Palaeogeographical reconstruction and hydrology of glacial Lake Purcell during MIS 2 and its potential impact on the Channeled Scabland, USA

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Large, ice-marginal lakes that were impounded by the maximally-extended Cordilleran Ice Sheet (CIS) provided source waters for the extraordinarily large floods that formed the Channeled Scabland of Washington and Idaho, USA. However, flood flows that drained CIS meltwater and contributed to landscape evolution during later stages of deglaciation have hitherto been poorly investigated. This paper provides the first evidence for such a late deglacial floodwater source: glacial Lake Purcell (gLP). Sedimentary evidence records the northward extension of gLP from Idaho, USA into British Columbia, Canada and establishes its minimum palaeogeographical extent. Sedimentary evidence suggests that the deglacial Purcell Lobe was a capable ice dam that impounded large volumes of gLP water. A review of glacioisostatically affected lakes during CIS deglaciation suggests that gLP could have been subjected to tilts ranging from 0 – >1.25 m km^{-1}. Sedimentary evidence suggests high lake plane tilts ($\gtrapprox$1.25 m km^{-1}) are the most likely to have affected gLP. Using this, the palaeogeography and volume of gLP are modelled, revealing that ~116 km^3 of water was susceptible to sudden drainage into the Channeled Scabland via the Columbia River system. This calculation is supported by sedimentary and geomorphic evidence compatible with energetic flood flows along the gLP drainage route and suggests gLP drained suddenly, causing significant landscape change.

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Growing concerns over the stability of future hydrosphere-cryosphere interactions and our ability to accurately predict the behaviour of modern glaciers and ice sheets (e.g. Bamber *et al*. 2009; Gardner *et al*. 2013) highlight the importance of a complete understanding of Cordilleran Ice Sheet (CIS) decay. Ice-marginal lakes play an important role in this improved understanding because they affect ice dynamics (Carrivick & Tweed 2013), are effective sediment traps that record detailed glacial histories (e.g. Larsen *et al.* 2011; Liermann *et al.* 2012), and are prone to catastrophic drainage that can influence regional sediment transport and drainage systems (Korup 2012). Considering the important effects that glacial lakes can impose on ice sheet decay and landscape evolution, and their increasing abundance and size along modern, deglaciating ice margins (Carrivick & Tweed 2013), the importance of developing a thorough understanding of their role during the deglaciation of the CIS is evident.

At the Local Last Glacial Maximum (LLGM) large glacial lakes, like the ~2 600 km^3 glacial Lake Missoula (O'Connor & Baker 1992; Miyamoto *et al.* 2006, 2007), formed when the southern margin of the CIS disrupted regional drainage patterns (Baker 2009). Some of these lakes drained catastrophically and contributed to the formation of the Channeled Scabland (Fig. 1A), a megaflood landscape that geomorphically and sedimentologically records flood flows of nearly unprecedented Earthly scale with maximum discharges of 10-20 Sverdrups (Benito & O'Connor 2003; Denlinger & O'Connor 2010). Whereas geological and sedimentological signatures of enormous jökulhlaups (glacial lake outburst floods) entering the Channeled Scabland are abundant and well documented (e.g. Bretz 1925, 1969; Baker 2009; Benito & O'Connor 2003), the potential for post-LLGM flood flows from the drainage of glacial lakes in British Columbia has been proposed (Shaw *et al.* 1999; Lesemann & Brennand 2009; Waitt *et al.* 2009; Waitt 2016) but remains relatively poorly understood.

The role of ice-marginal lake formation within the Purcell Trench during CIS deglaciation has received inconstant speculation. Alden (1953) first contemplated a glacial lake in the Purcell Trench and its possible drainage into the Columbia River system via the Kootenay River valley (Fig. 1B). Most researchers (e.g. Alden 1953; Johns 1970; R. Fulton, pers. comm. 2010) speculated that the glacial lake in the Purcell Trench was shallow and primarily ice marginal or supraglacial, owing to stagnant ice occupying the Purcell Trench. These authors also suggest that glacial lake water in the Purcell Trench likely drained gradually past a spillway in the south (the Elmira spillway) and the downwasting ice in the north (Fig. 1B). However, Waitt *et al*. (2009) and Waitt (2016) propose that more energetic drainage of a proglacial lake in the Purcell Trench may have supplied post-Missoula flood flows to the Columbia River.

This study provides the first comprehensive investigation of glacial lake evolution in the Purcell Trench. We use geological evidence and previous records of CIS glacioisostatic tilt to inform a palaeogeographic reconstruction of a large lake, named here glacial Lake Purcell (gLP). We explore evidence for ice damming of the lake and its drainage through the Kootenay River valley. These analyses are used to assess the potential for energetic flood flows from the Purcell Trench into the Channeled Scabland after the final drainage of glacial lakes Missoula and Columbia.

Previous work on Purcell Lobe ice-marginal lakes

Previous studies near the Purcell Trench have reconstructed glacial Lake Kootenai (gLK) from thick deposits of lake bed sediments (sand and silt) in valley systems in northern Idaho and northwestern Montana (Alden 1953; Johns 1970; Smith 2006; Fig. 1A). This lake formed when river systems were impounded by the retreating Purcell Lobe (Alden 1953; Johns 1970; Smith 2006; Fig. 2). The sediments recording glacial Lake Kootenai are over 90 m thick in some areas and record rapid deposition proximal to inflows (Alden 1953; Smith 2006). Valley-side benches composed of lake bed sediments attributed to gLK range in elevation from 700-740 m a.s.l. in Idaho and from 730-762 m a.s.l. in Montana due to different spillway heights (Alden 1953). The Bull River spillway (Fig. 1B) in Montana was the first flow to be activated and would have commenced following a lowering of the final stage of gLM in the Clark Fork River valley to the south of gLK (Alden 1953). After sufficient northward retreat of the Purcell Lobe, gLK decanted into the southern Purcell Trench, forming a large flood-related fan on the valley floor and an unnamed proglacial lake. Lake levels in the Purcell Trench were dictated by the Elmira spillway (ibid). The geomorphology of the Elmira spillway suggests that its original height was ~710 m a.s.l. and that incision from lake drainage is responsible for its current elevation of 655 m a.s.l. (ibid).

The naming conventions used by Alden (1953) and adopted by Johns (1970) and Smith (2006) are abandoned in this study because they ambiguously describe distinct water bodies with a single name (gLK). Furthermore, the name 'glacial Lake Kootenay' employed by Waitt *et al*. (2009) is not used, as its closeness to Alden's lake name is a potential source of confusion. Instead a naming system is employed that distinguishes the discrete and possibly contemporaneous lakes that occupied separate basins (Fig. 2). This new naming scheme retains Alden's glacial Lake Kootenai moniker in Montana, USA (where most of his research was conducted) but designates the unnamed lake and its northern expansion in the Purcell Trench "glacial Lake Purcell" (Fig. 2).

The volumes of these glacial lakes have also been speculated upon and several researchers have pointed out that volume was contingent on the style of CIS retreat through the Purcell Trench. If Purcell Lobe retreat was dominated by stagnation and downwasting, the ice would have likely displaced much of the volume available to any glacial lake. Fulton (1967, 1991) proposes a CIS deglacial model dominated by stagnant, residual ice occupying valley systems resulting from a rapid rise of the equilibrium line due to rapid climate amelioration. Sedimentary evidence for this stagnation, and resultant downwasting, has been reported in the interior of British Columbia (Eyles & Clague 1991; Ryder *et al*. 1991). During ice stagnation, glacial lake volume would have been minimized by valley occupying ice. However marginal areas of the CIS may have experienced a more complex pattern of decay (Fulton 1967) and these complications may have been further exacerbated in mountainous terrain by late deglacial alpine ice advances (Lakeman *et al*. 2008). Such complexities, along with potential inconsistencies in regional glacioisostatic response from crustal heterogeneities (cf. Thorson 1989), may have enabled the formation of a deep, high-volume gLP and highlight the need for investigations in the Purcell Trench.

Initial evidence for a high-volume, late-deglacial gLP has been supplied by Waitt *et al*. (2009) and Waitt (2016), who suggest that a glacial lake in the Purcell Trench was a potential water source for flood flow(s) in the Columbia River valley. Putative geomorphic evidence for post-Missoula, late-Wisconsin Glacial Lake Outburst Floods (GLOFs) in the Columbia River valley includes two megaflood bars marked by dune-scale bedforms ("giant current dunes") near Chelan Falls, Washington (Waitt *et al.* 2009; Fig. 1A). These dune-scale bedforms are tephrostratigraphically dated to <13.5 cal. ka BP (Kuehn et al. 2009), after the final drainage of glacial lakes Missoula and Columbia and Lake Bonneville (Waitt *et al.* 1994, 2009). Age constraints on the deglacial CIS are compatible with the tephrostratigraphic age of the dune-scale bedforms and place the Purcell Lobe ice margin near the Kootenay River valley by ~13.5 cal. ka BP (Dyke *et al*. 2003).

