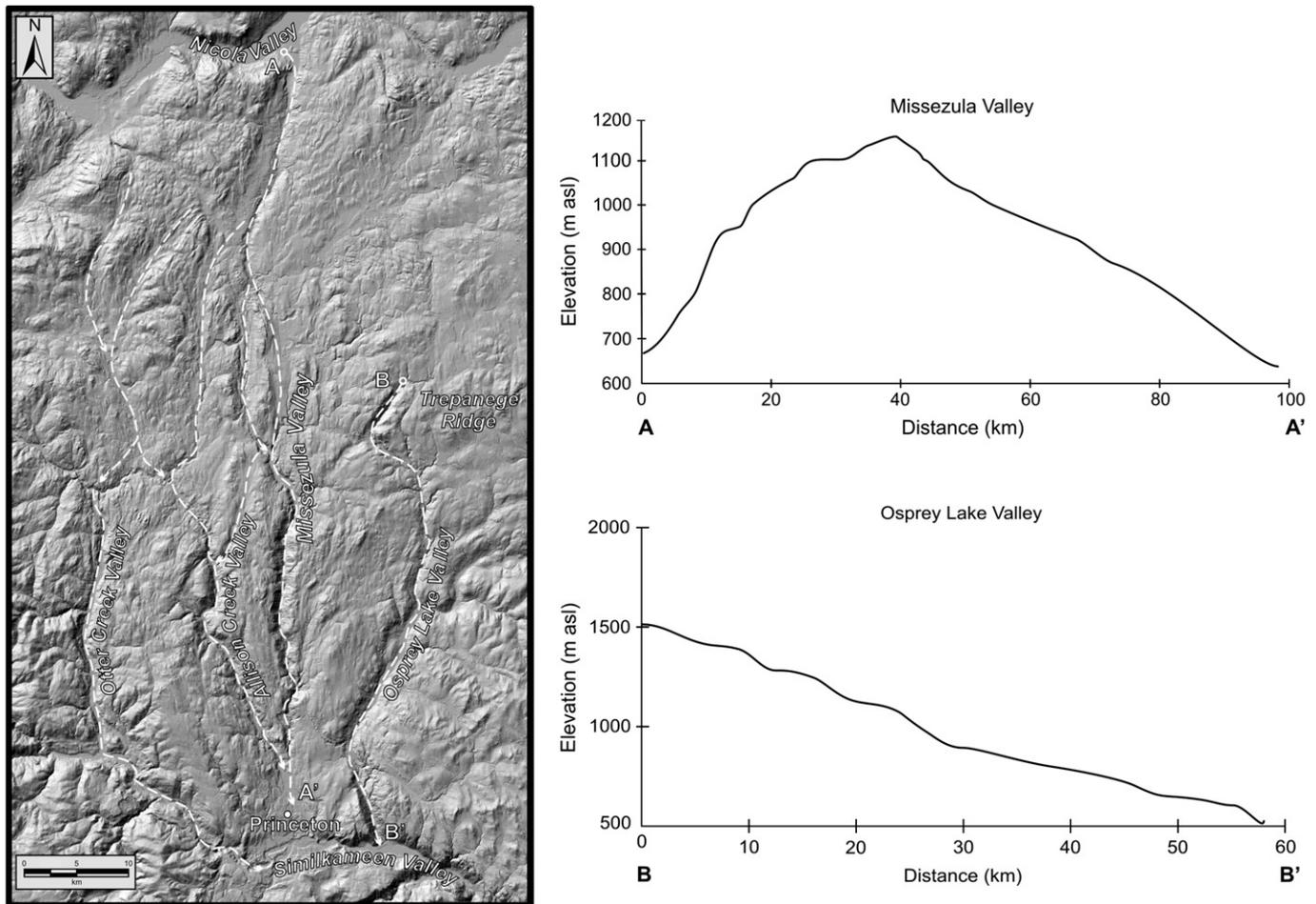
#### 4.2. Valleys

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. 1) (Eyles et al., 1990; 1991; Vanderburgh and Roberts, 1996). However, bedrock remains 300 m and over 150 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



**Fig. 5.** Morphological characteristics of drumlins on Thompson Plateau (see Fig. 3 for location). (A) Drumlins exhibit stoss-side crescentic troughs sometimes occupied by lakes, and lateral furrows. (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 form (dotted line; partially visible). A prominent crescentic trough occurs at the head of this composite form. Height of transverse asymmetric form is 6 m. BC Aerial photograph #15 BC96057-88 (copyright British Columbia Government, by permission).



**Fig. 6.** Hillshaded digital elevation model of mainly bedrock valleys (see Fig. 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 1 dataset, 25 m resolution. BC Albers projection, NAD 83.

Canyon east of Kalamalka Lake (Fig. 10A), Wildhorse Canyon and Goode Canyon traversing Squally Point (Fig. 12), Palmer Lake valley west of the Tonasket drumlins (Figs. 11, 13), and a number of bedrock coulees near Tonasket and Omak, Washington and on the flanks of Omak Plateau (Fig. 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. 11, 12; Table 2). Sediments of variable thickness partially fill some channels (Table 2). In some cases, convex longitudinal profiles describe the bedrock surface as well as the modern valley floor (Fig. 11).

A >25 km-long network of anabranching bedrock valleys, the Garnet Valley network, occurs west of Okanagan Lake and south of Mt. Kathleen Ridge (Fig. 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. 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.

