Glacial imprints of the Okanogan Lobe, southern margin of the Cordilleran Ice Sheet

DORI J. KOVANEN* and OLAV SLAYMAKER
Department of Geography, University of British Columbia, Vancouver, British Columbia, V6T 1Z2, Canada

ABSTRACT: The glacial geomorphology of the Waterville Plateau (ca. 55 km^2) provides information on the dynamics of the Okanogan Lobe, southern sector of the Cordilleran Ice Sheet in north-central Washington. The Okanogan Lobe had a profound influence on the landscape. It diverted meltwater and floodwater along the ice front contributing to the Channeled Scabland features during the late Wisconsin (Fraser Glaciation). The glacial imprint may record surge behaviour of the former Okanogan Lobe based on a comparison with other glacial landsystems. Conditions that may have promoted instability include regional topographic constraints, ice marginal lakes and dynamics of the subglacial hydrological system, which probably included a subglacial reservoir. The ice-surface morphology and estimated driving stresses (17–26 kPa) implied from ice thickness and surface slope reconstructed in the terminal area also suggest fast basal flow characteristics. This work identifies the location of a fast flowing ice corridor and this probably affected the stability and mass balance of the south-central portion of the Cordilleran Ice Sheet. Evidence for fast ice flow is lacking in the main Okanogan River Valley, probably because it was destroyed during deglaciation by various glacial and fluvial processes. The only signature of fast ice flow left is the imprint on the Waterville Plateau. Copyright © 2004 John Wiley & Sons, Ltd.
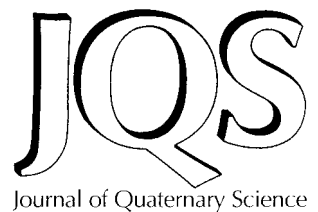
KEYWORDS: Okanogan Lobe; geomorphology; Cordilleran Ice Sheet; ice stream; glacial landsystem.

Introduction

Along the southern margin of the Cordilleran Ice Sheet (CIS), the growth and decay of the Okanogan Lobe, a large outlet glacier, formed an assortment of glacial imprints on the land surface. The Okanogan Lobe was the largest glacier east of the Cascade crest, and cataclysmic floods and meltwater routing events were governed by the topography and extent of ice cover, especially when ice blocked the Columbia River. This in turn affected the erosional development of major channel complexes in north-central Washington (Fig. 1A; e.g. Daly, 1912; Bretz, 1923, 1930, 1969; Waters, 1933; Flint and Irwin, 1939; Baker, 1973; Waitt, 1982; Atwater, 1986; Dawson, 1989). Although the Quaternary geology of the region has been the focus of research for a century, especially relating to Missoula floods and the carving of the Channeled Scablands, much remains to be elucidated with regard to the dynamics of the Okanogan Lobe. Detailed regional ice-flow patterns have not yet been reconstructed, although they have been inferred from the maximum ice-limit. The only ice-surface reconstruction comes from Waitt and Thorson (1983) with a 1000 m contour interval and is a generalisation (no ice-flow data). In this paper, we describe the geomorphology and associated sedi-

ments of the Okanogan Lobe in Washington State to provide a better understanding of the controls on ice sheet and glacier activity.

Study rationale and objectives

Glacial landforms and sediment characteristics provide information about the physical nature of former glaciers and ice sheets, including: ice-flow directions, transient conditions at the ice–bed interface, position(s) of former ice centre and volume change, and various mechanisms of ice-sheet growth and collapse. Each landform type with its associated sediments and structures will reflect certain subglacial sedimentary environments and may be used to infer subglacial processes and dynamics of ice flow (e.g. Boulton and Clark, 1990; Hicock and Dreimanis, 1992; Clark and Walder, 1994; Hicock et al., 1996; Clark, 1997; Benn and Evans, 1998; Evans, 2003).

Early published regional summaries (Richmond et al., 1965; Easterbrook, 1979; Waitt and Thorson, 1983; Clague, 1989) refer to the glacial landform types, but little information is presented with regard to the associated sediments and the glaciodynamic conditions they may represent. Some of these same studies refer to diamictons as 'till' that apparently caps some of the features, however, data pertaining to the diamicton genesis (interpretation of the various till types) have not been

* Correspondence to: Dori J. Kovanen, 3117 Chandler Parkway, Bellingham, WA 98226, USA. E-mail: dkovanen@geog.ubc.ca
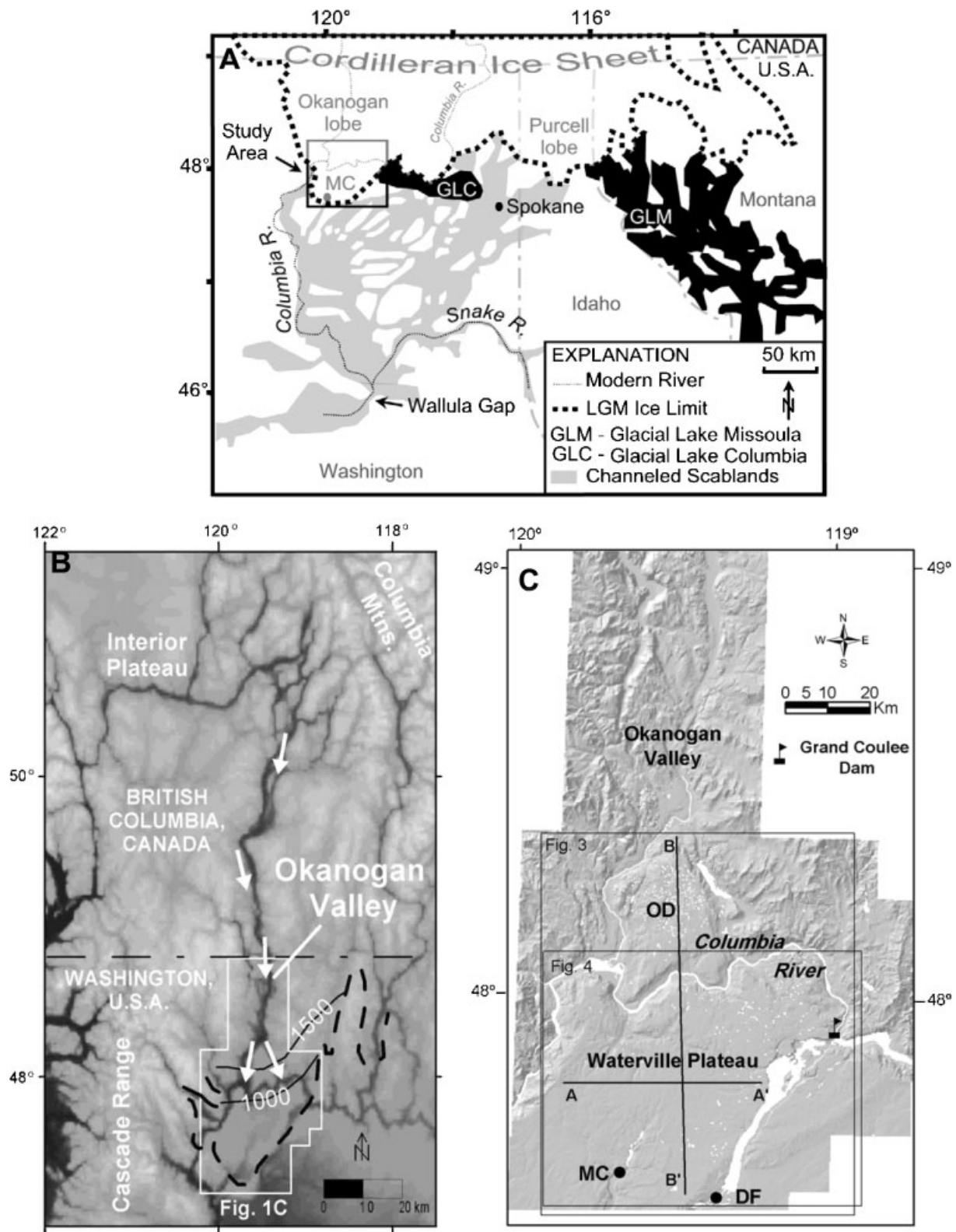
Figure 1 (A) Map of the late Pleistocene ice limit of the eastern portion of the Cordilleran Ice Sheet. Glacial lakes are shown approximately. (B) Simplified map of the maximum extent of ice on the Waterville Plateau and the upstream area with present-day topography and trunk drainage routes. (C) Shaded digital topographic model of the Okanogan valley and the Waterville Plateau. OD, Okanogan Dome; DF, Dry Falls; MC, Moses Coulee. Horizontal resolution is 10 m. For profiles A and B refer to Fig. 12

presented. This is probably because the importance of the rheology of the substrate in influencing ice flow and ice-sheet stability was not realised until recently. Criteria for identification of till types and the perceptions regarding the genesis of subglacial diamictons are given by Boulton and Hindmarsh (1987), Dreimanis (1989), Boyce and Eyles (1991), Hicock and Dreimanis (1992), Paterson (1994), Boulton (1996),

Hicock *et al.* (1996), Benn and Evans (1996), Ruszczyńska-Szenajch (2001) and Lian *et al.* (2003). In the context of our study, the latest assessment of the subglacial till and associated sediments indicates that central parts of the CIS in Canada experienced fast ice flow and progressive changes in thermal conditions at the glacier bed during glaciation (Lian and Hicock, 2000; Lian *et al.*, 2003). Additionally, recent analysis

of glacial depositional features in British Columbia and the Scablands erosional architecture suggests that many of the Missoula floods may have come from subglacial outbursts from the CIS (Shaw et al., 1999) and this has fueled recent debate on the source of the meltwater (Atwater et al., 2000; Komatsu et al., 2000; Shaw et al., 2000).

In this paper, we have mapped distinct glacial landforms (mostly drumlins and megaflutes, eskers and moraines) and have discussed the general characteristics of associated sediments from representative sites to provide data to explore the ice–bed conditions in the terminal area of the Okanogan Lobe. We then reconstruct the ice-flow system and ice surface. Finally, we compare the main depositional components with other glacial landsystems (e.g. Evans, 2003) to provide a better understanding of the overall ice lobe dynamics.

Physiographic setting and study area

The physiographic setting of south-central British Columbia and north-central Washington is critical to understanding the distribution of both glacial landforms and sediments. The region is situated on the leeside (east) of the Cascade Range and is arid to semi-arid. The present annual average temperature at Waterville is 7.3 °C and precipitation is 0.29 mm yr^{-1} (1931–2003; Waterville meteorological station, National Ocean Atmospheric Administration). The main physiographic elements of the study area reflect the bedrock topography (Fig. 1B and C). The Okanogan River Valley in Washington State is a north–south-trending depression, bounded to the west by the Okanogan Range with summit elevations up to 2400 m, and on the east by the Okanogan Highlands that generally lie below 1800 m. (The spelling for Okanogan varies; that is, Okanagan in British Columbia and Okanogan in Washington.) During the last glacial cycle, glacier outflow and meltwater was channelled south through the topographically controlled drainage from dispersal centres on the western flank of the Columbia Mountains and Interior Plateau in British Columbia (Fig. 1B; Willis, 1887; Dawson, 1898; Pardee, 1910, 1922; Daly, 1912; Bretz, 1923, 1925, 1959; Flint, 1935, 1937; Nasmith, 1962; Fulton and Smith, 1978). The channelling of ice and meltwater interactions has important implications for understanding the dynamics of ice flow because it provides lubrication at the ice–bed interface and this in turn affects the ice-sheet dynamics (e.g. Clarke et al., 1984a). A modern analogue for these physiographic conditions may be Antarctica with its ice streams (e.g. Alley et al., 1987). Many of those glaciers are predisposed to streaming of ice owing to their topographic disposition. The main difference here is that the Antarctic ice-streams have a tidewater margin, whereas the Okanogan Lobe was land-terminating.