Study area

Data were gathered for this study within the Purcell Trench, its high-relief tributary valleys, and along the Kootenay River valley (KRv; Fig. 1B). Much of the floor of the Purcell Trench in Canada is occupied by Kootenay Lake, which is a ribbon-shaped lake >100 km long with an average width of ~6.5 km (Fig. 1B). Kootenay Lake's water surface elevation is controlled by the Corra Linn Dam in the Kootenay River valley to an elevation of ~532 m a.s.l. (Davis 1920; Kyle 1938; Fig. 1B). Kootenay Lake marks a change in spelling from the Kootenai River to the Kootenay River (Fig. 1B) and is essentially a stagnation point in the flow of the Kootenai/y River along its circuitous westward route from the Rocky Mountain Trench, British Columbia through the Columbia Mountains. Kootenay Lake drains out of its West Arm via the Kootenay River, which is the first major tributary of the Columbia River. In this study, the West Arm of Kootenay Lake and the Kootenay River are jointly referred to as the Kootenay River valley (KRv; Fig. 1B).

Methods

Geomorphology and sedimentology

Geomorphic analyses and preliminary investigations to identify potential field sites were carried out using publicly available digital elevation models from Geobase (from Natural Resources Canada) and the National Elevation Database (NED, from the United States Geological Survey). The two datasets were compiled and re-gridded into a single, 25-m resolution Digital Elevation Model (DEM).

Sedimentary investigations entailed lithofacies identification, gravel fabric and ripple palaeoflow measurements, and centimetre-scale logging of exposures. Sedimentary data are presented as stereograms, rose diagrams, and exposure photographs. For each fabric analysis ≥30 clast a-b plane attitude measurements (maximum dip and down-dip direction of the a-b plane) were taken in <0.5 m^2 areas of exposures. The data were then plotted on lower hemisphere, equal area (Schmidt) diagrams as contoured stereonets using the cosine sums method (Stereo32 software) with a cosine exponent of 20 (cf. Roeller 2008). Fabrics from kame deposits are designated K1, K2, K3, K4, K5 and palaeoflow data recording drainage from the Purcell Trench are designated

Dr1, D1, D2 and D3 (Fig. 1B). Clast a-axis position relative to the direction of a-b plane maximum dip in gravel fabrics was used to determine the likely mode of clast mobilization. A dominance of clast a-axes transverse to dip direction (a(t)) suggests clasts rolled along the bed, whereas a dominance of clasts parallel to dip direction (a(p)) implies clast sliding across the bed or deposition from suspension in a hyperconcentrated flow (cf. Brennand 1994).

Palaeogeographical modelling of gLP

Typically, glacial lake extent is reconstructed from the distribution of lake bed sediments and by correlating water-plane indicators (e.g., deltas, shorelines; cf. Johnsen & Brennand 2004). However, a dearth of gLP water-plane indicators were identified on the steep bedrock valley walls of the Purcell Trench. There is also a poor potential for lacustrine sediment preservation within the floodplain of the modern Kootenai River and Kootenay Lake may cover significant areas of gLP sediment (because they share a common basin). Thus, after confirming minimum extents with sedimentary data, gLP extent and volume are estimated by assessing modelled lake surface planes against limited evidence and in comparison to contemporaneous and geographically close palaeolakes.

Two critical palaeogeographical elements enabled this modelling of gLP: (i) the reconstruction of the pre-incision lake bed and (ii) the application of an appropriate glacioiosostatic adjustment (GIA) to the water plane. These reconstructions were used to produce a combination of rasters that were used in conjunction with modern topography to define gLP palaeogeography and calculate its volume (cf. Leverington *et al*. 2002; DeVogel *et al*. 2004). The pre-incision palaeo-lake bed was reconstructed using an inverse distance weighting function to interpolate a series of pre-incision elevations measured using the composite DEM. In the model, the boundaries of this pre-incision lake bed surface were defined by its intersection with topographic highs on the composite DEM of the Purcell Trench. A plausible range of GIA was derived from a survey of previously reported glacioisostatic tilts for CIS glacial lake planes. The published ages of these lakes are also reported, which are derived using disparate methods with varying accuracy. These ages were recalibrated for this study with Calib software (Stuiver & Reimer 1993) to 2σ confidence using the IntCal13 radiocarbon curve (Reimer *et al.* 2013) to improve comparability of the ages (cf. Peters *et al.* 2016).

The resultant array of plausible lake plane tilts was projected along the Purcell Trench from the Elmira Spillway, which controlled gLP lake levels prior to drainage into the Kootenay River valley (Alden 1953; Johns 1970; Smith 2006). A DEM of modern Kootenay Lake bathymetry (bathymetric DEM) was constructed from interpolated individual soundings using a GIS and was used to estimate the total volume of gLP (i.e. the volume of modern Kootenay Lake was added to the calculated drainable volume of gLP based on topographic DEMs).

Results and interpretations

The northward extension of gLP

Evidence for the northward extension of gLP is mainly preserved in large (>60 km long and up to ~20 km wide) sediment benches that occupy the floor of the Purcell Trench adjacent to the Kootenai River (Fig. 1B). These benches extend from the Elmira spillway (~10 km south of Bonners Ferry, Idaho) to ~5 km north of Creston, British Columbia (Fig. 1B), and reach an elevation of up to 706 m a.s.l. (Fig. 3), >176 m above the modern water level of Kootenay Lake. Bench tread elevations are incrementally lower towards the Kootenai River floodplain (Fig. 3). Some treads exhibit channels on their surfaces (channels A and B, Fig. 3).

The benches are composed of massive or laminated silt and clay rhythmites (sites 3, 4, 5, 15; Fig. 4A) containing occasional pebble- to cobble-sized clasts (lonestones) that display striated facets and plucked ends (sites 5, 15). Lonestones occur in relatively high abundance within massive silt (Fig. 4B). Rare deposits of massive, silty coarse sand, interbedded with silt and clay laminae that drape lonestones are also present (Fig. 4C). Silt and clay rhythmites are exposed north of the contiguous benches at an elevation of 675 m a.s.l. (site 2, Fig. 1B). Gravel deposits composed of dipping (apparent 30° downwards dip towards 221°), normally-graded, tabular beds that reach thicknesses of ≥10 m are also exposed in the Idaho sediment benches (Fig. 4G). These gravel deposits are typically capped by laminated or massive silt.

The highest bench surfaces are composed of flat-topped silt deposits that share similar elevations, recording contiguous lake bed deposits (cf. Ryder *et al.* 1991; Johnsen & Brennand 2004). The flat-topped, occasionally channelised bench segments (e.g. Fig. 3) record remnant lake bed sediments following fluvial incision (cf. Clague 1986). This interpretation of terrace formation is supported by the close proximity of the Kootenai River.

Rhythmites record lake bed sedimentation and suggest a record of varying sediment influx from suspension settling and underflows (Smith & Ashley 1985). Clasts with striated facets and plucked ends are interpreted as glacigenically modified (Sharp 1982) and suggest an ice-proximal sediment source. Correspondingly, lonestones within laminated and massive silt and sand deposits (Fig. 4C) are interpreted as dropstones (Lønne 1995) and their occurrence indicates that lake bed sedimentation took place in an ice-marginal environment with water deep enough to induce calving (Pelto & Warren 1991; Boyce *et al.* 2007; Tsutaki *et al.* 2011). Abundant dropstones within massive silt (Fig. 4B) record iceberg rollover events (cf. Winsemann *et al.* 2004).

The inclined gravel deposits (Fig. 4G) are consistent with alluvial fan progradation (Blair & McPherson 1994) and their increased occurrence in the south of the study area (site 10, Figs 1B, 3) implies that they record deposition during the decanting of gLK into gLP (unnamed lake, Alden 1953). These inclined gravels may also record deltaic deposition, however no topsets have been identified, so an interpretation of alluvium, or potentially expansion bar deposits, is preferred. Thus, the lake bed benches and bench segments confirm the northward extension of gLP through the Purcell Trench and suggest that it was an ice-contact, proglacial lake.

The Purcell Lobe as an ice dam

A series of elevated gravel benches occupy interfluves along the valley walls of the Purcell Trench (kame sediment, Fig. 1B). The gravel benches are ≥6-16 m thick and reach elevations of 600-725 m a.s.l. They are typically composed of normally-graded beds of planar-stratified and trough cross-stratified sand and gravel lithofacies (Fig. 4E) that occasionally exhibit faulting. Small pebbles to large cobbles are typically well rounded and imbricated with gravel fabrics revealing predominantly a(t) pebble orientations and valley-parallel southerly and northerly palaeoflow directions (stereograms K1-K5, Fig. 1B). Unconsolidated, poorly sorted sand clasts are occasionally preserved in the gravel with little evidence of rounding (Fig. 4D).

The normally-graded, planar- and trough cross-stratified gravel beds (Fig. 4E) are consistent with fluvial deposition in a gravel-bed stream (Miall 1977). Gravel fabrics record valley-parallel stream flows and traction transport (rolling; Brennand 1994) that is anomalous to modern topography. Furthermore, the northward flows are also irreconcilable with an interpretation of remnant advance outwash deposits and the elevation of these deposits (up to 725 m a.s.l.) place them ~200 m above the surface of the modern Kootenay River valley where post-gLP damming is unlikely (Peters 2012). Thus, the gravels are interpreted as kame terraces deposited against the valley walls by ice-marginal meltwater streams (cf. Terpiłowski 2007; Pisarska-Jamroży *et al.* 2010). This interpretation is further supported by the incorporation of unconsolidated, unrounded sand-clasts (Fig. 4D), which may record rapid deposition by turbidites in a deltaic ice-marginal environment (cf. Winsemann *et al.* 2018); alternatively, the angularity of these deposits and lack of cohesive material (e.g. silt) suggest that they could have been preserved during mobilisation and incorporation into the gravel because they were frozen (Menzies 1990). The distribution of the kame deposits (Fig. 1B) suggests the Purcell Lobe dammed northern flow of gLP at its maximum extent.

