Canadian Journal of Earth Sciences
Revue canadienne des sciences de la Terre

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Journal:	<i>Canadian Journal of Earth Sciences</i>
Manuscript ID	cjes-2016-0198.R2
Manuscript Type:	Article
Date Submitted by the Author:	31-May-2017
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Is the invited manuscript for consideration in a Special Issue? :	N/A
Keyword:	Quaternary geology, Glacial history, Central British Columbia, Chronology, Stratigraphy

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Quaternary geology of part of the McLeod Lake map-area (NTS 093J), central British Columbia: lithostratigraphy, glacial history, and chronology

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Abstract

During the last (Fraser) glaciation, the coalescence and subsequent divergence of glaciers near the McLeod Lake map-area is recorded by landforms, ice-flow indicators and the distribution of sediments. Ice initially flowed into the study area from the northwest with at least one fluctuation in the ice-front position. Ice-flow during glacial maximum was generally to the northeast and transitioned to the east during deglaciation. The Quaternary stratigraphic record spans the Fraser Glaciation and is represented by a sequence of advance glaciolacustrine sediments, multiple till units, retreat glaciolacustrine and glaciofluvial sediments and associated post-glacial aeolian material. The surficial geology is dominated by thick, streamlined till that thins where relief is high. Glaciofluvial outwash occurs mostly in northeast and southeast trending meltwater channels, while ice-contact glaciofluvial deposits and ablation till occur in depressions throughout the region. Extensive glaciolacustrine deposits blanket low-lying regions in the southern parts of the study area. Meltwater and ice-flow features suggest that deglaciation in the region was dominantly frontal retreat and that ice was largely active as it retreated through the study area. Post-glacial aeolian activity was brief; optical dating on K-feldspar from aeolian landforms indicate that they had stabilized by between 8.71 – 10.71 and 12.3 – 14.3 ka, and provide minimum ages for ice retreat. The data from the study area are evaluated with similar data from adjacent regions to develop an ice-flow history, and refine the conceptual model of deglaciation for the northern Interior Plateau.

38 Introduction

39 The McLeod Lake map-area, National Topographic System (NTS) 093J, in central British
40 Columbia is situated near the geographic centre of the Cordilleran Ice Sheet (CIS); here glaciers
41 from the Coast, Skeena, Omineca, and Cariboo mountains coalesced, creating a dynamic glacial
42 environment that has previously only been described in limited detail (Tipper 1971*a, b*). The
43 region is characterized by thick drift consisting of sediments and landforms of the Fraser
44 Glaciation allowing a record of glacial events to be determined. Mineral exploration in the region
45 has been limited by the thick drift and infrequent, isolated bedrock exposures. Techniques that
46 utilize glacial deposits, such as till geochemical and mineralogical surveys, are best suited to
47 these environments (Bobrowsky et al. 1995; McClenaghan et al. 2000). Increased mineral
48 exploration interest in this area has necessitated a more detailed understanding of glaciation and
49 its impact on the landscape to plan, execute and interpret these surveys and to provide resources
50 for development.

51 The objective of this paper is to provide an overview of the surficial geology and Quaternary
52 history of the study area, and suggest some regional-scale implications for the understanding of
53 the Fraser Glaciation in central British Columbia. The glacial history of the area is based on the
54 distribution of surficial materials and landforms, lithostratigraphy, and geochronology. The
55 interpretations are extended into adjacent map sheets NTS 093F/G/K and /O (Figure 1), and used
56 to propose a model for deglaciation in the region that provides insight into the growth and decay
57 of the CIS.

58 *Previous Studies*

Only reconnaissance Quaternary work has been done in the McLeod Lake map area (Armstrong and Tipper, 1948; Tipper, 1971*a, b*); however, detailed studies have been conducted in adjacent regions. Regional glacial landform mapping was completed at 1:250 000 scale by Tipper (1971*a*). Prior to the flooding of the Williston reservoir to the north, Rutter (1974*a, b*) mapped the surficial geology and landforms, and documented stratigraphic data from the Peace, Parsnip and Finlay rivers (Rutter 1977). This was furthered with additions to understanding of the stratigraphic record by Bobrowsky and Rutter (1992). Hickin et al., (2015, 2016) discuss stratigraphy, ice dynamics and deglacial geomorphology around the Peace River, east of the Rocky Mountains. Kerr (1991) carried out surficial geology studies of the Mount Milligan area that straddles the east and west sides of map area NTS 093N/1 and 093O/4, respectively. Glacial history has been discussed south of the study area in parts of map sheets NTS 093A/B/G and /H by Clague (1988), Clague et al. (1990), and Levson et al. (1990); in the west and northwest adjacent map sheets NTS 093K/N by Plouffe (1992, 1997, 2000), Plouffe and Williams (1998), and Stumpf et al. (2000); about 125 km west in parts of map sheets NTS 093L/M by Levson (2001); and to the southwest in the northern part of map sheet NTS 093F Levson et al. (1998), Mate and Levson (2001), and Plouffe and Levson (2001*a, b*). For a comprehensive review of previous Quaternary studies and their extents see Sacco (2012).

