Late-glacial lakes in the Thompson Basin, British Columbia: paleogeography and evolution

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Abstract: During the decay of the Cordilleran Ice Sheet 10 000 – 13 000 BP, glacial lakes developed within valleys that dissect the Interior Plateau of British Columbia. In this paper, we (1) illustrate a procedure for assessing paleo water planes that has general application, (2) document lake paleogeography and evolution in the Thompson Valley, (3) provide new data on the glacio-isostatic response of the central Cordillera, and (4) present new evidence of its late-glacial environment. We employ geomorphology and sedimentology, digital elevation models, and new technologies (differential global positioning systems, ground penetrating radar, and geographic information systems) to refine paleogeographic reconstructions of glacial lakes. Glacial Lake Thompson and Glacial Lake Deadman were ribbon-shaped (width to length ratio ≈ 3:100), deep (>>140 to ~50 m) lakes that contained significant water volumes (84–24 km^3). They lengthened to the west and their water level lowered as ice decayed. Final ice dam failure resulted in an ~20 km^3 jökulhlaup that eroded bedforms and deposited flood eddy bars within the lake basin, travelled ~250 km along the Fraser River system, and may have deposited exotic mud offshore between 10 190 and 11 940 BP. Glacio-isostatic tilts of water planes are among the highest in the world (1.7–1.8 m km^{-1}). Their orientations suggest that ice sheet loads were greater or longer-lived to the north-northwest of the study area, lending support to the notion of an ice divide centred on the Fraser Plateau.

Résumé : Durant la décroissance de l’inlandsis de la Cordillère, 10 000 – 13 000 ans avant le présent, des lacs glaciaires se sont développés dans des vallées qui recoupent le plateau intérieur de la Colombie-Britannique. Dans le présent article, nous (1) décrivons une procédure d’application générale pour évaluer les paléonappes d’eau, (2) documentons la paléogéographie et l’évolution des lacs dans la vallée de la rivière Thompson, (3) fournissons de nouvelles données sur la réponse glacio-isostatique de la cordillère centrale et (4) présentons de nouvelles évidences de son environnement tardi-glaciaire. Nous utilisons la géomorphologie et la sédimentologie, des modèles numériques d’élévation et des nouvelles technologies (systèmes de positionnement global différentiels, géoradars et systèmes d’information géographiques) pour préciser les reconstructions paléogéographiques des lacs glaciaires. Les lacs glaciaires Thompson et Deadman avaient une forme rubanée (rapport largeur à longueur ~3:100), ils étaient profonds (>> 140 à ~50 m) et ils contenaient des volumes d’eau importants (84–24 km^3). Ils s’allongeaient vers l’ouest et leur niveau d’eau s’abaissait au fur et à mesure de la fonte de la glace. Une rupture finale de la digue de glace a causé un jökulhlaup ~20 km^3 qui a érodé les fonds rocheux et a déposé des barres de tourbillons de crue dans le bassin du lac; cette masse d’eau s’est déplacée sur ~250 km le long du système de la rivière Fraser et aurait pu avoir déposé des boues exotiques au large entre 10 190 et 11 940 ans avant le présent. Les basculements glacio-isostatiques des nappes d’eau sont parmi les plus élevés au monde (1,7–1,8 m km^{-1}). Leurs orientations suggèrent que les charges des feuillets glaciaires étaient plus grandes ou étaient de plus longue durée au nord-nord-ouest de la région étudiée, supportant la notion d’une ligne de partage glaciaire centrée sur le plateau Fraser.

[Traduit par la Rédaction]

Introduction

During the decay of the Cordilleran Ice Sheet 10 000 – 13 000 BP (Fulton 1969; Ryder et al. 1991; Clague and James 2002) many ribbon-shaped glacial lakes developed within valleys that dissect the Interior Plateau of British Columbia (Fig. 1; e.g., Fulton 1969). They acted as major sinks for sediments derived from plateau areas (e.g., Fulton 1965; Shaw 1977; Sawicki and Smith 1992; Eyles and Mullins 1997). Consequently, a rich geomorphic and sedimentary record remains from these lakes for study today. In this paper, we (1) illustrate a useful procedure for assessing paleo water planes that has general application, (2) document lake paleogeography and evolution in the Thompson Valley, (3) provide new data on the glacio-isostatic response of the central Cordillera (where such evidence is until now largely absent; Fulton and Walcott 1975), and (4) present new evidence of the late-glacial environment of the central

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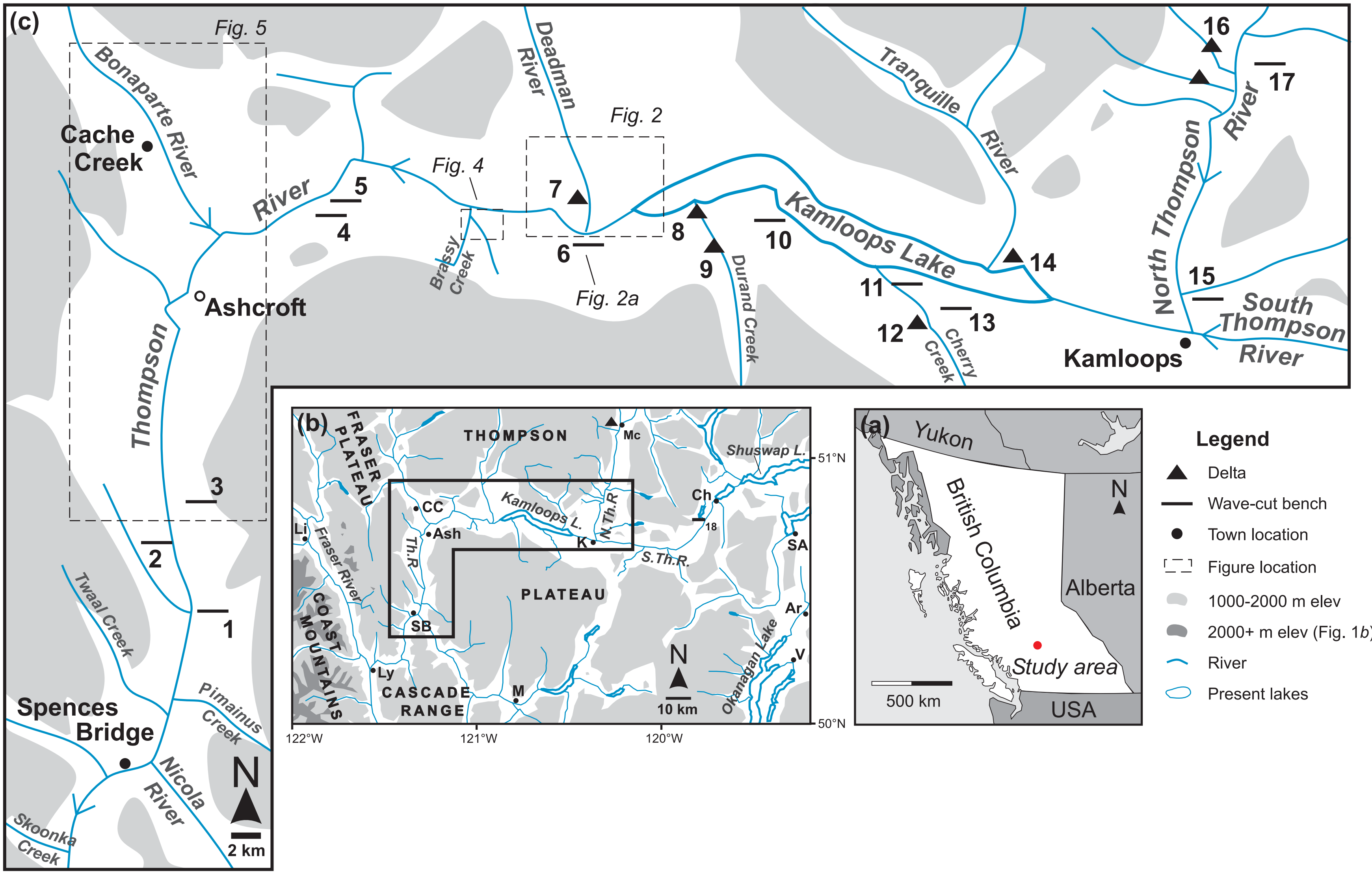
Fig. 1. (*a*) Location of study area in British Columbia, Canada. (*b*) The regional context of the study area showing location of wave-cut bench 18. (*c*) The study area within the Thompson Basin showing locations (and numbers) of wave-cut benches, deltas, and figures. Li, Lillooet; Ly, Lytton; SB, Spences Bridge; M, Merritt; Ash, Ashcroft; CC, Cache Creek; K, Kamloops; Mc, McClure; Ch, Chase; SA, Salmon Arm; Ar, Armstrong; V, Vernon. *Th.R*, Thompson River; *N.Th.R*, North Thompson River; *S.Th.R*, South Thompson River.



Cordillera. We employ traditional geomorphology and sedimentology, digital elevation models (DEMs), and new technologies (e.g., differential global positioning systems (DGPS), ground penetrating radar (GPR), and geographic information systems (GIS)) to improve upon earlier paleogeographic reconstructions of glacial lakes (Fulton 1969; Ryder 1976, 1981).

Study area and previous research

The study area is situated within the southern Interior Plateau of British Columbia and includes 220 km of valleys in the Thompson Basin (Fig. 1). Not included in this study, yet located within the Thompson Basin, are the Nicola River valley and the Merritt Basin (Fig. 1*b*), previously studied by Anderton (1970) and Fulton and Walcott (1975), respectively. Fluvial and glacial erosion have dissected the Thompson Plateau resulting in deep valleys (up to ~1500 m relief), today filled with thick (up to ~800 m) Quaternary deposits (Fulton 1965; Ryder 1976; Eyles and Mullins 1997). High (>100 m) cliffs of glaciolacustrine silt dominate the South Thompson and Thompson valleys and are by far the dominant Quaternary fill (Fulton and Armstrong 1965; Fulton and Smith 1978; Ryder 1976, 1981; Clague 2000; Clague and Evans 2003). West of Kamloops Lake to Spences Bridge, numerous fluvial terraces dominate the valley-sides and record up to 150 m of river incision.

Decay of the Cordilleran Ice Sheet in the interior of British Columbia was by regional stagnation rather than by active retreat (Fulton 1967, 1991). The ice sheet downwasted and backwasted toward ice divides, and dead ice masses stranded within the valleys dammed numerous glacial lakes (Fulton 1969). Lake basins lengthened coeval with the recession of the margins of these valley ice masses. Previous research presents a generalized picture of lake evolution for the Thompson Basin (Fulton 1969). Four stages of two eastward draining late-glacial lakes, Glacial Lake Thompson and Glacial Lake Deadman, are documented (Fulton 1969; Ryder 1976, 1981). Progressive westward lake extension and relocation and lake-level lowering characterize the lakes. This study refines our knowledge of lake paleogeography and evolution and extends our knowledge of the deglacial environment of the central Cordillera.

