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

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4 Peters, J. L. & Brennand, T. A.: Palaeogeographical reconstruction and hydrology of glacial Lake
5 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

15 deglaciation suggests that gLP could have been subjected to tilts ranging from $0 - >1.25$ m km^{-1}.

16 Sedimentary evidence suggests high lake plane tilts ($\gtrsim 1.25$ m km^{-1}) are the most likely to have

17 affected gLP. Using this, the palaeogeography and volume of gLP are modelled, revealing that

18 ~116 km^3 of water was susceptible to sudden drainage into the Channeled Scabland via the

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.

37   At the Local Last Glacial Maximum (LLGM) large glacial lakes, like the ~2 600 km^3 glacial Lake

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.

59   This study provides the first comprehensive investigation of glacial lake evolution in the Purcell

60   Trench. We use geological evidence and previous records of CIS glacioisostatic tilt to inform a

61   palaeogeographic reconstruction of a large lake, named here glacial Lake Purcell (gLP). We

62   explore evidence for ice damming of the lake and its drainage through the Kootenay River valley.

63   These analyses are used to assess the potential for energetic flood flows from the Purcell Trench

64   into the Channeled Scabland after the final drainage of glacial lakes Missoula and Columbia.

65   **Previous work on Purcell Lobe ice-marginal lakes**

66   Previous studies near the Purcell Trench have reconstructed glacial Lake Kootenai (gLK) from

67   thick deposits of lake bed sediments (sand and silt) in valley systems in northern Idaho and

68   northwestern Montana (Alden 1953; Johns 1970; Smith 2006; Fig. 1A). This lake formed when

69   river systems were impounded by the retreating Purcell Lobe (Alden 1953; Johns 1970; Smith

70   2006; Fig. 2). The sediments recording glacial Lake Kootenai are over 90 m thick in some areas

71   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).

90   The volumes of these glacial lakes have also been speculated upon and several researchers have

91   pointed out that volume was contingent on the style of CIS retreat through the Purcell Trench. If

92   Purcell Lobe retreat was dominated by stagnation and downwasting, the ice would have likely

93   displaced much of the volume available to any glacial lake. Fulton (1967, 1991) proposes a CIS

94   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**

115   Data were gathered for this study within the Purcell Trench, its high-relief tributary valleys, and

116   along the Kootenay River valley (KRv; Fig. 1B). Much of the floor of the Purcell Trench in

117   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

125   Lake and the Kootenay River are jointly referred to as the Kootenay River valley (KRv; Fig. 1B).

Methods

Geomorphology and sedimentology

128   Geomorphic analyses and preliminary investigations to identify potential field sites were carried

129   out using publicly available digital elevation models from Geobase (from Natural Resources

130   Canada) and the National Elevation Database (NED, from the United States Geological Survey).

131   The two datasets were compiled and re-gridded into a single, 25-m resolution Digital Elevation

132   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 ≥30 clast a-b

136   plane attitude measurements (maximum dip and down-dip direction of the a-b plane) were taken

137   in <0.5 m^2 areas of exposures. The data were then plotted on lower hemisphere, equal area

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

141   Dr1, D1, D2 and D3 (Fig. 1B). Clast a-axis position relative to the direction of a-b plane maximum

142   dip in gravel fabrics was used to determine the likely mode of clast mobilization. A dominance of

143   clast a-axes transverse to dip direction (a(t)) suggests clasts rolled along the bed, whereas a

144   dominance of clasts parallel to dip direction (a(p)) implies clast sliding across the bed or deposition

145   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

162   pre-incision elevations measured using the composite DEM. In the model, the boundaries of this

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 2σ confidence

168   using the IntCal13 radiocarbon curve (Reimer *et al.* 2013) to improve comparability of the ages

169   (cf. Peters *et al.* 2016).

170   The resultant array of plausible lake plane tilts was projected along the Purcell Trench from the

171   Elmira Spillway, which controlled gLP lake levels prior to drainage into the Kootenay River valley

172   (Alden 1953; Johns 1970; Smith 2006). A DEM of modern Kootenay Lake bathymetry

173   (bathymetric DEM) was constructed from interpolated individual soundings using a GIS and was

174   used to estimate the total volume of gLP (i.e. the volume of modern Kootenay Lake was added to

175   the calculated drainable volume of gLP based on topographic DEMs).

176   **Results and interpretations**

177   **The northward extension of gLP**

178   Evidence for the northward extension of gLP is mainly preserved in large (>60 km long and up to

179   ~20 km wide) sediment benches that occupy the floor of the Purcell Trench adjacent to the

180   Kootenai River (Fig. 1B). These benches extend from the Elmira spillway (~10 km south of

181   Bonners Ferry, Idaho) to ~5 km north of Creston, British Columbia (Fig. 1B), and reach an

182   elevation of up to 706 m a.s.l. (Fig. 3), >176 m above the modern water level of Kootenay Lake.