Physiography

The study area lies within the Interior Plateau of the Canadian Cordillera and includes parts of the Fraser Basin, Nechako Plain and Nechako Plateau physiographic subregions (Holland 1976). The majority of the study area is within the Fraser Basin and Nechako Plain, which are characterized by moderate to low relief between 675 and 1000 m above sea level (m a.s.l.). Surficial deposits are thick enough to mask most bedrock topography, and bedrock exposures are

rare. The northwestern part of the study area is within the Nechako Plateau. Here, the southern portion of the Wolverine Mountain Range is notably steep and rocky with some peaks over 1500 m a.s.l. Surficial deposits are generally less than a few metres thick and bedrock exposures are more numerous, except in valley bottoms where thicker deposits occur.

Methods

Terrain mapping of six 1:50 000-scale NTS map areas (093J05, 06, 11, 12, 13, and 14; Maynard et al. 2013*a, b, c*; Sacco et al. 2013*b, c*; Ward et al. 2013) used standardized terrain mapping methods based on sediment-landform associations (Resources Inventory Committee 1996; Howes and Kenk 1997) supplemented with detailed ground observations. Polygons were delineated and interpreted based on surficial material, surface expression and geomorphological processes using 1:40 000-scale, black-and-white aerial photographs acquired in 1996 and 1997. Point and line symbols were used to indicate features that could not be represented in the polygon labels (e.g. streamlined features, eskers, meltwater channels). Field verification was conducted by truck, ATV and foot traverses. Stratigraphic sections were described and units were characterized using colour, texture, sorting, stratification, sedimentary structures and contact relationships.

Dominant ice-flow directions were determined using streamlined macroforms that are interpreted to represent the dominant ice-flow direction (Figure 2b). These features, including drumlins, flutings, crag and tails, and streamlined bedrock, were identified on aerial photographs and a digital elevation model (DEM) from Canadian Digital Elevation Data (GeoBase© 2009), and supplemented with microform (i.e. striations, rat tails and grooves) and till clast fabric data measured in the field. Relative ice-flow chronologies were determined from microforms based

on truncated rat tails, striations that are formed within grooves, and striations on lee surfaces that would have been protected from later ice-flow. The trends of vertical series of clast a-axis till fabrics were also used to infer potential variations in ice-flow direction throughout the deposition of the till unit. These inferences were used with caution when determining the glacial history if the trends were not corroborated by other evidence as the clast orientations can be influenced by other processes.

Meltwater channels, ice-contact deposits, and glaciolacustrine sediments and landforms identified in the terrain mapping were used to determine the morphology of the glacier margin and drainage systems during deglaciation (*cf.* Fulton 1967; Plouffe 2000; Figure 2b-f). Similar landforms and sediments outside the terrain mapping extent were identified using aerial photography, the DEM, and where possible, verified with ground observations. Lateral meltwater channels were used to determine the shape and successive positions of the ice margin. The minimum extents of glacial lakes were inferred from the maximum elevations of aerially extensive glaciolacustrine deposits and delta topset beds and the minimum elevations of spillways measured from the DEM. The potential sizes and positions of glacial lakes as they evolved during deglaciation are reconstructed by digitally ‘flooding’ the DEM surface to match significant landforms.

Minimum ages for deglaciation were determined using optical dating of aeolian sediments that directly overlie glacial sediments (see lithostratigraphy section for description of aeolian sediments). Optical dating measures the time elapsed since grains of quartz or feldspar were last exposed to adequate sunlight or heat, which for aeolian sediment, corresponds to the time since those grains were buried. The possibility of inadequate sunlight exposure prior to burial is the most significant factor that can lead to an inaccurate optical age. Thus, aeolian sediment is a

good candidate for optical dating because it is typically exposed to sufficient sunlight before burial. Samples were collected by inserting an opaque 15x8 cm ABS tube into a freshly exposed vertical section. Samples were collected near the base of the deposits in an effort to obtain material that was deposited soon after ice retreat. Zones with evidence of bioturbation or other post-depositional disturbances were avoided as these processes can introduce grains with different light exposure histories. All sample tubes were completely filled with material to keep the contents immobile and the slip caps were reinforced with duct tape to provide an air-tight seal to retain the as-collected water contents. Details of the method are discussed in Lian and Roberts (2006) and Lian (2013), and experimental details are reported in Appendix A.