Stratigraphic relationships suggest that the lakes are late glacial in age (Clague 2000; Clague and Evans 2003). The absolute timing of lake initiation and duration are uncertain due to the rarity of datable organics. Deglaciation of plateau areas was likely well advanced by 12 000 BP (Fulton 1971; Clague 1981). A 9740–10 210 BP date from bog sediments in a spillway of Glacial Lake Shuswap near Armstrong (Dyck et al. 1965) provides the minimum age for all late-glacial lakes in the southern Interior (Fulton 1969). Lakes in the Thompson Basin may have been short-lived (~80 years), based on varve counting (Fulton 1965, 2000). They may have lasted as long as 540–1130 years, based on limited radiocarbon ages (Fulton 1969). This short duration for late-glacial ribbon lakes is consistent with findings elsewhere (e.g., Eyles et al. 1991; Mullins and Eyles 1996).

The only robust study of glacio-isostatic rebound in the Cordilleran interior was completed around Merritt (Fig. 1*b*; Fulton and Walcott 1975) where very high isostatic tilts (reliable values of 1.6–1.9 m km^{-1}) were documented. This study provides new data on the isostatic response of the central Cordillera.

Approach and methods

A five-step process was used for paleogeographic reconstruction: (1) identify and inventory all landforms that developed with respect to paleo water planes (Table 1), (2) determine the genesis of landforms using GPR, geomorphic, and sedimentologic data, (3) survey the position and elevation of landforms using a DGPS, (4) discover paleo water planes using three-dimensional (3-D) graphical assessment of water-plane indicator data and trend surface analysis (i.e., regression analysis; Fulton and Walcott 1975; Gray 1983; Burrough 1986), and (5) produce a map of lake stages and calculate lake parameters (e.g., area, volume) using DEMs, formulae describing paleo water-plane geometries, and geomorphic inference of lake extents and ice dams (Johnsen 2004).

In this study, planar (i.e., first-order) modelling of tilted water planes proved adequate. The regression procedure provides a planar surface described by

$$[1] \qquad Z = a + bX + cY$$

where *Z*, *X*, and *Y* are the elevation (CVD28 datum), easting and northing (UTM, Zone 10 coordinates, NAD83 datum, respectively, and *a*, *b*, and *c* are parameters determined by the least-squares procedure.

The upslope direction of the plane of best fit (*W*) in degrees azimuth from grid north is derived using

$$[2] \qquad W = \tan^{-1}(b/c)$$

and converted to true north by correcting for grid north (adding 1.75°). The slope of the plane (*S*) in mm^{-1} is derived using

$$[3] \qquad S = \tan^{-1}(b/\sin(W))$$

The error associated with *W* and *S* was also calculated.

Landform identification and genesis

Both primary (wave-cut benches, Gilbert-type deltas) and secondary (lake drainage bedforms, lake-bottom sediments, subaqueous deltas, paleo-channels, and river terraces) water-plane indicators were used for paleogeographic reconstruction (Table 1). The application of wave-cut benches and Gilbert-type deltas to the problem of reconstructing past lake stages is presented in the following two subsections; secondary water-plane indicators are used to constrain these reconstructions.

Wave-cut benches

Wave-cut benches develop in areas of adequate fetch by the action of waves eroding and transporting sediments in the nearshore (Fulton and Walcott 1975; Allan et al. 2002). They are identified on the basis of their appearance (Table 1, Fig. 2*a*), consistent geometry (Johnsen 2004), and regional correlation. In the study area, 16 wave-cut benches are used in regional water-plane reconstructions (Fig. 1; Johnsen 2004). All are located well above the elevation of the highest

Table 1. Paleo water-plane indicators.

Indicator[a]	Appearance	Application	Explanation and limitations
Primary indicators: approximation of paleo water plane			
Wave-cut benches	Distinct horizontal benches along hillsides (100–2000 m long); consistent bench and riser slope morphology (Johnsen 2004); crosscutting relationships with gullies (Fig. 2*a*)	Best approximation of water plane	Elevation records wave base; elevation may over-estimate average wave base if bench forms during high wind events (e.g., storms); Formation requires sufficient fetch, lake-level stability, and (or) time. The lowest slope-break of a bench was consistently used to estimate the paleo water plane as it (1) was the most consistently distinguishable morphological element, (2) agrees with modern environments (e.g., Allan et al. 2002), and (3) is an appropriate method for regional correlations (J. Teller, personal communication, 2001)
Gilbert-type deltas	Large (2–13 km^2) bodies of gravel, located where tributaries meet main valleys; generally fan-shaped and gently sloping from tributary mouth; often composed of multiple inset surfaces	Approximation of water plane	The topset–foreset contact is used for approximation; water plane may lie 1 to 4 m above this contact (Gustavson et al. 1975, Thorson 1989); elevation may be seasonally biased (lake levels may have been highly variable as even under the current regime the unmanaged level of Kamloops Lake varies up to 7 m (Pharo and Carmack 1979); delta levels may be produced by autogenic processes independent of lake-level changes (Muto and Steel 2004); upper surface is longitudinally sloped and can produce significant error; distal truncation of delta leads to over-estimation of water plane. The topset thickness was subtracted from the elevation of the delta surface to derive an estimate of lake level. The topset thickness was determined from either sedimentary exposures, GPR profiles, or estimated from other delta surfaces in the study
Secondary indicators: over-estimation or under-estimation of paleo water plane.			
Drainage bedforms	Large bedforms (>100 m wavelength, >2 m amplitude), steeper stoss slopes; GPR shows erosional origin; found on Deadman delta (within the lake basin; Fig. 3)	Under-estimation of water plane	Produced during final drainage of the last lake to occupy the valley; location influenced by confinement in flow (i.e., distal portion of sloped delta surface); associated with erosional surfaces on Deadman delta
Lake-bottom sediments	Large (50–100 m tall) cliffs of pale-yellow, fine-grained, waterlain sediment; often capped by imbricate gravel and mapped as river terraces (Ryder 1976, 1981)	Under-estimation of water plane; approximation of lake bottom	Elevation of lake-bottom sediments is an under-estimation of the water level of the lake in which the sediments were deposited; at their highest elevation, there is no capping gravel, and thus they record lacustrine benches that approximate the elevation of the former lake bottom. In this study, all large, continuous bodies of lake-bottom sediment were below the lowest paleo water plane
Subaqueous deltas	Large (tens of metres tall), wedge-shaped, body of medium- to coarse-grained, waterlain sediment; located where tributaries meet main valleys	Under-estimation of water plane	These underflow-dominated deltas have no topsets and were deposited subaqueously (Nemec 1990)
Paleo-channels	Perched, dry, incised, river channel; gravel lined	Over-estimation or under-estimation of water plane	Represents either former river flow (1) into a lake (over-estimation), or (2) post-lake fluvial incision (over- or under-estimation — see River terraces below)
River terraces	Broad, near-flat, imbricate gravel surfaces frequently capping cliffs of glaciolacustrine sediment	Over-estimation or under-estimation of water plane	After the lake drains, a fluvial system develops. The highest fluvial terraces remaining after incision record an under-estimate of a paleo lake level. However, if post-lake aggradation exceeded the paleo-lake level, this principle cannot be applied. In this study, all terraces were below the lowest paleo water plane. Related to this indicator are ice-marginal terraces that form from the deposits of lake-terminating ice marginal streams that run along valley-occupying ice tongues. Ice-marginal terraces over-estimate a paleo lake-water plane

[a]Indicators are listed in decreasing order of certainty for estimation of a paleo water plane. Water-plane indicators were identified from aerial photographs, a digital elevation model, previous research (Fulton 1965, 1967, 1969; Anderton 1970; Ryder 1970, 1976, 1981), and by visual field identification. The geomorphology, sedimentology, and GPR character of some landforms were examined closely to confirm genesis (see text for discussion).

Fig. 2. (*a*) Oblique view of wave-cut bench 6 (see Fig. 1 for location). (*b*) Perspective (5 × vertoca; exaggeration (VE)) amd (*c*) plan view hillshade DEMs of the Deadman River delta showing GPR survey locations (G1–G4; Figs. 2*e*, 3). Note multiple inset delta surfaces ((*d*), shaded areas 1, 2, 3a, 3b), delta dissection, kettle holes, and drainage bedforms (Fig. 3). Lake stage modeled in this study are projected across the DEM. Arrows in (*b*) indicate present river flow direction. (*e*) Portion of 350 m-long GPR profile G1 (location in (*c*) showing topset and foreset architecture of this Gilbert-type delta.