Glacial Lake Purcell reconstruction

Glacioisostatic adjustments (GIA) derived from CIS palaeolake plane data range from horizontal to ~2.1 m km^{-1} (Table 1). This dataset was assessed for outliers using 1.5x the inter-quartile range (low cut-off -0.5, high cut-off 3.1), which revealed that all the tilts assessed were mathematically relevant (all values fall between the limits defining outliers). GIA data were plotted against time (Fig. 5) to elucidate possible patterns in glacioisostatic behaviour during MIS 2 deglaciation. Glacial lakes Arrow and Invermere formed closest to gLP (within ~50 and 70 km of the Purcell Trench, respectively; Fulton *et al.* 1989; Sawicki & Smith 1992) and thus are more likely to have had rates of glacioisostatic rebound governed by similar lithospheric properties (Clague & James 2002) and the Clayhurst stage of glacial Lake Peace was chronologically nearest to the proposed dates for gLP (Mathews 1978; Table 1; Fig. 5). Together, these records suggest that potential GIA of gLP water planes ranged from 0 m km^{-1} to ~1 m km^{-1} but could have been as high as 2 m km^{-1} (Fig. 5). A projected array of tilted water planes (0.0, 0.5, 0.75, 1.0 and 1.25 m km^{-1}) produces a range of modelled gLP extents (Fig. 6) that encompass all but three of the previously reported GIAs with age constraints (Fig. 5). The calculated volume of gLP for the modelled array of tilts (assuming a steep Purcell Lobe ice margin and a position consistent with estimates by Dyke *et al.* 2003) ranges from 40-142 km^3 (Table 2).

The intersection of the modelled lake extent and the geomorphology of the Kootenai River flood-related fan suggests that with low GIAs, gLP could have drained past the Elmira Spillway via a south-flowing stream that drained along the west side of the Purcell Trench (Fig. 6). However, sedimentary evidence suggests a steeper GIA may be more accurate. Exposures of the Kootenai River flood-related fan in the southern Purcell Trench (site 10, Figs 1B, 3) are topped by silt deposits (e.g. Fig. 4G) that suggest a lacustrine environment. Furthermore, the contiguous lake sediment benches recording the minimum extent of gLP (Fig. 1B) cover a larger area than modelled extents with low GIAs (Fig. 6). Thus, a steep GIA is deemed most likely to have influenced gLP's lake plane (i.e. at least 1.25 m km^{-1}; Fig. 6). A GIA of 1.25 m km^{-1}, like the tilt that affected glacial Lake Bretz (Table 1), would have resulted in a total gLP volume of 142 km^3 (Table 2). Its surface elevation against a northern ice dam would have been >800 m a.s.l. and it would have reached depths of >400 m (Table 2, Fig. 7).

GLOF evidence in the Kootenay River valley

The Kootenay River valley contains elevated (>100 m above the modern Kootenay River floodplain) sediment benches or terraces (GLOF sediment, Fig. 1B), and alluvial fans, which are remnants of a thicker valley fill (Figs 1B, 8). The highest truncated and bisected alluvial fan remnants reach elevations of up to ~675 m a.s.l. (Fig. 8B). The highest terrace occupies both sides of the Kootenay River valley with tread elevations from 642 m a.s.l. near the Purcell Trench to ~600 m a.s.l. near the Kootenay River valley confluence with the Columbia River valley (Fig. 8A, B) and a down-valley slope of ~1.5 m km^{-1}. The terraces typically contain normally-graded planar-stratified and trough cross-stratified sand, imbricated gravel with occasional massive gravel beds, and, occasionally, diffusely graded, sinusoidally stratified sand beds (Fig. 4I). Trough-cross stratified gravel is most common near the top of the valley fill, where smoothed and potholed bedrock also exists at elevations up to 624 m a.s.l. (Figs 1B, 8A). Type-a ripples overlain by type-s ripples (Ashley *et al.* 1982; Fig. 4F) are exposed in inset terraces lower than 560 m a.s.l. in the Kootenay River valley valley fill and towards the valley centreline (Fig. 1B). Gravel fabrics from planar-stratified gravels and type-a ripple measurements record westward palaeoflows through the Kootenay River valley (Fig. 1B, Stereograms D1, D2 and Dr1).

The normally-graded, planar and cross-stratified beds of sand and imbricated gravel with westward palaeoflows in the terraces are compatible with observations of sand and/or gravel dune or bar formation (e.g. Carrivick *et al.* 2004; Rushmer 2006; Russell 2009) during a confined GLOF from gLP. Diffusely graded, sinusoidally stratified sand records stationary and breaking antidunes, deposited rapidly from supercritical hyperconcentrated GLOF flows (cf. Lang & Winsemann 2013; Lang *et al.* 2017). Occasional massive gravel beds are associated with deposition of traction load during GLOFs (Carrivick *et al.* 2004; Russell 2009) or rapid deposition (Rushmer 2006). Ripples in fine sediments have been interpreted to record deposition during waning GLOF flows (Russell 2009) or hydraulic damming during the latter stages of floods (cf. Touchet Beds in the Channeled Scabland, Waitt 1980). Together, this sedimentary sequence is interpreted to record a period of sand and gravel aggradation during energetic GLOF flows through the Kootenay River valley, followed by fine sediment deposition during waning flood flows or in areas of local hydraulic damming (cf. Winsemann *et al.* 2016).

The Kootenay River valley highest terrace, just below the alluvial fan remnants, is interpreted to record the pre-GLOF valley bottom because the relative prevalence of alluvial fans that have aggraded onto it (Fig 8B) suggests this surface was exposed for a significant amount of time. This terrace's elevation suggests a pre-incision down-valley slope of ~1.5 m km^{-1} and a pre-flood valley bottom elevation of ~640 m a.s.l. near the Purcell Trench (i.e., Kootenay Lake, Fig. 8A). Terrace geomorphology indicates that the valley fill (pre-GLOF and GLOF sediments) has experienced ~110 m of incision near the Purcell Trench and ~150 m near the Columbia River valley (Fig. 8).

Although it is likely that the highest terraces were formed during erosive stages of a gLP GLOF, it is unknown exactly which terraces formed by channelising GLOF flows, and which by postglacial fluvial incision. However, recent geomorphic examinations suggest the bedrock around Cora Linn Dam (~530 m a.s.l., Figs 1B, 8) was sculpted by energetic flows (Waitt 2016). The potholed bedrock near the top of the valley fill sequence (624 m a.s.l., Fig 8A) also suggests bedrock erosion during a period of energetic flow that likely removed pre-GLOF valley fill sediment (Zen & Prestegaard 1994; Fig. 8A). Taken together, these observations suggest that the rising stage of a gLP GLOF was responsible for alluvial fan truncation, the creation of elevated potholes (624 m a.s.l.), removal of pre-GLOF sediment to bedrock (at least 90 m thickness at Corra Linn Dam), and bedrock sculpting around Corra Linn Dam (Waitt 2016). In this scenario, the terraces formed after GLOF aggradation during incision by waning-stage channelising GLOF flows or post-glacial river flows. More conservatively, the rising stage of the GLOF may only have been responsible for alluvial fan truncation, elevated pothole formation and partial incision through valley fill. In this scenario, waning-stage channelizing GLOF flows and post-glacial river flows may have been responsible for the bulk of sediment removal, terracing and the bedrock sculpting around Corra Linn Dam.

The large, fan-shaped landform situated at the confluence of the Kootenay River valley and the Columbia River valley (Playmor Junction, Fig. 1B) exhibits multiple terraces (Fig. 8C) and reaches an elevation of approximately 490 m a.s.l., which is ~110 m lower than the highest terrace treads recording the pre-GLOF valley bottom in the western Kootenay River valley (Fig. 8A). It is largely composed of very coarse, poorly-sorted, clast-supported massive gravel with abundant well-rounded boulders (up to 0.5 m b-axes) (Fig. 4H). Clast a-axes are randomly oriented and gravel fabric measurements have a polymodal distribution (D3, Fig. 1B). The coarse, fan-building gravels are capped by ~10 m of trough cross-stratified sand and cobble gravel (Peters 2012).