#### 4.3. Bedrock outcrops and forms

Between drumlin fields (described above), extensive portions of Okanagan Valley (up to 70 km in length) appear to 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 ~340 and 400 m asl along Okanagan Valley. Till is mostly 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. 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 ephemeral. 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.

**Table 2**

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

Name	Coordinates <sup>a</sup>	Length (km)	Width (m)	Depth (m)	Elevation gain, loss (m) <sup>b</sup>	Channel boundary; Fill
<i>Thompson Plateau</i>						
Miszezula 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 boundaries and fill at upflow and downflow ends, bedrock middle section with no fill
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	"
<i>Okanagan Valley</i>						
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
<i>Okanagan Valley</i>						
Omak Lake Valley	48°17' N, 119°24' W	42	600–5000	80–200	+325, –300	Bedrock; sand, boulder gravel and lacustrine silt
Soap Lake Valley	48°14' N, 119°37' W	25	400–3000	80–130	+430, –500	Bedrock; none

<sup>a</sup> Approximate midpoint valley coordinates.<sup>b</sup> Most valleys have convex longitudinal profiles rising to a crestpoint before dropping. Elevation gain refers to maximum elevation increase along upflow segment. Elevation drop refers to maximum elevation decrease along downflow segment.

#### 4.4. Moraines

Some landforms are notable by their absence from the landscape. In particular, major end moraines are not present in Okanagan Valley or on the 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 extensive channelization breaches the moraine in many places.