183   Bench tread elevations are incrementally lower towards the Kootenai River floodplain (Fig. 3).

184   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 ≥10 m are also exposed in the Idaho sediment benches (Fig. 4G). These gravel

193   deposits are typically capped by laminated or massive silt.

194   The highest bench surfaces are composed of flat-topped silt deposits that share similar elevations,

195   recording contiguous lake bed deposits (cf. Ryder *et al.* 1991; Johnsen & Brennand 2004). The

196   flat-topped, occasionally channelised bench segments (e.g. Fig. 3) record remnant lake bed

197   sediments following fluvial incision (cf. Clague 1986). This interpretation of terrace formation is

198   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).

207 The inclined gravel deposits (Fig. 4G) are consistent with alluvial fan progradation (Blair &

208 McPherson 1994) and their increased occurrence in the south of the study area (site 10, Figs 1B,

209 3) implies that they record deposition during the decanting of gLK into gLP (unnamed lake, Alden

210 1953). These inclined gravels may also record deltaic deposition, however no topsets have been

211 identified, so an interpretation of alluvium, or potentially expansion bar deposits, is preferred.

212 Thus, the lake bed benches and bench segments confirm the northward extension of gLP through

213 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 ≥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).

223 The normally-graded, planar- and trough cross-stratified gravel beds (Fig. 4E) are consistent with

224 fluvial deposition in a gravel-bed stream (Miall 1977). Gravel fabrics record valley-parallel stream

225 flows and traction transport (rolling; Brennand 1994) that is anomalous to modern topography.

226 Furthermore, the northward flows are also irreconcilable with an interpretation of remnant advance

227 outwash deposits and the elevation of these deposits (up to 725 m a.s.l.) place them ~200 m above

228 the surface of the modern Kootenay River valley where post-gLP damming is unlikely (Peters

229 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 ~2.1 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

249 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}

250 (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

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^3 (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

263   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

264   that affected glacial Lake Bretz (Table 1), would have resulted in a total gLP volume of 142 km^3

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   **<u>GLOF evidence in the Kootenay River valley</u>**

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 ~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,

274   B) and a down-valley slope of ~1.5 m km^{-1}. The terraces typically contain normally-graded planar-

275   stratified and trough cross-stratified sand, imbricated gravel with occasional massive gravel beds,

276   and, occasionally, diffusely graded, sinusoidally stratified sand beds (Fig. 4I). Trough-cross

277   stratified gravel is most common near the top of the valley fill, where smoothed and potholed

278   bedrock also exists at elevations up to 624 m a.s.l. (Figs 1B, 8A). Type-a ripples overlain by type-

279   s ripples (Ashley *et al.* 1982; Fig. 4F) are exposed in inset terraces lower than 560 m a.s.l. in the

280   Kootenay River valley valley fill and towards the valley centreline (Fig. 1B). Gravel fabrics from

281   planar-stratified gravels and type-a ripple measurements record westward palaeoflows through the

282   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).

296   The Kootenay River valley highest terrace, just below the alluvial fan remnants, is interpreted to

297   record the pre-GLOF valley bottom because the relative prevalence of alluvial fans that have

298   aggraded onto it (Fig 8B) suggests this surface was exposed for a significant amount of time. This

299   terrace's elevation suggests a pre-incision down-valley slope of ~1.5 m km^{-1} and a pre-flood valley

300   bottom elevation of ~640 m a.s.l. near the Purcell Trench (i.e., Kootenay Lake, Fig. 8A). Terrace

301   geomorphology indicates that the valley fill (pre-GLOF and GLOF sediments) has experienced

302   ~110 m of incision near the Purcell Trench and ~150 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

313   terraces formed after GLOF aggradation during incision by waning-stage channelising GLOF

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.

319   The large, fan-shaped landform situated at the confluence of the Kootenay River valley and the

320   Columbia River valley (Playmor Junction, Fig. 1B) exhibits multiple terraces (Fig. 8C) and reaches

321   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 largely

323   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 gravel

325   fabric measurements have a polymodal distribution (D3, Fig. 1B). The coarse, fan-building

326   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

344 difference between its highest surface (~490 m a.s.l.) and the local pre-GLOF valley bottom (~600

345 m a.s.l.; Fig. 8A) indicates deposition after an initial period of erosive GLOF flows.

Discussion

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.