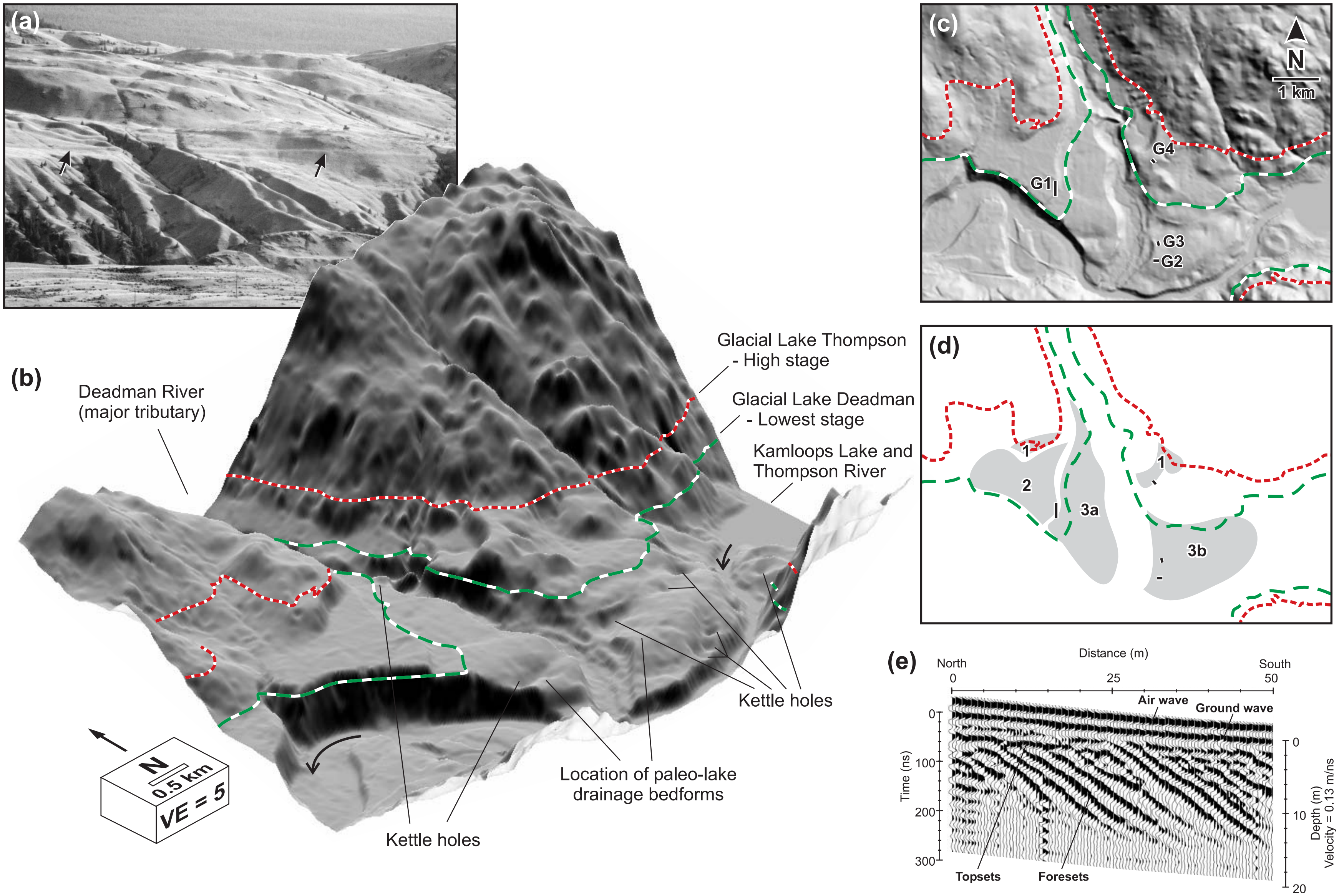
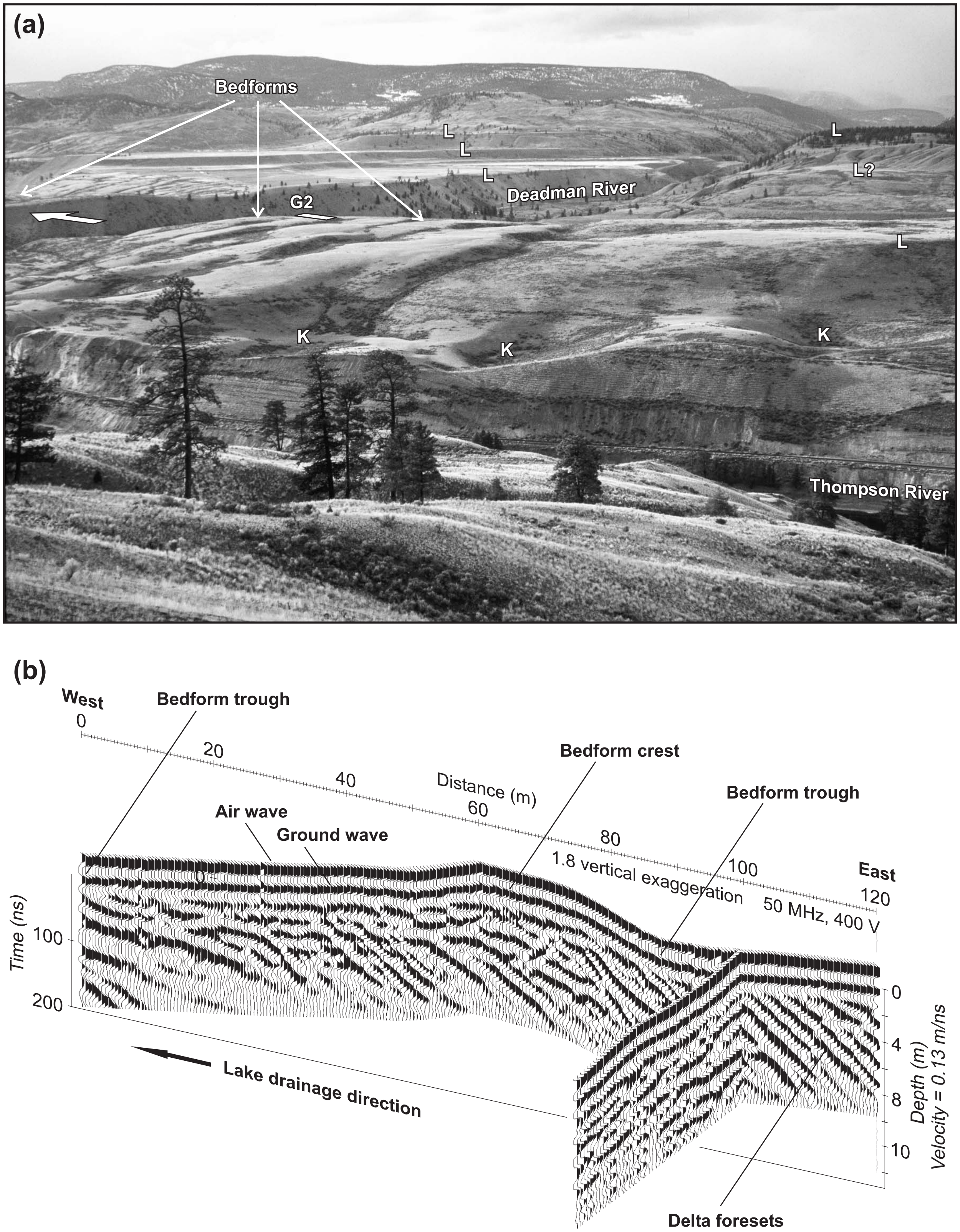


Fig. 3. Undulating terrain on Deadman delta is interpreted as a field of erosional bedforms (Pickering 1995; Massaria and D'Alessandrob 2000) produced during drainage of Glacial Lake Deadman - Lowest stage (see Fig. 2*b* for location). (*a*) View of bedforms looking northwest from wave-cut bench 6 (Fig. 1). Kettles, K; inset delta levels, L; G2, GPR profile location. Large arrow, final lake drainage direction. (*b*) GPR profile (G2) across an erosional bedform. An erosional origin is inferred because (1) inclined radar reflections (dip is 28° at 100 m on profile) extend well below the bedform and record delta foresets truncated at the land surface, and (2) topsets (subparallel, subhorizontal reflections) are largely absent.



fluvial terraces and appear to be cut into till-covered slopes, similar to those near Merritt (Fulton and Walcott 1975).

Gilbert-type deltas

Gilbert-type deltas developed where rivers delivered large quantities of sediment to the lake and aggradation of these sediments reached the elevation of a lake water level (Table 1, Fig. 2). They have upper surfaces that gently slope into the main valleys and often contain several inset surfaces (Fig. 2). Post-lake incision has truncated and deeply dissected these landforms. The location and elevation of the topset–foreset contact of 12 delta surfaces (Fig. 1, Table 1; Johnsen 2004) are used to reconstruct paleo water planes in the study area. They are located between McClure and the Deadman – Thompson River confluence (Fig. 1).

Delta identification is confirmed by geomorphology (Table 1), sedimentology, and (or) GPR surveys; one example is presented. The largest (13 km^2) Gilbert-type delta in the region is the Deadman delta (7, Figs. 1–3; Fulton 1987). Three dominant delta surfaces (~540, 480, and 450 m above sea level (asl) topset–foreset contact, Figs. 2*d*, 3*a*) and several intermediate delta surfaces are visible. Sediment exposures show inclined beds of sand and imbricate gravel (foresets) underlying horizontally stratified imbricate gravel (topsets). Both GPR profiles (Figs. 2*e*, 3*b*; Johnsen 2004) and sediment exposures confirm the origin of this landform as a Gilbert-type delta. Steady or episodic lake-level lowering initiated delta dissection and formed nested delta surfaces (Muto and Steel 2004).

Reinterpreted water-plane indicators

The correct identification of water-plane indicators is critical; incorrect interpretations of individual landforms can corrupt paleogeographic reconstructions. This investigation has resulted in a reinterpretation of several landforms previously studied in the Thompson Basin (Ryder 1976, 1981): Brassy Creek terrace (Fig. 4), Brassy Creek bench, and the paleo-Bonaparte River system (Fig. 5).

Brassy Creek terrace

West of the Deadman – Thompson River confluence is a prominent terrace, referred to here as Brassy Creek terrace, that has previously been interpreted as a delta that prograded from Brassy Creek during the Glacial Lake Deadman - Tranquille stage (Figs. 1, 4; Ryder 1976). A reexamination of the landform based on aerial photographs, GPR surveys, and field observations indicates that it is likely a fluvial terrace.

Brassy Creek terrace (Fig. 4) is re-interpreted as a fluvial terrace because (1) the upper surface of the landform is not sloped down toward the valley axis like other deltas or alluvial fans in the region, but is near-flat like fluvial terraces (at 420 m asl), (2) the gullies above the landform are too small to have transported the cobble-sized gravel that compose the landform (Fig. 4), (3) there is no source of gravel within the catchment area of these gullies, (4) nearby glaciolacustrine sediments are close to the elevation of the landform, which indicates that the basin was too shallow for a delta of this thickness (~60 m) to develop, and (5) fluvial terraces nearby are at a similar elevation. Local valley side and bedrock topography (Fig. 4) facilitated its preservation during Holocene incision.

In addition, GPR surveys indicate that this landform is not a delta. Two radar facies are separated by an angular unconformity (Fig. 4). The upper facies (1) is composed of near-horizontal continuous reflections onlapping from the north (Fig. 4*c*) and northeast (Fig. 4*b*), i.e., accreting onto the valleyward sloping upper boundary of radar facies 2. An exposure of imbricate gravel correlative with the position of radar facies 1 indicates westward paleoflow, consistent with a floodplain gravel interpretation for radar facies 1. Onlapping argues against delta topsets building to the north and northeast (Ryder 1976) and is consistent with a Thompson River terrace interpretation.

Radar facies 2 is correlative with cobbles that are exposed along the terrace riser (Fig. 4). Its upper boundary dips gently down toward the valley axis and is an unconformity. Internal reflections are variable — they dip gently toward the valley axis (Fig. 4*b*) and are concave and inclined toward the valley wall (Fig. 4*c*). Unlike all other fluvial terraces observed in the study area that are underlain by lacustrine sediments, this terrace overlies cobble gravel. Sediments of radar facies 2 may have been created by preferential deposition of sediments that were eroded from the nearby Deadman delta (~ 5 km to the east) during catastrophic westward drainage of the last lake to occupy the valley (discussed later in the text; Johnsen 2004). This is supported by (1) the proximity of this landform to Deadman delta (~5 km), (2) the position of this landform east of a bedrock protuberance, which may have aided in the preferential deposition of sediments transported along the valley to the west (Fig. 4; this is the first major bedrock protuberance found along the valley west of Deadman delta), (3) the wedge-like shape of the deposit with its highest end adjacent to the bedrock protuberance, (4) the lower elevation of this landform compared with eroded portions of Deadman delta (Fig. 3), (5) the similarity in grain size (cobble-sized) of both the eroded topsets of the Deadman delta and the sediments of radar facies 2, and (6) the possibility that taken together radar reflections of facies 2 may represent gravel bedforms or barforms (Beres et al. 1999) tentatively interpreted as flood eddy deposits (cf. Baker 1973). The unconformity between radar facies 1 and 2 may have been created by river erosion proceeding aggradation at this location or may be inherent in the creation of flood eddy bars.

