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

#### 3 JARED L. PETERS AND TRACY A. BRENNAND

Peters, J. L. & Brennand, T. A.: Palaeogeographical reconstruction and hydrology of glacial Lake
Purcell during MIS 2 and its potential impact on the Channeled Scablands.

6 Large, ice-marginal lakes that were impounded by the maximally-extended Cordilleran Ice Sheet 7 (CIS) provided source waters for the extraordinarily large floods that formed the Channeled 8 Scabland of Washington and Idaho, USA. However, flood flows that drained CIS meltwater and 9 contributed to landscape evolution during later stages of deglaciation have hitherto been poorly 10 investigated. This paper provides the first evidence for such a late deglacial floodwater source: 11 glacial Lake Purcell (gLP). Sedimentary evidence records the northward extension of gLP from 12 Idaho, USA into British Columbia, Canada and establishes its minimum palaeogeographical 13 extent. Sedimentary evidence suggests that the deglacial Purcell Lobe was a capable ice dam that 14 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<sup>-1</sup>. 15 16 Sedimentary evidence suggests high lake plane tilts ( $\geq 1.25$  m km<sup>-1</sup>) are the most likely to have 17 affected gLP. Using this, the palaeogeography and volume of gLP are modelled, revealing that ~116 km<sup>3</sup> of water was susceptible to sudden drainage into the Channeled Scabland via the 18 19 Columbia River system. This calculation is supported by sedimentary and geomorphic evidence 20 compatible with energetic flood flows along the gLP drainage route and suggests gLP drained 21 suddenly, causing significant landscape change.

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26 Growing concerns over the stability of future hydrosphere-cryosphere interactions and our ability 27 to accurately predict the behaviour of modern glaciers and ice sheets (e.g. Bamber et al. 2009; 28 Gardner et al. 2013) highlight the importance of a complete understanding of Cordilleran Ice Sheet 29 (CIS) decay. Ice-marginal lakes play an important role in this improved understanding because 30 they affect ice dynamics (Carrivick & Tweed 2013), are effective sediment traps that record 31 detailed glacial histories (e.g. Larsen et al. 2011; Liermann et al. 2012), and are prone to 32 catastrophic drainage that can influence regional sediment transport and drainage systems (Korup 33 2012). Considering the important effects that glacial lakes can impose on ice sheet decay and 34 landscape evolution, and their increasing abundance and size along modern, deglaciating ice 35 margins (Carrivick & Tweed 2013), the importance of developing a thorough understanding of 36 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<sup>3</sup> glacial Lake 37 38 Missoula (O'Connor & Baker 1992; Miyamoto et al. 2006, 2007), formed when the southern 39 margin of the CIS disrupted regional drainage patterns (Baker 2009). Some of these lakes drained 40 catastrophically and contributed to the formation of the Channeled Scabland (Fig. 1A), a 41 megaflood landscape that geomorphically and sedimentologically records flood flows of nearly 42 unprecedented Earthly scale with maximum discharges of 10-20 Sverdrups (Benito & O'Connor 43 2003; Denlinger & O'Connor 2010). Whereas geological and sedimentological signatures of 44 enormous jökulhlaups (glacial lake outburst floods) entering the Channeled Scabland are abundant 45 and well documented (e.g. Bretz 1925, 1969; Baker 2009; Benito & O'Connor 2003), the potential 46 for post-LLGM flood flows from the drainage of glacial lakes in British Columbia has been 47 proposed (Shaw et al. 1999; Lesemann & Brennand 2009; Waitt et al. 2009; Waitt 2016) but 48 remains relatively poorly understood.

49 The role of ice-marginal lake formation within the Purcell Trench during CIS deglaciation has 50 received inconstant speculation. Alden (1953) first contemplated a glacial lake in the Purcell 51 Trench and its possible drainage into the Columbia River system via the Kootenay River valley 52 (Fig. 1B). Most researchers (e.g. Alden 1953; Johns 1970; R. Fulton, pers. comm. 2010) 53 speculated that the glacial lake in the Purcell Trench was shallow and primarily ice marginal or 54 supraglacial, owing to stagnant ice occupying the Purcell Trench. These authors also suggest that 55 glacial lake water in the Purcell Trench likely drained gradually past a spillway in the south (the 56 Elmira spillway) and the downwasting ice in the north (Fig. 1B). However, Waitt et al. (2009) 57 and Waitt (2016) propose that more energetic drainage of a proglacial lake in the Purcell Trench 58 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.