Surficial geology

Most sediments and landforms in the study area are products of the Fraser Glaciation. Figure 2 is a simplified surficial map that represents six 1:50 000 terrain maps (Maynard et al. 2013*a, b, c*; Sacco et al. 2013*b, c*; Ward et al. 2013).

Till is the dominant surficial material in the study area (Figure 2) and is generally a highly compacted matrix-supported diamicton with clast contents ranging from 10 – 30%. The thickest till occurs in low-relief areas where it masks the underlying topography, and is commonly streamlined. In uplands, it mantles the underlying topography as blankets and veneers. Bedrock is exposed on the steepest slopes commonly with colluvial veneers and blankets composed of a mixture of remobilized till and angular bedrock fragments. Colluvial fans composed of stratified sand, gravel and diamictons occur at the mouths of steep drainages. Bedrock is only exposed in isolated outcrops in the rest of the study area.

148 Glaciofluvial sand and gravel deposits occur in a complex of subglacial, ice-marginal and
149 proglacial meltwater channels. Subglacial channels are generally oriented to the northeast,
150 aligned parallel to the deglacial ice-flow direction. Lateral channels are more common, occurring
151 as a series of sub-parallel northwest-southeast oriented features over most of the area. Proglacial
152 channels vary in direction as they were controlled by local topography. West of Carp Lake, thick
153 units of sand and gravel were deposited in a series of proglacial and lateral channels that formed
154 large outwash plains. These and other large plains along the McLeod, MacDougall and Salmon
155 rivers were later incised forming terraces. Glaciofluvial sediment was also deposited in eskers
156 and other hummocky ice-contact features. Small eskers on the flanks of valleys or depressions
157 commonly lead to hummocky glaciofluvial and ablation till deposits, and indicate local ice
158 stagnation. Larger esker systems are commonly associated with subglacial drainage networks.

159 Thick blankets of glaciolacustrine sediments occur in large river valleys and low-lying areas to
160 the south. These deposits are generally composed of laminated beds of fine sand, silt and clay.
161 Holocene gully erosion has created steep slopes in some of these deposits, making them
162 susceptible to mass wasting. Glaciolacustrine sediments in the northern part of the study area are
163 rare, generally occurring in valleys as thin, coarse-grained sand, gravel, or diamict veneers,
164 typical of more ice-proximal deposition. Glacial lake outlets with minimum elevations of 1020 m
165 and 1100 m, and deltaic deposits with a topset elevations of 690 m and 750 m were identified
166 associated with some of these deposits (Figure 2). Glacial lake outlets that occur beyond the
167 extent of mapping, but impounded glacial lakes within the study area, are identified on the
168 appropriate figure in the discussion section. Aeolian veneers and dunes occur overlying and
169 adjacent to glaciofluvial and glaciolacustrine deposits. These landforms are composed of fine

170 sand and silt, and are typically massive or show minor cross-bedding. Although thin, these
171 aeolian deposits are common throughout the study area, and can be aerially extensive.

172 Modern fluvial deposits composed of sand and gravel are typically restricted to floodplains
173 within meltwater channels as most modern drainage follows systems created during deglaciation.

174 Organic deposits are common around slow moving streams, abandoned meltwater channels and
175 small lakes, and occur in depressions on a many poorly drained substrates such as till, fine-
176 grained glaciolacustrine material, or bedrock.

177 **Lithostratigraphy**

178 Numerous stratigraphic sections with varied exposure were described in the study area. Six
179 sections that are representative of the Quaternary stratigraphy of the study area are described
180 (Figures 2, 3). Sections DS-01, DS-02 and DS-03 are natural bluffs on the McDougall River.

181 Section DS-01 reveals the internal strata of a roughly 15 m high, northwest oriented drumlinoid
182 feature, and an additional ~50 m of sediment below the drumlinoid. Sections DS-02 and DS-03
183 expose the stratigraphy of glaciofluvial terraces in the river valley. Section DS-04 occurs within
184 a meltwater channel, and DS-05 on the east shore of McLeod Lake. Section DS-06 occurs
185 adjacent to several southeast trending meltwater channels and was created during the excavation
186 of a gravel pit.