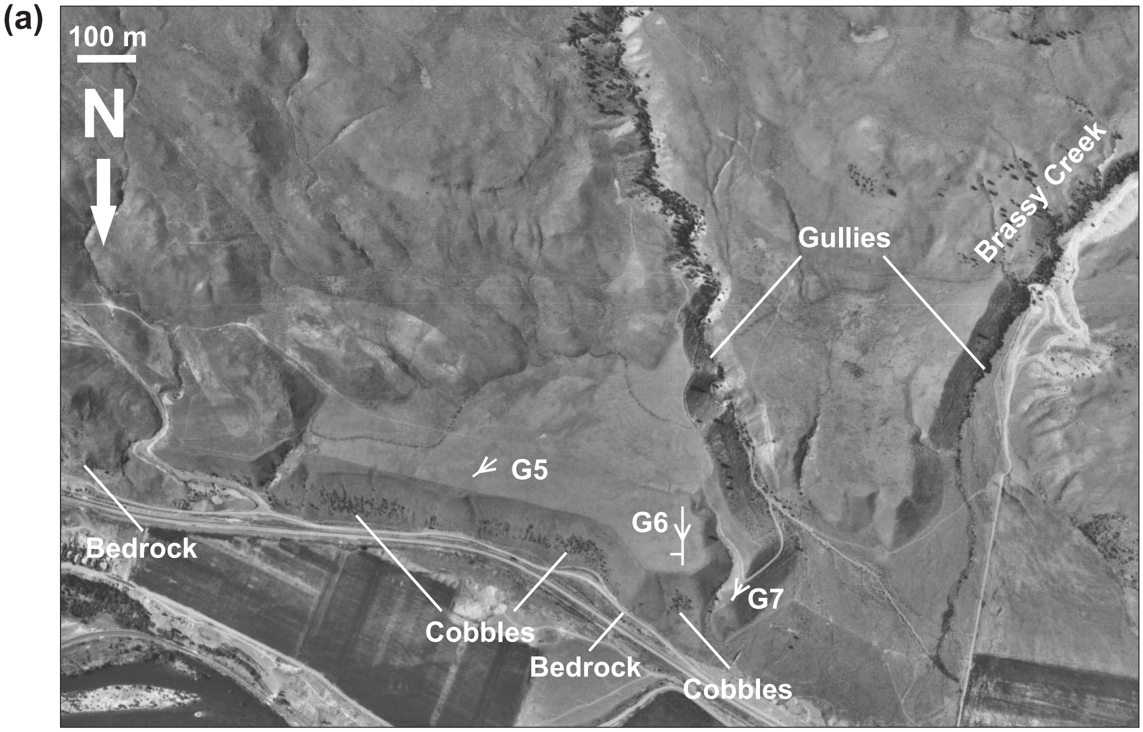
Brassy Creek bench

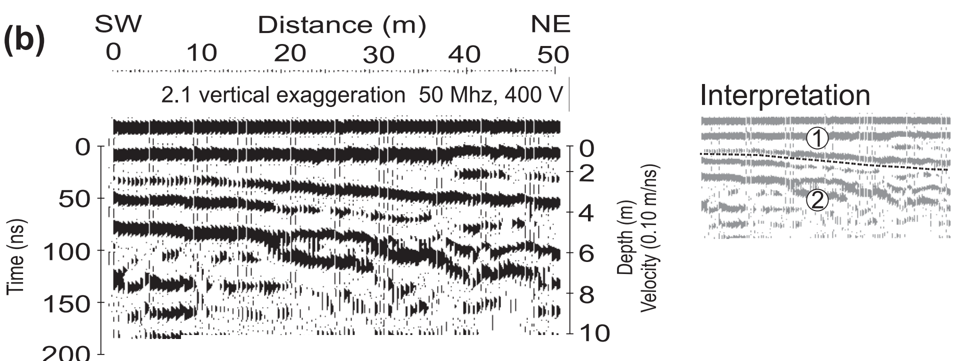
Brassy Creek bench is a subtle, laterally discontinuous bench located 1 km east of Brassy Creek terrace and at the same elevation (Fig. 1). Ryder (1976) interprets it as a wave-cut bench recording the Tranquille stage of Glacial Lake Deadman. We interpret Brassy Creek bench as a fluvial terrace remnant because (1) it is at the same elevation as the Brassy Creek terrace to the east and a dominant fluvial terrace to the west, and (2) the bench occurs and is broader where it intersects gullies, unlike other wave cut benches in the region that do not occur in gullies (cf. Fig. 2*a*). The bench is a product of gully infilling during aggradation of the Thompson River followed by incision and preservation.

Paleo-Bonaparte River system

Gravel benches between three valley-parallel hills west of Ashcroft (EH, CH, RH, Fig. 5) have been interpreted as pos-

Fig. 4. Brassy Creek river terrace (Fig. 1). (*a*) Geomorphic context and GPR survey locations (Aerial photograph BC86038: 146, 1986, British Columbia Government). GPR profiles (*b*) G5 and (*c*) G6. Radar facies 1 and 2 are interpreted as river terrace gravel and flood eddy deposits, respectively. See text for discussion.





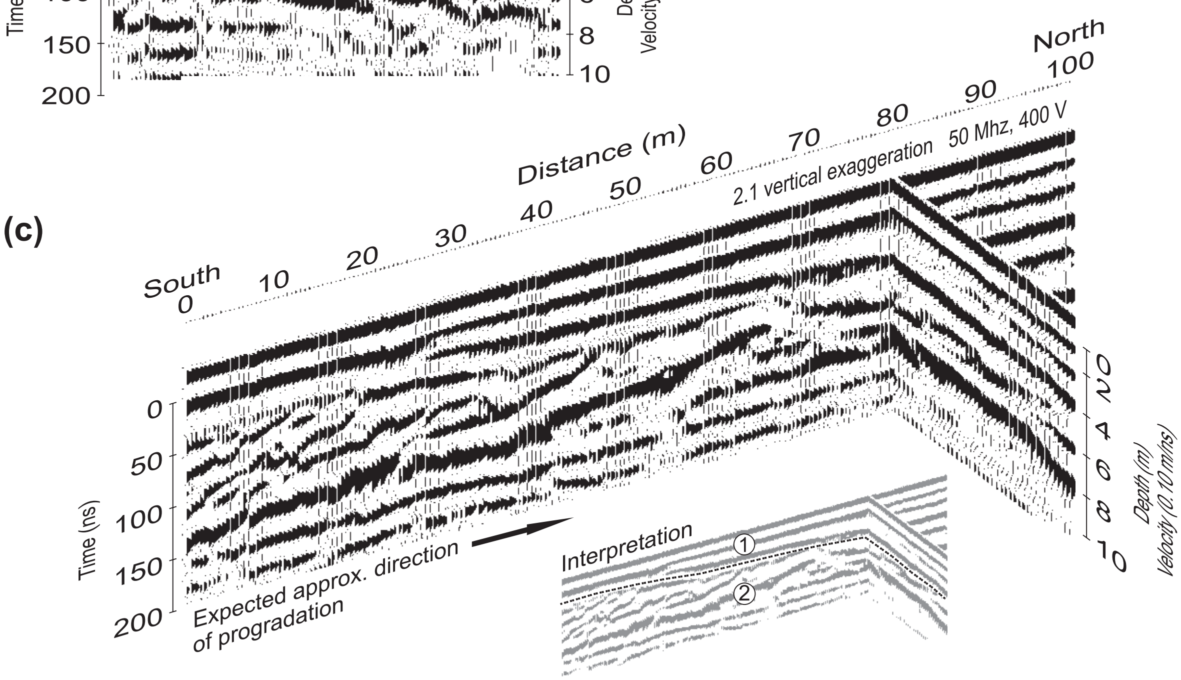
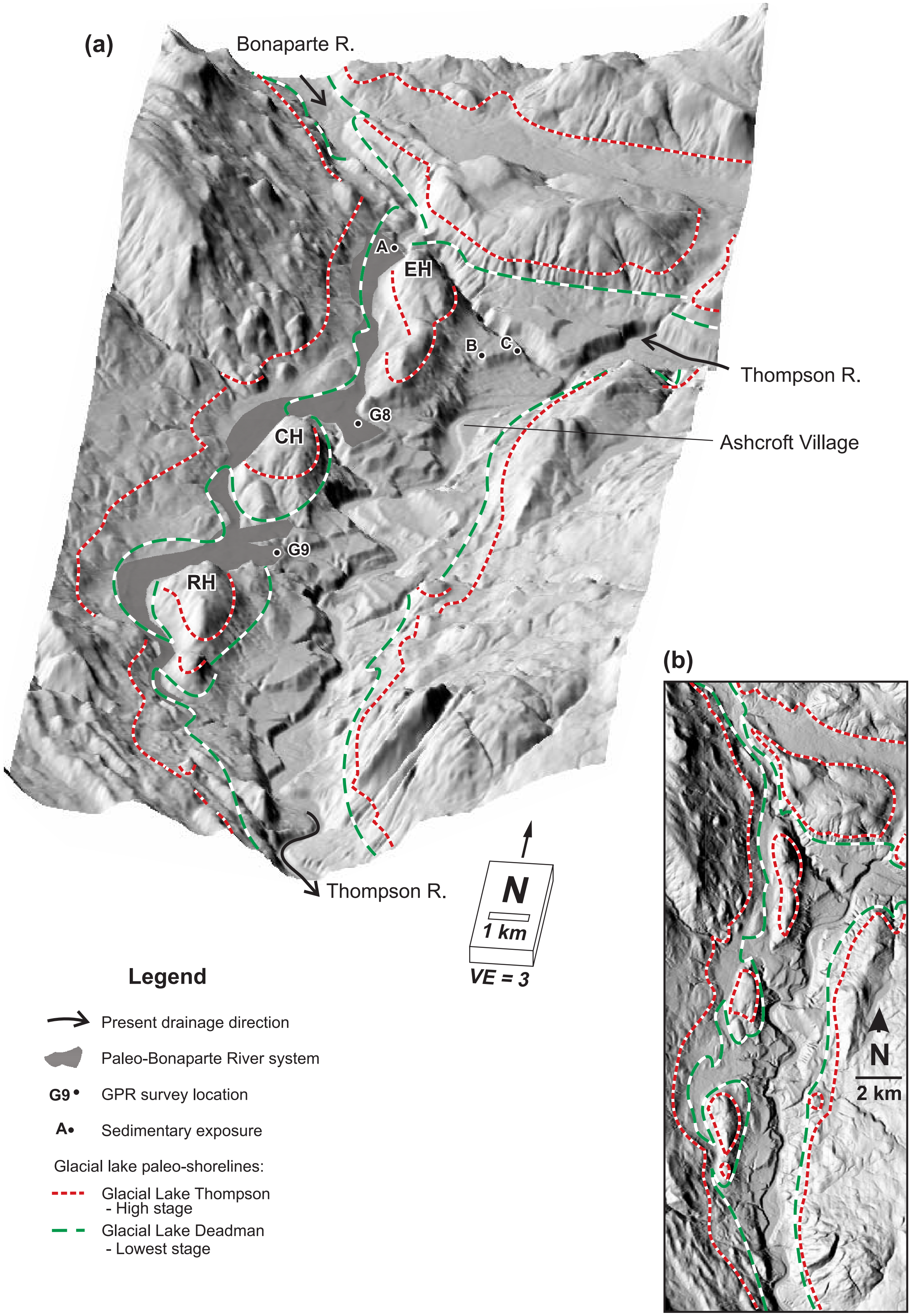


Fig. 5. (*a*) Oblique and (*b*) plan view hillshade DEMs of the Paleo-Bonaparte River system near Ashcroft (see Fig. 1 for location) comprising paleo-channels and high terraces adjacent to Elephant Hill (EH), Coyote Hill (CH), and Red Hill (RH). Sedimentary exposures, GPR survey locations, and paleo water-plane projections from this study are also shown. See text for explanation.