#### 65 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 72 composed of lake bed sediments attributed to gLK range in elevation from 700-740 m a.s.l. in 73 Idaho and from 730-762 m a.s.l. in Montana due to different spillway heights (Alden 1953). The 74 Bull River spillway (Fig. 1B) in Montana was the first flow to be activated and would have 75 commenced following a lowering of the final stage of gLM in the Clark Fork River valley to the 76 south of gLK (Alden 1953). After sufficient northward retreat of the Purcell Lobe, gLK decanted 77 into the southern Purcell Trench, forming a large flood-related fan on the valley floor and an 78 unnamed proglacial lake. Lake levels in the Purcell Trench were dictated by the Elmira spillway 79 (ibid). The geomorphology of the Elmira spillway suggests that its original height was ~710 m 80 a.s.l. and that incision from lake drainage is responsible for its current elevation of 655 m a.s.l. 81 (ibid).

82 The naming conventions used by Alden (1953) and adopted by Johns (1970) and Smith (2006) are 83 abandoned in this study because they ambiguously describe distinct water bodies with a single 84 name (gLK). Furthermore, the name 'glacial Lake Kootenay' employed by Waitt et al. (2009) is 85 not used, as its closeness to Alden's lake name is a potential source of confusion. Instead a naming 86 system is employed that distinguishes the discrete and possibly contemporaneous lakes that 87 occupied separate basins (Fig. 2). This new naming scheme retains Alden's glacial Lake Kootenai 88 moniker in Montana, USA (where most of his research was conducted) but designates the unnamed 89 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

95 rapid rise of the equilibrium line due to rapid climate amelioration. Sedimentary evidence for this 96 stagnation, and resultant downwasting, has been reported in the interior of British Columbia (Eyles 97 & Clague 1991; Ryder et al. 1991). During ice stagnation, glacial lake volume would have been 98 minimized by valley occupying ice. However marginal areas of the CIS may have experienced a 99 more complex pattern of decay (Fulton 1967) and these complications may have been further 100 exacerbated in mountainous terrain by late deglacial alpine ice advances (Lakeman et al. 2008). 101 Such complexities, along with potential inconsistencies in regional glacioisostatic response from 102 crustal heterogeneities (cf. Thorson 1989), may have enabled the formation of a deep, high-volume 103 gLP and highlight the need for investigations in the Purcell Trench.

104 Initial evidence for a high-volume, late-deglacial gLP has been supplied by Waitt et al. (2009) and 105 Waitt (2016), who suggest that a glacial lake in the Purcell Trench was a potential water source 106 for flood flow(s) in the Columbia River valley. Putative geomorphic evidence for post-Missoula, 107 late-Wisconsin Glacial Lake Outburst Floods (GLOFs) in the Columbia River valley includes two 108 megaflood bars marked by dune-scale bedforms ("giant current dunes") near Chelan Falls, 109 Washington (Waitt et al. 2009; Fig. 1A). These dune-scale bedforms are tephrostratigraphically 110 dated to <13.5 cal. ka BP (Kuehn et al. 2009), after the final drainage of glacial lakes Missoula 111 and Columbia and Lake Bonneville (Waitt et al. 1994, 2009). Age constraints on the deglacial 112 CIS are compatible with the tephrostratigraphic age of the dune-scale bedforms and place the 113 Purcell Lobe ice margin near the Kootenay River valley by ~13.5 cal. ka BP (Dyke *et al.* 2003).

#### 114 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 118 average width of ~6.5 km (Fig. 1B). Kootenay Lake's water surface elevation is controlled by the 119 Corra Linn Dam in the Kootenay River valley to an elevation of ~532 m a.s.l. (Davis 1920; Kyle 120 1938; Fig. 1B). Kootenay Lake marks a change in spelling from the Kootenai River to the 121 Kootenay River (Fig. 1B) and is essentially a stagnation point in the flow of the Kootenai/y River 122 along its circuitous westward route from the Rocky Mountain Trench, British Columbia through 123 the Columbia Mountains. Kootenay Lake drains out of its West Arm via the Kootenay River, 124 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). 125

126 Methods

### 127 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).

133 Sedimentary investigations entailed lithofacies identification, gravel fabric and ripple palaeoflow 134 measurements, and centimetre-scale logging of exposures. Sedimentary data are presented as 135 stereograms, rose diagrams, and exposure photographs. For each fabric analysis  $\geq 30$  clast a-b 136 plane attitude measurements (maximum dip and down-dip direction of the a-b plane) were taken in  $<0.5 \text{ m}^2$  areas of exposures. The data were then plotted on lower hemisphere, equal area 137 138 (Schmidt) diagrams as contoured stereonets using the cosine sums method (Stereo32 software) 139 with a cosine exponent of 20 (cf. Roeller 2008). Fabrics from kame deposits are designated K1, 140 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).