The fan shape and composition of the gravel deposit at Playmor Junction (Fig. 1B) suggests deposition as a large alluvial fan (Blair & McPherson 1994) or GLOF expansion bar (e.g. Baker 1984; Benito 1997). The well-rounded gravel that comprises the fan indicates fluvial mobilisation and its massive, poorly-sorted structure with boulder-sized clasts is consistent with deposition during high-energy flows associated with GLOFs (Cenderelli & Wohl 2003; Harrison *et al.* 2006; Russell 2009). The polymodal fabric (D3, Fig. 1B; cf. Morison & Hein 1987; Meetei *et al.* 2007) with randomly-oriented clast a-axes, poorly sorted texture, and massive structure of the gravel suggests rapid deposition in a turbulent, possibly hyperconcentrated flow (cf. Brennand & Shaw 1996). This boulder gravel is coarser than any gravels observed in the Kootenay River valley, suggesting deposition during the most energetic GLOF flows responsible for pothole erosion, alluvial fan truncation, and pre-GLOF valley-fill incision in the Kootenay River valley. The capping, trough cross-stratified sand and gravel indicates a subsequent period of braided stream or alluvial fan deposition (cf. Allen 1983; Blair 1987; Kjær *et al.* 2004) or possibly deposition during waning flood flows (Winsemann *et al.* 2016) and the multiple terraces preserved on the fan suggest incremental incision by fluvial erosion over time (Fig. 8C). Thus, we interpret the Playmor Junction deposit to most likely have been deposited as a GLOF expansion bar that was capped by alluvium and subsequently incised by the postglacial Kootenay River. The ~110 m height difference between its highest surface (~490 m a.s.l.) and the local pre-GLOF valley bottom (~600 m a.s.l.; Fig. 8A) indicates deposition after an initial period of erosive GLOF flows.

Discussion

<u>Summary of gLP evolution</u>

Sedimentary and geomorphic evidence records gLP evolution from a small ice-contact proglacial lake (described by Alden 1953) to a large, valley-filling ice-contact proglacial lake that drowned >100 km length of the Purcell Trench (Fig. 2). At one stage in its evolution, gLK (Alden 1953) decanted much of its volume into gLP (Fig. 2B). This drainage is recorded by the flood-related fan (Alden 1953) that has since been incised by the Kooteani River (Figs. 1B, 6). This fan formed the southern shore of gLP and forced southward drainage towards the Elmira spillway against the western valley wall (Fig. 6). The northern extent of gLP was dictated by the northward-retreating Purcell Lobe of the CIS (Figs 2, 6).

No direct geochronological ages are available for gLP or gLK. However, tephrostratigraphic ages that constrain the geomorphic evidence for Columbia River valley flood flows provided by Waitt *et al*. (2009) can provide some tentative, preliminary gLP chronology. The dune-scale bedforms described by Waitt *et al*. (2009) and Waitt (2016) were deposited <13.5 cal. ka BP, which is younger than the final drainage of other regional palaeolakes (Missoula, Columbia, and Bonneville), but compatible with our reconstructions of gLP and reconstructions of the deglacial Purcell Lobe's ice margin, which was near the Kootenay River valley by ~12 cal. ka BP (Dyke *et al*. 2003; Fig. 6). Thus, in the absence of independent dating, gLP sedimentation is assumed to have occurred during MIS 2 with a maximum possible age between 14 750±375 and 15 200±400 cal. a (LLGM; Atwater 1987). Drainage most likely occurred after ~13.5 cal. ka BP (Waitt *et al.* 1994, 2009; Dyke *et al*. 2003).

GLP volume and local glaciological impacts

Most of the lake bed sediments that comprise the contiguous benches in the southern Purcell Trench are not significantly deformed, suggesting deposition in an ice-distal environment without remobilisation from removal of buried ice. This suggests previous hypotheses of a low-volume, ice-marginal, or supraglacial lake that formed during CIS downwasting in the Purcell Trench are incorrect (e.g. Alden 1953; Eyles & Clague 1991; Ryder *et al*. 1991). Furthermore, iceberg rain out sediment interpreted in this study suggests that gLP's lake depth (>400 m, Table 2) was enough to force ice-marginal flotation and induce calving retreat through the Purcell Trench (Carrivick & Tweed 2013). Thus, the Purcell Lobe would have likely formed a steep terminus (Fig. 7) and been unable to displace significant amounts of gLP volume. This interpretation is compatible with the lack of ice-marginal landforms (kame terraces and moraines) in the southern Purcell Trench. Furthermore, kame terrace deposits (Figs 1B, 6D, E) confirm that the Purcell Lobe was sufficiently sealed to the Purcell Trench valley-wall for a period that allowed at least 16 m of glaciofluvial deposition. The relatively low position of the kame terraces within the valley (600-725 m a.s.l.) indicates that the seal existed late in the deglaciation of the Purcell Lobe. This seal, although not likely to be concurrent with the lacustrine deposits of gLP (based on elevation, Table 2), provides evidence that the Purcell Lobe could have dammed large volumes of water long after the CIS margin retreated northward into British Columbia. This evidence of a high-volume gLP elucidates important potential for large flood flows late in CIS deglaciation.

A deep, high-volume gLP (>400 m, almost 150 km^3, respectively, Table 2) would have held a similar amount of water as modern Lake Tahoe, or ~30% more than the Dead Sea. Such a lake would have mechanically exacerbated local CIS mass loss through calving, thereby steepening the ice margin causing increased ice flow velocities (Carrivick & Tweed 2013). Mass loss would also have been accelerated in the Purcell Trench by thermal erosion, because ice-marginal lakes deliver heat to glacier termini. Such thermal erosion can undercut the ice margin at the water line (e.g. Kirkbride & Warren 1999; Röhl 2006) further steepening the terminus and intensifying calving retreat. These feedbacks suggest that the deglacial Purcell Lobe would have had a steep ice-front prior to gLP drainage (Fig. 7).

GLP drainage and impacts on landscape evolution

GLP was confined to the Purcell Trench until the Purcell Lobe's calving margin retreated sufficiently northward to allow drainage into the Kootenay River valley (Figs 1B, 2). At this time, in order to drown gLP lake bed sediments and the flood-related fan in the southern Purcell Trench (Alden 1953; Fig. 2B), the gLP water surface was most likely tilted ~1.25 m km^{-1}, relative to the modern landscape (Fig. 6). Applying this tilt to a modelled gLP lake plane results in a surface elevation against the ice dam near the Kootenay River valley of 817 m a.s.l. and a drainable volume of ~116 km^3 (Table 2). This elevated lake surface is ~180 m above the pre-GLOF valley bottom in the Kootenay River valley (Table 2, Fig. 8A), suggesting that gLP water likely drained suddenly into the Kootenay River valley following catastrophic ice-dam failure.

The ~180 m elevation difference between the gLP water surface and the pre-GLOF valley bottom in the Kootenay River valley (Table 2) suggests that this sudden drainage would have generated extremely high specific and total stream powers, capable of eroding large amounts of boulder-sized sediment (cf. Cenderelli & Wohl 2003). This erosive GLOF is recorded by truncated alluvial fans and potholed bedrock ~140 m and ~90 m, respectively, above the modern Kootenay River valley floor (cf. Winsemann *et al.* 2016). If the fluvially-eroded bedrock described by Waitt (2016) at the Corra Linn Dam is attributed to a gLP GLOF, >90 m depth of sediment would have been removed by the GLOF at this location, which is comparable to previous models of GLOF erosion (e.g. Winsemann *et al.* 2016; Lang *et al.* 2019; Fig. 8A). After incising the Kootenay River valley fill, the flood flows debouched into the larger Columbia River valley at Playmor Junction, where a large, fan-shaped expansion bar was formed from cobble- and boulder-sized bedload (cf. Baker 1984; Benito 1997; Figs 1B, 4H, 8C).

The flood flows generated by gLP drainage would have entered the Columbia River valley via the Kootenay River valley (Figs 1B, 8C). Whether or not these flows would have been capable of enough geomorphic work to have formed dune-scale bedforms in the Channeled Scabland near Chelan Falls, Washington (Waitt *et al.* 1994, 2009) depends on flow attenuation along the ~500-km long flood route (defined by the lengths of the modern Kootenay River and Columbia River from the Kootenay confluence). However, because gLP likely drained after ~13.5 cal. ka BP (Kuehn *et al.* 2009; Waitt *et al.* 2009), drained a large volume of water (~116 km^3, Table 2), and likely drained suddenly (based on the modelled ~180 m elevation difference between the gLP water surface and the top of the highest terrace in the Kootenay River valley), it is possible that gLP flood flows induced late-Pleistocene geomorphic changes in the Channeled Scabland (Fig. 1A; cf. Waitt *et al.* 2009); however, hydraulic modelling should be performed to assess this (e.g. Winsemann *et al.* 2016).

Regardless of the role the gLP GLOF may have played in the Channeled Scabland, its regional effects on postglacial fluvial systems are evidenced geomorphically and sedimentologically. The low (<560 m a.s.l.) terraces located along the Kootenay River valley (Fig. 8A, B) likely record postglacial (late-Pleistocene and Holocene) fluvial incision by the Kootenay River, which remobilised the waning flood-flow deposits towards the Kootenay River valley confluence with the Columbia River valley (Figs 1B, 8C). As the sediment-laden Kootenay River exited the narrow Kootenay River valley, it deposited its bedload as ~10 m of trough cross-stratified sand and gravel alluvium (Peters 2012) over the surface of the boulder-gravel expansion bar (cf. Kehew *et al.* 2010). Finally, when the Kootenay and Columbia rivers neared their modern elevations and the Kootenay River reached the Kootenay River valley's bedrock and/or its specific sediment yield relaxed following postglacial incision (Church & Slaymaker 1989), alluvial deposition over the expansion bar was replaced with incision, forming extensive fluvial terraces (Fig. 8C).