sible delta surfaces because they are at a similar elevation to Fulton's (1969) projected Glacial Lake Deadman - Tranquille stage water plane (Ryder 1976). Leading to these benches are small, abandoned, gravel-lined river channels (i.e., paleo-channels; A, Fig. 5); together, the benches and paleo-channels compose the paleo-Bonaparte River system (Fig. 5). We interpret the benches as shared high fluvial terraces of the paleo-Bonaparte and paleo-Thompson River systems because (1) the benches are not sloped like other delta surfaces in the study area but are near-flat like fluvial terraces, (2) the benches are at a similar elevation to river terraces composed of imbricate, horizontally stratified gravel (B, C, Fig. 5) at the mouth of Bonaparte River, (3) GPR surveys of limited length and penetration (Johnsen 2004) did not find dipping reflections characteristic of Gilbert-type deltas (G8, G9, Fig. 5), (4) the benches lie below the lake levels reconstructed in this study, (5) the paleo-channels indicate that the Bonaparte River once had three alternative routes adjacent to and between the three valley-parallel hills, and thus, a fluvial aggradation origin for the between-hill terraces is compatible with multiple routes of the paleo-Bonaparte River, and (6) this area is favourable for significant post-lake fluvial aggradation as the Thompson Valley narrows to the south and the Bonaparte River tributary enters the valley from the north. In addition, (7) the benches did not form as kame terraces, as (*i*) the paleo-Bonaparte surface is not pitted, (*ii*) lithofacies in the adjacent Thompson Valley do not record ice collapse, and (*iii*) the presence of an ice mass in this area of the Thompson Valley is inconsistent with late-glacial lake reconstructions presented here. These fluvial terraces are the highest in the area and have thicker gravel caps than lower fluvial terraces, as they represent the maxima of fluvial aggradation (~455 m asl).

Lake reconstruction

We reconstruct the paleogeography of deglacial lakes by correlating wave-cut benches and Gilbert-type deltas, superimposing water planes onto a land surface DEM and documenting dams and outlets in the geomorphic and sedimentary record. Lake statistics are derived from analysis of DEMs (paleo-water surface elevations, land surface elevation, and elevation of lake-bottom sediments).

Correlation of water-plane indicators

Wave-cut benches and deltas define two distinct lake-levels in the study area: (1) the upper water plane (Fig. 6*a*) and (2) the lower water plane (Fig. 6*b*).

The upper water plane (Fig. 6*a*) is defined by a tight grouping of wave-cut benches along a first-order surface with an upslope bearing of 332° ± 7.7° and slope of 1.8 ± 0.6 m km^{-1} (number of observations (n) = 8, coefficient of determination (r^2) = 0.97, at 99.99% confidence; error of bearing and slope at two times the standard error; Fig. 6*a*). The standard deviation of the residuals is 4.2 m over an ~ 45 km baseline (Fig. 6*c*). The excellent fit of these water plane indicators along the first-order surface confirms that a higher order trend surface is not warranted. Elevations of this water plane along the Thompson Valley are shown in Fig. 7. The upper water plane is described by

[4] $$Z = -7528.9 - 0.00087855 \cdot X + 0.0015381 \cdot Y$$

The upper water plane (Fig. 6*a*) is best defined by completely excluding delta surfaces and only correlating wave-cut benches. This is expected given seasonal and inferred water-plane elevation biases in the formation of deltas (Table 1). Other researchers have also found that deltas are unreliable for accurate water-plane reconstructions compared with wave-cut benches (Rayburn 1997; R.J. Fulton, personal communication, 2001).

The lower water plane is defined by a first-order surface with an upslope bearing of 321° ± 3.8° and slope of 1.7 ± 0.2 m km^{-1} (n = 6, r^2 = 0.99, at 99.85% confidence; error of bearing and slope at two times the standard error, Fig. 6*b*). The standard deviation of the residuals is 2.4 m over an ~30 km baseline. Elevations of this water plane along the Thompson Valley are shown in Fig. 7. The lower water plane is defined by

[5] $$Z = -5923.59 - 0.0011026 \cdot X + 0.001261 \cdot Y$$

Three wave-cut benches and three delta surfaces were used in this correlation. As deltas are less reliable indicators of lake level (Rayburn 1997; R.J. Fulton, personal communication, 2001; Muto and Steel 2004), less confidence is given to the accuracy of this correlation than for the upper water plane. Elevations of this water plane along the Thompson Valley are shown in Fig. 7.

Compared with the upper water plane, the lower water plane has a similar orientation and a lesser slope (Fig. 6). A decrease in slope is expected because the younger, lower water plane would have experienced less differential rebound than the older, upper water plane (Walcott 1970). The strength of the water-plane correlations as records of former lake levels rests on (1) their statistical significance and (2) the similarity of their orientations and slopes to those reported from the nearby Merritt Basin (upslope direction of 350° and slopes of 1.6–1.9 m km^{-1}, Fulton and Walcott 1975; Fig. 1*b*).

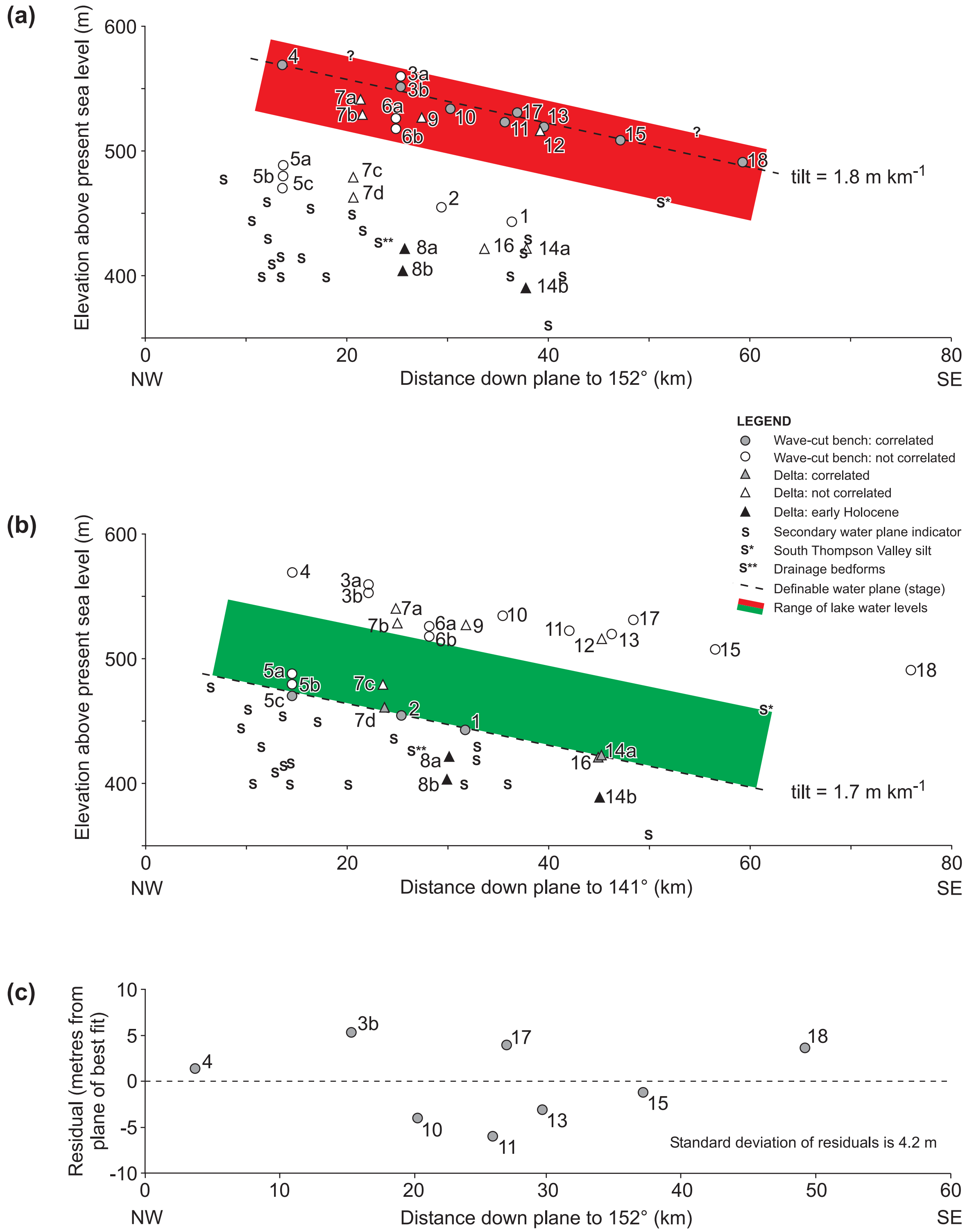
Lake and stage naming

Lake water levels that can be defined by correlating water-plane indicators are termed lake stages (e.g., Fulton 1969). A new lake name is used when (1) the lakes are in separate locations or (2) the lake outlet position changed significantly (Fulton 1969).

Projections of the upper and lower water planes defined in our research confirm the existence of two late-glacial lakes with separate outlets (Fulton 1969): (1) Glacial Lake Thompson (Figs. 6*a*, 8) had its outlet near Chase, and (2) Glacial Lake Deadman (Figs. 6*b*, 8) had its outlet through high elevation lake-bottom sediments east of Kamloops (Fig. 7).

Each lake exhibited a range of water levels (Fig. 6). The relative elevation of the water planes within these ranges directs our naming of stages. The upper water plane defines a High stage of Glacial Lake Thompson, and the lower water plane defines the Lowest stage of Glacial Lake Deadman in the study area (Figs. 6, 8). The elevation of the South Thompson Valley silts (s*, Fig. 6) marks the lowest elevation of possible water levels for Glacial Lake Thompson and the highest elevation of possible water levels for Glacial Lake Deadman (Fulton 1969). As the orientation and tilt of our water planes differ from those proposed by Fulton (1969; based on the three-point geometric method) for the Thompson Basin and our reconstructions correlate more water-plane in-

Fig. 6. Best-fit (first-order trend surface) glacial lake water-plane projections for (*a*) an upper water plane (Glacial Lake Thompson - High stage) and (*b*) the lowest water plane (Glacial Lake Deadman - Lowest stage) in the Thompson Basin. The shaded bars indicate the range of water levels for each lake. See text for discussion. (*c*) Plot of residuals from modeling results for the upper water plane indicates that a first-order trend surface is adequate. Water-plane indicators located in Fig. 1.



dicators, we do not apply Fulton's stage names to our results.

Some wave-cut benches and Gilbert-type deltas are located above, between, and below defined paleo water planes. Only indicator 3a and isolated deposits of lake-bottom sediments (not plotted) are located above Glacial Lake Thompson - High stage. These features indicate that higher water levels of Glacial Lake Thompson existed (Figs. 6*a*, 8*a*). The presence

Fig. 7. Down-valley cross-section of lakes (stages) from Spences Bridge to Chase (Fig. 1). Changes in slopes to the water plane result from changes in valley orientation. The North Thompson Valley is not shown. Approximately 20.0 km^3 of water (Table 2) in the lake forming the lowest stage of Glacial Lake Deadman drained catastrophically once the ice dam south of Spences Bridge failed (bold arrow). An additional ~3.6 km^3 of water dammed behind Deadman delta (a) was likely not part of this lake drainage.