The Columbia River separates the Okanogan River Valley and Waterville Plateau by deeply incising the bedrock along terrain boundaries. This is marked by major escarpments with relief up to ca. 500 m. The Waterville Plateau is situated between ca. 500 and 915 m a.s.l. (mean 671 m), with a mean regional slope of 5° (8%) toward the central portion of the plateau. The plateau is a broad concaved-shaped basin; the rim of the plateau rises from ca. 50–200 m above the mean elevation (Fig. 2); and is entrenched at Foster and Moses coulees. Presently these coulees are essentially dry channels. The bedrock of the Waterville Plateau consists mostly of Miocene, horizontally bedded, tholeiitic basalt, of the Columbia River Basalt Group (17–6 Ma; Hooper and Swanson, 1987). On a local scale, it is jointed and fractured, making it susceptible to erosion by plucking. Bedrock in the Okanogan River Valley consists generally of granitic batholiths, low-grade metamorphic rocks and the crystalline Okanogan dome (Cheney, 1980). A discontinuous blanket of drift mantles bedrock on the Waterville Plateau, and thick valley fills (>300 m) exist in the Okanogan River Valley (in British Columbia, Nasmith, 1962; in Washington, Pine, 1985).

Previous work and Quaternary history

Early studies focused on particular aspects of Quaternary geology and geomorphology. Bretz (1923, 1925, 1959), Pardee (1910, 1922), Waters (1933), Flint (1935), Flint and Irwin (1939) and Bretz et al. (1956) recognised the role of glaciation in modifying the landscape. Flint (1937) and Flint and Irwin (1939) described a thick sequence of sediments that fill the valley near Grand Coulee (Fig. 1C) and suggested that two (early and middle Wisconsin) advances of the Okanogan Lobe occurred. These advances blocked the Columbia River and created glacial lake Columbia. This glacial lake stretched across much of northern Washington and was a major Pleistocene physiographic feature. The combined flows of the drainage of glacial lakes Missoula and Columbia, and the diversion of meltwater along the margin of the Okanogan Lobe, have been proposed for the origin of Lower Grand Coulee and the Channeled Scabland (Bretz, 1923; Flint and Irwin, 1939; Fig. 1).

Later studies applied modern generalised sedimentological, geomorphological and dating methods to this region (e.g. Hanson, 1970). They include regional syntheses that focused on the relationships of the Okanogan Lobe and other sublobes of the CIS (Richmond et al., 1965; Easterbrook, 1979, 1992; Waitt and Thorson, 1983; Clague, 1989), and ice-dammed lakes, especially glacial lake Missoula and its many outburst floods (Baker, 1973; Mullineaux et al., 1978; Clarke et al., 1984b; Baker and Bunker, 1985; Waitt, 1985; Atwater, 1986; O'Connor and Baker, 1992; Smith, 1993; to name only a few).

The timing of the advance and retreat sequence of the Okanogan Lobe is not well constrained. The main chronological control comes from: (i) the presence or absence of volcanic ash in deposits, which provide limiting ages (e.g. Mullineaux et al., 1978; Porter, 1978; Mack et al., 1979; Waitt and Thorson, 1983; Carrara et al., 1996); (ii) the correlation with other major ice lobes, some of which may (Puget Lobe) or may not (Columbia River and Purcell Trench lobes) be well-dated (i.e. Richmond et al., 1965; Easterbrook, 1979); and (iii) varve chronology based on dated lacustrine sediments (e.g. Waitt, 1985; Atwater, 1986). Build-up of ice and inundation of the landscape probably did not occur until after ca. 17 500 ^{14}C yr BP (Fulton and Smith, 1978; Clague et al., 1980) and is thought to be broadly synchronous with the Puget Lobe, west of the Cascades (Waitt and Thorson, 1983; Easterbrook, 1992). The Puget Lobes' maximum extent was achieved at ca. 14 500 ^{14}C yr BP, ca. 3500 yr after the global glacial maximum, indicating that portions of the CIS responded extremely fast to environmental changes. The absence of Glacier Peak tephra layer G from the lower Okanogan Valley suggested to researchers that the valley was still ice covered at the time of eruption ca. 11 200 ^{14}C yr BP (Porter, 1978; Mehringer et al., 1984).

Traditionally, formation of the Channeled Scablands has been associated with catastrophic draining of glacial lake Missoula. Waitt (1985) advocated the time period for repeated drainage from glacial lake Missoula to be between 15 300 and 12 700 ^{14}C yr BP, but it was found recently that flooding occurred earlier than 15 500 ^{14}C yr BP (Brunner et al., 1999).
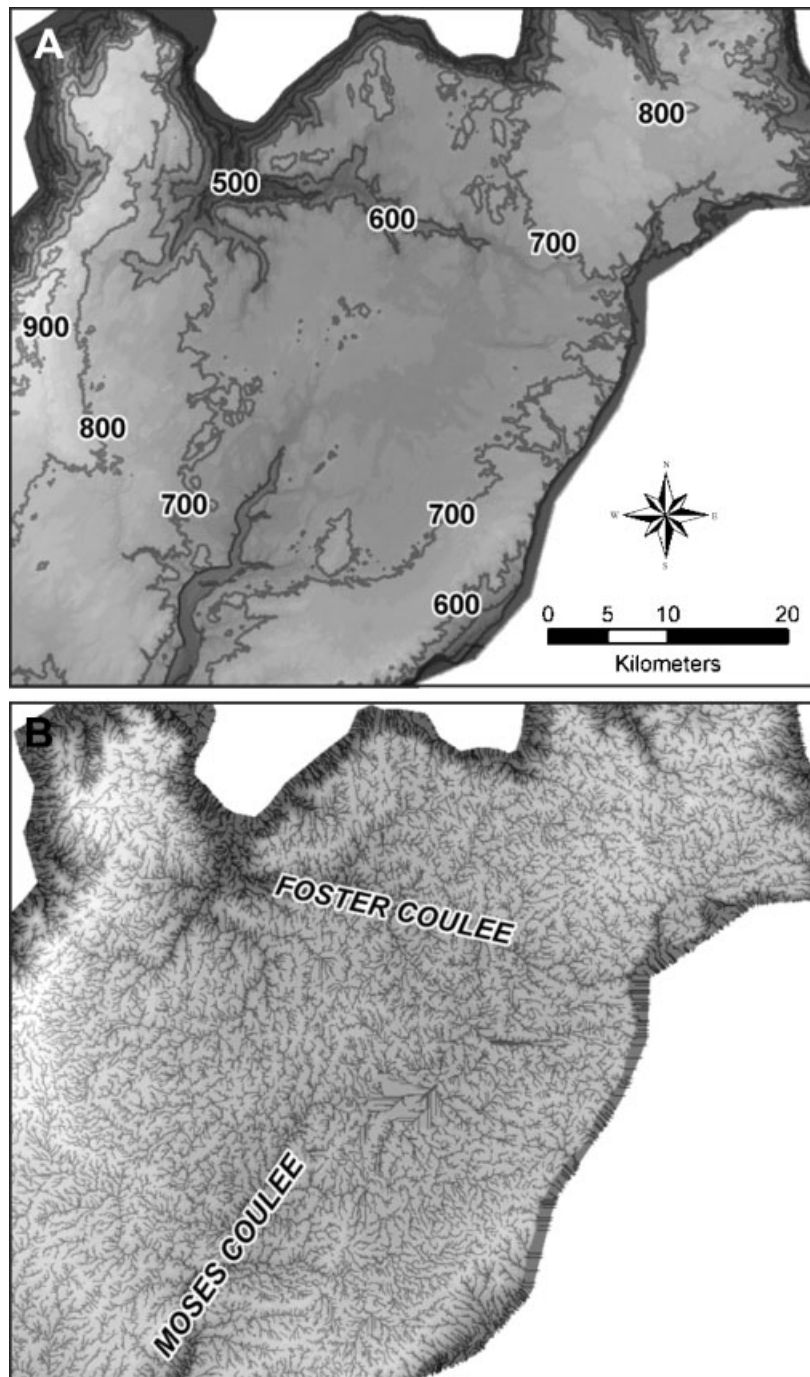
Figure 2 (A) Topographic characteristics of the Waterville Plateau. (B) Drainage network on the Waterville Plateau extracted from the DEM

The Okanogan Lobe probably dammed the Columbia River between about $15\,550 \pm 400$ and $13\,050 \pm 650$ ^{14}C yr BP (Atwater, 1986).

Database and methods

The distribution of glacial features was extracted from 1:15 000, black-and-white aerial photographs, 1:24 000 topographic maps, and 10 m digital elevation models (DEMs). Eighty-five USGS (U.S. Geological Survey) 7.5 minute quadrangles were acquired from the Washington State Geospatial Data Archive (http://wagds.lib.washington.edu) and merged together in a geographical information system (GIS; ESRI software,

ArcView 3.2, ArcGIS 8.2, and Golden software, Surfer 8.0) for seamless topographic representation of the area. The DEM was smoothed with a 3×3 filter. This data base was used to delineate trends of features that were identified on the aerial photographs. The direction of ice flow was inferred from the orientation of streamlined features (mainly drumlin, macroflutes, till lineations). Where exposures were available, sections were cleaned with a shovel to reveal glacigenic structures and the general character of the sediments was described.

In addition, information was compiled from earlier studies (Easterbrook, 1979; Hanson et al., 1979) and 1:100 000 scale surficial maps by the Washington State Department of Natural Resources, Division of Geology and Earth Resources (Gulick and Korosec, 1990; Waggoner, 1992).

The drainage network shown in Fig. 2B was extracted from the DEM using common hydrological analysis tools in the GIS

software. The extraction process included: removal of spurious minima (pit filling); calculation of flow direction (average slope in direction of flow) and flow accumulation (or commonly referred to as expected upslope area); and derivation of the flow path and stream network from the flow accumulation grid.

Terminology

Morphology and sedimentology have been used traditionally to reconstruct the glacial history of formerly glaciated areas. Therefore, before we can begin our analysis and discussion, we must define the terminology used in this paper. We have adopted the terminology for the various ice lobes used by other authors (Richmond *et al.*, 1965; Waitt and Thorson, 1983). Sublobes of the Okanogan Lobe are known as the Columbia River and Omak sublobes, named after drainages that were occupied by ice at the inferred maximum extent. To date, only limited geomorphological data have been presented for the margins of these sublobes.

Drumlins are elongate to oval hills, commonly tear-drop shaped at a range of sizes, their genesis includes erosion and pervasive deformation of ice–bed sediments (Boulton, 1987). These processes reflect ice mass-balance destabilisation, strength of the underlying sediments, ice-marginal oscillation and, hypothetically, ice-flow velocities. Drumlins can be distinguished by the composition of their cores: (i) till drumlins; (ii) glaciofluvial drumlins covered by thin layer of till; (iii) bedrock-cored drumlins. It has been suggested that catastrophic floods that drain subglacial meltwater can erode the ice–bed, creating cavities where sediment is deposited in the drumlin shape or reshaped by the above-mentioned processes (e.g. Shaw, 1983; Shaw and Gilbert, 1990). Drumlin genesis and their glaciodynamic significance are widely debated (e.g. Menzies and Rose, 1987) and the glacial and fluvial hypotheses both involve water; free water along the ice–bed interface and/or porewater in saturated sediments.

Flutes are distinctive bedforms composed of till, streamlined parallel to the direction of former ice flow. The models proposed for flute formation include, lodgement processes, post-depositional plastic deformation in the lee of an embedded boulder, primary deposition and erosion. Drumlins and flutes are probably part of a continuum of forms in a landform assemblage. Because the distinction between morphometric parameters (length and width) are somewhat arbitrary, we mapped drumlins and flutes as streamlined forms. Many low-amplitude flutes were visible on aerial photographs, but were not mapped as they did not make a topographic expression on the 10-m DEMS. However, they did provide supporting information for the inferred ice-flow direction.