Conclusions

- GLP was a large (~1 152 km^2, 142 km^3) ice-contact proglacial lake that most likely reached water depths of >400 m. This deep water induced calving retreat along the Purcell Lobe terminus, evidenced by iceberg rain-out deposits and dropstones within the gLP lakebed sediments. This evidence contradicts previous hypotheses that propose stagnant ice filled the valley limiting lake volume.

- Kame terraces were formed by ice-marginal stream deposition along the flanks of the deglacial Purcell Lobe north of the Kootenay River valley, indicating that an ice-valley wall seal was maintained throughout much of CIS deglaciation in the Purcell Trench. This suggests that the Purcell Lobe could have effectively dammed gLP within the Purcell Trench without allowing significant gradual drainage into the Kootenay River valley.

- The Purcell Lobe's terminus was altered mechanically by its calving margin and thermally by heat exchange with gLP. These processes likely exacerbated the northward rate of Purcell Lobe retreat and formed a steep ice front in the Purcell Trench. This steep ice margin dammed the northern extent of gLP prior to its drainage into the Kootenay River valley.

- The gLP lake surface was likely >800 m a.s.l. against its dam prior to its final drainage into the Kootenay River valley after 13.5 cal. ka BP, which is ~180 m above the top of the Kootenay River valley's pre-GLOF valley fill. This height discrepancy suggests gLP could have drained 116 km^3 of water into the Columbia River via the Kootenay River valley. This large volume of water likely drained suddenly following catastrophic ice dam failure.

- The initial flood flows caused by the gLP GLOF may have eroded up to ~150 m of pre-existing sediment from the Kootenay River valley, scouring to bedrock in places and producing an expansion bar at its junction with the Columbia River, before depositing GLOF sand and gravel in the Kootenay River valley.

- The timing of gLP drainage into the Kootenay River valley (based on CIS reconstructions) is compatible with tephrostratigraphic age constraints from dune-scale bedforms along the Columbia River at Chelan in the Channeled Scabland, Washington. Considering that gLP drainage likely supplied a >100 km^3 pulse of water into the Columbia River system, it is conceivable that this GLOF formed these Channeled Scabland dune-scale bedforms; however hydraulic modelling of flow attenuation should be performed to verify this hypothesis.

- Following the catastrophic drainage of gLP, the Kootenay River incised into the GLOF sediments, leaving a series of terraces formed by the GLOF and later, postglacial fluvial incision at elevations from ~600 – 550 m a.s.l. (~20 – 60 m above the modern river). A ~10-m thick deposit of alluvium was deposited over the expansion bar at the confluence of the Kootenay River valley with the Columbia River valley, which was also incised as the Kootenay River approached its modern elevation, leaving a series of fluvial terraces.

- Overall, these findings suggest that previous hypotheses favouring stagnant ice during CIS deglaciation may underestimate the potential hydrological impacts of transient, late-deglacial lakes. Furthermore, it seems likely that CIS GLOFs may have effected changes in the Channeled Scabland after glacial lakes Missoula and Columbia had drained.

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Tables

Table 1: Previously reported differential glacioisostatic adjustments associated with the CIS during MIS 2 assessed for this study.

Tilt[1] (m km^{-1})	Location (glacial lake name)	Distance from LLGM limit[2] (km)	Upslope direction	Reported age	Calibrated age range (median) cal. ka BP[3]	Reference
0	Rocky Mountain Trench (Invermere)	120	NA	10±0.14 (ka BP)	11.2 – 12.0 (11.6)	Sawicki & Smith (1992)
0.2	SW Yukon Territory (Champagne)	<100	South	Unknown	NA	Gilbert & Desloges (2006)
~0.4	Peace Basin (Peace, Clayhurst stage)	100	West	11.6 (ka BP)	13.1 – 13.7 (13.4)[4]	Mathews (1978)
0.46	Peace Basin (Peace Phase II)	100	230.9°±5°	<16.0±2.5 (optical)	NA	Hickin *et al.* (2015)
0.6	Shuswap Basin (Shuswap, Rocky Point stage)	200	East	Unknown	NA	Fulton (1969)
0.7	Okanogan Valley (Penticton, B.X. stage)	140	North	8.41±0.1 (ka BP)	9.1 – 9.5 (9.4)	Nasmith (1962)
0.85	Puget Lowland (Russell-Hood)	20	North	13.7 (ka BP)	16.1 – 17.0 (16.6)[4]	Thorson (1989)
0.9	Columbia River valley ("glacial Lake Arrow")	100	North	10.1±0.15 (ka BP)	11.2 – 12.2 (11.7)	Fulton *et al.* (1989)
1.15	Puget Lowland (Bretz, western)	10	North	13.5 (ka BP)	15.8 – 16.8 (16.3)[4]	Thorson (1989)
1.6 ± 0.7	Nicola Basin (Hamilton, lower stage)	200	North (347°±7°)	Unknown	NA	Fulton & Walcott (1975)
1.7 ± 0.4	Thompson Basin (Deadman, lowest stage)	200	Northwest (321°±6.1°)	~10.9 – 11.1 (cal. ka BP)	10.9 – 11.1 (11.0)	Johnsen & Brennand (2004)
1.8 ± 0.6	Nicola Basin (Merritt)	200	Northwest (341°±18°)	Unknown	NA	Fulton & Walcott (1975)
1.8 ± 0.6	Nicola Basin (Hamilton, upper stage)	200	North (354°±11°)	Unknown	NA	Fulton & Walcott (1975)
1.8 ± 0.7	Thompson Basin (Thompson, high stage)	200	Northwest (332°±9.9°)	>10.9 – 11.1 (cal. ka BP)	10.9 – 11.1 (11.0)	Johnsen & Brennand (2004)
~2.1	Nicola Basin (Quilchena)	200	Northwest	≳8 900	9.5 – 10.3 (10.0)	Fulton & Walcott (1975)

[1] Land surface tilt caused by differential glacioisostatic adjustment, as recorded by palaeo-lake-level indicators.
[2] Approximate distances between the glacial lake's nearest margin of the CIS during the LLGM from Fulton *et al.* (2004). GLP was ~40 km from the LLGM limit, perhaps <13.5 cal. ka BP (Waitt *et al.* 2009).
[3] Radiocarbon ages calibrated for this study with Calib software version 7.10 (Stuiver & Reimer 1993) using the IntCal13 radiocarbon curve (Reimer *et al.* 2013). Reported as the 2σ median probability (e.g. Peters *et al.* 2016).
[4] Calibration was performed using an assumed standard error of ±160 (the highest reported in this review, to avoid spurious precision) because insufficient information was reported.

Table 2: Dimensions (area and volume) of gLP and its ice dam for the tested range of glacioisostatic tilts.

Water-plane tilt (m km^{-1})	Lake area[1] (km^2)	Lake volume[1] (km^3)	Maximum drainable volume[2] (km^3)	Water surface elevation[3] (m a.s.l.)	Water depth[4] (m)	Minimum ice dam thickness[5] (m)	Height of water surface above pre-GLOF valley bottom[6] (m)
0	600	40	14	655	257	286	15
0.50	663	69	43	723	325	361	83
0.75	667	73	47	756	358	398	116
1.00	671	93	67	788	390	433	148
1.25	1152	142	116	817	419	466	177

[1] Location of ice dam placed at 49°36’55.7”N, 116°52’21.9”W (~132 km north of the Elmira spillway); a calving margin is assumed (see text).

[2] Lake volume minus the volume of Kootenay Lake within the palaeolake extent.

[3] Maximum elevations, as measured against the ice dam.

[4] Maximum depths, measured against the ice dam.

[5] Clean ice thickness required to resist flotation at the dam following the 9/10[ths] ratio of ice to water densities (Thorarinsson 1939; Fowler 1999).

[6] Height of water surface above the height of the highest terrace in the Kootenay River valley (640 m a.s.l.).