386   A deep, high-volume gLP (>400 m, almost 150 km^3, respectively, Table 2) would have held a

387   similar amount of water as modern Lake Tahoe, or ~30% more than the Dead Sea. Such a lake

388   would have mechanically exacerbated local CIS mass loss through calving, thereby steepening the

389   ice margin causing increased ice flow velocities (Carrivick & Tweed 2013). Mass loss would also

390   have been accelerated in the Purcell Trench by thermal erosion, because ice-marginal lakes deliver

391   heat to glacier termini. Such thermal erosion can undercut the ice margin at the water line (e.g.

392   Kirkbride & Warren 1999; Röhl 2006) further steepening the terminus and intensifying calving

393   retreat. These feedbacks suggest that the deglacial Purcell Lobe would have had a steep ice-front

394   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

399   (Alden 1953; Fig. 2B), the gLP water surface was most likely tilted ~1.25 m km^{-1}, relative to the

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 ~116 km^3 (Table 2). This elevated lake surface is ~180 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 ~140 m and ~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

423 (Kuehn *et al.* 2009; Waitt *et al.* 2009), drained a large volume of water (~116 km^3, Table 2), and

424 likely drained suddenly (based on the modelled ~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).

429 Regardless of the role the gLP GLOF may have played in the Channeled Scabland, its regional

430 effects on postglacial fluvial systems are evidenced geomorphically and sedimentologically. The

431 low (<560 m a.s.l.) terraces located along the Kootenay River valley (Fig. 8A, B) likely record

432 postglacial (late-Pleistocene and Holocene) fluvial incision by the Kootenay River, which

433 remobilised the waning flood-flow deposits towards the Kootenay River valley confluence with

434 the Columbia River valley (Figs 1B, 8C). As the sediment-laden Kootenay River exited the narrow

435 Kootenay River valley, it deposited its bedload as ~10 m of trough cross-stratified sand and gravel

436 alluvium (Peters 2012) over the surface of the boulder-gravel expansion bar (cf. Kehew *et al.*

437 2010). Finally, when the Kootenay and Columbia rivers neared their modern elevations and the

438 Kootenay River reached the Kootenay River valley's bedrock and/or its specific sediment yield

439 relaxed following postglacial incision (Church & Slaymaker 1989), alluvial deposition over the

440 expansion bar was replaced with incision, forming extensive fluvial terraces (Fig. 8C).

441 **Conclusions**

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

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

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

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

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

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

473   •   Following the catastrophic drainage of gLP, the Kootenay River incised into the GLOF
474   sediments, leaving a series of terraces formed by the GLOF and later, postglacial fluvial
475   incision at elevations from ~600 – 550 m a.s.l. (~20 – 60 m above the modern river). A
476   ~10-m thick deposit of alluvium was deposited over the expansion bar at the confluence of
477   the Kootenay River valley with the Columbia River valley, which was also incised as the
478   Kootenay River approached its modern elevation, leaving a series of fluvial terraces.

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

487   Data availability – We agree to make data available upon request.

488   Author contributions – JLP collected the data used for this research with occasional supervision
489   from TAB. JLP digitised and processed the data. Both authors interpreted the results, discussed
490   their overarching scientific relevance, and contributed to the writing of this manuscript.

491

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710   the paleo-potomac river. *Geology 22*, 47-50.

711

712   **List of captions**

713   **Tables**

714   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)

715 [1] Land surface tilt caused by differential glacioisostatic adjustment, as recorded by palaeo-lake-level indicators.
716 [2] 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 [3] 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 2σ median probability (e.g. Peters et al. 2016).
720 [4] Calibration was performed using an assumed standard error of ±160 (the highest reported in this review, to avoid
721 spurious precision) because insufficient information was reported.

722

723   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

724   [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
725   margin is assumed (see text).

726   [2] Lake volume minus the volume of Kootenay Lake within the palaeolake extent.

727   [3] Maximum elevations, as measured against the ice dam.

728   [4] Maximum depths, measured against the ice dam.

729   [5] Clean ice thickness required to resist flotation at the dam following the 9/10^{ths} ratio of ice to water densities
730   (Thorarinsson 1939; Fowler 1999).

731   [6] Height of water surface above the height of the highest terrace in the Kootenay River valley (640 m a.s.l.).