Eskers in a morphological sense are commonly recognised as elongate, straight or sinuous, and continuous or segmented/beaded glaciofluvial ridges, which approximate ice-flow directions at the margin of retreating warm-based glaciers. Esker formation in tunnel, tunnel-mouth, re-entrant and proglacial environments is well established in the literature (e.g. Brennand, 2000; references therein). Modes of esker deposition involve a complex origin as a result of deposition in different depositional environments and show a large variation in facies characteristics (Brennand and Shaw, 1996). In our field area, eskers were easily mapped because of modern land-use practices; they rise above the ploughed field surface or are found as ridges in valleys.

Ice streams are characterised by abundant subparallel erosional and depositional flow-parallel features. Ice streams experience enhanced ice-flow rates that result from decoupled sliding over the ice–bed or from failure and deformation of subglacial sediments. Flow may be transitory or continuous. Flow mechanisms are driven at the bed and governed by time-dependent thermal and hydrological regimes beneath the ice, geology and topography, and require a temperate bed that permits free water (e.g. Clarke, 1987; Paterson, 1994). Ice streams drain large portions of ice sheets and therefore play a critical role in controlling ice-sheet geometries and their pattern of change.

Results: sediment zones and geomorphology

In this section the distribution and relative age of glacial landforms are described. The sediments associated with representative features are also described. The geomorphology is subdivided into features that relate to the subsequent detailed reconstruction of the ice-flow system and ice surface in the terminal area of the Okanogan Lobe: streamlined forms (drumlins and macroflutes), moraines and hummocks, eskers, and meltwater channels.

Erosion and deposition zones

Meltwater release at the base of a glacier can be characterised as a thin layer, channellised flow, and by discharge of meltwater through subglacial aquifers. An important aspect of subglacial meltwater drainage, whether catastrophic or steady-state flow, is that it influences the behaviour of glaciers and ice sheets. Hence, the partitioning of the erosion and deposition areas may provide insights into the ice motion attributed to different mechanisms including basal motion, sediment deformation and changes in the subglacial water pressure or effective pressure distributions (e.g. Boulton, 1987, 1996; Paterson, 1994).

The large-scale pattern of exposed bedrock (erosion) and sediment-rich (deposition) areas is shown in Fig. 3. In our immediate field area, the subglacial sediment sequence is generally thin, less than 15 m, resting on a rigid bed (basalt). The sediments associated with individual glacial features may be of greater thickness, but the fill in swales between features is thin and not well exposed. This suggests that deposits may have been removed by the subglacial meltwater discharge in the melting zone. A higher proportion of drift is found concentrated along the western and south-central portion of the Waterville Plateau (sediments associated with the Withrow moraine are described in the next section). In contrast, a larger fraction of bedrock is exposed to the east and forms vertical walls along drainages. A broad swath of scoured bedrock extends for 20 km along the eastern border of the plateau (see Fig. 3) and is about 150 m lower than the western boundaries of the plateau. This suggests that the elevation and slope of the bedrock surface may have been a controlling factor on erosion. The exact mode of erosion is not identified, although evidence for several generations of erosion include glacial striations on the bedrock surface in the direction of ice flow, and interfluves that have become incised by inner channels indicative of meltwater erosion.

A relict scoured landscape is observed behind the Withrow moraine. Drumlins and eskers drape discordantly over the scoured surface, post-dating the initial erosion (streamline forms and eskers are described below). The spatial continuity of the scoured surface is complex and is also probably
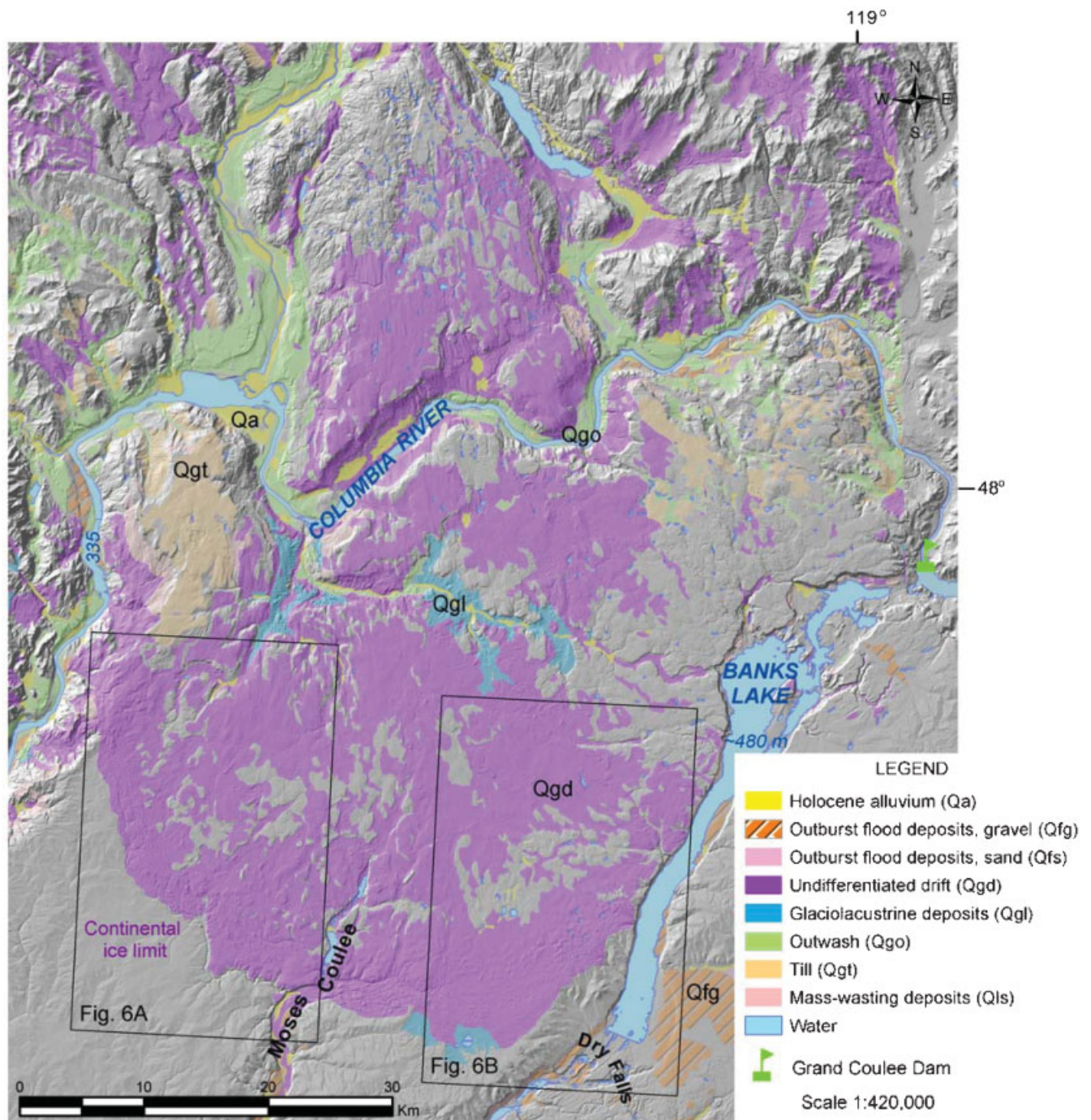
Figure 3 Distribution of glacial deposits on the Waterville Plateau and Okanogan uplands. The surficial geology is draped on to a shaded digital topographic model. Refer to Fig. 1B for location. The grey shading is exposed bedrock (data from Hanson *et al.*, 1979; Waggoner, 1992)

polygenetic. For instance, Foster and Moses coulees were probably inundated by meltwater at various times during the Pleistocene, but now are dry. Foster Coulee in now partially filled with glaciofluvial and glaciolacustrine sediments that were deposited during deglaciation. Sediment-free areas exist behind the Withrow moraine and appear to be linked together. Therefore, the nature of the bedrock surface is attributed to a combination of processes: subglacial and proglacial meltwater erosion, and/or changes between different aspects of the sub-glacial environment (glacial erosion). The meltwater from sub-glacial and englacial storage sites and basal melt could have drained through steady-state flow and/or by catastrophic out-burst. Because the drift cover is relatively thin, essentially impermeable over large areas, the pressure owing to the ice and volume of basal water probably had a strong influence

on friction and glacier dynamics (e.g. Boulton and Hindmarsh, 1987). This in turn affects the way we interpret sediment–pro-cess relationships in a glacial landsystem model and the ice-surface reconstruction (below).

Streamlined forms (drumlins and macroflutes)