Figures

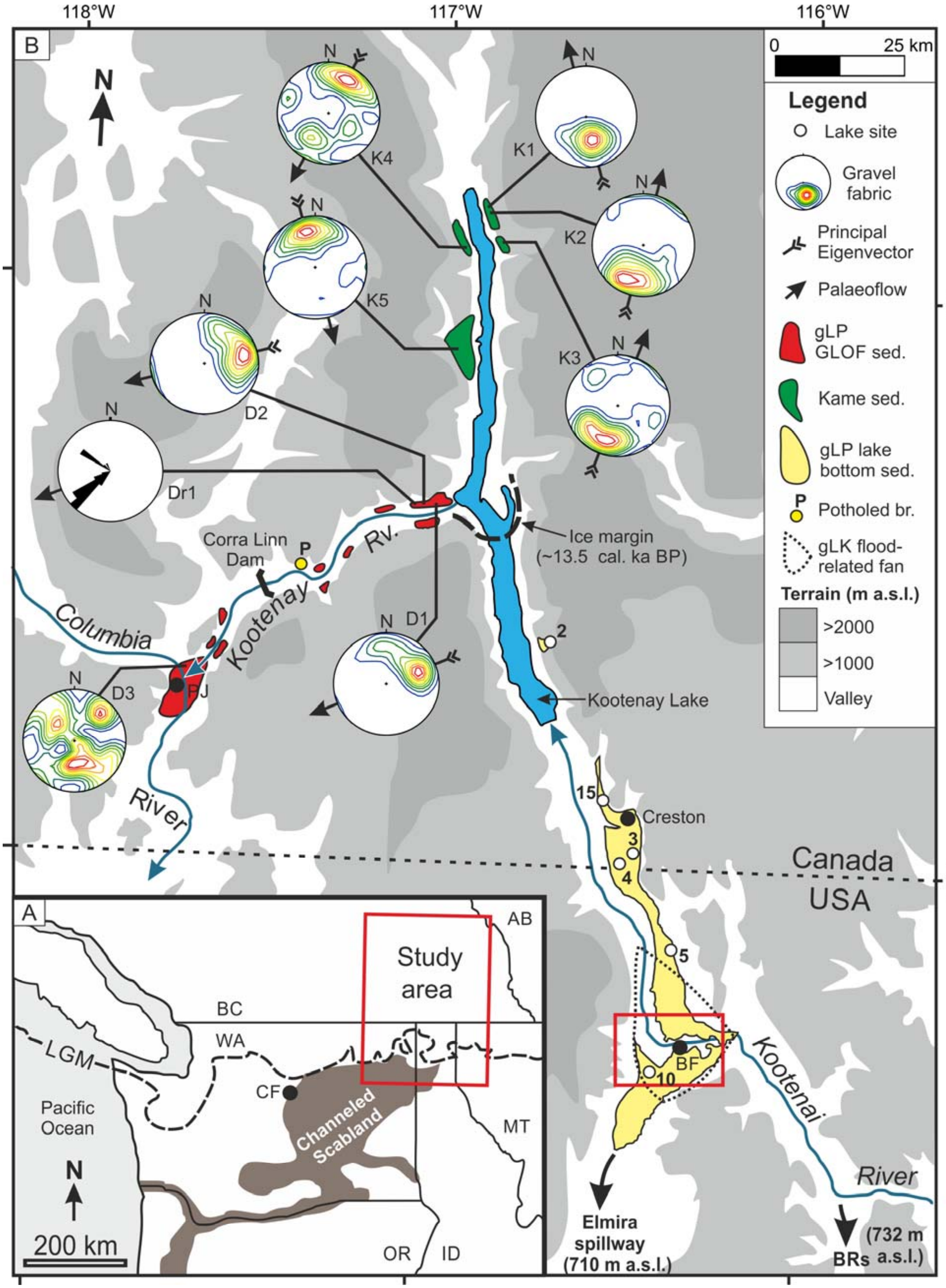

118°W
117°W
116°W
50°N
49°N
B
N
0
25 km
Legend
Lake site
Gravel fabric
Principal Eigenvector
Palaeoflow
gLP GLOF sed.
Kame sed.
gLP lake bottom sed.
P
Potholed br.
gLK flood-related fan
Terrain (m a.s.l.)
>2000
>1000
Valley
K1
K2
K3
K4
K5
D1
D2
D3
Dr1
Ice margin (~13.5 cal. ka BP)
Corra Linn Dam
Kootenay Rv.
Columbia
River
RJ
Kootenay Lake
2
15
Creston
3
4
5
10
BF
Canada
USA
Kootenai
River
Elmira spillway (710 m a.s.l.)
BRs
(732 m a.s.l.)
A
AB
BC
WA
LGM
Study area
CF
Channeled Scabland
Pacific Ocean
MT
OR
ID
200 km

Fig. 1: A. Study area (red box) at the western USA-Canada border showing the Channeled Scabland and CIS extent at the Last Glacial Maximum (LGM; Fulton *et al*. 2004). The Purcell Lobe's southern terminus at ~13.5 ka BP (dashed black line) is mapped after Dyke *et al*. (2003). CF = Chelan Falls. B. Schematic map of the study area depicting the locations of sedimentary deposits (kame; glacial Lake Purcell (gLP) lake bed sediments; gLP drainage sediments; gLP Glacial Lake Outburst Flood (GLOF) sediments; glacial Lake Kootenai flood-related fan), palaeoflow measurements (lower hemisphere, equal area projection and rose diagram) and other locations discussed in the text. Note that GLOF sediments are present throughout the Kootenay River valley, but only major exposures are mapped. Also, the West Arm of Kootenay Lake (east of the Corra Linn Dam) and the Kootenay River are jointly referred to in the text as the Kootenay River valley (KRv) and changes in river width between these water bodies are not represented to enhance illustrative clarity. Red box delineates the area shown in Fig. 3. BF = Bonners Ferry; PJ = Playmor Junction.

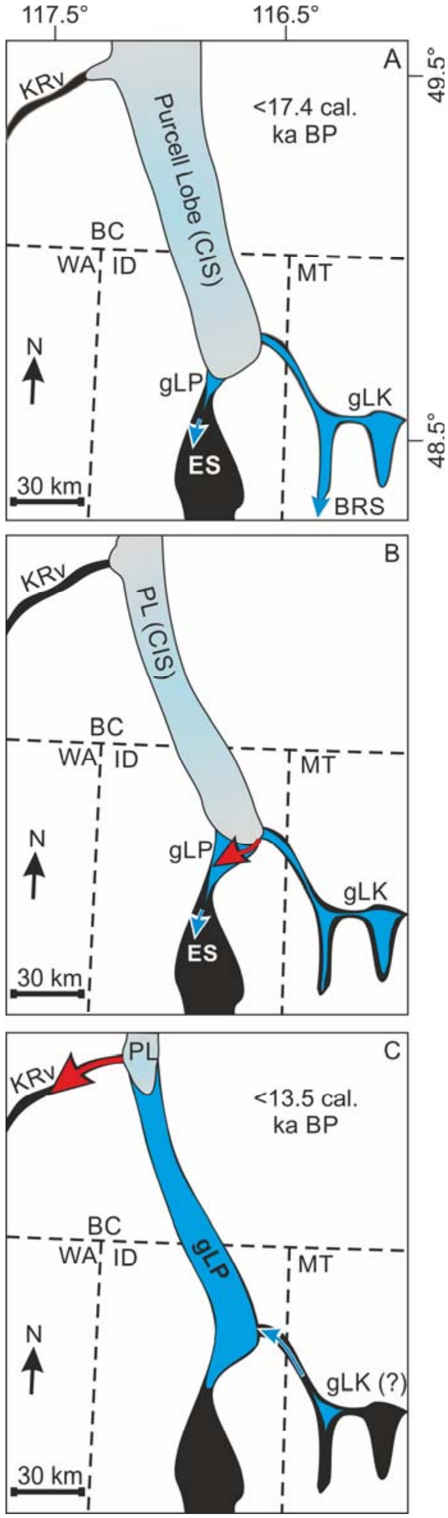


Fig. 2: Schematic representation of glacial lake evolution showing. A. Discrete stages of glacial Lakes Purcell (gLP) and Kootenai (gLK) (dark blue) impounded behind the Purcell Lobe (light blue, PL) sometime after the Last Late Glacial Maximum (LLGM) (likely <17.4 cal. ka BP;

Atwater 1987; Porter & Swanson 1998; Clague & James 2002). Both the Elmira spillway (ES) at 710 m a.s.l. and Bull River spillway (BRS) at 732 m a.s.l. are active (shown as blue arrows). During this stage Alden (1953) considered gLP an unnamed proglacial lake. B. Continued retreat of the Purcell Lobe causes the growth of gLP and its northward expansion. Eventually the portion of gLK above 710 m a.s.l. (the water above the Elmira spillway) decanted into gLP (shown as red arrow), rapidly reducing gLK's volume. C. Glacial Lake Kootenai has largely drained into gLP and may not exist at all. The late deglacial Purcell Lobe retreats northward towards the Kootenay River valley (KRv) until it fails to dam gLP. GLP debouches into the Kootenay River valley (red arrow) and eventually its floodwaters reach the Channeled Scabland via the Columbia River (Fig. 1). Note that the naming scheme of Alden (1953) is abandoned and replaced with the one depicted by this schematic because it ignores gLP's nascent formation and considers later stages of gLP to be the same lake as gLK. BC = British Columbia; WA = Washington; ID = Idaho; MT = Montana.

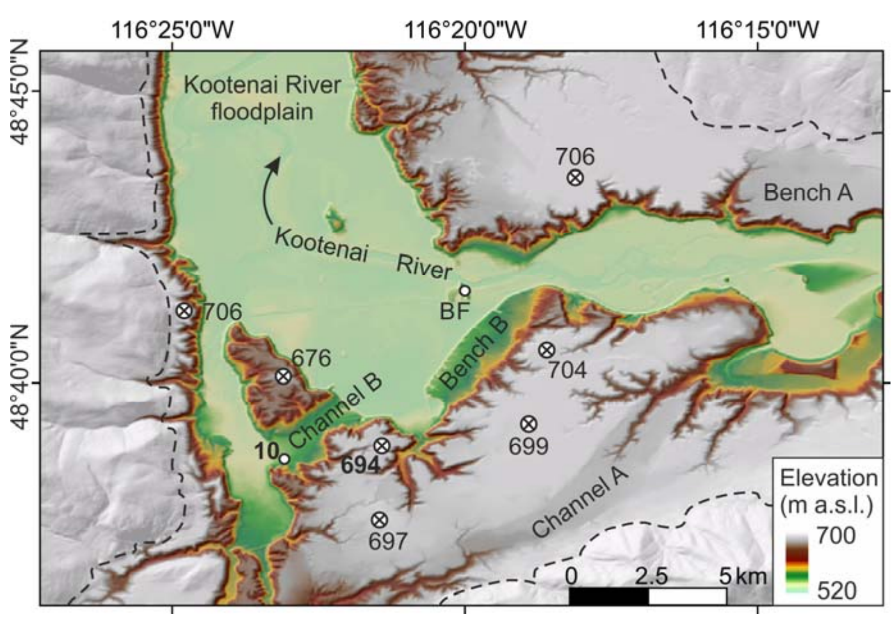


Fig. 3: Hillshaded DEM from a composite of Geobase (Government of Canada 2019) and National Elevation Dataset (United States Geological Survey) data (USGS 2019) revealing benches A and

B incised into the gLP lake bed deposits. Dashed lines delineate the contact between the lake bed sediment and the valley walls. Point elevations (white dots with Xs) that typify elevation data used to reconstruct the pre-incision gLP lake bed are provided and highlight the relative flatness of the deposit's surface across the Purcell Trench. Channels 'A' and 'B' correspond in elevation with benches 'A' and 'B', respectively. Site 10 is shown as a labelled white dot within Channel B.