### 5. 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.

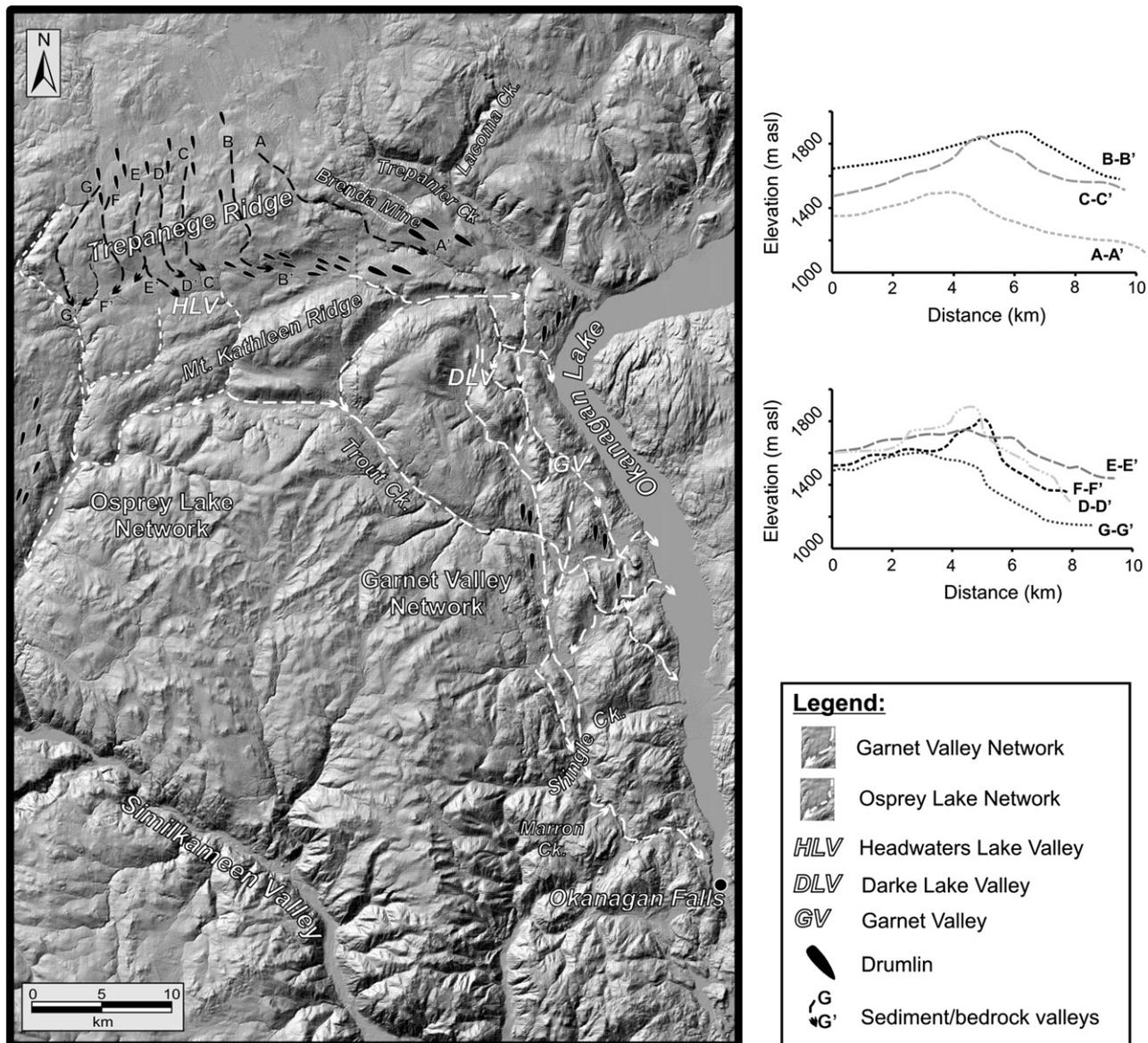
#### 5.1. Drumlins

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. Deforming material can accrete in low-pressure zones leeward of bed obstructions to create a *depositional drumlin*. Finally, streaming deforming bed material can erode a drumlin from pre-existing substrate. 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 Breidamerkurjökull, Iceland showing metres-thick bed deformation (Boulton, 1987). Associations of thick deforming beds with modern ice streams rely on seismic interpretation of metre-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 Breidamerkurjökull 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 metre-thick deforming beds are not representative of true conditions (Alley, 2000). Instead, deformation takes place over a thin centimetre-scale layer close to the ice bed (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 centimetre-scale deformation (Iverson and Iverson, 2001; Tulaczyk, 1999). Although there is little doubt that bed deformation can occur under elevated porewater pressure conditions, new findings on the mechanics of bed deformation make us question its effectiveness for generating metre- to decametre-scale drumlins.

##### 5.1.1. Thompson Plateau drumlins

Based on individual drumlin characteristics and on regional landform associations, drumlins are predominantly erosional forms. Within the same swarm, drumlins composed of bedrock, bedrock and diamict, and diamict and undeformed sorted and stratified sediments truncated at the drumlin surface (Fig. 4), preclude a depositional origin for these forms (Lesemann and Brennand, 2005). Sediment within drumlins record pre-drumlin deposition within subglacial water-filled cavities and between

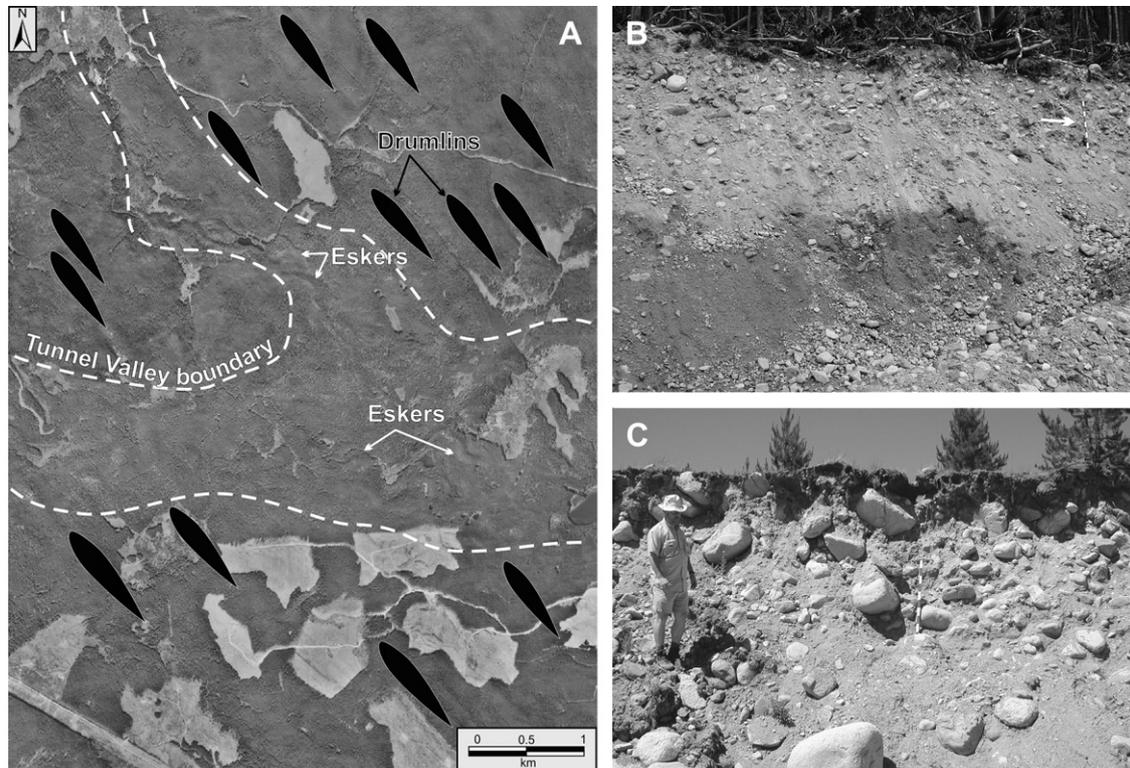


**Fig. 7.** Mainly bedrock valleys along the west side of Okanagan Lake (see Fig. 1 for location). Valleys (A–G) traversing Trepanege Ridge exhibit convex longitudinal profiles (constructed from digital elevation data: BC Government TRIM 1 dataset, 25 m resolution). Valley walls are composed of sediment 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 anabranching 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.

cavities under more grounded conditions. These sediments include deformed lodgement tills, debris flow diamictons, and stratified glaciofluvial and glaciolacustrine sediments. Shallow and patchy deformation affected a portion of these sediments in zones of high shear stress between cavities and around cavity margins. Clast fabrics record regional ice build-up and compression against Trepanege Ridge, rather than drumlin-scale deformation. Drumlins were eroded into these sediments (Lesemann and Brennand, 2005).