The spatial distribution of streamlined forms, eskers and mor-aines is shown in Fig. 4. Streamlined forms begin at the With-row moraine and continue up ice into British Columbia, making a bottleneck flow pattern. The streamlined features exist at varying elevation on both topographic highs and lows. Drumlins vary in size, 0.1–2 km long, 0.1–0.5 km wide, with a
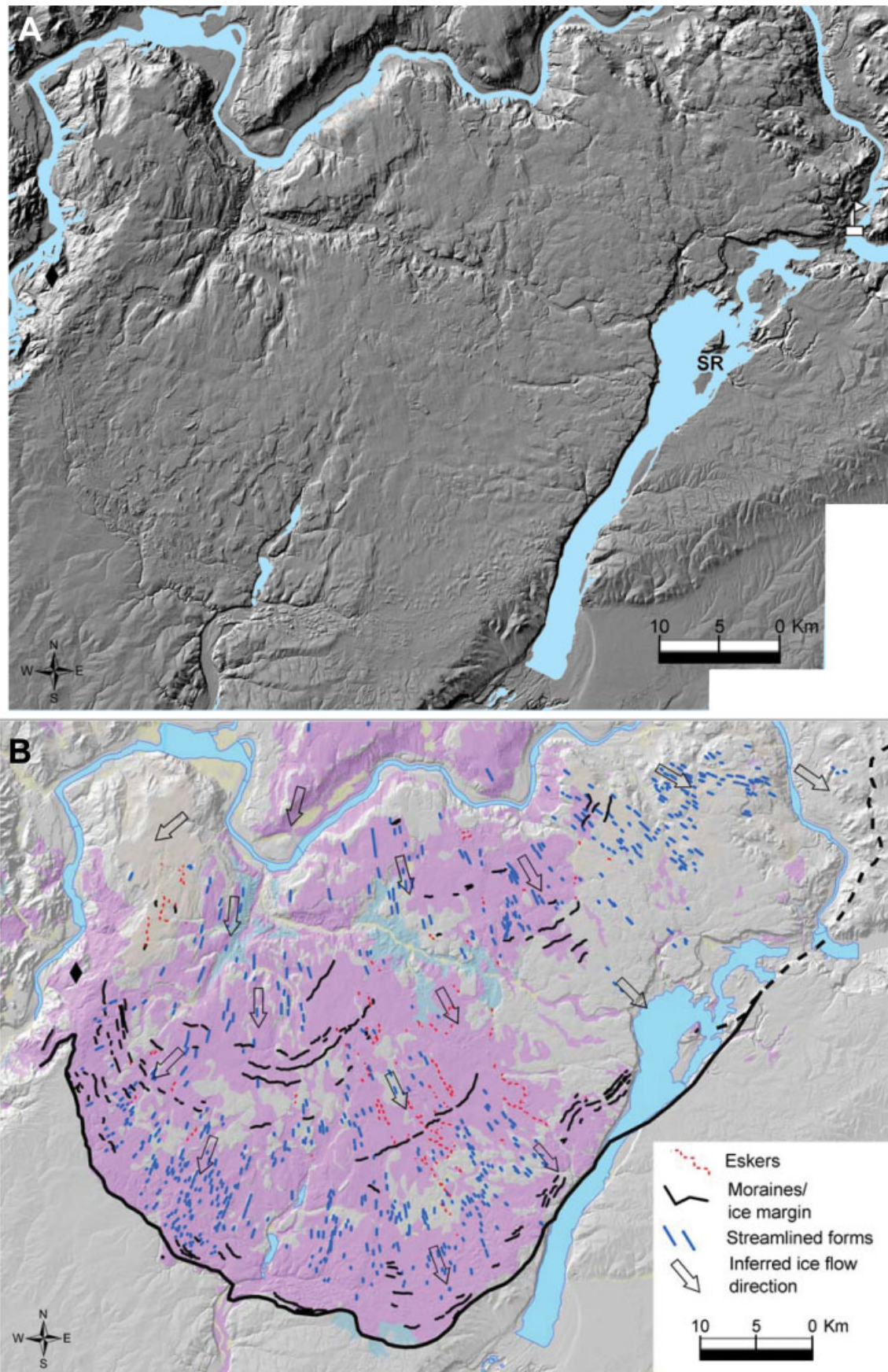
Figure 4 (A) Shaded digital topographic model showing the terminal area of the Okanogan lobe on the Waterville Plateau. Refer to Fig. 1B for location. (B) Distribution of large-scale glacial landforms and inferred ice-flow directions. Eskers formed during the deglaciation. Successive ice margins during retreat are also shown. From other light source angles (not shown here) used to construct the digital model these features and others may be more (or less) apparent. SR, Steamboat Rock. ◆ Indicates location of Fig. 11. This figure appears in color on the journal's website: www.interscience.wiley.com/journal/jqs
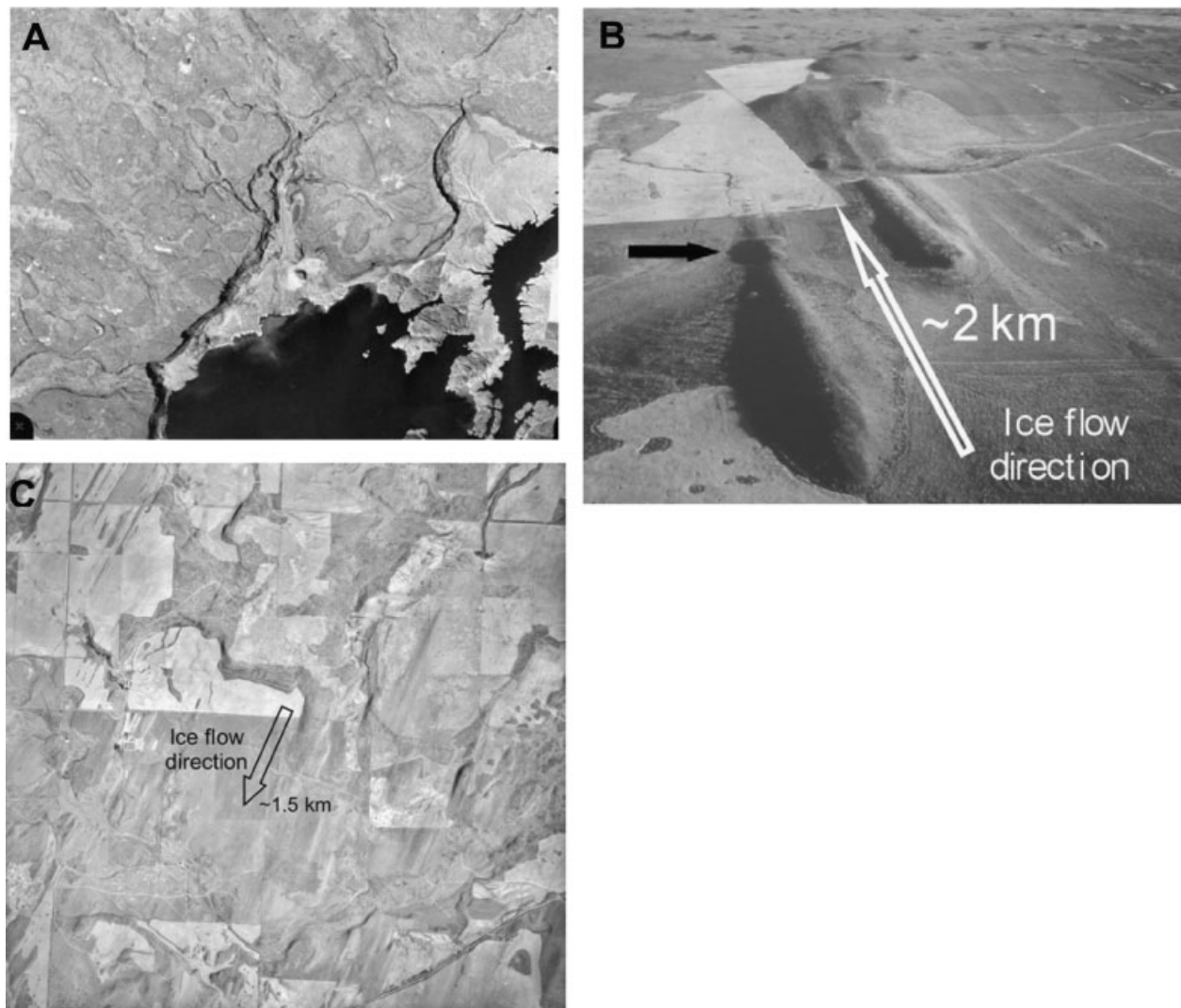
Figure 5 Oblique aerial photographs of landforms representative of the area. (A) Scoured bedrock surface along the eastern margin of the plateau. A network of channels is incised into the bedrock. (B) Drumlins near the terminal zone. Note overlapping feature (arrowed). (C) Lineations and esker (upper right-hand corner)

relative height up to 30 m. On the Waterville Plateau, their length generally increases with distance from the Withrow moraine, with width-to-length ratios ranging from 1:3 to 1:15. Drumlin density is higher over the sediment-rich areas, the crests of drumlins form a radiating pattern which emanate from the Okanogan Valley to the north (Figs 1 and 4B). They commonly occur in the lee of bedrock highs or escarpments. Smaller drumlins are found interspersed with the large drumlins and none appear to overlap. Their shapes range from parabolic to teardrop-shaped (Fig. 5B). Large drumlins are clearly older than retreat moraines. At many locations, recessional moraines drape over large drumlins and macroflutes. Eskers are also superimposed upon large features. Low-relief flutes with elongation ratios >1:15 are situated between drumlins (Fig. 5C). The internal structures of these large streamlined forms have not been well exposed. At one locality, coarse-grained, stratified, gravel and sand glaciofluvial beds are overlain by a thin matrix-supported diamicton cover. Drumlin-forming processes included erosion or streamlining and remoulding of pre-existing sediments caused by sliding at the ice–bed interface. The range in size of the streamlined forms results from variations in flow velocity (Chorley, 1959; Hart, 1999), length of time of formation, sediment availability and strength (Boulton and Hindmarsh, 1987; Boyce and Eyles, 1991). An explanation for the large drumlins that were left

unmodified by the retreating/thinning glacier is sustained frozen-bed conditions and/or absence of basal sliding. Additional systematic study is needed to determine downstream variations in mean streamlined form elongation ratio, which may hypothetically indicate changes in ice-flow velocity.

Moraines and hummocks

The glacial-maximum moraine of the Okanogan Lobe is known as the Withrow moraine (e.g. Bretz, 1923; Flint, 1935; Richmond et al., 1965). It is a 5–9 km wide (Figs 4 and 6) morainal complex consisting of isolated ridges and hummocks composed of sorted sediments, probably glaciofluvial in origin. The relief of these features ranges from 30 to 50 m and they are generally situated transverse to the ice-flow direction. Many of the hummocks merge to form moraine ridges and some appear stacked (Fig. 6). Along the western margin of the plateau (Fig. 6A) the closely spaced stacked moraine ridges may account for sediment thickening in this area. Along the eastern margin, where the relief is lower, individual asymmetric curvilinear ridges appear more subdued and distributed more or less en échelon, indicating oscillations of the glacier (Figs 6B and 7). Some of the hummocks are streamlined in

the down-ice direction (Fig. 7). It is not known whether the hummocks were overridden by ice, whether they formed by pushing of proglacial material at the edge of active ice, or whether they are products of ice-marginal squeezing of water-soaked subglacial sediments. The streamlined form on some of the crests (cupola hills) may indicate that the production of the bedforms and moraines may be related.

Steamboat Rock (Fig. 4A; elevation 213 m) is covered with some drift in distinct ridges and mounds that are interpreted as the continuation of the Withrow Moraine. One interpretation is that the Okanogan Lobe overrode Steamboat Rock after Grand Coulee was formed (Bretz, 1959, 1969).

The Withrow morainal complex drapes into Moses Coulee and overlies a giant flood bar. This indicates that the moraine is younger and the glaciofluvial activity or flood events that incised the coulee ceased sometime prior to the ice advance to its maximum extent (Bretz, 1923; Flint and Irwin, 1939; Hanson, 1970; Easterbrook, 1979). The large size of the terminal zone was inferred by previous researchers (Easterbrook, 1979, 1992; Waitt and Thorson, 1983) to indicate that the Okanogan lobe remained at its terminal position for a prolonged period. Patterned ground is occasionally observed near the terminal margin and this may have influenced the subglacial hydrology by reducing the hydraulic conductivity of the drift.