## 146 Palaeogeographical modelling of gLP

147 Typically, glacial lake extent is reconstructed from the distribution of lake bed sediments and by 148 correlating water-plane indicators (e.g., deltas, shorelines; cf. Johnsen & Brennand 2004). 149 However, a dearth of gLP water-plane indicators were identified on the steep bedrock valley walls 150 of the Purcell Trench. There is also a poor potential for lacustrine sediment preservation within 151 the floodplain of the modern Kootenai River and Kootenay Lake may cover significant areas of 152 gLP sediment (because they share a common basin). Thus, after confirming minimum extents 153 with sedimentary data, gLP extent and volume are estimated by assessing modelled lake surface 154 planes against limited evidence and in comparison to contemporaneous and geographically close 155 palaeolakes.

156 Two critical palaeogeographical elements enabled this modelling of gLP: (i) the reconstruction of 157 the pre-incision lake bed and (ii) the application of an appropriate glacioiosostatic adjustment 158 (GIA) to the water plane. These reconstructions were used to produce a combination of rasters 159 that were used in conjunction with modern topography to define gLP palaeogeography and 160 calculate its volume (cf. Leverington et al. 2002; DeVogel et al. 2004). The pre-incision palaeo-161 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 162 163 pre-incision lake bed surface were defined by its intersection with topographic highs on the

164 composite DEM of the Purcell Trench. A plausible range of GIA was derived from a survey of 165 previously reported glacioisostatic tilts for CIS glacial lake planes. The published ages of these 166 lakes are also reported, which are derived using disparate methods with varying accuracy. These 167 ages were recalibrated for this study with Calib software (Stuiver & Reimer 1993) to  $2\sigma$  confidence 168 using the IntCal13 radiocarbon curve (Reimer *et al.* 2013) to improve comparability of the ages 169 (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).

#### 176 **Results and interpretations**

#### 177 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). 185 The benches are composed of massive or laminated silt and clay rhythmites (sites 3, 4, 5, 15; Fig. 186 4A) containing occasional pebble- to cobble-sized clasts (lonestones) that display striated facets 187 and plucked ends (sites 5, 15). Lonestones occur in relatively high abundance within massive silt 188 (Fig. 4B). Rare deposits of massive, silty coarse sand, interbedded with silt and clay laminae that 189 drape lonestones are also present (Fig. 4C). Silt and clay rhythmites are exposed north of the 190 contiguous benches at an elevation of 675 m a.s.l. (site 2, Fig. 1B). Gravel deposits composed of 191 dipping (apparent 30° downwards dip towards 221°), normally-graded, tabular beds that reach 192 thicknesses of  $\geq 10$  m are also exposed in the Idaho sediment benches (Fig. 4G). These gravel 193 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.

199 Rhythmites record lake bed sedimentation and suggest a record of varying sediment influx from 200 suspension settling and underflows (Smith & Ashley 1985). Clasts with striated facets and plucked 201 ends are interpreted as glacigenically modified (Sharp 1982) and suggest an ice-proximal sediment 202 source. Correspondingly, lonestones within laminated and massive silt and sand deposits (Fig. 203 4C) are interpreted as dropstones (Lønne 1995) and their occurrence indicates that lake bed 204 sedimentation took place in an ice-marginal environment with water deep enough to induce calving 205 (Pelto & Warren 1991; Boyce et al. 2007; Tsutaki et al. 2011). Abundant dropstones within 206 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.

## 214 The Purcell Lobe as an ice dam

215 A series of elevated gravel benches occupy interfluves along the valley walls of the Purcell Trench 216 (kame sediment, Fig. 1B). The gravel benches are  $\geq 6-16$  m thick and reach elevations of 600-725 217 m a.s.l. They are typically composed of normally-graded beds of planar-stratified and trough 218 cross-stratified sand and gravel lithofacies (Fig. 4E) that occasionally exhibit faulting. Small 219 pebbles to large cobbles are typically well rounded and imbricated with gravel fabrics revealing 220 predominantly a(t) pebble orientations and valley-parallel southerly and northerly palaeoflow 221 directions (stereograms K1-K5, Fig. 1B). Unconsolidated, poorly sorted sand clasts are 222 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-

230 marginal meltwater streams (cf. Terpiłowski 2007; Pisarska-Jamroży et al. 2010). This 231 interpretation is further supported by the incorporation of unconsolidated, unrounded sand-clasts 232 (Fig. 4D), which may record rapid deposition by turbidites in a deltaic ice-marginal environment 233 (cf. Winsemann *et al.* 2018); alternatively, the angularity of these deposits and lack of cohesive 234 material (e.g. silt) suggest that they could have been preserved during mobilisation and 235 incorporation into the gravel because they were frozen (Menzies 1990). The distribution of the 236 kame deposits (Fig. 1B) suggests the Purcell Lobe dammed northern flow of gLP at its maximum 237 extent.