Multiple lines of evidence argue against drumlin erosion by a deforming bed on the 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

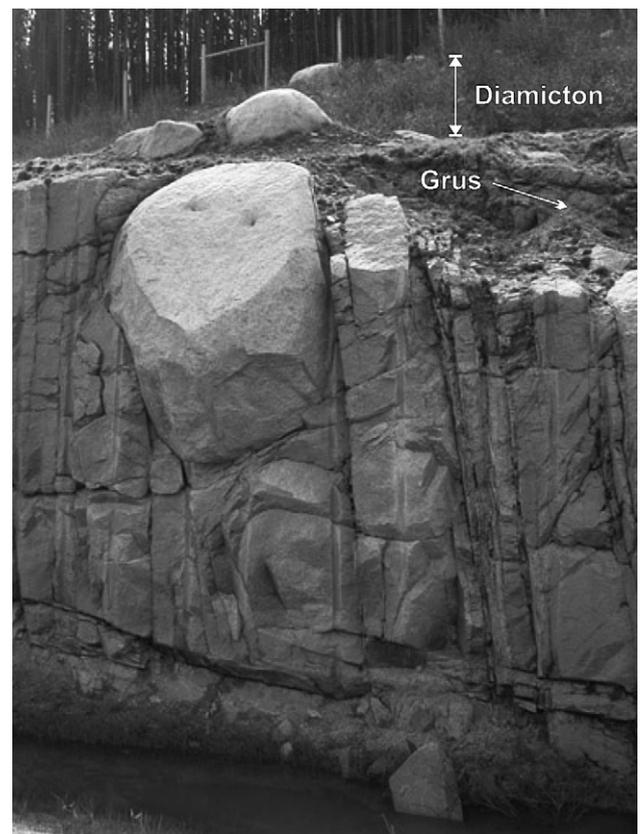
mobile and erosive (Boyce and Eyles, 1991). Under such circumstances the till carapace could be absent over sediment drumlins and be replaced by striae and other forms of abrasion on bedrock drumlins. The surfaces of some incipient bedrock drumlins in the Thompson Plateau drumlin swarm are striated. 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. Striated surfaces and incipient bedrock drumlins may also be palimpsest features from previous glaciations. 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. Such



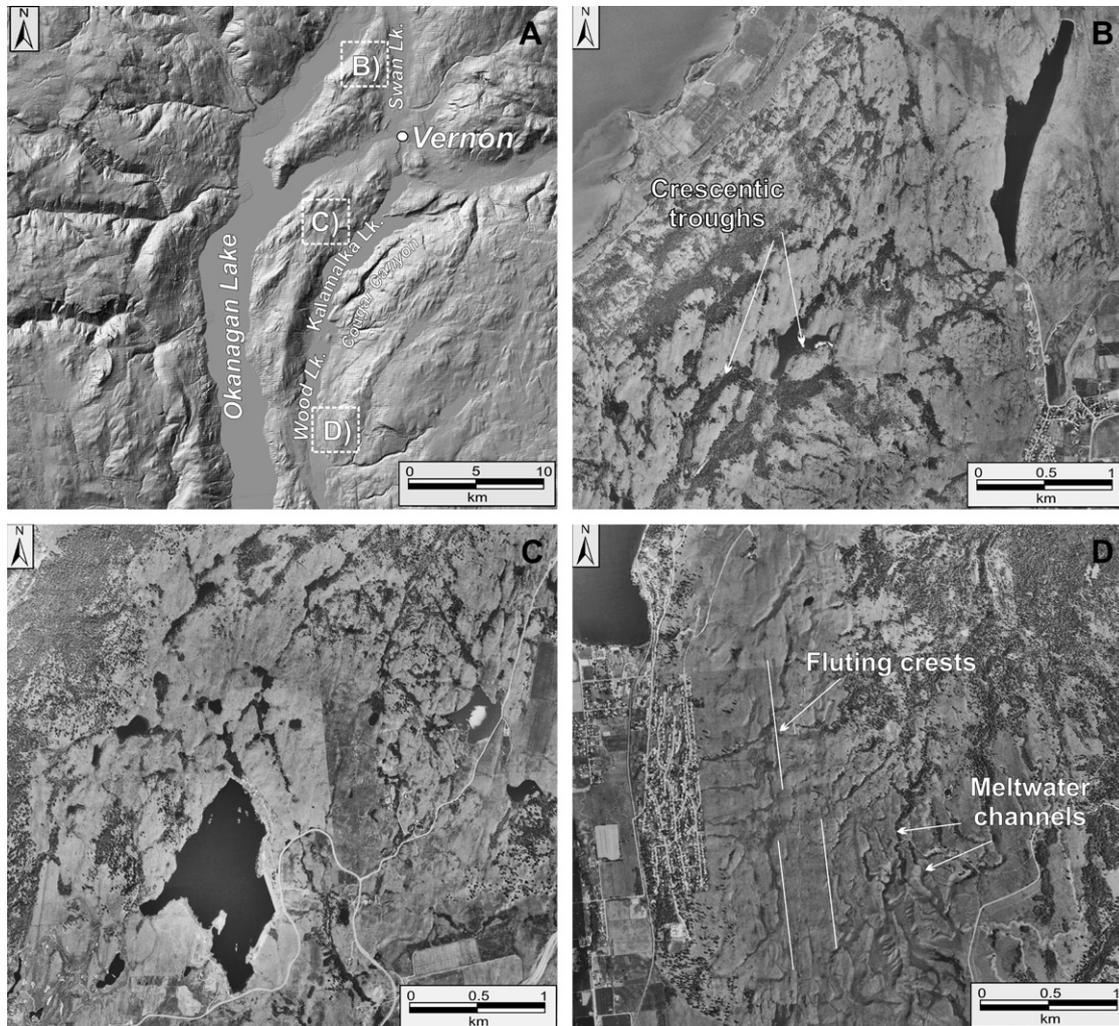
**Fig. 8.** (A) Aerial photograph showing a segment of a sediment-walled tunnel valley cross-cutting the Thompson Plateau drumlin swarm. Eskers occupy the valley floor. (B) Sorted and imbricate cobble gravel partially fills valleys and creates low-relief undulations. (C) Heterogeneous gravel also occurs within valleys and forms irregular mounds (rod is 1 m long). BC Aerial photograph #15BCB96056-185 (copyright British Columbia Government, by permission). See Fig. 3 for locations.