A sequence of heavily deformed coarse- to fine-grained, sorted sediments is exposed in a gravel pit along the western margin of the Withrow moraine (Fig. 8). This sequence of glaciofluvial sediments is up to 8 m thick and is overlain by a thin (1–3 m thick), stratified, coarse-grained gravel and sand. A glaciotectonic thrust structure was observed. The displacement plane appears between the fine-grained sand and silt layers and coarse-grained gravel and sand with sand interlayers. The displacement plane dips steeply (nearly vertical) and the 'ramp' is not well exposed. Hinges of small isoclinal folds near the displacement zone in the fine-grained sand and silt are steeply dipping and orientated subparallel to the ice-flow direction. With distance from the displacement plane the folds become disharmonic to asymmetric. These types of glaciotectonic structures are commonly associated with wet-based ice. In contrast, the coarse-grained gravel and sand layers form more coherent blocks with the macrofabric of the layers dipping toward the displacement plane and showing no intrafacies deformation. This unit becomes more gravelly and massive away from the displacement plane and gives the appearance of thickening. These complexities probably arise where different sedimentary layers respond differently to the applied stresses and the time-dependent basal thermal evolution. An interpretation for this sequence is that shear failure of the sediments resulted from the drag of the overriding glacier. Or the sediments could have been deformed by tensile fracturing parallel to the direction of stress, when the stress exceeded the sediment strength. Based on the nearby geomorphology,
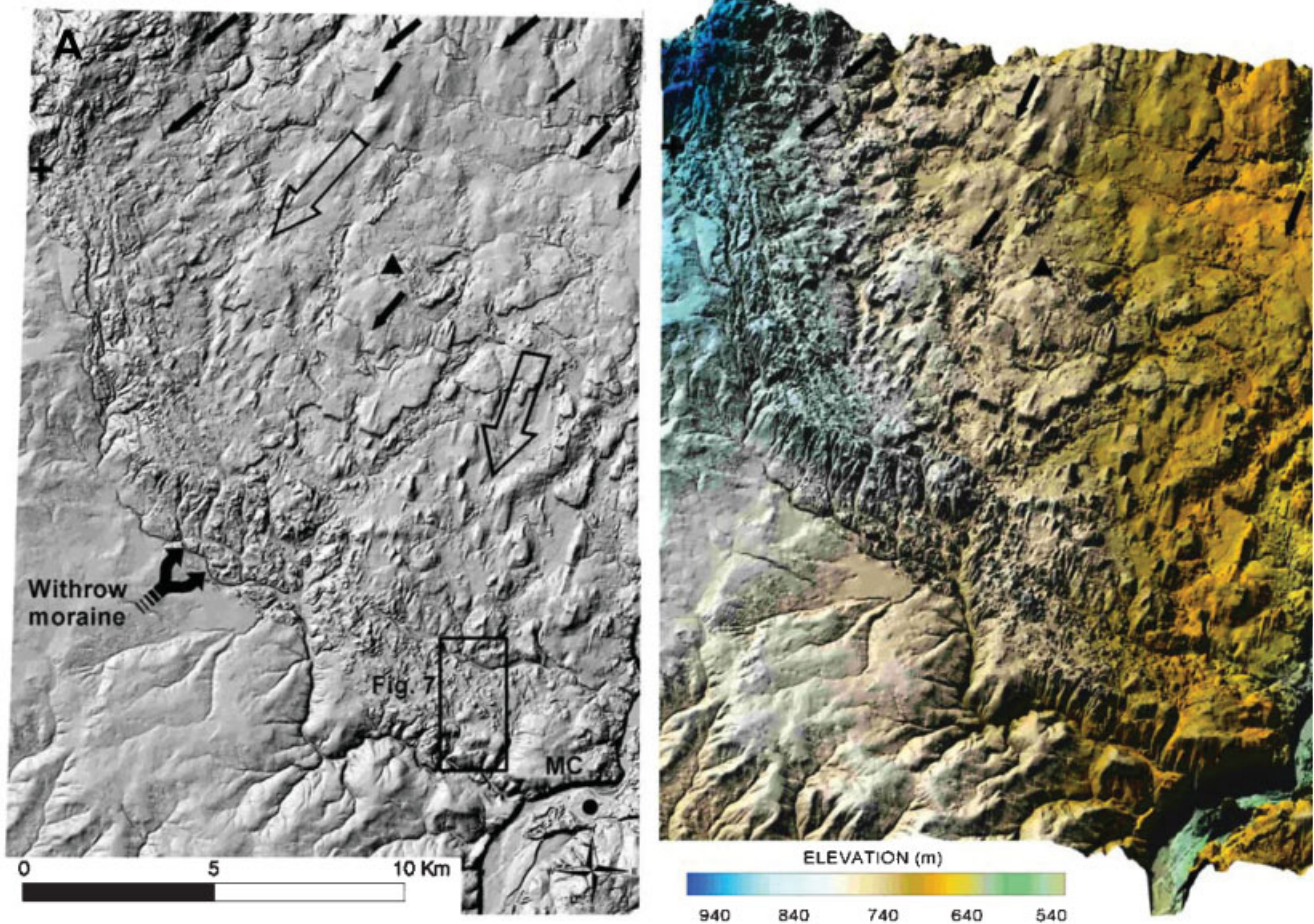
Figure 6   (A) Relationship between Withrow morainal complex, streamlined forms and recessional end-moraines along western portion of the Waterville Plateau (from USGS 1:24,000 Farmer, Withrow, Jameson Lake SW, SE, W, E, Mud, Mansfield, Piersol Hills Quadrangles). Left-hand panel is a three-dimensional perspective of the right-hand panel. Note the 'stacked' moraine ridges. Scale varies with view. ▲ is location of Fig. 9. (B) Digital topographic model showing relationship between Withrow morainal ridges, streamlined forms and eskers along eastern portion of the Waterville Plateau. Note that some of the asymmetric curvilinear forms in the Withrow moraine overlap (from USGS 1:24,000 Sims Corner, Barnes Butte, St Andrews, Mold, Park Lake, Coulee City Quadrangles). Open black arrows indicate inferred ice-flow direction; solid black arrows indicate recessional moraines; SC, Sims Corner; DF, Dry Falls; MC, Moses Coulee. Refer to Fig. 3 for location, + is location of Fig. 8
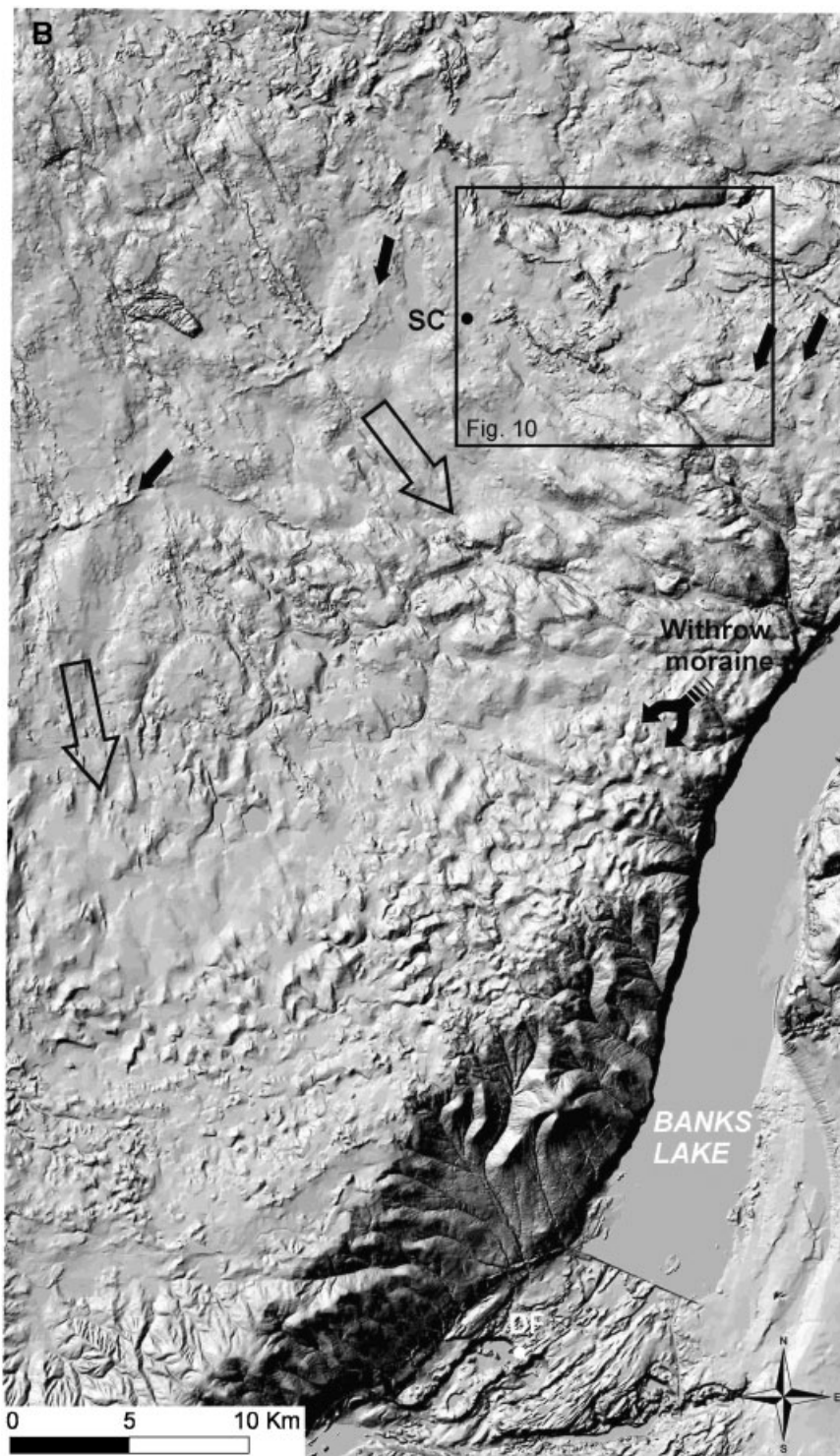
Figure 6 Continued

the deformation is likely related to compressive flow against the slope of the underlying bedrock.

Successive ice margins during the deglaciation are marked by many segmented moraines that are subparallel, and show a fan-shaped pattern (Fig. 4B). These moraines represent still-stands and possibly readvances (push moraines?; Kruger, 1996) of ice. Some can be traced almost continuously across the plateau and some small ice-marginal meltwater channels are apparent. A morainal segment may be up to ca. 60 m wide and rise 15–20 m above the local topography. Large erratic boulders, up to 15 m high, litter the surface of some of these moraine ridges. These ice margins have not been dated. Therefore, ice margin rates of change across the plateau are not known. Because the configuration of the marginal retreat looks

similar to that at the maximum extent, we believe that the ice-flow system (direction of flow-parallel features) is rather uniform and therefore the streamlined forms can be used for the reconstruction of the ice surface. However, flow lines of different age (younger) can be inferred from the ice-marginal position during the deglaciation and in general were likely to be more affected by the local topography (i.e. Okanogan dome to the north).

In contrast to the glaciofluvial sediments of the Withrow moraine, at many localities behind the Withrow moraine, road-cut exposures through hummocks reveal a hard, poorly sorted, matrix-supported diamicton that is generally massive (Dmm) and occasionally crudely stratified (Dml, Fig. 9) and clast rich. Clasts in the Dmm are often coated with a white
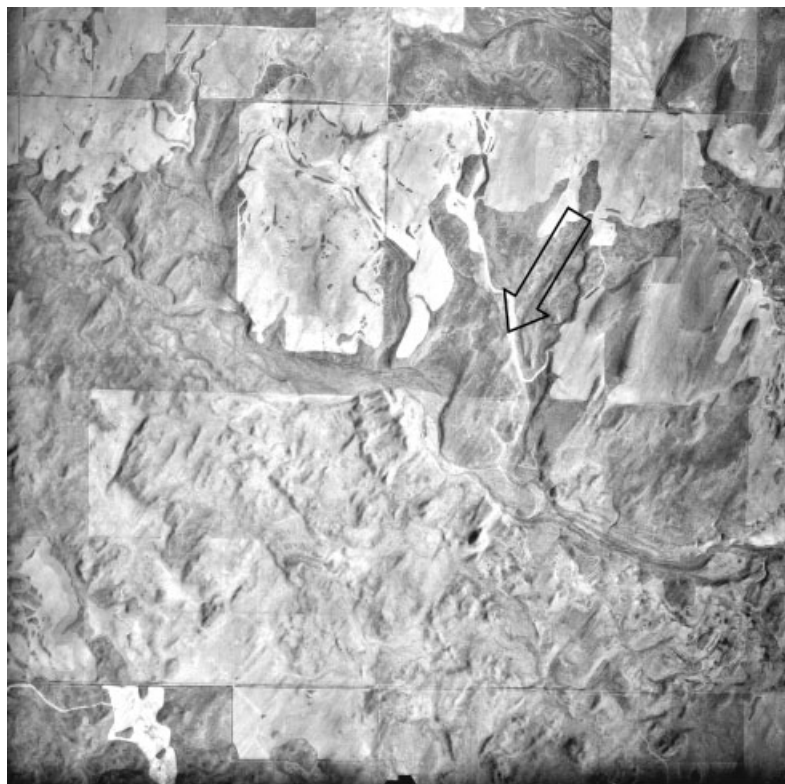
Figure 7 Aerial photograph of the complex Withrow moraine and associated features (photograph no. AAQ-3FF-202, U.S. Department of Agriculture, 1965). Refer to Fig. 6A for approximate location. Arrow indicates ice-flow direction

calcium-carbonate precipitate. At several places this Dmm has a concrete-like appearance and resembles viscous slurry. The striae on these clasts were highly varied. At another place, the Dmm/Dml had a granular structure, contains clasts with a moderate preferred orientation in the inferred ice-flow direction with slope conformable dip. The Dmm often reveals wavy streaks (thin friable laminae) of the alkaline cement and the base at one location is erosional. The variable orientation of clasts and slurry appearance suggests viscous deformation during periods of pore-water drainage. This Dmm/Dml could have been deposited following a period of meltwater discharge or could be the result of meltout or mass flow in a meltout sequence (e.g. Hicock and Dreimanis, 1992). This interpretation should be considered tentative if extended to similar features because of the small number of exposures available.