732

733 **Figures**

734
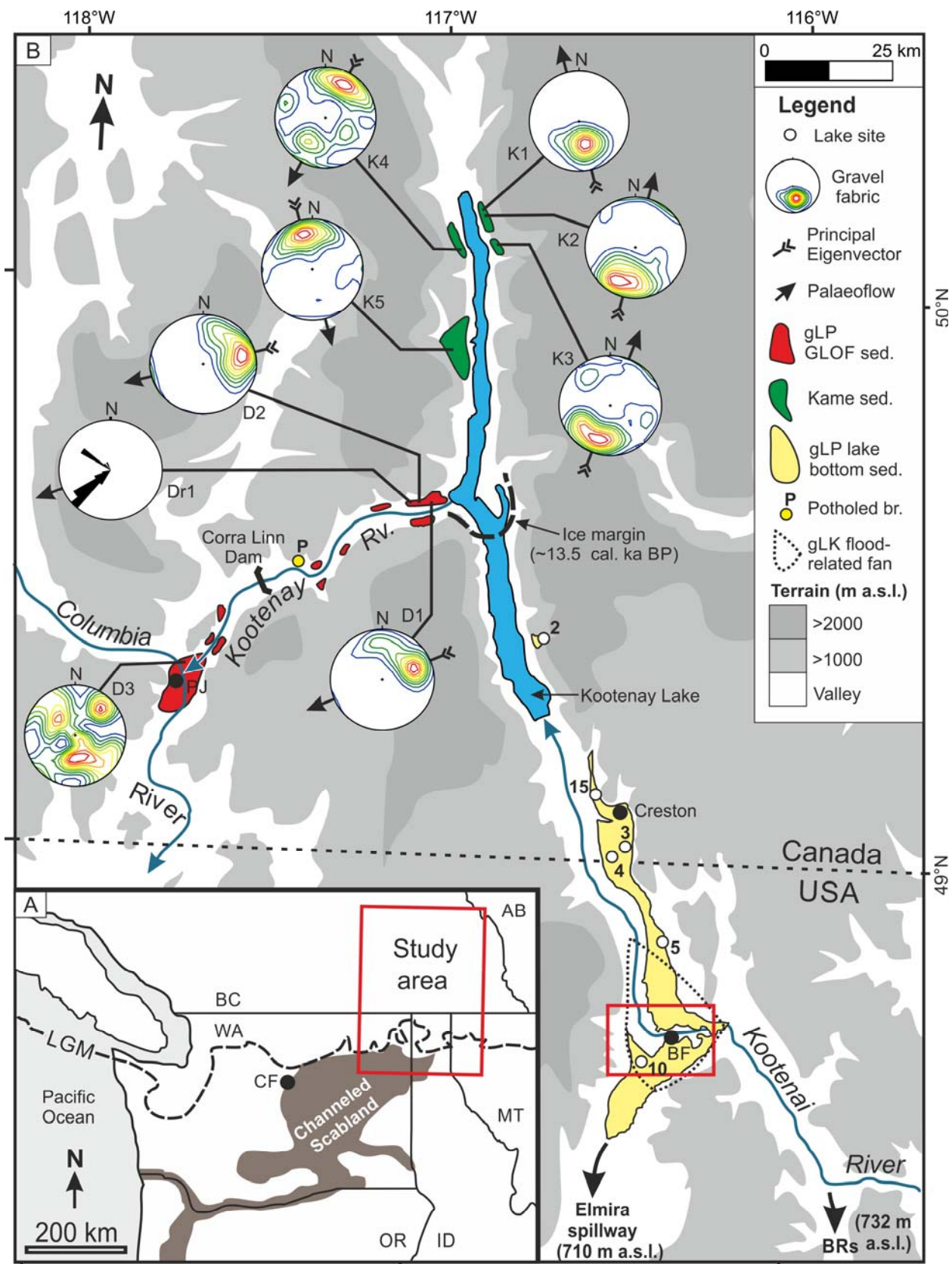
735   Fig. 1: A. Study area (red box) at the western USA-Canada border showing the Channeled

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.