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 drumlin forms. 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. 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. 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



**Fig. 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. See Fig. 3 for location.



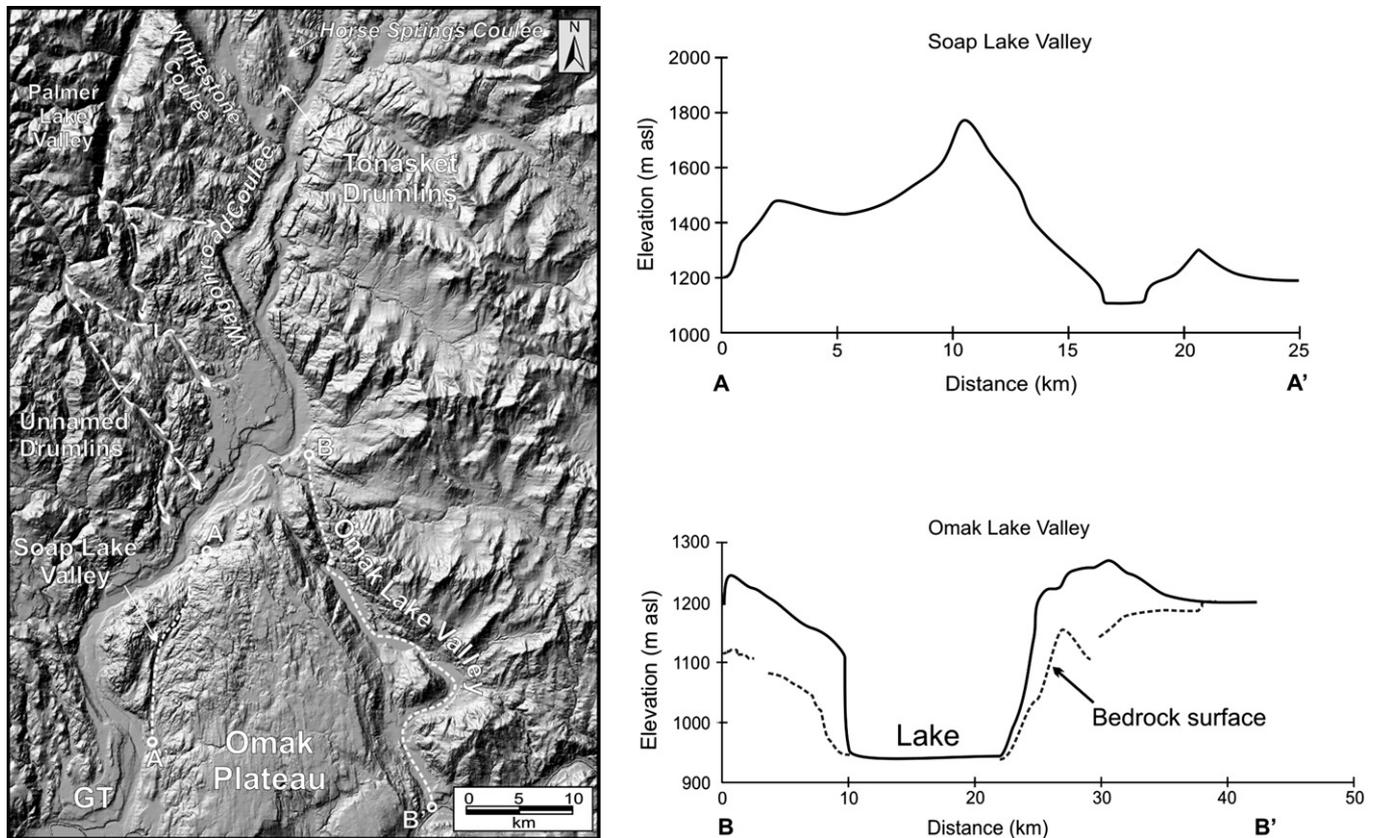
**Fig. 10.** (A) Bedrock drumlins and tunnel valleys in Okanagan Valley (see Fig. 1 for location). Cougar Canyon is a prominent bedrock valley on the east side of Kalamalka Lake. (B) Swan lake drumlins exhibit stoss-side crescentic troughs. (C) Ellison Ridge drumlins are arranged in a 'rhomboidal' pattern 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 1 dataset, 25 m resolution. BC Albers projection, NAD 83. BC Aerial photographs #15BC97023-126, 15BC97027-43, 45 (copyright British Columbia Government, by permission).

Thompson Plateau drumlins by an abrasive mobile deforming sediment sheet cannot be supported. We develop sediment flux arguments more comprehensively in Section 5.3.