## 238 Glacial Lake Purcell reconstruction

239 Glacioisostatic adjustments (GIA) derived from CIS palaeolake plane data range from horizontal 240 to  $\sim 2.1 \text{ m km}^{-1}$  (Table 1). This dataset was assessed for outliers using 1.5x the inter-quartile range 241 (low cut-off -0.5, high cut-off 3.1), which revealed that all the tilts assessed were mathematically 242 relevant (all values fall between the limits defining outliers). GIA data were plotted against time 243 (Fig. 5) to elucidate possible patterns in glacioisostatic behaviour during MIS 2 deglaciation. 244 Glacial lakes Arrow and Invermere formed closest to gLP (within ~50 and 70 km of the Purcell 245 Trench, respectively; Fulton et al. 1989; Sawicki & Smith 1992) and thus are more likely to have 246 had rates of glacioisostatic rebound governed by similar lithospheric properties (Clague & James 247 2002) and the Clayhurst stage of glacial Lake Peace was chronologically nearest to the proposed 248 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<sup>-1</sup> to  $\sim$ 1 m km<sup>-1</sup> but could have been as high as 2 m km<sup>-1</sup> 249 250 (Fig. 5). A projected array of tilted water planes (0.0, 0.5, 0.75, 1.0 and 1.25 m km<sup>-1</sup>) produces a 251 range of modelled gLP extents (Fig. 6) that encompass all but three of the previously reported 252 GIAs with age constraints (Fig. 5). The calculated volume of gLP for the modelled array of tilts

253 (assuming a steep Purcell Lobe ice margin and a position consistent with estimates by Dyke et al. 254 2003) ranges from 40-142 km<sup>3</sup> (Table 2).

255 The intersection of the modelled lake extent and the geomorphology of the Kootenai River flood-256 related fan suggests that with low GIAs, gLP could have drained past the Elmira Spillway via a 257 south-flowing stream that drained along the west side of the Purcell Trench (Fig. 6). However, 258 sedimentary evidence suggests a steeper GIA may be more accurate. Exposures of the Kootenai 259 River flood-related fan in the southern Purcell Trench (site 10, Figs 1B, 3) are topped by silt 260 deposits (e.g. Fig. 4G) that suggest a lacustrine environment. Furthermore, the contiguous lake 261 sediment benches recording the minimum extent of gLP (Fig. 1B) cover a larger area than 262 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<sup>-1</sup>; Fig. 6). A GIA of 1.25 m km<sup>-1</sup>, like the tilt 263 264 that affected glacial Lake Bretz (Table 1), would have resulted in a total gLP volume of 142 km<sup>3</sup> 265 (Table 2). Its surface elevation against a northern ice dam would have been >800 m a.s.l. and it 266 would have reached depths of >400 m (Table 2, Fig. 7).

267

#### **GLOF** evidence in the Kootenay River valley

268 The Kootenay River valley contains elevated (>100 m above the modern Kootenay River 269 floodplain) sediment benches or terraces (GLOF sediment, Fig. 1B), and alluvial fans, which are 270 remnants of a thicker valley fill (Figs 1B, 8). The highest truncated and bisected alluvial fan 271 remnants reach elevations of up to  $\sim 675$  m a.s.l. (Fig. 8B). The highest terrace occupies both sides 272 of the Kootenay River valley with tread elevations from 642 m a.s.l. near the Purcell Trench to 273 ~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  $\sim 1.5$  m km<sup>-1</sup>. The terraces typically contain normally-graded planar-274 275 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 types 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).

283 The normally-graded, planar and cross-stratified beds of sand and imbricated gravel with westward 284 palaeoflows in the terraces are compatible with observations of sand and/or gravel dune or bar 285 formation (e.g. Carrivick et al. 2004; Rushmer 2006; Russell 2009) during a confined GLOF from 286 gLP. Diffusely graded, sinusoidally stratified sand records stationary and breaking antidunes, 287 deposited rapidly from supercritical hyperconcentrated GLOF flows (cf. Lang & Winsemann 288 2013; Lang et al. 2017). Occasional massive gravel beds are associated with deposition of traction 289 load during GLOFs (Carrivick et al. 2004; Russell 2009) or rapid deposition (Rushmer 2006). 290 Ripples in fine sediments have been interpreted to record deposition during waning GLOF flows 291 (Russell 2009) or hydraulic damming during the latter stages of floods (cf. Touchet Beds in the 292 Channeled Scabland, Waitt 1980). Together, this sedimentary sequence is interpreted to record a 293 period of sand and gravel aggradation during energetic GLOF flows through the Kootenay River 294 valley, followed by fine sediment deposition during waning flood flows or in areas of local 295 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  $\sim 1.5 \text{ m km}^{-1}$  and a pre-flood valley bottom elevation of  $\sim 640 \text{ m}$  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  $\sim 110 \text{ m}$  of incision near the Purcell Trench and  $\sim 150 \text{ m}$  near the Columbia River valley (Fig. 8).