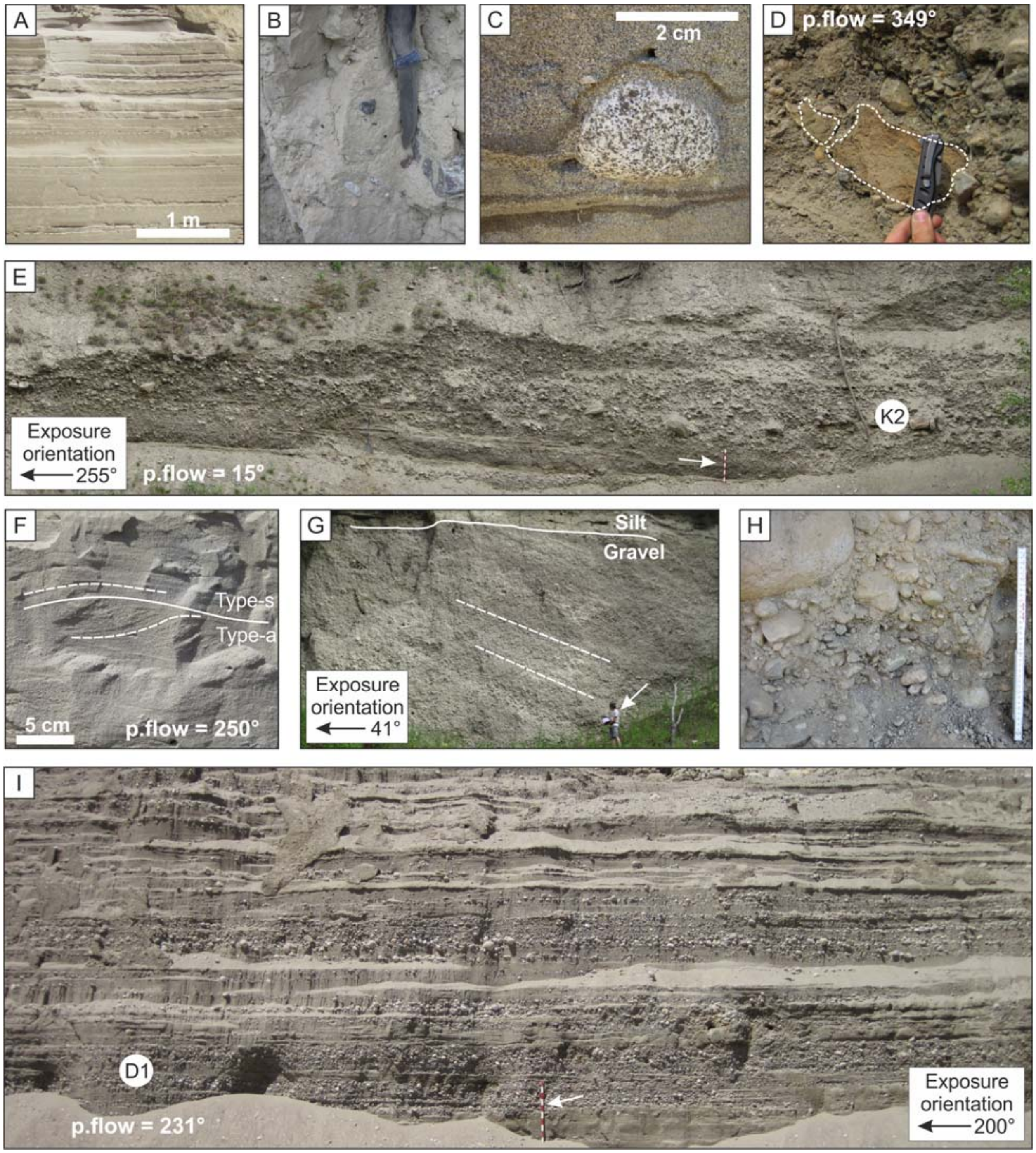


Fig. 4: Examples of sediment deposits. Sites are located in Fig. 1B. A. Massive and laminated silt deposits are common throughout the lake bed bench (photograph is from site 5). B. Massive silt with abundant, outsized clasts (lonestones) at site 15. Knife blade is ~9 cm long. C. Massive, coarse sand laminae, interlaminated with silt and clay laminae that conformably overlie a well-

rounded, granitic lonestone at site 5. D. Angular, unconsolidated sand clasts (outlined by white dashed lines) in an interfluve-occupying, valley-wall deposit (kame) near gravel fabric K1. Knife handle is ~9 cm long. E. Planar-stratified sand and gravel in a kame terrace. Arrow marks metre stick for scale. Location of gravel fabric K2 is shown as a labelled white dot. F. Climbing ripples (after Ashley *et al.* 1982) measured for palaeoflow Dr1 from a ~560 m a.s.l. gLP GLOF terrace in the Kootenay River valley. G. Inclined gravel beds (white dashed lines highlight two lower contacts) overlain by massive silt at site 10 (Fig. 3). Arrow points to a person for scale. H. Poorly sorted cobble and boulder gravel at ~490 m a.s.l. in a gLP GLOF expansion bar measured for gravel fabric D3. Ruler is 36 cm long. I. Planar-stratified sand and gravel and diffusely graded, sinusoidally stratified sand composes a ~640 m a.s.l. gLP GLOF terrace in the Kootenay River valley. Location of gravel fabric D1 shown as a labelled white dot. Arrow marks metre stick for scale. p.flow is palaeoflow.

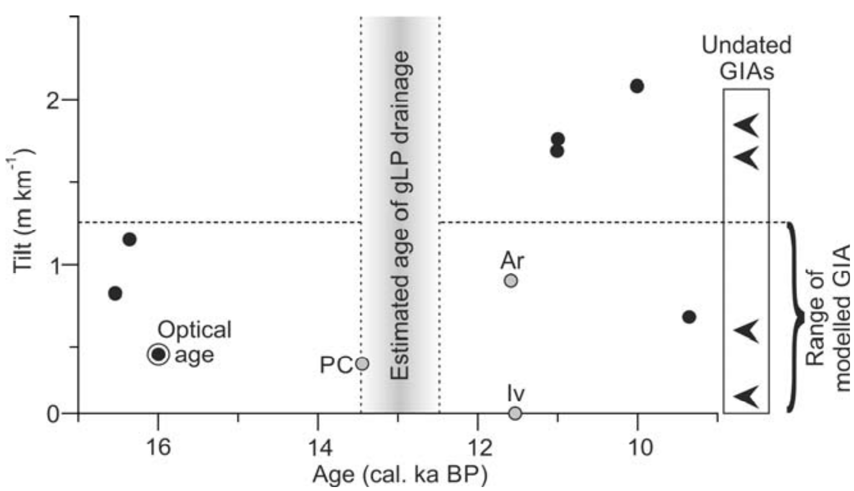


Fig. 5: Plot of the ten glacioisostatic tilts derived from CIS palaeolake planes with reported ages (black and grey dots, Table 1). Glacial lakes Arrow (Ar) and Invermere (Iv) are highlighted grey because they are geographically near the the Purcell Trench and Glacial Lake Peace, Clayhurst

stage (PC), because it's chronologically closest to putative ages for gLP (delineated by the labelled grey bar; Dyke *et al.* 2003; Waitt *et al.* 2009; Table 1). The four undated GIAs reviewed are marked as chevrons. The estimated age of gLP drainage is derived from previous CIS reconstructions (Dyke *et al.* 2003) and tephrochronologic ages (Waitt *et al.* 2009). The distribution of the previously reported lake tilts reveals the propriety of the modelled GIAs.