Drumlin genesis has been attributed to erosion and/or deposition by turbulent meltwater flows associated with subglacial underbursts (e.g., Shaw, 1994a,b; 1996; 2002). In this meltwater erosion model, drumlins are upstanding residuals eroded by horseshoe vortices developed around bed obstacles such as bedrock knobs or other bed defects (Shaw and Sharpe, 1987; Sharpe et al., 2004). 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. 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.

### 5.1.2. 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. Such forms could be indicative of sediment squeezing into the lee of an obstacle (Hart, 1995), or, theoretically, of erosion by deforming sediment sheets (e.g., Boulton, 1987). However, their 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 sedimentary exposures. However, the regional context of erosion constrains interpretations. For example, drumlins to the north and south of the Wood Lake forms are eroded in bedrock. If we invoke streaming excavational till sheets 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



**Fig. 11.** Hillshaded digital elevation model illustrating drumlins, coulees and valleys of Okanogan Valley, USA (see Fig. 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 Okanogan Valley and the Columbia River is composed of sand and gravel.

a range of substrates within the same swarm, consistent with drumlin characteristics in Okanogan Valley. In addition, sediments removed by meltwater may be stored within the thick valley fill of Okanogan 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 because 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 Okanogan Valley, drumlin fields throughout the valley probably record erosion by subglacial meltwater flows.

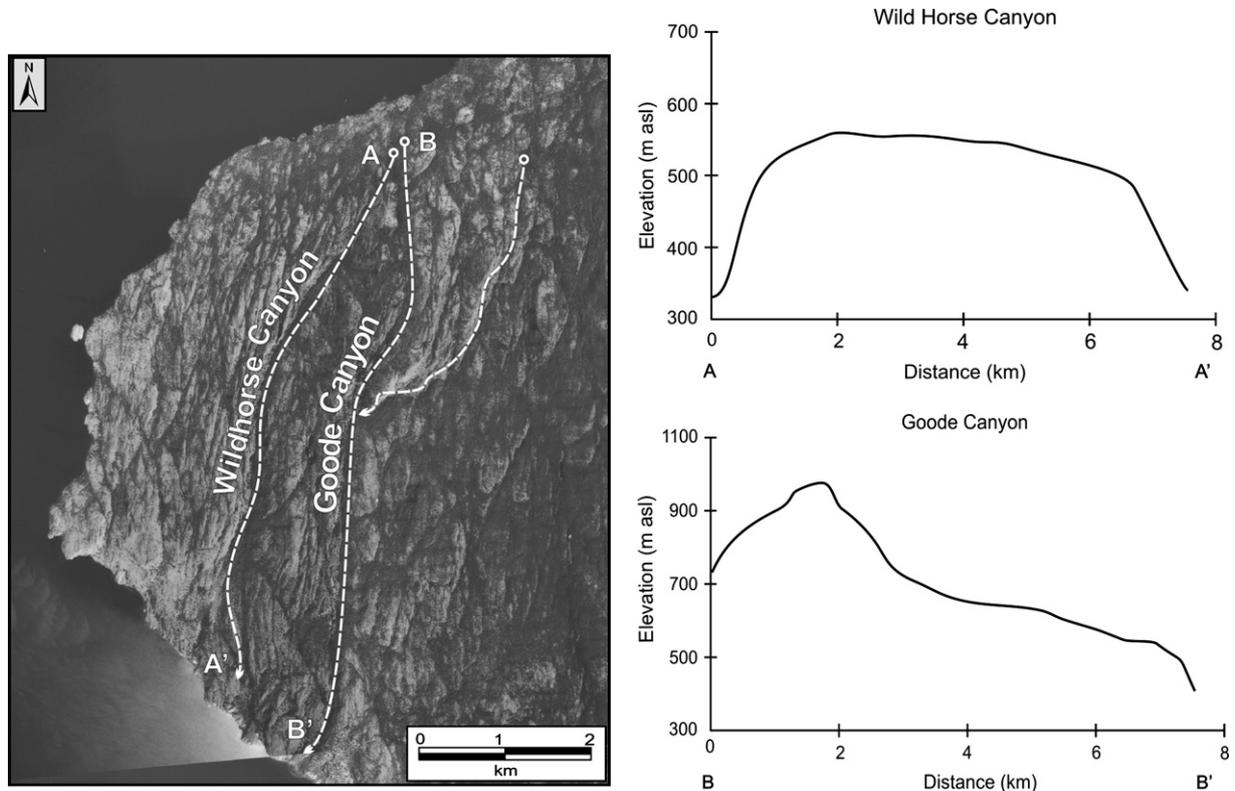
## 5.2. Valleys

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 erosion by pressurized subglacial

meltwater flows. 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'Cofoigh, 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 (Brennand and Shaw, 1994), suggesting common genetic agents.

Most valleys of Thompson Plateau and Okanogan 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. Our use of the term tunnel valley, rather than tunnel channel, reflects this uncertainty. However, we cannot rule out the possibility that some were tunnel channels at least at some point in their development.

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



**Fig. 12.** Aerial photograph of bedrock forms at Squally Point in Okanogan Valley (see Fig. 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, by permission).

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). However, low flow velocities in canals limit clast entrainment and they are unlikely

to erode large tunnel valleys. 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).



**Fig. 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 ephemeral streams. It may have operated as a tunnel valley during glaciation. Smaller bedrock coulees like Whitestone Coulee branch off toward Okanogan valley.