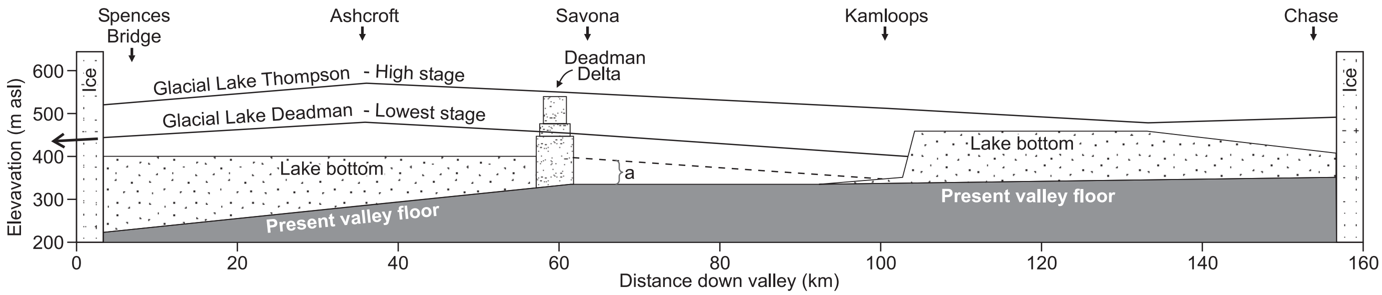
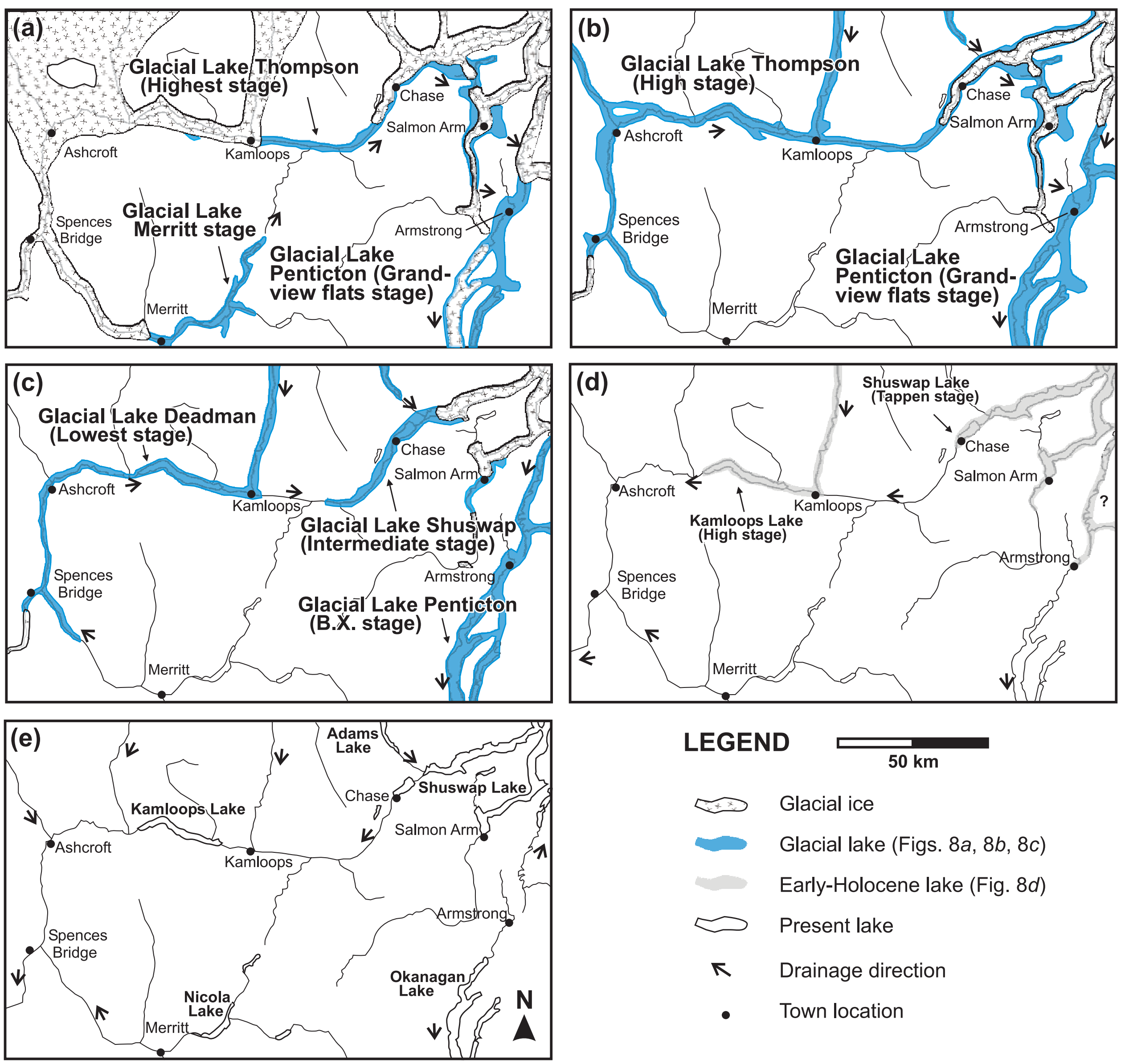


Fig. 8. Late-glacial lake evolution in the southern interior of British Columbia derived from this investigation. Two glacial lakes with three glacial lake stages (*a–c*) are identified in the study area. The extent of Glacial Lake Thompson - Highest stage is after Fulton (his South Thompson stage; 1965, 1969). Glacial lakes and naming outside the Thompson Basin are applied from Fulton (1969). Earlier stages of Glacial Lake Merritt and Glacial Lake Penticton are not shown. Plateau ice is not shown in (*b*) and (*c*) as ice-marginal positions remain uncertain. (*d*) Early-Holocene High stage of Kamloops Lake. (*e*) Present lakes and drainage directions.



of delta surfaces and wave-cut benches between defined water planes (5a, 5b, 6a, 6b, 7a, 7b, 7c and 9, Fig. 6) suggests steady or step-wise lowering of the water surface between defined stages (Muto and Steel 2004).

Delta surfaces below Glacial Lake Deadman - Lowest stage (8a, 8b, 14b, Fig. 6) are only located east of Deadman delta (7, Fig. 1). They were likely formed in more extensive early-Holocene stages of Kamloops Lake, dammed behind deposits of the Deadman delta after the drainage of Glacial Lake Deadman (Fig. 8). These high stages of Kamloops Lake are not late glacial because their delta surfaces are lower than lake-bottom sediments west of Deadman delta (Fig. 6*b*) and drainage bedforms on Deadman delta (Figs. 2, 3, 6*b*). Glacial Lake Deadman - Lowest stage records the lowest lake level west of Deadman delta because wave-cut benches and deltas below this stage are absent here. The fact that lake-bottom sediments, subaqueous deltas (Nemec 1990), and river terraces (s, Fig. 6*b*) lie below Glacial Lake Deadman - Lowest stage is consistent with this interpretation. Fluvial aggradation did not exceed the elevation of Glacial Lake Deadman - Lowest stage.

Lake extents and dams

Lake extents are defined by the intersection of water planes, topography, and paleo-dams (ice and (or) sediment). The extents of late-glacial lake stages in the Thompson Basin discovered in this study are presented in Fig. 8. Our reconstruction implies that (1) both Glacial Lake Thompson and Glacial Lake Deadman were more extensive than previously thought (Fulton 1969; Ryder 1981) and (2) Kamloops Lake had an early-Holocene High stage. Paleo-dams and outlets were confirmed near Spences Bridge (Ryder 1970; Anderton 1970), Kamloops and Chase (Fulton 1969; Johnsen 2004).

Regional topography causes the modern South Thompson and Thompson rivers to flow west and then south into the Fraser River (Fig. 1). Thus, a dam in the lower Thompson or Fraser valleys was required for late-glacial lakes to have existed within the Thompson Basin. DEM-constrained extrapolation of lake levels place the southern extent of Glacial Lake Thompson - High stage and Glacial Lake Deadman - Lowest stage lakes near Skoonka Creek, 7 km south of Spences Bridge (Fig. 1). At Skoonka Creek, lake-bottom sediments terminate abruptly and are completely absent in the lower Thompson Valley between Skoonka Creek and Lytton (Fig. 1*b*). These observations suggest that late-glacial lakes were dammed in this section of valley by (1) glacial or pre-glacial fill, (2) a large landslide, or (3) an ice mass. A downwasting ice remnant seems the most likely dam (Ryder 1970; Anderton 1970) because (1) Fraser Glaciation till overlies pre-Fraser sediments (Anderton 1970; Ryder 1981) and is considerably lower than the defined lake stages, (2) major landslide scars or landslide deposits higher than the elevation of the defined lake stages do not exist in this section of the valley (Ryder 1981), (3) thick, downwasting ice once existed in the Thompson Valley based on the distribution of kettled topography and regional deglacial reconstructions (Fulton 1969, 1991; Johnsen 2004; Ryder 1976), (4) this section of valley is narrow and with steep mountainsides that would have provided shade and possibly colluvium, both facilitating ice preservation, and (5) tributaries to the lower Thompson Valley contain elevated lake-bottom sediments (~525 m asl; Ryder 1981) necessitating an ice dam in the Thompson Valley.

Previous research identified Glacial Lake Deadman - Durand stage as the lowest late-glacial lake level in the study area (Fulton 1969; Ryder 1976, 1981). The Durand stage had no measurable isostatic tilt (Fulton 1969) and, therefore, it was inferred to have developed after the period of major glacio-isostatic adjustment following unloading of the Cordilleran Ice Sheet (Fulton 1969). A projection of this water plane (375 m asl, Fulton 1969) through the western portion of the study area intersects continuous deposits of lake-bottom sediments (tops ~400 m asl) and lies well below the elevation of lake drainage bedforms on Deadman delta (427 m asl, s**, Fig. 6). If the water level in Glacial Lake Deadman fell to this stage an eastward-flowing river would have formed in the western portion of the study area, not a lake. No gravel terraces record eastward paleoflows. These inconsistencies suggest that the Durand stage of Glacial Lake Deadman should be abandoned.

Lake paleogeography and statistics

The paleogeography of Glacial Lake Thompson - High stage and of Glacial Lake Deadman - Lowest stage was mapped by integrating inferred lake extents with DEMs of their stages (from eqs. [4], [5]) and the land surface (British Columbia Government 1996) in a GIS. A downvalley cross-section through these lakes is presented in Fig. 7. Quantitative parameters for these late-glacial lakes and their subsequent incision were determined by differencing DEMs of paleolake stages, present topography, and estimated lake bottom (Fig. 7). Lake-bottom elevation was estimated from the highest occurrences of lake-bottom sediments. The results are presented in Table 2.