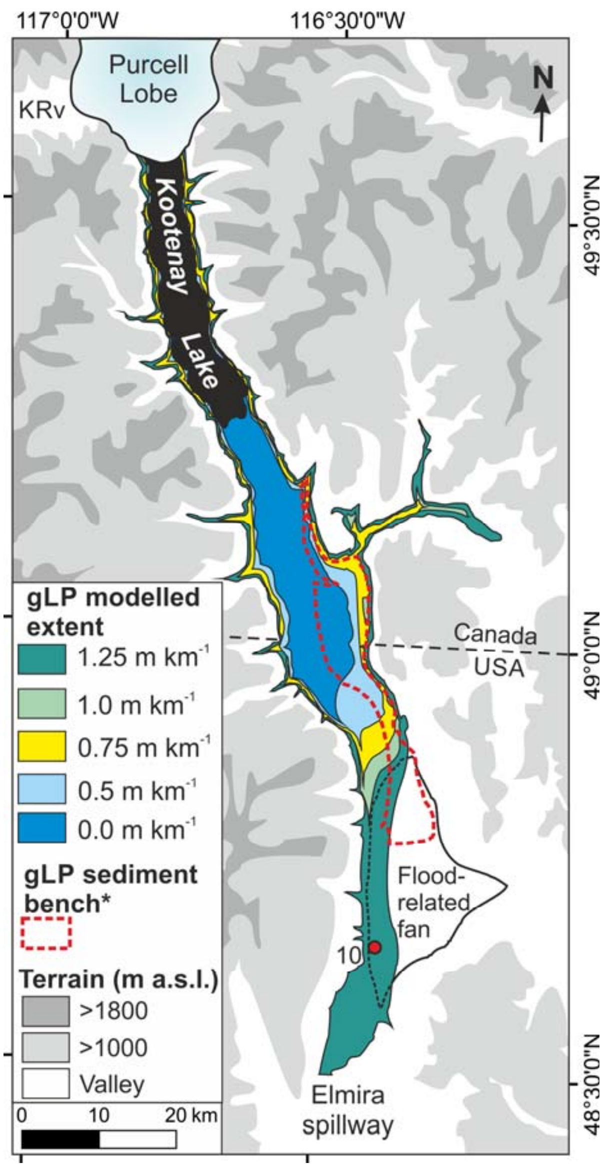


Fig. 6: Extent of gLP based on the array of tested glacioisostatic tilts (Table 2). Note that only the steepest tilt (1.25 m km^{-1}) covers the gLP lake bed sediment bench (red dashed line; Fig. 1A) and allows gLP to overtop the flood-related fan formed by gLK's drainage (Alden 1953) which is capped by silty lake bed sediments at site 10 (red dot). This suggests that the fan was partially inundated by gLP. The locations of modern Kootenay Lake (black), the Kootenay River valley (KRv), and the Purcell Lobe <13.5 cal. ka BP (Dyke *et al.* 2003) are also shown. Only the northernmost extent of the gLP lakebed is outlined for visual clarity with other aspects of this figure; see Fig. 1A for the complete extent.

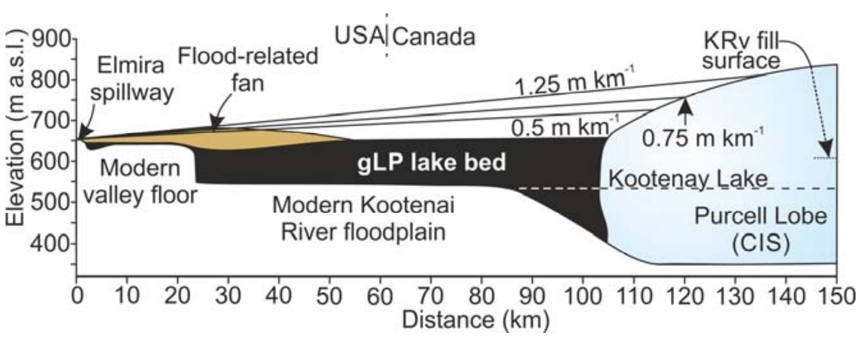
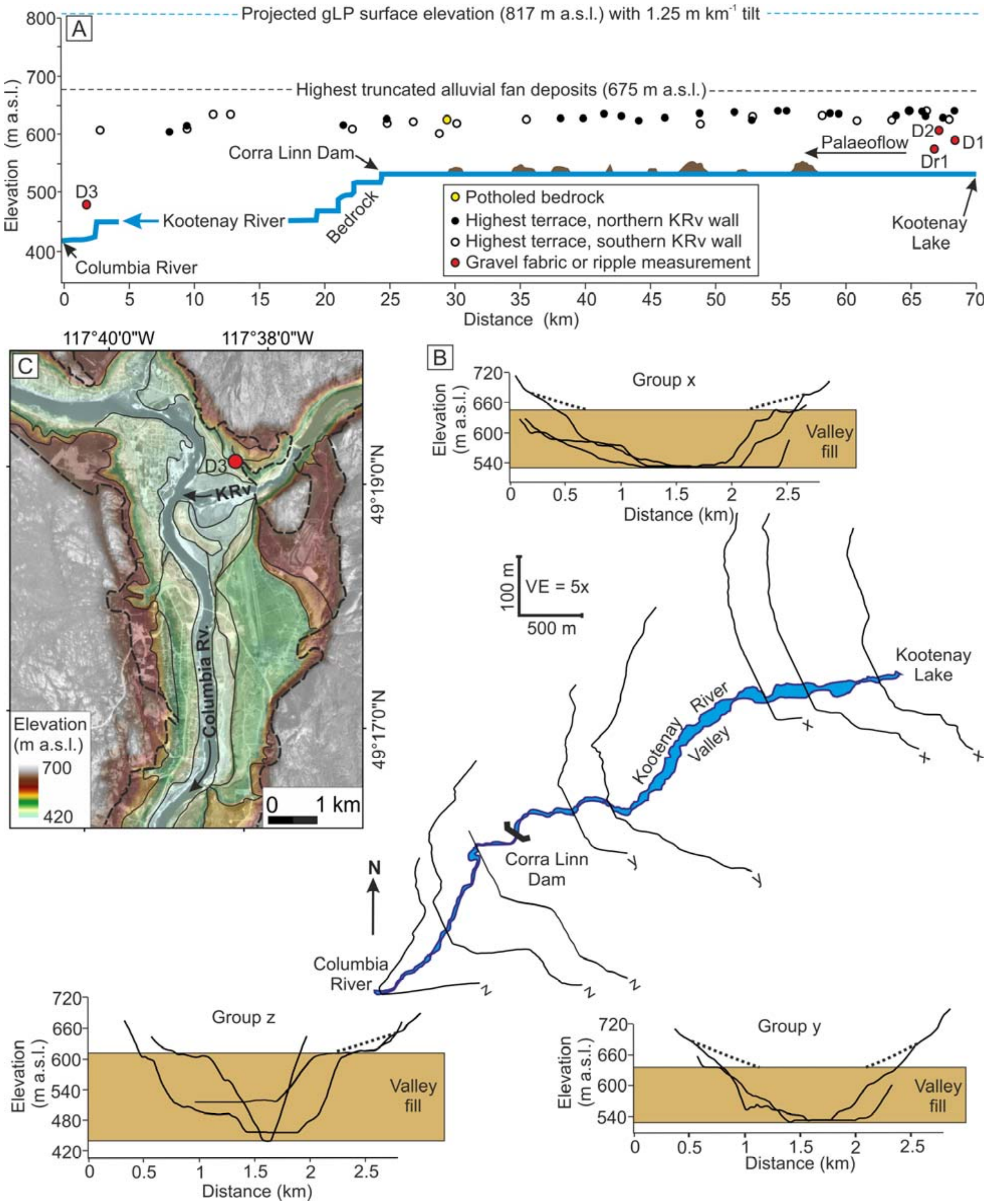


Fig. 7: Schematic representation of three modelled gLP water surface tilts (Table 2) projected northwards from the Elmira spillway, past the alluvial fan deposited by the drainage of gLK (Alden 1953) and over the reconstructed gLP lake bed. The elevations of the pre-incision Kootenay River valley sediment fill and modern Kootenay Lake are also depicted (note the depth of Kootenay Lake during glaciation is schematically represented).



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822 Fig. 8: Kootenay River valley (KRv; Fig. 1B) geomorphology. A. Kootenay River valley long

823 profile cartoon depicting the relationship between the most likely gLP elevation prior to drainage

(817 m a.s.l. at a 1.25 m km^{-1} tilt), the highest truncated alluvial fan deposits (~675 m a.s.l.), the highest valley-flanking terrace (depicting the pre-GLOF valley bottom), and the modern Kootenay River (bold blue line). Low-elevation, untruncated alluvial fans aggrading into the Kootenay River valley are shown as brown polygons. Locations of GLOF palaeoflow measurements and potholes are provided (see Fig. 1B for map view). B. Eight Kootenay River valley, cross-sectional profiles (from Geobase, Natural Resources Canada DEM; Government of Canada 2019), grouped to represent three reaches of the Kootenay River valley. The groups are shown with stacked profiles that reveal trends in bench and remnant fan elevations and allow estimations of pre-incision valley-fill elevations. Estimated extents of truncated alluvial fan remnants are shown as dotted lines. C. Terrace treads incised into an expansion bar at the Kootenay River valley (KRv)-Columbia River valley (Rv) confluence (see Fig. 1B for regional map) (DEM (Geobase, Natural Resources Canada) overlain by a georeferenced aerial photograph (National Aerial Photograph Library, Environment Canada)). Contacts between the valley-fill sediment and the bedrock valley walls are shown as dashed lines. Solid black lines outline individual terraces identified using the DEM. The location of gravel fabric D3 (Figs 1B, 6H) is shown as a labelled red dot.