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 (Brennand and Shaw, 1994; Fisher and Taylor, 2002; Wright, 1973; Wingfield, 1990; 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). In these cases, one or a few overlapping fans could occur at conduit mouths where flow expands. Sedimentation can also take place within the channel itself (Russell et al., 2003) and/or subaerially (Cutler et al., 2002). These flows can be fully contained within channel boundaries (e.g., Wright, 1973) or they can 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 jökulhlaup transitioned 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 Okanogan Valley and 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 might be expected during time-transgressive development. These valleys likely formed during a single event.

### 5.2.1. 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 meters 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 they operated contemporaneously and subglacially as tunnel valleys (cf. Brennand and Shaw, 1994).

### 5.2.2. Squally Point: residual bedrock highs flanked by bedrock troughs

At Squally Point, bedrock troughs flank upstanding bedrock residuals (Fig. 11). 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 (figures 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, figure 6E) 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 (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 may have been 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) (Section 5.2.3) 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.

### 5.2.3. 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. From drill cores 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.

### 5.3. Sediment flux continuity and landform associations

The principle of sediment flux continuity 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, we examine sediment flux continuity in the context of a subglacial meltwater model for landscape evolution.

Drumlins on Thompson Plateau and in Okanogan Valley are residuals of a surface layer that acted primarily as a sediment source. In terms of sediment continuity, 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 swarm. 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. Okanogan Valley may be a more adequate sediment sink. However, if we consider drumlin alignment to be indicative of the general direction of sediment flux, only a fraction of sediment transport was directed toward Okanogan Valley from Thompson Plateau. Over most of Thompson Plateau there is a discrepancy in sediment continuity 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 often associated landscape elements. They can form sequentially when sheetflows eroding drumlins collapse into discrete, linear flows to erode valleys (Brennand and Shaw, 1994; Fisher et al., 2005). If we invoke underbursts capable of eroding broad portions of the landscape, water discharge continuity must also be considered along with sediment flux continuity. In the case of the Headwaters Lake area, we surmise that single bedrock conduits cutting through Mt. Kathleen Ridge and connecting to the Garnet Valley network (Fig. 7) were 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 Okanogan Valley via bedrock valleys such as Trepanier and Lacoma Creeks (Fig. 7). Flushing of meltwater and sediment toward Okanogan Valley occurred through its tributary valleys. For sediments to be drawn toward Okanogan Valley, hydraulic potential must have been lower toward the valley and accommodation space for water and sediment must have been available. Therefore, we consider that drumlin formation on Thompson Plateau was contemporaneous with the proposed subglacial scouring of Okanogan Valley (Vanderburgh and Roberts, 1996), and formation/retouching of streamlined bedrock forms in the Okanogan Valley corridor. In addition, the Garnet Valley anabranching network and bedrock

tunnel valleys at Cougar Canyon, Squally Point and possibly in Washington State operated at this time. Thus, storage of sediment in Okanogan 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 Okanogan Valley. Instead, it ends abruptly in the Similkameen Valley (Fig. 3), ultimately connecting to the Okanogan Valley corridor via Palmer Lake Valley (Figs. 1, 11, 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 interfluvies 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. 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. 1). Sand and gravel deposits in the Great Terrace (Fig. 11) (Flint, 1935) 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). Calculated water volumes of glacial Lake Missoula appear insufficient to account for some of the high water marks within the Channeled Scabland tracts (Komatsu et al., 2000). The Okanogan underburst would be an additional meltwater source for the Channeled Scabland formation during the Late Pleistocene and could account for some of the "missing" water volume (Lesemann and Shaw, 2000). In terms of sediment continuity, a meltwater underburst/outburst could transport sediment far beyond the ice margin (e.g., Leventer et al., 1982). Sediments from the Okanogan 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 indicate a source area that includes portions of southern British Columbia and possibly Okanagan Valley (Prytulak et al., 2006).

### 5.3.1. 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<sup>2</sup>) 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 transient water-filled cavities at the ice-bed interface and poreflow within the sediment prior to drumlin formation (Lesemann and Brennand, 2005). 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 palaeogeology implied by associated landforms and sediments? We note that anchored corestones and large boulders resting on bedrock surfaces (Fig. 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, we suggest that 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 interflues 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.

### 5.3.2. 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.

## 5.4. 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 occurred. The shape and fill characteristics of Okanagan Valley suggest voluminous and energetic subglacial meltwater flows. For example, till and pre-glacial sediments are largely absent from the Okanagan Valley sides and fill (Vanderburgh and Roberts, 1996). This general absence, and the over-deepening 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). We propose that Icelandic valleys overdeepened by jökulhlaups are appropriate analogues for Okanagan Valley. We suggest that the 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. 7, 10–13). 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).

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

### 6.1. 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; Bartholomäus 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. Modelling 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 centreline. 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 (Thorarinsson, 1969). However, jökulhlaups do not necessarily lead to ice fracturing. Modelling 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 (Patyn, 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 5.3.2).