Eskers

Eskers are formed by running water in basal ice tunnels and record drainage beneath an ice sheet during deglaciation, when melting and down-wasting are greatest. Therefore, esker patterns can be used as a measure of bed transmissivity, glacial activity and basal meltwater production in a drainage network over an area. Eskers on the Waterville Plateau are well preserved, discrete, radiate toward the glacier margins, have tributaries as high as 3rd order, are up to 15 m in relative relief and 5 km long, and are often associated with outwash fans (Figs 6B and 10). Generally, two types of eskers were noted (i) those that are found associated with channels that are cut downward into bedrock and are spatially fixed (i.e. 'Nye Channels' or 'N channels'; Nye, 1976); and (ii) those that were formed when channels were melted upward into the glacier ice (i.e. 'Röthlisberger channels' or 'R channels'; Röthlisberger, 1972).

In the few exposures available for examination, the eskers consisted of crudely stratified, coarse gravel and sand. Well-sorted sandy gravel occurs near the top of some eskers. Some eskers east of Sims Corner (Fig. 6B) are covered with a matrix-supported diamicton. Several of the eskers drape over small drumlins and others are in between them. Many of the eskers have been deposited against the slope of the bedrock topography, indicating that the ice surface slope exerted stronger control on the hydraulic gradient. Well-developed, steep-fronted fans were formed at the mouth of several R-channel type eskers indicating the time-transgressive formation during ice-front retreat (Fig. 10). There appears to be a higher concentration of eskers along the east-central portion of the plateau. This concentration occurs where the bedrock topography is lower than the rim of the plateau.

As with the excellent preservation of the streamlined forms, the esker preservation suggests limited basal sliding at the time of deposition. Because many eskers are associated with transient ice-frontal positions, we suppose that initiation of esker deposition is associated with the deglaciation melt episode. An analogue for the preservation of the eskers may be that of the Keewatin sector of the Laurentide Ice Sheet (Aylsworth and Shilts, 1989; Brennand, 2000). There, the esker pattern is thought to reflect the widespread integrated surface drainage system, which was situated below the equilibrium line during deglaciation. The premise was that glacier retreat was achieved by wasting from the surface downward by fairly inactive ice.

Meltwater channels

Meltwater channels, most of them presently dry, are generally of two generations: (i) those that border the terminal morainal belt; and (ii) those that post-date the phase of drumlinisation
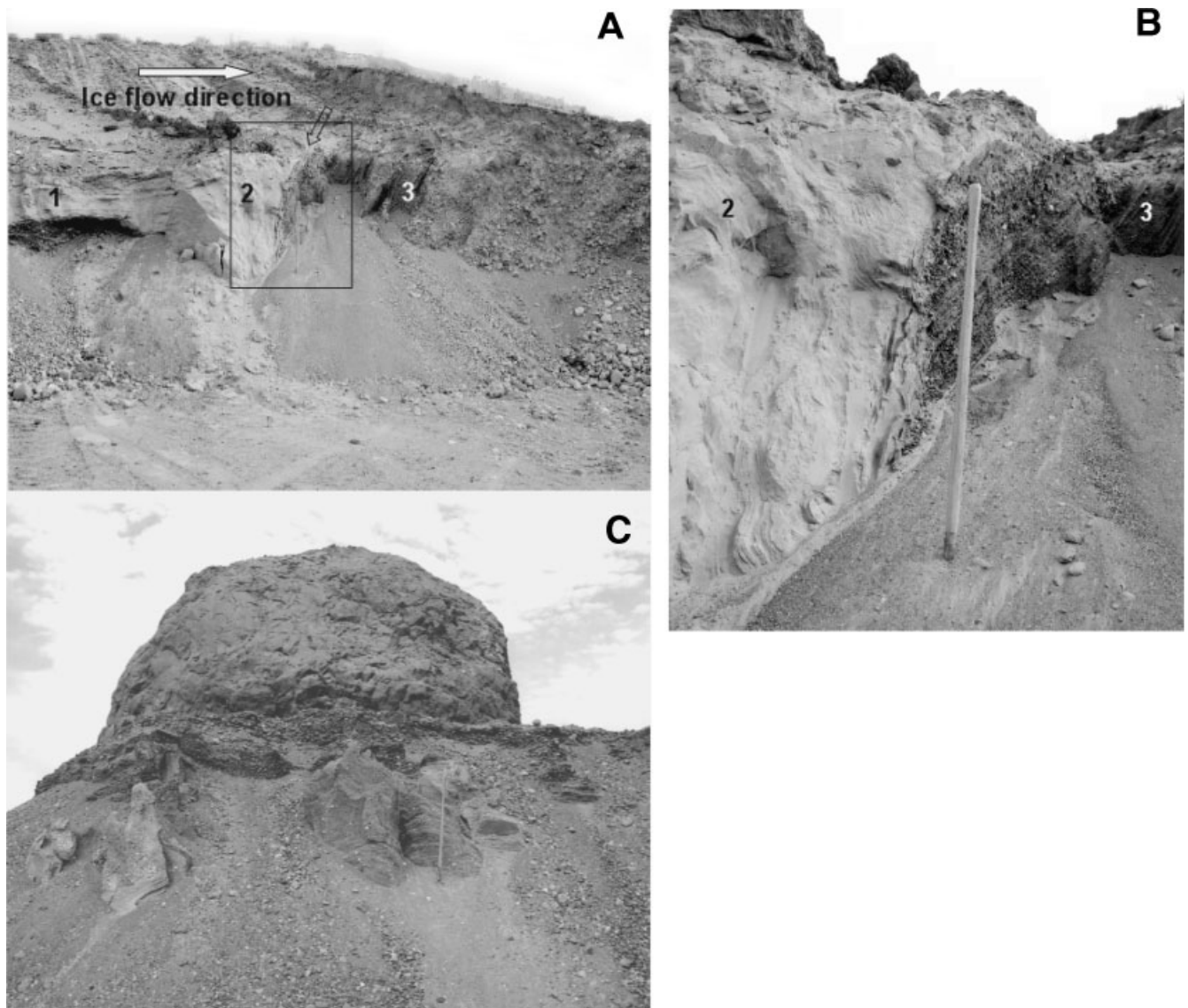
Figure 8 Sedimentary details of a reference section in the western margin of the Withrow moraine showing glaciotectonic structures. (A) Facies of the section: 1, the fine-grained sand and silt; 2, overfold in fine-grained sand and silt; 3, dipping coarse-grained gravel and sand with sand interlayers. Black arrow points to main displacement plane. (B) Facies exposed in box from (A). Refer to Fig. 6A for location. (C) Basalt erratic on the surface of the deformed sediments, adjacent to the location shown in (A). Shovel handle is 1.2 m long

and recessional moraine deposition, and appear to flow around individual features and therefore record late-stage drainage during deglaciation. The drainage of proglacial streams was channelled along the terminal margin and also around individual hummocks (Figs 6 and 7). The upper end of Moses Coulee appears to be buried by the glaciofluvial sediments of the complex Withrow moraine. The routeways formed during deglaciation are not especially extensive and the lack of an expansive sandur and/or pitted outwash surface is curious and suggests that meltwater was channelled away. The northern portion of the plateau is incised by multiple channels, which are roughly parallel to the ice-flow direction, indicating that meltwater was probably derived from the former glacier, rather than by drainage of glacial lake Columbia, the ancient Columbia River and/or Missoula flood events.

In the Okanogan trench, a series of local ice-marginal lakes, abandoned meltwater channels, prominent kame terraces and fjord-lake basins exist. They indicate complex meltwater routing events associated with retreat of ice from the valley (Nasmith, 1962; Fulton and Smith, 1978; Pine, 1985; Eyles *et al.*, 1990, 1991).

Other features

With the northerly retreating ice margin, an ice-marginal lake developed in Foster Coule (Fig. 2). A thick-bedded rhythmic fill of lacustrine clay and silt occurs up to an elevation of ca. 640 m (Hanson, 1970), giving a maximum lake depth of ca. 320 m. The water volume extracted from the 10-m gridded data is ca. 2800 km^3.

Published generalised sources (Richmond *et al.*, 1965; Waitt and Thorson, 1983) have referred to sublobes of the Okanogan Lobe as the Columbia River and Omak sublobes for drainages that were occupied by ice at the inferred maximum extent. No data have been presented for ice margins of these sublobes. Here, we have discerned a large moraine segment banked against the plateau along the Columbia River (Fig. 11). Based on the height of this morainal segment, ice was of sufficient thickness to flow westward (15 km) into the Lake Chelan basin to the west (e.g. Richmond, 1986). The interplay between the regional and bed topography, ice thickness and ice-surface height of the glacier on the Waterville Plateau and along the Columbia River is not known. But during deglaciation it is
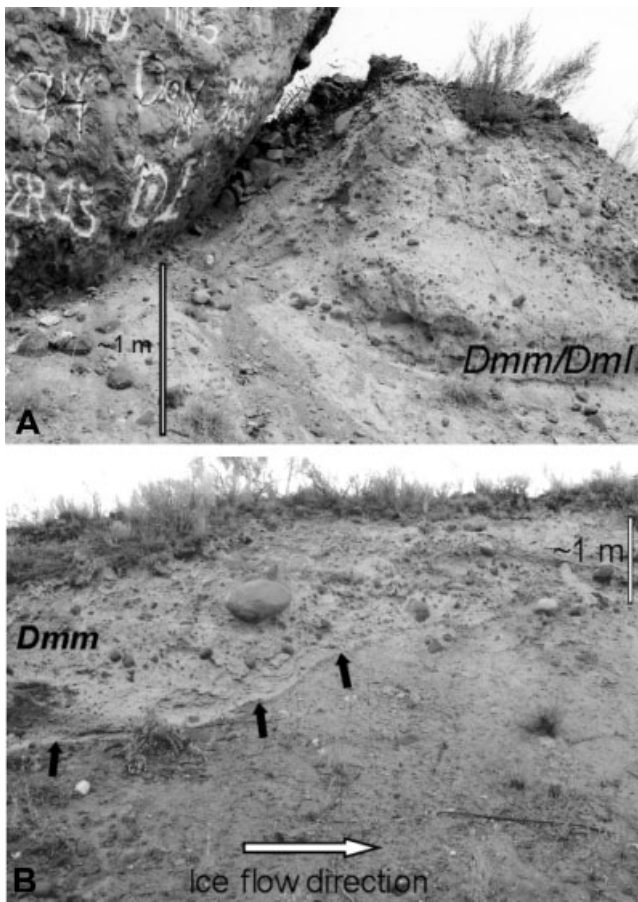
Figure 9 Photographs of matrix-supported diamicton behind the Withow moraine. (A) Basalt erratic resting on the massive to crudely stratified diamicton (Dmm/Dml) near Mansfield. (B) Roadcut exposure showing matrix-supported diamicton. The erosional contact is arrowed. Refer to Fig. 6A for location

probable that an interlobate area formed on the northwestern part of the Waterville Plateau as ice thinned. The ice-flow pattern reconstruction found in the next section shows a flow-line representing the Columbia River sublobe, although we do not know its maximum extent.