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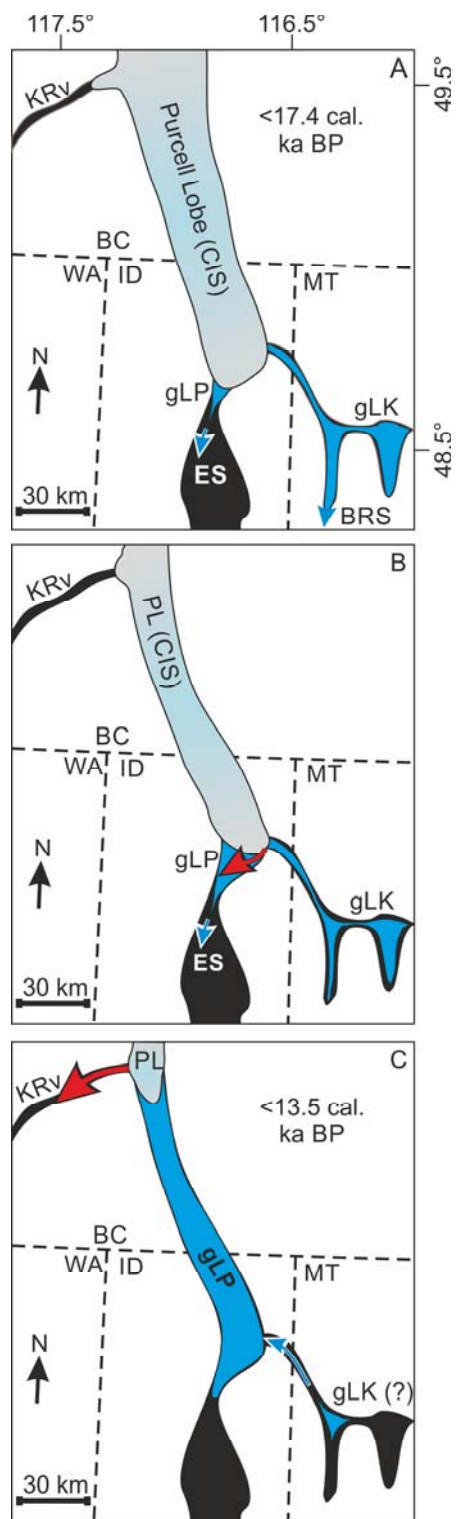
749

750   Fig. 2: Schematic representation of glacial lake evolution showing. A. Discrete stages of glacial

751   Lakes Purcell (gLP) and Kootenai (gLK) (dark blue) impounded behind the Purcell Lobe (light

752   blue, PL) sometime after the Last Late Glacial Maximum (LLGM) (likely <17.4 cal. ka BP;

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
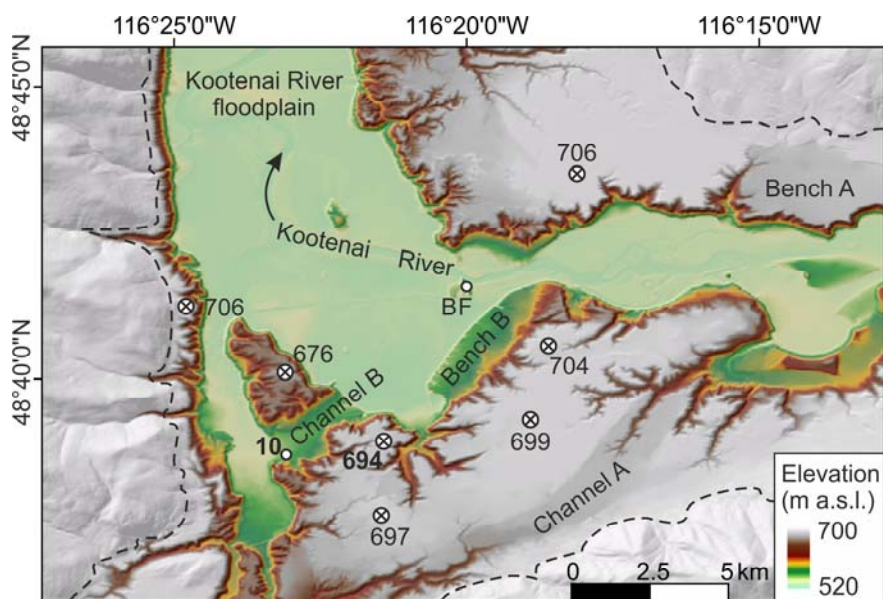
766

767   Fig. 3: Hillshaded DEM from a composite of Geobase (Government of Canada 2019) and National

768   Elevation Dataset (United States Geological Survey) data (USGS 2019) revealing benches A and

769   B incised into the gLP lake bed deposits. Dashed lines delineate the contact between the lake bed

770   sediment and the valley walls. Point elevations (white dots with Xs) that typify elevation data

771   used to reconstruct the pre-incision gLP lake bed are provided and highlight the relative flatness

772   of the deposit's surface across the Purcell Trench. Channels 'A' and 'B' correspond in elevation

773   with benches 'A' and 'B', respectively. Site 10 is shown as a labelled white dot within Channel