Evolution of late-glacial lakes in the Thompson Basin

The paleogeography of late-glacial lakes in the Thompson Basin has been inferred from a synthesis of information (Fig. 8). No dates are available for individual lakes or their stages. However, we suggest that higher stages were formed before lower stages because (1) lake base levels were controlled by stagnant, downwasting ice masses and readily eroded lake-bottom sediment (Fulton 1967, 1991; Figs. 7, 8), (2) lower delta surfaces are nested within higher delta surfaces (Fig. 2; Muto and Steel 2004), (3) lake drainage bedforms lie below the elevation of the lowest late-glacial lake level (Glacial Lake Deadman - Lowest stage; Figs. 2, 3, 6; Johnsen 2004), and (4) lower water planes exhibit less tilt and have thus experienced less glacio-isostatic deformation.

The South Thompson Valley became ice free first and was occupied by Glacial Lake Thompson with an eastern outlet along ice margins and topographic divides. The highest stage of this lake was restricted to the South Thompson Valley, impounded by ice on its eastern and western ends (Fig. 8*a*; Fulton 1969). This lake extended westward and lowered as ice backwasted into the Thompson Valley. Lake stabilization and (or) high-energy conditions allowed for the development of wave-cut benches that define the later high stage of this lake (Figs. 8*a*, 8*b*). Glacial Lake Thompson - High stage

Table 2. Quantitative parameters of late-glacial lakes and incision in the Thompson Basin.

Property[a]	Units	Stage (and water plane): Glacial Lake Thompson - High stage (Upper)	Stage (and water plane): Glacial Lake Deadman - Lowest stage (Lower)
Water plane			
Orientation (up-plane)	degrees	332	321
Tilt	m km^{-1}	1.8	1.7
Volume			
Of lake	km^3	83.6	23.6
Of final drainage event[b]	km^3	—	20.0
Surface area	km^2	801	414
Width	km	1.5–6.7	1.5–4.0
Length	km	220	160
Maximum depth	m	198	80
Typical depth	m	~140	~50
Incision volume[c]	km^3	—	14.1

[a]All calculations account for the isostatic deformation of representative water planes. The Kamloops Lake area may have been occupied by ice during the time of these lakes, and therefore the volume (3.7 km^3) and depth (71 m mean depth, 143 m maximum depth) of Kamloops Lake were not used in the calculations. Errors in estimates: As the land surface DEM is of the present landscape, some relatively minor error was introduced from Holocene river incision and post-lake valley fills (fluvial gravel and paraglacial fans capping lake-bottom sediments). Incision of the main valley was eliminated by using a crude lake-bottom DEM. Lake bottom could only be estimated from valley-side exposures, which likely gave a shallow estimate of lake bottom and thus an under-estimate of lake volume. Minor elevation error introduced from the land surface DEM (±10 m at 90% confidence for interpolated points, British Columbia Government 1992). The same lake-bottom DEM was used for both lake stage reconstructions, although the lake-bottom surface was likely an unknown amount lower for the upper, earlier lake stage due to continual sediment accumulation in the basin. Since the northern extent of both lakes in the North Thompson Valley is unknown beyond McClure, it is possible that volumes for these lakes could be higher.

[b]Based on elevation considerations 3.6 km^3 of Glacial Lake Deadman that was likely retained behind Deadman Delta (a, Fig. 7) when the lake drained catastrophically.

[c]Significant incision occured following the final drainage of Glacial Lake Deadman during the Holocene. This is the volume of sediment below the estimated lake bottom that has since been removed by river incision. This is a low estimate as not all of the North Thompson Valley was included in the calculations.

contained ~84 km^3 of water (Table 2). Continued downwasting and backwasting of ice resulted in continued westward lake growth and lake-level lowering. Once the lake level lowered to expose the South Thompson silt, lakes in the Thompson Basin were separated from those in the Shuswap Basin. Glacial Lake Deadman formed with a new eastern outlet near Kamloops — a river incised into South Thompson silt (Fig. 8c). The lowest stage of Glacial Lake Deadman occurred after substantial incision (~65 m) into the South Thompson silt (s*, Figs. 6, 7, 8*c*). Glacial Lake Deadman - Lowest stage contained ~24 km^3 of water (Table 2). Lake-level lowering between Glacial Lake Thompson - High stage and Glacial Lake Deadman - Lowest stage was likely gradual rather than abrupt because (1) primary water-plane features and inset delta surfaces are found between these two water planes (Figs. 2, 6; Muto and Steel 2004), and (2) base level was controlled by readily eroded South Thompson silt.

Catastrophic lake drainage

Glacial Lake Deadman - Lowest stage drained westward catastrophically as the ice dam south of Spences Bridge failed. Catastrophic drainage is supported by (1) the absence of wave-cut benches and Gilbert-type deltas below the Lowest stage (Fig. 6), (2) the presence of long-wavelength (>100 m) erosional bedforms (Pickering 1995; Massaria and D'Alessandrob 2000) on the distal portion of Deadman delta (Figs. 2, 3; Johnsen 2004) ~26 m below the lowest lake level, and (3) theory on the drainage of ice-dammed lakes (e.g., Clague and Evans 1994).

Various draining mechanisms have been proposed for ice-dammed lakes, including overspill, tunnel enlargement, ice-dam flotation, and seismic events (Walder and Costa 1996; Tweed and Russell 1999, and references therein). It is difficult to discern which of these mechanisms resulted in the final failure of the ice dam south of Spences Bridge. Tunnel enlargement is preferred given that the glaciers were likely warm-based during deglaciation (Lian and Hicock 2000), and an exponentially increasing discharge explains the erosional bedforms on Deadman delta. Failure by dam flotation may not have been possible, as colluvium from the adjacent, steep slopes may have caused the weight of the dam to exceed the buoyant force. Only a small amount of rock debris (as little as 4% by volume) is needed to prevent ice flotation (Tweed 2000). Failure of the ice dam by overspill may have occurred as the ice dam was lowered by ablation.

An estimated 20 km^3 of water (Table 2) drained catastrophically into the Fraser River system when the ice dam in the lower Thompson Valley failed, sometime before 9740 – 10 210 BP (GSC-193, Dyck et al. 1965; Fulton 1969). It is possible that this event may have triggered the failure of other glacial lakes downstream or upstream in the Fraser River system. Eventually the floodwaters would have reached the coast and flowed into the Strait of Georgia, a total distance of ~250 km. This event resulted in the regional drainage reversal of the Thompson and South Thompson rivers and the capture of the Thompson Basin (presently about 54 000 km^2 in area, Environment Canada 1989) by the Fraser River system.

Offshore outburst flood deposits

Exotic muds containing reworked Tertiary microfossils have been identified in drill cores from the Strait of Georgia and Saanich Inlet (Blais-Stevens et al. 2001, 2003; Conway et al. 2001). These muds are inferred to record a major flood event or events from the British Columbia mainland that occurred between 10 190 and 11 940 BP (Blais-Stevens et al. 2001, 2003; Conway et al. 2001) because altogether (1) they contain abundant reworked Tertiary microfossils of mainland provenance, (2) texturally and structurally the muds are anomalous in cores — they are normally graded and have a sharp, non-erosional basal contact and gradational upper contact, (3) they lack bioturbation structures, except at the base of one deposit (possible invertebrate escape structures), (4) they have a well-developed brackish-water signal in the diatom assemblage, and (5) they were deposited during the same period that late-glacial lakes on the mainland disappeared (Blais-Stevens et al. 2001, 2003; Conway et al. 2001). Cores in the Saanich Inlet show there are two exotic mud deposits separated by tens of centimetres of sediment (Blais-Stevens et al. 2001, 2003). The upper deposit has a gradational upper and lower contact. Cores in the Strait of Georgia show only one exotic mud deposit. Some variation in the dates between the Saanich Inlet and Strait of Georgia deposits (Blais-Stevens et al. 2001; Conway et al. 2001) may indicate that these deposits record separate flood events. Exotic muds may have been deposited by jökulhlaups resulting from the final drainage of Glacial Lake Deadman (Johnsen and Brennand 2001) or Glacial Lake Fraser (Clague 1987; Huntley and Broster 1997).

Post-glacial lake events

Following the drainage of Glacial Lake Deadman and the regional drainage reversal, progressively lower stages of Kamloops Lake developed behind Deadman delta (Fig. 8*d*). The ancestral Thompson River incised down through ~150 m of valley fill to bedrock, deposited gravel, and developed numerous fluvial terraces. Incision to within a few metres of present river level was achieved by the mid-Holocene (Ryder 1981; 7510–7670 BP, Hallett et al. 1997). During and since the Glacial Lake Deadman jökulhlaup, > 14.1 km^3 of valley fill has been eroded from the study area (Table 2). Most of this eroded fill was glaciolacustrine silt and fine sand. Differential glacio-isostatic rebound has caused the former shorelines of late-glacial lakes in the Thompson Basin to have among the highest tilts in the world (1.8–1.7 m km^{-1}). For example, lakes associated with the Laurentide Ice Sheet typically have their shorelines tilted 0.5 to 1.0 m km^{-1} (Teller and Thorleifson 1983). The tilt directions of water planes in the Thompson and Merritt basins suggest that crustal load was greatest to the north-northwest (330°–350° azimuth) and less to the south-southeast. This direction corresponds with a previously inferred Cordilleran Ice Sheet divide on the Fraser Plateau (e.g., Ryder et al. 1991).

Conclusion

As the Cordilleran Ice Sheet stagnated, two late-glacial, ice-dammed lakes developed within the Thompson Basin: Glacial Lake Thompson and Glacial Lake Deadman. Our understanding of the paleogeography and evolution of these lakes has been improved by applying a robust paleogeographic reconstruction procedure; this procedure has general application. The lakes were ribbon-shaped (width to length ratio of ~3:100), deep (>140 to ~50 m) and contained significant water volumes (84–24 km^3). We conclude that these lakes were more extensive than previously thought and that they lengthened and lowered as ice decayed. The final failure of an ice dam south of Spences bridge resulted in an ~20 km^3 jökulhlaup that eroded bedforms and deposited flood eddy bars within the lake basin, travelled ~250 km along the Fraser River system and may have deposited exotic mud offshore sometime between 10 190 to 11 940 BP.

We define the geometry of two paleo water planes. Their glacio-isostatic tilts are among the highest measured in the world (1.7–1.8 m km^{-1}) and their orientations suggest that ice sheet loads were greater or longer-lived to the north-northwest of the study area. These glacio-isostatic results support those from the nearby Merritt Basin (Fig. 1*b*; Fulton and Walcott 1975) and lend support to the notion of an ice divide centred north-northwest of the study area on the Fraser Plateau (e.g., Ryder et al. 1991).

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References

Allan, J.C., Stephenson, W.J., Kirk, R.M., and Taylor, A. 2002. Lacustrine shore platforms at lake Waikaremoana, North Island, New Zealand. Earth Surface Processes and Landforms, **27**: 207–220.

Anderton, L.J. 1970. Quaternary stratigraphy and geomorphology of the lower Thompson Valley, British Columbia. MSc. Thesis, The University of British Columbia, Vancouver, B.C., Canada.

Baker, V.R. 1973, Paleohydrology and sedimentology of Lake Missoula flooding in eastern Washington. Geological Society of America, Special Paper 144.