Ice-flow system and ice-surface reconstruction

Based on the streamlined forms and height of the Withow moraine the ice-flow system and ice surface are reconstructed. Moraine elevations were taken from USGS 1:24 000 topographic maps and the DEMs to constrain the elevation of the margin. We placed the northeastern margin along a kame terrace with confined marginal drainage. Because the retreating glacier margin exhibits a configuration similar to the Withow moraine, we assume that the streamline forms closely approximate the ice-flow system, even though we recognise that some of the features could have formed during deglaciation (Moores, 1989). The elevation of the ice is derived from highest elevations on the moraines and the form lines are then drawn normal to the ice-flow lines. This translates into the reconstructed configuration of the ice-surface at the maximum extent (Fig. 12) and this procedure is similar to that completed on the various lobes of the Laurentide Ice Sheet (Mathews, 1974; Clark, 1994).

Figure 13 shows the reconstructed ice-surface profiles in the terminal area. The highest moraine elevation along the western

margin was used to correspond to the minimum ice-surface elevation. The geometry of the ice lobe conforms well with the bedrock topography. Specifically, extending flow areas are observed where the bedrock is relatively low. A relatively broad shallow depression in the regional topography creates a concave topographic expression (Figs 2A and 13). Consequently, the ice would have been thickest here and the highest flow velocities would be expected here. Furthermore, if this is correct, owing to the overlying weight of the ice and depending on the availability of meltwater, the overburden pressure and potentiometric surface could have been relatively high in this area, and this is coincident with the distribution of eroded bedrock surface. Compressive flow is inferred along the western margin where the bedrock topography is highest.

Ice-surface contour of 1000 m is estimated from the height of marginal features along the valley sides and extrapolated to the Waterville Plateau. For comparison, the summary paper of Waitt and Thorson (1983; their fig. 3.1) is the only other reconstruction of the Okanogan Lobe and was a generalised image of the southern sector of the CIS in Washington, Idaho and Montana states. Their reconstruction does not show ice flowlines and the ice-surface contours are at 1000 m intervals. Details of their reconstruction were not presented and were not the focus of their paper, but we have used it as a basis for comparison below.

Many marginal lobes of the Laurentide and Fennoscandian ice sheets are thought to have moved fast under relatively low driving stresses and have low ice profiles (e.g. Mathews, 1974; Boulton, 1987, 1996; Clark, 1992; Hicock and Dreimanis, 1992). Their motion was thought to be accomplished by basal sliding and/or by deformation within the underlying material. Figure 13 shows a steep frontal edge, which is typical of ice advancing on a rigid bed. Steep fronts have also been documented where the margin could have been frozen but up-glacier the basal ice was at the pressure melting point (Clarke et al., 1984a; Kleman and Hätterstrand, 1999).

Driving stress at the maximum extent

Nye (1952) showed that ice-flow is driven by a shear stress acting down-glacier and is related primarily to the ice-surface slope. The first-order estimate of stress conditions distributed over the bed of the ice may be obtained from: $\tau = \rho g h(t) \sin \alpha$; where τ is the shear stress, ρ is constant ice density (900 kg m^{-3}); g is gravitational acceleration (9.81 m s^{-2}); $h(t)$ is the ice thickness at some time, here the maximum extent; and α is ice-surface slope (Paterson, 1994, chapter 11).

For an implied ice thickness of 225 m (Fig. 13), the normal stress would be ca. 1987 kPa. From the reconstructed ice-surface along a 50-km central flowline (Figs 3B and 13B) the minimum basal shear stresses are ca. 17–26 kPa (ice thickness = 225–425 m; ice-surface slope = 0.26–0.76°). This is comparable with the range of 20–25 kPa that has been reconstructed for the upper regions of some ice streams draining the West Antarctic ice sheet (Alley et al., 1987).

Discussion

In this section we limit our discussion to (i) the relation of the Okanogan Lobe reconstruction at its southernmost extent to adjacent work in the British Columbia Interior; (ii) inferred subglacial conditions; and (iii) glacial landsystem evolution.
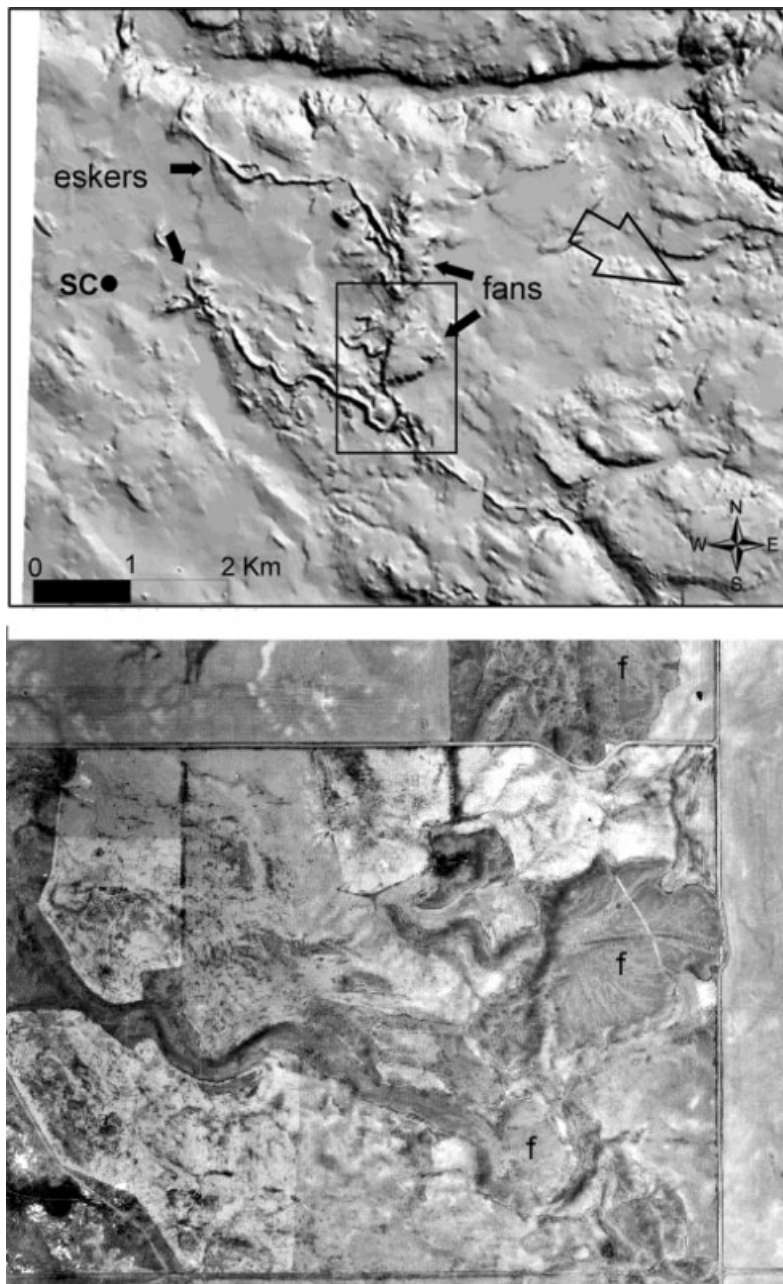
Figure 10   Digital topographic image and photograph showing eskers with fans near Sims Corner (SC). Open arrow is the inferred ice-flow direction; box in upper panel indicates area of photograph in lower panel; f indicates fans in photograph. Refer to Fig. 6B for location (photograph no. AAQ-2FF-43, U.S. Department of Agriculture, 1965)


Figure 11   Moraine segment of the Columbia River sublobe. Refer to Fig. 4B for location

Relation of the Okanogan Lobe reconstruction with adjacent work

At the climax of the Fraser Glaciation, outflow from ice centres on the western flank of the Columbia Mountains and Interior Plateau was channelled south through the Okanogan River Valley. At the southernmost extent the Okanogan Lobe flowed on to the Waterville Plateau creating a bottleneck pattern. The ice surface on the Waterville Plateau is now well constrained. The ice attained a minimum surface elevation of >700 m a.s.l. (minimum thickness of ~225 m). To the north along the Okanogan River Valley ice overtopped parts of the Okanogan Highlands (>1800 m a.s.l.). During deglaciation, climate amelioration triggered rapid downwasting and thinning of the CIS. Ice in the Okanogan River Valley dammed tributary drainages at various stages/heights when ice thinned below the mountainous topography (Nasmith, 1962; Fulton, 1969; Blais-Stevens, 2003; Lesemann and Brennand, 2003). As ice
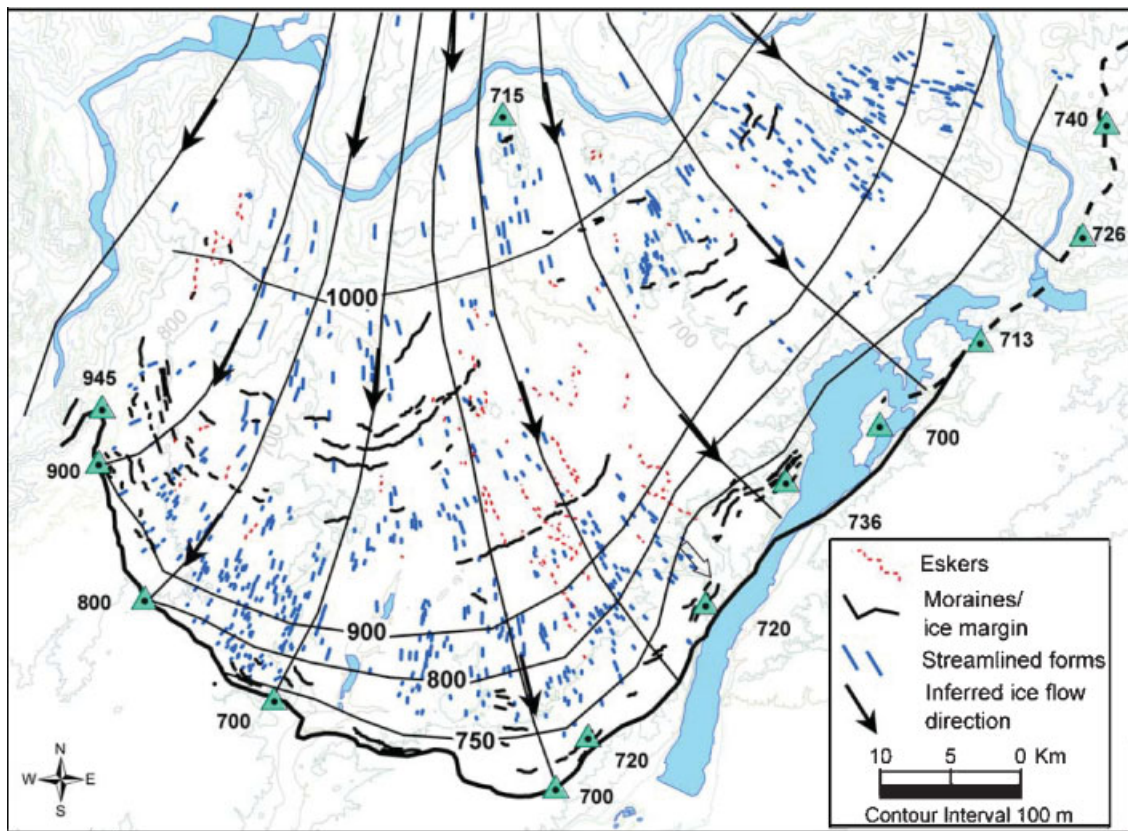
Figure 12 Configuration of the ice-flow system and ice-surface in the terminal area of the Okanogan Lobe

retreated these ice-dammed lakes and deep lake basins (600 m below modern sea-level) may have discharged catastrophically (e.g. Eyles *et al.*, 1990; Johnsen and Brennand, 2003). Given the regional topography, we speculate that similar lakes (and subglacial reservoirs) could have existed early in the glacial cycle and that these could have triggered a corridor of fast ice flow. The effect of fast flowing ice is to thin and lower the ice-sheet profile, possibly even below the equilibrium line. To this end, we reviewed other recent reconstructions of the CIS (Lian and Hicock, 2000, 2001; Stumpf *et al.*, 2000; McCuaig