774   B.

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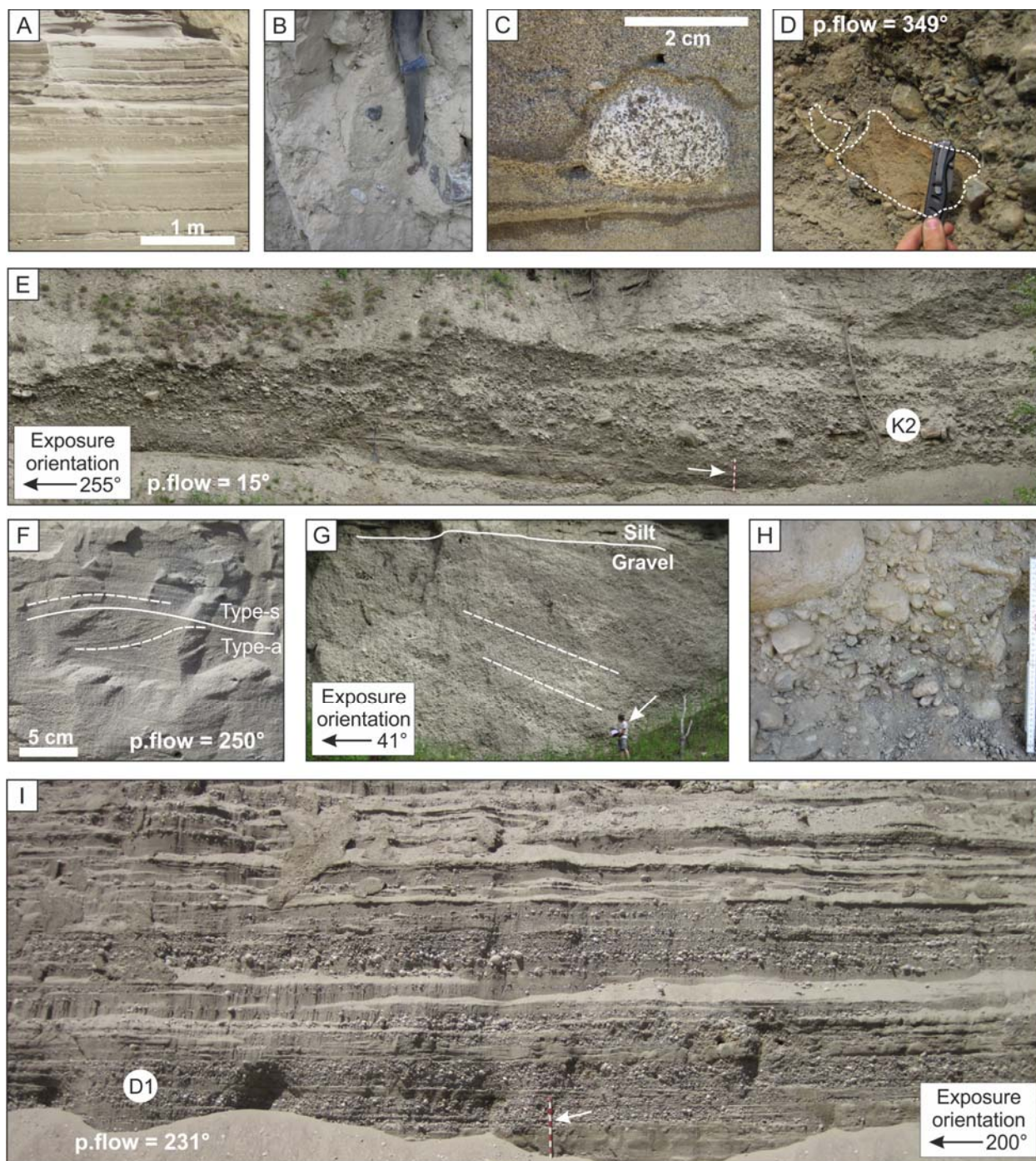
Fig. 4: Examples of sediment deposits. Sites are located in Fig. 1B. A. Massive and laminated silt deposits are common throughout the lake bed bench (photograph is from site 5). B. Massive silt with abundant, outsized clasts (lonestones) at site 15. Knife blade is ~9 cm long. C. Massive, coarse sand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

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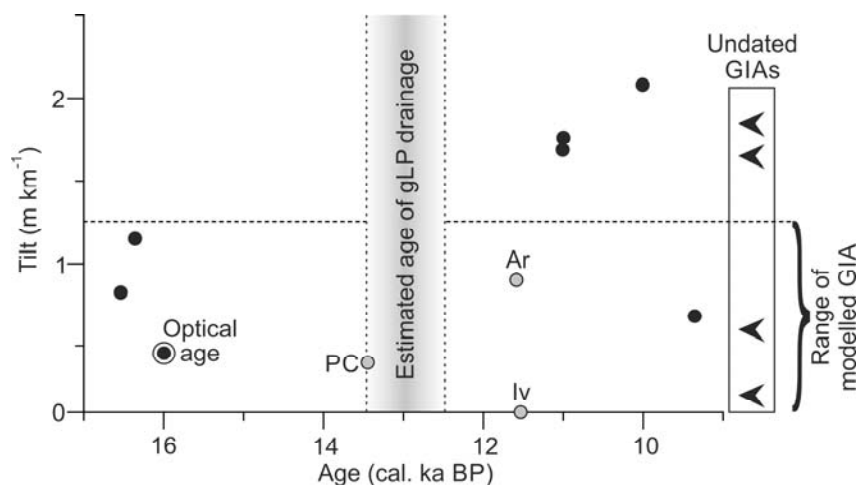
794