Bell, T., Liverman, D.G.E., Batterson, M.J., and Sheppard, K. 2001. Late Wisconsinan stratigraphy and chronology of southern St. George's Bay, Newfoundland: a re-appraisal. Canadian Journal of Earth Sciences, **38**: 851–869.

Beres, M., Huggenberger, P., Green, A.G., and Horstmeyer, H. 1999. Using two- and three-dimensional georadar methods to characterize glaciofluvial architecture. Sedimentary Geology, **129**: 1–24.

Blais-Stevens, A., Bornhold, B.D., Kemp, A.E.S., Dean, J.M., and Vaan, A.A. 2001. Overview of the Late Quaternary stratigraphy in Saanich Inlet, British Columbia: results of Ocean Drilling Program Leg 169S. Marine Geology, **174**: 3–26.

Blais-Stevens, A., Clague, J.J., Hebda, R.J., Mathewes, R.W., and Bornhold, B.D. 2003. Record of large, late Pleistocene outburst floods preserved in Saanich Inlet sediments. Quaternary Science Reviews, **21**: 2327–2334.

British Columbia Government. 1992. British Columbia specification and guidelines for geomatics, content series 3, digital base mapping at 1 : 20,000, release 2.0. Ministry of Environment, Lands and Parks, Victoria, B.C., Canada.

British Columbia Government. 1996. Digital elevation model, 25 m grid, TRIM1, 1 : 250,000, version 1.1. Ministry of Environment, Lands and Parks, Victoria, B.C., Canada.

Burrough, P.A. 1986. Principles in geographic information systems for land resources assessment. Oxford University Press, Oxford, UK.

Cheel, R.J., and Rust, B.R. 1980. Coarse-grained facies of glacio-marine deposits near Ottawa, Canada. *In* Research in glacial, glacio-fluvial, and glacio-lacustrine systems. *Edited by* R. Davidson-Arnott, W. Nickling, and B.D. Fahey. Proceedings of the 6th Guelph Symposium on Geomorphology, Guelph, Ont., 1980, pp.279–304.

Clague, J.J. 1981. Late Quaternary geochronology of British Columbia. Part 2: summary and discussion of radiocarbon-dated Quaternary history. Geological Survey of Canada, Paper 80-35.

Clague, J.J. 1987. Quaternary stratigraphy and history, Williams Lake, British Columbia. Canadian Journal of Earth Sciences, **24**: 147–158.

Clague, J.J. 2000. Recognizing order in chaotic sequences of Quaternary sediments in the Canadian Cordillera. Quaternary International, **68–71**: 29–38.

Clague, J.J., and Evans, S.G. 1994. Formation and failure of natural dams in the Canadian Cordillera. Geological Survey of Canada, Bulletin, 464.

Clague, J.J., and Evans, S.G. 2003. Geologic framework of large historic landslides in Thompson River Valley, British Columbia. Environmental and Engineering Geoscience, **9**: 201–212.

Clague, J.J., and James, T.S. 2002. History and isostatic effects of the last ice sheet in southern British Columbia. Quaternary Science Reviews, **21**: 71–87.

Conway, K.W., Barrie, J.V., and Hebda, R.J. 2001. Evidence for a Late Quaternary outburst flood event in the Georgia Basin, British Columbia. Geological Survey of Canada, Current Research 2001-A13.

Dyck, W., Fyles, J.G., and Blake, W., Jr. 1965. Geological Survey of Canada radiocarbon dates IV. Radiocarbon 7.

Environment Canada. 1989. Surface water data: reference index 1988. Water Survey of Canada, Ottawa, Ont.

Eyles, N., and Mullins, H.T. 1997. Seismic-stratigraphy of Shuswap Lake, British Columbia, Canada. Sedimentary Geology, **109**: 283–303.

Eyles, N., Mullins, H.T., and Hine, A.C. 1991. The seismic stratigraphy of Okanagan Lake, Brtiish Columbia; a record of rapid deglaciation in a deep 'fiord-lake' basin. Sedimentary Geology, **73**: 13–41.

Fulton, R.J. 1965. Silt deposition in late-glacial lakes of the southern British Columbia. America Journal of Science, **263**: 553–570.

Fulton, R.J. 1967. Deglaciation studies in Kamloops region, an area of moderate relief, British Columbia. Geological Survey of Canada, Bulletin 154.

Fulton, R.J. 1969. Glacial lake history, southern Interior Plateau, British Columbia. Geological Survey of Canada, Paper 69-37.

Fulton, R.J. 1971. Radiocarbon geochronology of southern British Columbia. Geological Survey of Canada, Paper 71-37.

Fulton, R.J. 1987. Stop 6-5 Deadman River. *In* Quaternary geology of the southern Canadian Cordillera. *Edited by* J.J. Clague, S.G. Evans, R.J. Fulton, J.M. Ryder, and A.H. Stryd. XII INQUA (International Union for Quarternary Research) congress field excursion, guide book A-18.

Fulton, R.J. 1991. A conceptual model for the growth and decay of the Cordilleran Ice Sheet. Géographie Physique et Quaternaire, **45**: 281–286.

Fulton, R.J., and Armstrong, J.E. 1965. Day 12, stop 12–3. *In* Guidebook for field conference J, Pacific Northwest, INQUA, VIIth congress, pp. 101–102.

Fulton, R.J., and Smith, G.W. 1978. Late Pleistocene stratigraphy of south-central British Columbia. Canadian Journal of Earth Sciences, **15**: 971–980.

Fulton, R.J., and Walcott, R.I. 1975. Lithospheric flexure as shown by deformation of glacial lake shorelines in southern British Columbia. In Quantitative studies in the geological sciences. *Edited by* E.H.T. Whitten. Geological Society of America, Boulder, CO., Memoir 142, pp. 163–173.

Gray, J.M. 1983. The measurement of relict shoreline altitudes in areas affected by glacio-isostasy, with particular reference to Scotland. *In* Shorelines and isostasy. *Edited by* D.E. Smith and A.G. Dawson. Institute of British Geographers, Special Publication 16, pp. 97–127.

Gustavson, T.C., Ashley, G.M., and Boothroyd, J.C. 1975. Depositional sequences in glaciolacustrine sedimentation. *In* Glaciofluvial and glaciolacustrine sedimentation. *Edited by* A.V. Jopling and B.C. MacDonald. Society of Economic Paleontologists and Mineralogists, Special Publication 23, pp. 264–280.

Hallett, D.J., Hills, L.V., and Clague, J.J. 1997. New accelerator mass spectrometry radiocarbon ages for the Mazama tephra layer from Kootenay National Park, British Columbia. Canadian Journal of Earth Sciences, **34**: 1202–1209.

Huntley, D.H., and Broster, B.E. 1997. The Late Wisconsinan deglacial history of the east-central Taseko Lakes area, British Columbia. Canadian Journal of Earth Sciences, **34**: 1510–1520.

Johnsen, T.F. 2004. Late glacial lakes of the Thompson Basin, southern interior of British Columbia: paleography and paleoenvironment. Unpublished M.Sc. Thesis. Simon Fraser University, Burnaby, B.C., Canada.

Johnsen, T.F., and Brennand, T.A. 2001. Glacial Lake Deadman, southern interior of British Columbia: paleogeography and paleoenvironment. Occasional papers in Earth Sciences, Canadian Quaternary Association, Annual Meeting 2001, Whitehorse, Yukon Territory, August 20–24, 2001, Program and Abstracts, p. 42.

Lian, O.B., and Hicock, S.R. 2000. Thermal conditions beneath parts of the last Cordilleran Ice Sheet near its centre as inferred from subglacial till, associated sediments, and bedrock. Quaternary International, **68–71**: 147–162.

Massaria, F., and D'Alessandrob, A. 2000. Tsunami-related scour-and-drape undulations in Middle Pliocene restricted-bay carbonate deposits (Salento, south Italy). Sedimentary Geology, **135**: 265–281.

Mullins, H.T., and Eyles, N. 1996. Subsurface geologic investigations of New York Finger Lakes; implications for late Quaternary deglaciation and environmental change. Geological Society of America, Special Paper 311.

Muto, T., and Steel, R.J. 2004. Autogenic response of fluvial deltas to steady sea-level fall: implications from flume-tank experiments. Geology, **32**: 401–404.

Nemec, W. 1990. Deltas — remarks on terminology and classification. *In* Coarse-grained deltas. *Edited by* A. Colella and B.B. Prior. International Association of Sedimentologists, Special Publication 10, Blackwell, Oxford, U.K., pp. 3–12.

Pharo, C.H., and Carmack, E.C. 1979. Sedimentation processes in a short residence-time intermontane lake, Kamloops, British Columbia. Sedimentology, **26**: 523–541.

Pickering, K.T. 1995. Are enigmatic erosional sandy wavelike bedforms in Jurassic Bridport Sands, Dorset, due to standing waves? Journal of the Geological Society (of London), **152**: 481–485.

Rayburn, J.A. 1997. Correlation of the Campbell strandlines along the northwestern margin of glacial Lake Agassiz. MSc. thesis, University of Manitoba, Winnipeg, Man., Canada.

Ryder, J.M. 1970. Alluvial fans of post-glacial environments within British Columbia. Ph.D. thesis, The University of British Columbia, Vancouver, B.C., Canada.

Ryder, J.M. 1976. Terrain inventory and Quaternary geology, Ashcroft, British Columbia. Geological Survey of Canada, Paper 74-79.

Ryder, J.M. 1981. Terrain inventory and Quaternary geology, Lytton, British Columbia. Geological Survey of Canada, Paper 79-25.

Ryder, J.M, Fulton, R.J., and Clague, J.J. 1991. The Cordilleran Ice Sheet and the glacial geomorphology of southern and central British Columbia. Géographie physique et Quaternaire, **45**: 365–377.

Sawicki, O., and Smith, D.G. 1992. Glacial Lake Invermere, upper Columbia River, British Columbia: a paleogeographic reconstruction. Canadian Journal of Earth Sciences, **29**: 687–692.

Shaw, J. 1977. Sedimentation in an alpine lake during deglaciation, Okanagan Valley, British Columbia, Canada. Geografiska Annaler, **59A**: 221–240.

Teller, J.T., and Thorleifson, L.H. 1983. The Lake Agassiz – Lake Superior Connection. *In* Glacial Lake Agassiz. *Edited by* J.T. Teller and L. Clayton. Geological Association of Canada, Special Paper 26, pp. 261–290.

Thorson, R.M. 1989. Glacio-isostatic response of the Puget Sound area, Washington. Geological Society of America, Bulletin, **101**: 1163–1174.

Tweed, F.S. 2000. Jökulhlaup initiation by ice-dam floatation: the significance of glacier debris content. Short communication. Earth Surfaces Processes and Landforms, **25**: 105–108.

Tweed, F.S., and Russel, A.J. 1999. Controls on the formation and sudden drainage of glacier-impounded lakes: implications for jökulhlaup characteristics. Progress in Physical Geography, **23**: 79–110.

Walcott, R.I. 1970. Isostatic response to loading of the crust in Canada. Canadian Journal of Earth Sciences, **7**: 716–727.

Walder, J.S., and Costa, J.E. 1996. Outburst floods from glacier-dammed lakes: the effect of mode of lake drainage on flood magnitude. Earth Surfaces Processes and Landforms, **21**: 701–723.