and Roberts, 2002; Lian *et al.*, 2003) and found that none related directly to the flow system of the Okanogan Lobe. However, work in British Columbia on the deep 'fjord lake' basins of the Okanagan shows an extreme thickness of glacio-fluvial and glaciolacustrine sediments that were rapidly deposited during deglaciation (Shaw, 1977; Eyles *et al.*, 1990, 1991). These basins and the dynamics of the subglacial hydrological system would have affected the overall glaciodynamic behaviour of the Okanogan Lobe by possibly enhancing ice-flow rates.
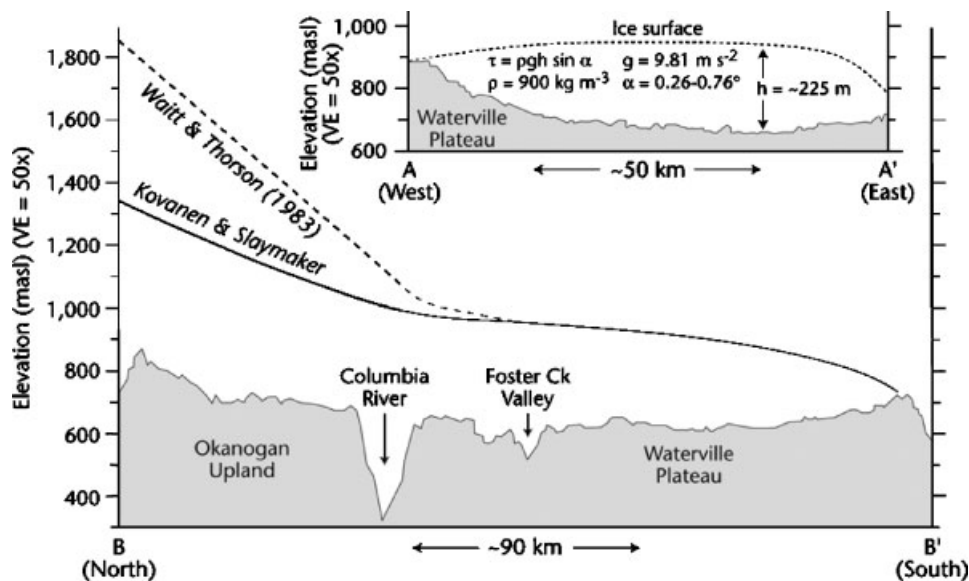
Figure 13 Reconstruction of the ice-surface profile in the terminal area of the Okanogan lobe at the maximum extent. (A) Cross-ice profile. Refer to Fig. 1C for profile locations. (B) Longitudinal profile

Inferred subglacial conditions

The distribution of water and the effect of water pressure at the glacier bed is an important factor in the control of temporal variations in ice flow. During the Fraser Glaciation, subglacial, ice-marginal and regional topographic conditions could have facilitated rapid flow of ice along the Okanogan River Valley. However, internal ice-sheet thermomechanical feedbacks and climate warming may also trigger fast ice flow. Whatever the flow mechanism, fast ice flow promotes increased meltwater production, which reduces basal friction, producing rapid glacier drawdown.

On the Waterville Plateau the glacial sediments are relatively thin and patchy overlying a basalt bed. Where subglacial drainage was poor, high water pressure probably developed and this could have further increased ice-flow velocities resulting in lower ice-surface profiles. The susceptibility of sediment to deform (internal yield strength) is also strongly influenced by the effective pressures. Low effective pressures inhibit shear deformation and this in turn influences the dynamic properties of the overlying ice sheet (Boulton and Hindmarsh, 1987). Based on the regional topography, our ice-surface reconstruction, the characteristics of the diamictons *behind* the Withrow moraine, the attenuated bedforms, and calculated low basal shear stresses, we believe that the glacier moved by a combination of basal processes and, at times, possibly by ice-bed separation rather than by internal deformation.

During deglaciation, however, basal processes were sluggish as indicated by the preservation of the large streamlined forms with moraines draped over them. The many recessional (push?) moraines over 40 km suggests active retreat rather than overall stagnation and that the Okanogan Lobe was a fairly stable system.

A regional perspective with regard to the extent of permafrost, its temporal evolution and relationship to the subglacial environment is still needed. It is possible that drainage of basal meltwater was blocked when the glacier margin was frozen, decreasing the effective pressure where drainage was poor or complex and subsequently the basal sliding increased. Future work includes consolidation tests on the basal diamictons to determine if the basal water pressures were near the ice-overburden pressure.

Glacial landsystem evolution

Earlier, we described the morphology, sediments and relative age of features associated with the Okanogan Lobe to serve as constraints for an evolving glacial landsystem model. The following depositional components are recognized: (i) the distinctive marginal push, thrusted and/or stacked glaciofluvial hummocks comprising a complex terminal moraine; (ii) subglacial landforms including drumlins and flutes; (iii) glaciofluvial forms, primarily eskers and ice-contact fans; and (iv) terraced glaciolacustrine sediments in incised bedrock channels. The conceptual glacial landsystem is shown in Fig. 14 and is similar to that based on the landsystem model of surging contemporary temperate Bruarjökull and Eyjabakkajökull in Iceland (Evans et al., 1999, their fig. 5; Evans and Twigg, 2002) where glacier flow has been enhanced by a deforming substrate (Boulton and Hindmarsh, 1987). Their model was based on specific landform–sediment associations and consisted of three zones: (i) an outer zone of proglacially thrust pre-surge sediments; (ii) a zone of weakly developed or patchy hummocky moraine deposited by supraglacial meltout and flowage of debris derived from the glacier bed; (iii) a zone of fluting produced by subsole deformation during the surge and crevasse-squeeze ridges. Eskers are draped over the crevasse-
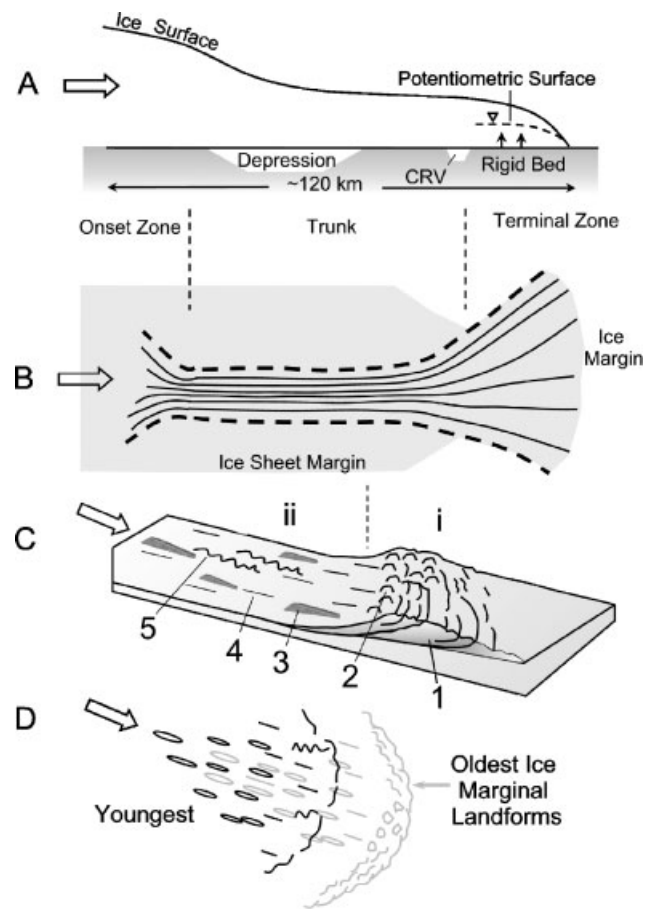
Figure 14 Schematic diagram showing the conceptual glacial landsystem model for the Okanogan Lobe. (A) Ice surface profile of simplified geometry of the ice-flow system. CRV, Columbia River Valley; significant depressions (Okangan and Penticten lakes basin) in the trough; rigid bed in the terminal zone that elevates the potentiometric surface. (B) Plan view of the ice system. A convergent onset zone and divergent terminal zone. (C) Glacial landsystem: (i) outer zone of proglacially thrust glaciofluvial sediments and juxtaposed hummocks; (ii) inner zone of flutes, eskers and drumlins. 1, complex thrust moraine and hummocky belts; 2, hummocky moraine; 3, drumlins; 4, flutes; 5, eskers (after Evans *et al.*, 1999). (D) The glacial imprint with an overprinted system that develops during deglaciation (after Clark and Stokes, chapter 9 in Evans, 2003)

squeeze ridges and fluting after meltout. There are some differences between our landsystem models, mainly that we have not observed crevasse-squeeze ridges or concertina eskers (e.g., Knudsen, 1995), that our bed is essentially rigid, and we lack a distinct zone of weakly developed or patchy hummocky moraine deposits. Additionally, our outer zone appears to be a combination of proglacially thrust pre-surge(?) sediments and hummocky moraine deposits. Based on our elongation ratios, the attenuated characteristics of the bedform, and the similarities between other surging glacier systems, fast ice flow and/or surge behaviour has been inferred (e.g. Boyce and Eyles, 1991; Stokes and Clark, 2002) during the advance stages. The convergent flow pattern leading to the trunk is characteristic of an ice stream, but additional work is necessary to characterise the margins of the flow system. No abrupt basal sliding boundaries were observed. This has been described in association with other ice sheets that experience ice-stream activity (e.g. Boulton and Clark, 1990; Kleman, 1994; Clark, 1997; Stokes and Clark, 1999) and have been linked to sharp lateral contrasts between bedform patterns that were created by discontinuity of basal thermal regime or ice velocity.

The glacial imprints on the Waterville Plateau and Okanogan trench provide evidence for dynamic links between the morphology, meltwater discharge and regional ice-sheet dynamics. On the Waterville Plateau the geomorphological evidence indicates that subglacial and ice-marginal conditions were favourable for rapid ice flow. It is not clear whether a significant event of fast ice flow or enhanced meltwater production occurred. Although, if a catastrophic episode of subglacial meltwater release did occur from an up-glacier reservoir, it could have been channelled down the Okanogan River and Columbia River valleys through a tunnel system. The only imprints on the Waterville Plateau are features inferred to represent fast basal flow velocities and the scoured bedrock. It is premature to speculate on the origin of the large drumlins in our field area, as we have not observed the sediment associations. The flutes, however, reflect moulding of the thin unlithified glacial substrate (e.g. Clark, 1999).

Conclusions

The Okanogan Lobe drained an area of ca. 4800 km^2 and was probably dependent on both topographic constraints and changes in the precipitation pattern, aridity and temperature depression feedbacks. The regional topographic conditions and relief are ideal for fast flowing ice and indeed, the glacial imprint is indicative of glaciation by a fast flowing glacier lobe or ice-stream. Based on the geomorphology and a comparison with other landsystems (Evans et al., 1999, their figs 3 and 7; Evans and Twigg, 2002), the former Okanogan Lobe may record surge behaviour. The continuous transition between flow traces suggests wet-based conditions. In order for the glacier to flow fast, upstream marginal and/or proglacial lake basins, and/or subglacial reservoirs probably provided sufficient lubrication at the base, which at times promoted instability.

The glacial geomorphology and reconstruction presented here may provide evidence for fast ice flow and inactive ice-marginal retreat over the Waterville Plateau. The location and vigour of fast flowing ice corridors are key to understanding the overall stability, mass balance and predictability of the CIS (e.g. Hicock and Fuller, 1995; Lian et al., 2003). Catastrophic outbursts are associated with the individual ice lobes along the southeastern margin of the CIS, but little is known about their dynamic behaviour and the relative consequence of the dominant regional controls (glaciodynamic or glacioclimate).

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