303 Although it is likely that the highest terraces were formed during erosive stages of a gLP GLOF, 304 it is unknown exactly which terraces formed by channelising GLOF flows, and which by 305 postglacial fluvial incision. However, recent geomorphic examinations suggest the bedrock around 306 Cora Linn Dam (~530 m a.s.l., Figs 1B, 8) was sculpted by energetic flows (Waitt 2016). The 307 potholed bedrock near the top of the valley fill sequence (624 m a.s.l., Fig 8A) also suggests 308 bedrock erosion during a period of energetic flow that likely removed pre-GLOF valley fill 309 sediment (Zen & Prestegaard 1994; Fig. 8A). Taken together, these observations suggest that the 310 rising stage of a gLP GLOF was responsible for alluvial fan truncation, the creation of elevated 311 potholes (624 m a.s.l.), removal of pre-GLOF sediment to bedrock (at least 90 m thickness at Corra 312 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 313 314 flows or post-glacial river flows. More conservatively, the rising stage of the GLOF may only have 315 been responsible for alluvial fan truncation, elevated pothole formation and partial incision through 316 valley fill. In this scenario, waning-stage channelizing GLOF flows and post-glacial river flows 317 may have been responsible for the bulk of sediment removal, terracing and the bedrock sculpting 318 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 322 recording the pre-GLOF valley bottom in the western Kootenay River valley (Fig. 8A). It is largely323 composed of very coarse, poorly-sorted, clast-supported massive gravel with abundant well-324 rounded boulders (up to 0.5 m b-axes) (Fig. 4H). Clast a-axes are randomly oriented and gravel325 fabric measurements have a polymodal distribution (D3, Fig. 1B). The coarse, fan-building326 gravels are capped by ~10 m of trough cross-stratified sand and cobble gravel (Peters 2012).

327 The fan shape and composition of the gravel deposit at Playmor Junction (Fig. 1B) suggests 328 deposition as a large alluvial fan (Blair & McPherson 1994) or GLOF expansion bar (e.g. Baker 329 1984; Benito 1997). The well-rounded gravel that comprises the fan indicates fluvial mobilisation 330 and its massive, poorly-sorted structure with boulder-sized clasts is consistent with deposition 331 during high-energy flows associated with GLOFs (Cenderelli & Wohl 2003; Harrison et al. 2006; 332 Russell 2009). The polymodal fabric (D3, Fig. 1B; cf. Morison & Hein 1987; Meetei et al. 2007) 333 with randomly-oriented clast a-axes, poorly sorted texture, and massive structure of the gravel 334 suggests rapid deposition in a turbulent, possibly hyperconcentrated flow (cf. Brennand & Shaw 335 1996). This boulder gravel is coarser than any gravels observed in the Kootenay River valley, 336 suggesting deposition during the most energetic GLOF flows responsible for pothole erosion, 337 alluvial fan truncation, and pre-GLOF valley-fill incision in the Kootenay River valley. The 338 capping, trough cross-stratified sand and gravel indicates a subsequent period of braided stream or 339 alluvial fan deposition (cf. Allen 1983; Blair 1987; Kjær et al. 2004) or possibly deposition during 340 waning flood flows (Winsemann et al. 2016) and the multiple terraces preserved on the fan suggest 341 incremental incision by fluvial erosion over time (Fig. 8C). Thus, we interpret the Playmor 342 Junction deposit to most likely have been deposited as a GLOF expansion bar that was capped by 343 alluvium and subsequently incised by the postglacial Kootenay River. The ~110 m height 345 m a.s.l.; Fig. 8A) indicates deposition after an initial period of erosive GLOF flows.

#### 346 **Discussion**

#### 347 Summary of gLP evolution

348 Sedimentary and geomorphic evidence records gLP evolution from a small ice-contact proglacial 349 lake (described by Alden 1953) to a large, valley-filling ice-contact proglacial lake that drowned 350 >100 km length of the Purcell Trench (Fig. 2). At one stage in its evolution, gLK (Alden 1953) 351 decanted much of its volume into gLP (Fig. 2B). This drainage is recorded by the flood-related 352 fan (Alden 1953) that has since been incised by the Kooteani River (Figs. 1B, 6). This fan formed 353 the southern shore of gLP and forced southward drainage towards the Elmira spillway against the 354 western valley wall (Fig. 6). The northern extent of gLP was dictated by the northward-retreating 355 Purcell Lobe of the CIS (Figs 2, 6).