### 6.2. Underburst operation and collapse

Critics 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 obstacle (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. Sheetflow 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 Trepane 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 drumlin swarm curving along the low-lying western flank leading to Headwaters Lake Valley (Fig. 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. 7, 10A, 11, 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. 9), Tonasket and Omak Plateau (Fig. 11) were eroded and/or enhanced by the underburst. Missezula valley and the Osprey Lake Valley (Fig. 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. 13) also indicates that subglacial flows were directed southward and joined Okanogan Valley via bedrock coulees debouching at Omak and Tonasket, WA (Figs. 11, 13). Thick basal sequences of boulder gravel in the fill of Okanogan 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 Okanogan Valley opens onto the Columbia Plateau. Erosion during flow expansion may seem counterintuitive. However, it is not if we consider 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 of metres to much larger features such as Moses Coulee, reaching 10s of km in 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, possibly as far as the Astoria fan on the coast of Oregon and California (Brunner et al., 1999; Normark and Reid, 2003).

### 6.3. Meltwater reservoir development

The proposed events for the Thompson Plateau and Okanogan 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. They 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). 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. 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

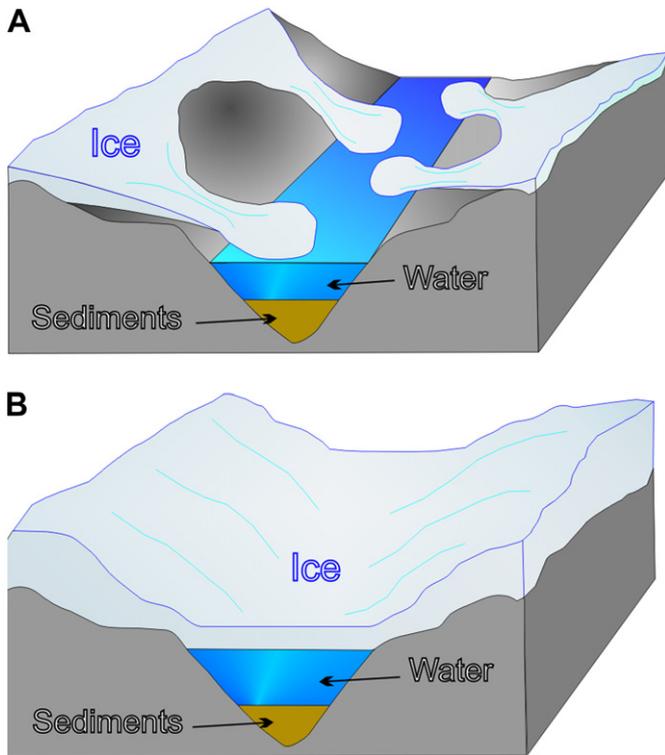
valleys of British Columbia. Once initiated, lakes can expand by further melting of the ice lid, by meltwater inflow from the surrounding area depending on regional head, or from supraglacial sources.

#### 6.3.1. Subglacial lake development in Okanogan Valley and adjacent valleys

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 Okanogan 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. We have discussed evidence for warm-based conditions around Thompson Plateau and Okanogan Valley. This is potentially problematic for lake sealing. However, there are mitigating factors that still allow ice shelves to develop. For example, the proposed subglacial lake length (within the deep, narrow trough of Okanogan 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 Okanogan 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 Okanogan 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. 14A). Finally, the role of seasonal lake ice must be considered. If we envisage calving into a subaerial lake prior to capture, 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 Okanogan Valley also occur in neighbouring valleys. Presently, not all valleys are hydrologically connected but they could be, given minor (metre-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 Okanogan Valley (Vanderburgh and Roberts, 1996; Eyles and Mullins, 1997). 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



**Fig. 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.

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 the lack of 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 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 exceeds 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 overcome.

### 6.3.2. 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 heat flux in Okanagan Valley and southern British Columbia is very high (Fairbank and Faulkner, 1992). Geothermal springs with waters reaching 137 °C emerge into Okanagan Lake near Kelowna (Grasby and Hutcheon, 2001) (Fig. 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). We postulate that ice over Okanagan Valley likely thinned and may have flowed rapidly during subglacial lake development (Sections 6.3.1, 6.4) 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. 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) (Hoskuldsson 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. 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.

### 6.3.3. 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) drumlinized 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. 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. We explore these more speculative regional connections.

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. 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. 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 are possible through the bedrock valley of Mahood and Canim Lakes (Fig. 2). Consequently, we speculate 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.

### 6.4. Effects of subglacial lake development on CIS geometry

The possibility that rapid ice flow occurred over wide areas during underbursts was largely rejected earlier (Section 5.3.2). 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 (Section 6.3.1).

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 numerical ice sheet models.

## 7. 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. Geodynamic conditions (thermal springs and subglacial volcanic eruptions) enhanced subglacial lake development and are 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 could have become the sites for supra-glacial 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. 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. We do not question processes of sediment deformation or abrasion *per se*. However, we do question their ability to produce the drumlins in our region. 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.

Our interpretations of the Thompson Plateau and Okanagan Valley landsystems also have repercussions for ice-sheet modelling 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.

## Acknowledgements

Financial support was provided by a Natural Sciences and Engineering Research Council of Canada *Discovery Grant* and *Research Tools and Instruments Grant* to T.A.B., and *Geological Society of America Student Research Grants* to J.-E.L. We offer thanks to field assistants Amanda Denney and Steve Momeyer. We also extend our gratitude to the British Columbia Integrated Land Management Bureau and its Base Mapping and Geomatic Services for providing digital elevation datasets for this project. We thank Glenn Thackray and an anonymous reviewer for their constructive comments, which improved the manuscript.

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