795 Fig. 5: Plot of the ten glacioisostatic tilts derived from CIS palaeolake planes with reported ages

796 (black and grey dots, Table 1). Glacial lakes Arrow (Ar) and Invermere (Iv) are highlighted grey

797 because they are geographically near the the Purcell Trench and Glacial Lake Peace, Clayhurst

798 stage (PC), because it's chronologically closest to putative ages for gLP (delineated by the labelled

799 grey bar; Dyke *et al.* 2003; Waitt *et al.* 2009; Table 1). The four undated GIAs reviewed are

800 marked as chevrons. The estimated age of gLP drainage is derived from previous CIS

801 reconstructions (Dyke *et al.* 2003) and tephrochronologic ages (Waitt *et al.* 2009). The distribution

802 of the previously reported lake tilts reveals the propriety of the modelled GIAs.

803
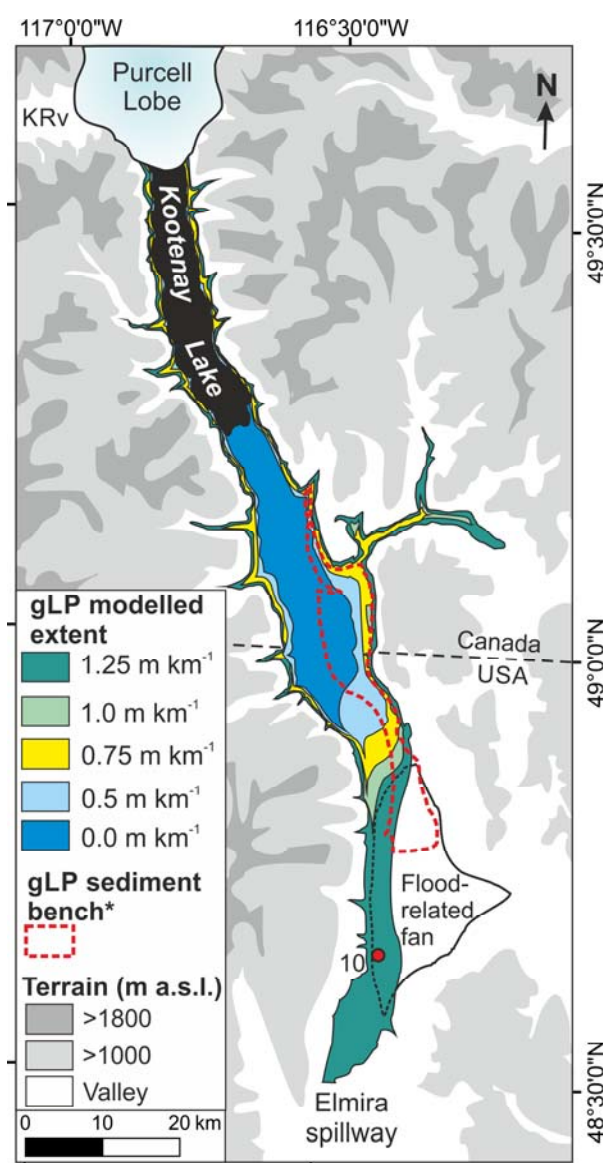
804

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^{-1}) 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.

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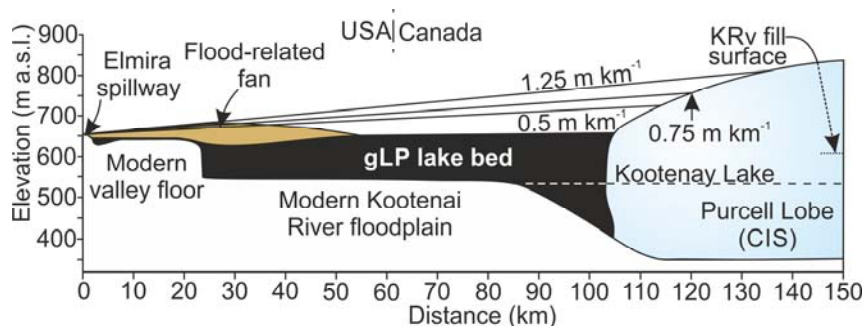
814