356 No direct geochronological ages are available for gLP or gLK. However, tephrostratigraphic ages 357 that constrain the geomorphic evidence for Columbia River valley flood flows provided by Waitt 358 et al. (2009) can provide some tentative, preliminary gLP chronology. The dune-scale bedforms 359 described by Waitt et al. (2009) and Waitt (2016) were deposited <13.5 cal. ka BP, which is 360 younger than the final drainage of other regional palaeolakes (Missoula, Columbia, and 361 Bonneville), but compatible with our reconstructions of gLP and reconstructions of the deglacial 362 Purcell Lobe's ice margin, which was near the Kootenay River valley by ~12 cal. ka BP (Dyke et 363 al. 2003; Fig. 6). Thus, in the absence of independent dating, gLP sedimentation is assumed to 364 have occurred during MIS 2 with a maximum possible age between 14 750±375 and 15 200±400 365 cal. a (LLGM; Atwater 1987). Drainage most likely occurred after ~13.5 cal. ka BP (Waitt et al. 366 1994, 2009; Dyke et al. 2003).

#### 367 GLP volume and local glaciological impacts

368 Most of the lake bed sediments that comprise the contiguous benches in the southern Purcell 369 Trench are not significantly deformed, suggesting deposition in an ice-distal environment without 370 remobilisation from removal of buried ice. This suggests previous hypotheses of a low-volume, 371 ice-marginal, or supraglacial lake that formed during CIS downwasting in the Purcell Trench are 372 incorrect (e.g. Alden 1953; Eyles & Clague 1991; Ryder et al. 1991). Furthermore, iceberg rain 373 out sediment interpreted in this study suggests that gLP's lake depth (>400 m, Table 2) was enough 374 to force ice-marginal flotation and induce calving retreat through the Purcell Trench (Carrivick & 375 Tweed 2013). Thus, the Purcell Lobe would have likely formed a steep terminus (Fig. 7) and been 376 unable to displace significant amounts of gLP volume. This interpretation is compatible with the 377 lack of ice-marginal landforms (kame terraces and moraines) in the southern Purcell Trench. 378 Furthermore, kame terrace deposits (Figs 1B, 6D, E) confirm that the Purcell Lobe was sufficiently 379 sealed to the Purcell Trench valley-wall for a period that allowed at least 16 m of glaciofluvial 380 deposition. The relatively low position of the kame terraces within the valley (600-725 m a.s.l.) 381 indicates that the seal existed late in the deglaciation of the Purcell Lobe. This seal, although not 382 likely to be concurrent with the lacustrine deposits of gLP (based on elevation, Table 2), provides 383 evidence that the Purcell Lobe could have dammed large volumes of water long after the CIS 384 margin retreated northward into British Columbia. This evidence of a high-volume gLP elucidates 385 important potential for large flood flows late in CIS deglaciation.

A deep, high-volume gLP (>400 m, almost 150 km<sup>3</sup>, 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).

## 395 **GLP drainage and impacts on landscape evolution**

396 GLP was confined to the Purcell Trench until the Purcell Lobe's calving margin retreated 397 sufficiently northward to allow drainage into the Kootenay River valley (Figs 1B, 2). At this time, 398 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<sup>-1</sup>, relative to the 399 400 modern landscape (Fig. 6). Applying this tilt to a modelled gLP lake plane results in a surface 401 elevation against the ice dam near the Kootenay River valley of 817 m a.s.l. and a drainable volume 402 of  $\sim 116 \text{ km}^3$  (Table 2). This elevated lake surface is  $\sim 180 \text{ m}$  above the pre-GLOF valley bottom 403 in the Kootenay River valley (Table 2, Fig. 8A), suggesting that gLP water likely drained suddenly 404 into the Kootenay River valley following catastrophic ice-dam failure.

405 The ~180 m elevation difference between the gLP water surface and the pre-GLOF valley bottom 406 in the Kootenay River valley (Table 2) suggests that this sudden drainage would have generated 407 extremely high specific and total stream powers, capable of eroding large amounts of boulder-408 sized sediment (cf. Cenderelli & Wohl 2003). This erosive GLOF is recorded by truncated alluvial 409 fans and potholed bedrock  $\sim$ 140 m and  $\sim$ 90 m, respectively, above the modern Kootenay River 410 valley floor (cf. Winsemann et al. 2016). If the fluvially-eroded bedrock described by Waitt (2016) 411 at the Corra Linn Dam is attributed to a gLP GLOF, >90 m depth of sediment would have been 412 removed by the GLOF at this location, which is comparable to previous models of GLOF erosion 413 (e.g. Winsemann *et al.* 2016; Lang *et al.* 2019; Fig. 8A). After incising the Kootenay River valley
414 fill, the flood flows debouched into the larger Columbia River valley at Playmor Junction, where
415 a large, fan-shaped expansion bar was formed from cobble- and boulder-sized bedload (cf. Baker
416 1984; Benito 1997; Figs 1B, 4H, 8C).