815 Fig. 7: Schematic representation of three modelled gLP water surface tilts (Table 2) projected

816 northwards from the Elmira spillway, past the alluvial fan deposited by the drainage of gLK (Alden

817 1953) and over the reconstructed gLP lake bed. The elevations of the pre-incision Kootenay River

818 valley sediment fill and modern Kootenay Lake are also depicted (note the depth of Kootenay

819 Lake during glaciation is schematically represented).

820
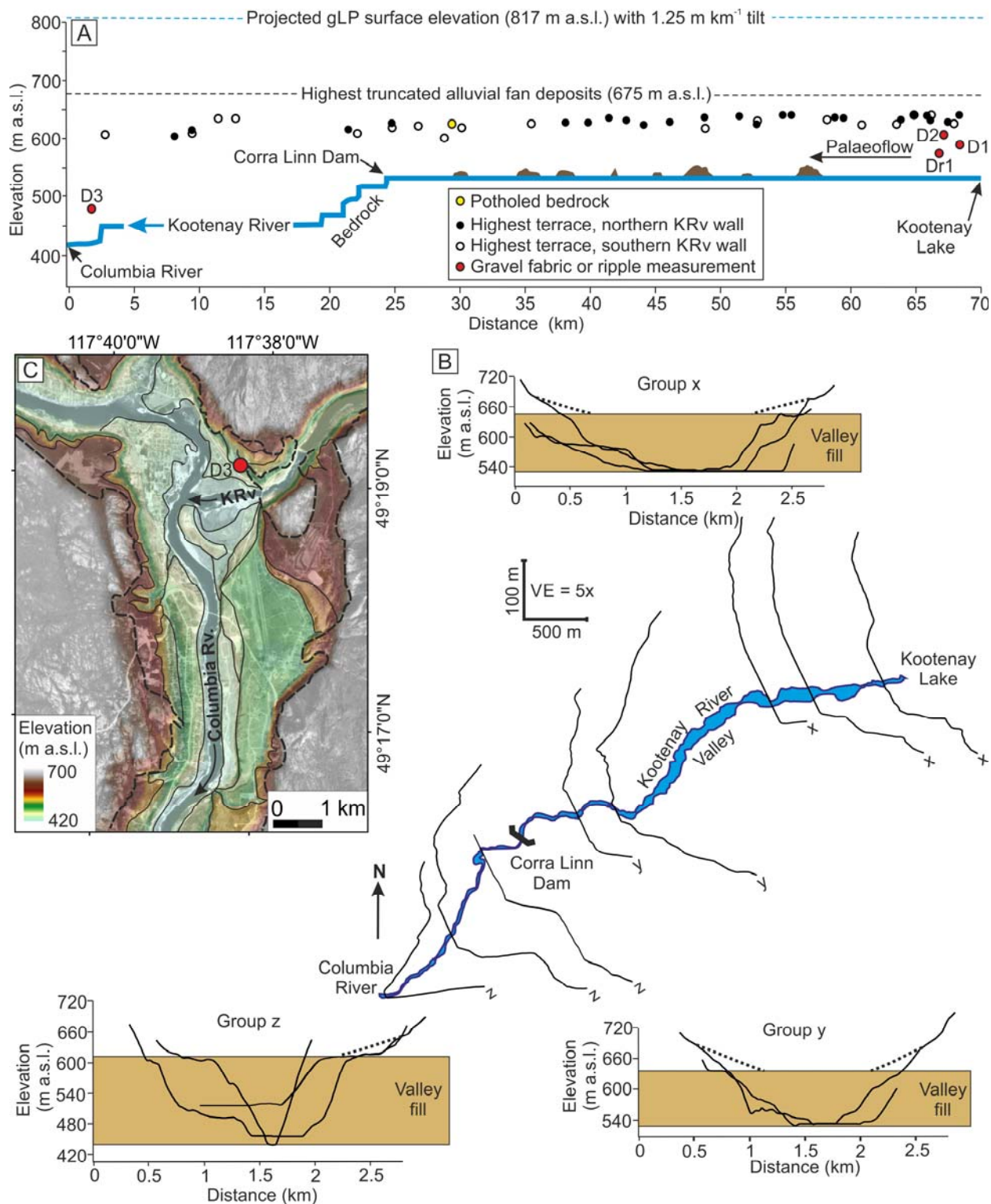
821

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

824   (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

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.

839