417 The flood flows generated by gLP drainage would have entered the Columbia River valley via the 418 Kootenay River valley (Figs 1B, 8C). Whether or not these flows would have been capable of 419 enough geomorphic work to have formed dune-scale bedforms in the Channeled Scabland near 420 Chelan Falls, Washington (Waitt et al. 1994, 2009) depends on flow attenuation along the ~500-421 km long flood route (defined by the lengths of the modern Kootenay River and Columbia River 422 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<sup>3</sup>, Table 2), and 423 424 likely drained suddenly (based on the modelled  $\sim 180$  m elevation difference between the gLP 425 water surface and the top of the highest terrace in the Kootenay River valley), it is possible that 426 gLP flood flows induced late-Pleistocene geomorphic changes in the Channeled Scabland (Fig. 427 1A; cf. Waitt et al. 2009); however, hydraulic modelling should be performed to assess this (e.g. 428 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).

## 441 **Conclusions**

- GLP was a large (~1 152 km<sup>2</sup>, 142 km<sup>3</sup>) 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<sup>3</sup> 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<sup>3</sup> 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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- 491

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  the paleo-potomac river. *Geology 22*, 47-50.
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## 712 List of captions

- 713 **<u>Tables</u>**
- 714 Table 1: Previously reported differential glacioisostatic adjustments associated with the CIS during MIS 2 assessed for this study.

Tilt <sup>1</sup> (m km <sup>-1</sup> )	Location (glacial lake name)	Distance from LLGM limit <sup>2</sup> (km)	Upslope direction	Reported age	Calibrated age range (median) cal. ka BP <sup>3</sup>	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) <sup>4</sup>	Mathews (1978)
0.46	Peace Basin (Peace Phase II)	100	230.9°±5°	<16.0±2.5 (optical)	NA	Hickin <i>et al.</i> (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) <sup>4</sup>	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 <i>et al.</i> (1989)
1.15	Puget Lowland (Bretz, western)	10	North	13.5 (ka BP)	15.8 - 16.8 (16.3) <sup>4</sup>	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)
$\begin{array}{c} 1.8 \pm \\ 0.6 \end{array}$	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)

715 <sup>1</sup> Land surface tilt caused by differential glacioisostatic adjustment, as recorded by palaeo-lake-level indicators.

716 <sup>2</sup> Approximate distances between the glacial lake's nearest margin of the CIS during the LLGM from Fulton et al. 717

(2004). GLP was ~40 km from the LLGM limit, perhaps <13.5 cal. ka BP (Waitt et al. 2009).

718 <sup>3</sup> Radiocarbon ages calibrated for this study with Calib software version 7.10 (Stuiver & Reimer 1993) using the

719 IntCal13 radiocarbon curve (Reimer et al. 2013). Reported as the 25 median probability (e.g. Peters et al. 2016).

720 <sup>4</sup> Calibration was performed using an assumed standard error of  $\pm 160$  (the highest reported in this review, to avoid

721 spurious precision) because insufficient information was reported.

Water- plane tilt (m km <sup>-1</sup> )	Lake area <sup>1</sup> (km <sup>2</sup> )	Lake volume <sup>1</sup> (km <sup>3</sup> )	Maximum drainable volume <sup>2</sup> (km <sup>3</sup> )	Water surface elevation <sup>3</sup> (m a.s.l.)	Water depth <sup>4</sup> (m)	Minimum ice dam thickness <sup>5</sup> (m)	Height of water surface above pre- GLOF valley bottom <sup>6</sup> (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

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

<sup>1</sup> 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).

- 726 <sup>2</sup> Lake volume minus the volume of Kootenay Lake within the palaeolake extent.
- <sup>3</sup> Maximum elevations, as measured against the ice dam.
- <sup>4</sup> Maximum depths, measured against the ice dam.

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

- <sup>6</sup> Height of water surface above the height of the highest terrace in the Kootenay River valley (640 m a.s.l.).
- 732





735 Fig. 1: A. Study area (red box) at the western USA-Canada border showing the Channeled 736 Scabland and CIS extent at the Last Glacial Maximum (LGM; Fulton et al. 2004). The Purcell 737 Lobe's southern terminus at ~13.5 ka BP (dashed black line) is mapped after Dyke et al. (2003). 738 CF = Chelan Falls. B. Schematic map of the study area depicting the locations of sedimentary 739 deposits (kame; glacial Lake Purcell (gLP) lake bed sediments; gLP drainage sediments; gLP 740 Glacial Lake Outburst Flood (GLOF) sediments; glacial Lake Kootenai flood-related fan), 741 palaeoflow measurements (lower hemisphere, equal area projection and rose diagram) and other 742 locations discussed in the text. Note that GLOF sediments are present throughout the Kootenay 743 River valley, but only major exposures are mapped. Also, the West Arm of Kootenay Lake (east 744 of the Corra Linn Dam) and the Kootenay River are jointly referred to in the text as the Kootenay 745 River valley (KRv) and changes in river width between these water bodies are not represented to 746 enhance illustrative clarity. Red box delineates the area shown in Fig. 3. BF = Bonners Ferry; PJ 747 = 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;</li>

753 Atwater 1987; Porter & Swanson 1998; Clague & James 2002). Both the Elmira spillway (ES) at 754 710 m a.s.l. and Bull River spillway (BRS) at 732 m a.s.l. are active (shown as blue arrows). 755 During this stage Alden (1953) considered gLP an unnamed proglacial lake. B. Continued retreat 756 of the Purcell Lobe causes the growth of gLP and its northward expansion. Eventually the portion 757 of gLK above 710 m a.s.l. (the water above the Elmira spillway) decanted into gLP (shown as red 758 arrow), rapidly reducing gLK's volume. C. Glacial Lake Kootenai has largely drained into gLP 759 and may not exist at all. The late deglacial Purcell Lobe retreats northward towards the Kootenay 760 River valley (KRv) until it fails to dam gLP. GLP debouches into the Kootenay River valley (red 761 arrow) and eventually its floodwaters reach the Channeled Scabland via the Columbia River (Fig. 762 1). Note that the naming scheme of Alden (1953) is abandoned and replaced with the one depicted 763 by this schematic because it ignores gLP's nascent formation and considers later stages of gLP to 764 be the same lake as gLK. BC = British Columbia; WA = Washington; ID = Idaho; MT = Montana.

765



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-

781 rounded, granitic lonestone at site 5. D. Angular, unconsolidated sand clasts (outlined by white 782 dashed lines) in an interfluve-occupying, valley-wall deposit (kame) near gravel fabric K1. Knife 783 handle is ~9 cm long. E. Planar-stratified sand and gravel in a kame terrace. Arrow marks metre 784 stick for scale. Location of gravel fabric K2 is shown as a labelled white dot. F. Climbing ripples 785 (after Ashley et al. 1982) measured for palaeoflow Dr1 from a ~560 m a.s.l. gLP GLOF terrace in 786 the Kootenay River valley. G. Inclined gravel beds (white dashed lines highlight two lower 787 contacts) overlain by massive silt at site 10 (Fig. 3). Arrow points to a person for scale. H. Poorly 788 sorted cobble and boulder gravel at ~490 m a.s.l. in a gLP GLOF expansion bar measured for 789 gravel fabric D3. Ruler is 36 cm long. I. Planar-stratified sand and gravel and diffusely graded, 790 sinusoidally stratified sand composes a ~640 m a.s.l. gLP GLOF terrace in the Kootenay River 791 valley. Location of gravel fabric D1 shown as a labelled white dot. Arrow marks metre stick for 792 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.

803



805 Fig. 6: Extent of gLP based on the array of tested glacioisostatic tilts (Table 2). Note that only the 806 steepest tilt (1.25 m km<sup>-1</sup>) covers the gLP lake bed sediment bench (red dashed line; Fig. 1A) and 807 allows gLP to overtop the flood-related fan formed by gLK's drainage (Alden 1953) which is 808 capped by silty lake bed sediments at site 10 (red dot). This suggests that the fan was partially 809 inundated by gLP. The locations of modern Kootenay Lake (black), the Kootenay River valley 810 (KRv), and the Purcell Lobe <13.5 cal. ka BP (Dyke et al. 2003) are also shown. Only the 811 northernmost extent of the gLP lakebed is outlined for visual clarity with other aspects of this 812 figure; see Fig. 1A for the complete extent.

813



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).





Fig. 8: Kootenay River valley (KRv; Fig. 1B) geomorphology. A. Kootenay River valley long
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<sup>-1</sup> tilt), the highest truncated alluvial fan deposits (~675 m a.s.l.), the 824 825 highest valley-flanking terrace (depicting the pre-GLOF valley bottom), and the modern Kootenay 826 River (bold blue line). Low-elevation, untruncated alluvial fans aggrading into the Kootenay River 827 valley are shown as brown polygons. Locations of GLOF palaeoflow measurements and potholes 828 are provided (see Fig. 1B for map view). B. Eight Kootenay River valley, cross-sectional profiles 829 (from Geobase, Natural Resources Canada DEM; Government of Canada 2019), grouped to 830 represent three reaches of the Kootenay River valley. The groups are shown with stacked profiles 831 that reveal trends in bench and remnant fan elevations and allow estimations of pre-incision valley-832 fill elevations. Estimated extents of truncated alluvial fan remnants are shown as dotted lines. C. 833 Terrace treads incised into an expansion bar at the Kootenay River valley (KRv)-Columbia River 834 valley (Rv) confluence (see Fig. 1B for regional map) (DEM (Geobase, Natural Resources Canada) 835 overlain by a georeferenced aerial photograph (National Aerial Photograph Library, Environment 836 Canada)). Contacts between the valley-fill sediment and the bedrock valley walls are shown as 837 dashed lines. Solid black lines outline individual terraces identified using the DEM. The location 838 of gravel fabric D3 (Figs 1B, 6H) is shown as a labelled red dot.