187 Nine lithostratigraphic units are designated to facilitate description and interpretation of the
188 Quaternary succession (Table 1). The oldest units are described first. The relative chronology for
189 the stratigraphic units is determined by relative stratigraphic position.

190 Unit 1 was observed at section DS-01 and DS-04, which occur in the same north-draining
191 watershed. This unit is divided into three sub-units. Unit 1a (DS-04) consists of >1.5 m of

massive, moderately well sorted, medium and coarse sand. The lower contact of this unit is covered. Unit 1b (DS-04) is a 5.5-m coarsening-upward sequence of laminated and finely bedded clay, silt and sand with localized faulting and at least one clastic dyke. The lowest portion of unit 1b is composed of laminated clay and minor silt that grade into dominantly silt laminations near the middle, and finely bedded silt and fine sand at the top. Localized faulting is concentrated in the upper metre and around the clastic dyke. The exposed portion of the dyke is 10 – 15 cm thick and composed mainly of laminated clay that generally parallels the dyke walls. It penetrates obliquely to a depth of about 1 m over a length of about 3 m. Unit 1c (DS-01) comprises about 11 m of interstratified sand, gravel and diamict beds with an apparent dip of 10° to the east. The lowermost 2 m is a dark grey, silty, matrix-supported diamict containing massive to weakly stratified muddy sand lenses. It has 15% subrounded to rounded clasts, and the upper portion passes into 20 cm of normally graded fine sand and silt. The lower diamict is conformably overlain by 5 – 30 cm undulating and discontinuous beds of fine to coarse sand, well rounded pebble gravel and minor lenses of diamict (Figure 4b). Fine sand beds in the upper part of this exposure contain soft-sediment deformation and dewatering structures (Figure 4c). The lower contact is covered.

Unit 1 is interpreted as advance-phase glacial lake sediments. In unit 1a, the sediment size, sorting and lack of structure suggest that this unit was deposited rapidly in a shallow lake, although a fluvial genesis is possible. In unit 1b, the fine-grained, laminated sediments are consistent with sedimentation in deep water. The coarsening upwards indicates increasing energy levels, likely due to higher flow regimes as a result of advancing ice. Localized faulting in the upper section is interpreted as glacial tectonization. The brittle deformation around the clastic dyke and laminations parallel to the dyke walls suggests it was injected into the substrate (*cf.* van

215 der Meer et al. 2009), possibly when ice was overriding the site. The sequence of sediments and
216 dewatering structures in unit 1c are consistent with a prograding subaqueous fan (Reading 2009).
217 The gently dipping, discontinuous sand and gravel beds were likely deposited as the fan
218 aggraded into a glacial lake. Diamict lenses are likely sediment gravity flows down the fan
219 surface.

220 Unit 2 is observed at DS-01 where it comprises 15 – 17 m of compact, grey, sandy silty, matrix-
221 supported diamict with good fissility and jointing. Unit 2 may also occur in the north central part
222 of the study area where southeastward and northeastward trends were measured in till clast a-axis
223 fabric 09-01 (see ice-flow history section; Figure 7). It has 10 – 20% clasts of mixed lithologies.
224 These clasts are commonly striated and faceted. In the lower 40 cm of unit 2, the matrix is
225 coarser which is interpreted as a gradational contact with the underlying unit 1c.

226 Unit 2 is interpreted as basal till based on compaction, sorting, clast morphology, and the
227 occurrence of fissility and jointing caused by glacitectonic shear (Benn and Evans 2010). Its
228 stratigraphic location below another glaciogenic diamict interpreted as basal till (unit 4) and
229 interbedded sand, silt and diamict (unit 3) indicate this till was likely deposited during an early
230 glacial advance. The beds dipping to the east in the underlying subaqueous outwash (unit 1c)
231 suggest that the advance came from a source area to the west. The exact age of units 1c and 2 are
232 unknown, but the lack of oxidation suggests these were deposited during the Fraser Glaciation.

233 Unit 3 only occurs at section DS-01. Its lower contact with unit 2 is irregular, consisting of
234 chaotically interbedded diamict and sand. It is 20 – 25 m thick and dominated by ~1-m thick
235 diamict layers and interbeds of coarse sand and moderately to well rounded pebble gravel

between 0.5 – 1 m thick. Laminated and thin-bedded silt and fine sand beds occur near the top of the unit.

The interbedded sand and diamict layers of Unit 3 are consistent with a subaqueous outwash fan, and the fine-grained laminations indicate the sediment was likely deposited in a glacial lake. The high frequency of diamict beds, which are interpreted as subaqueous debris flows from the glacier, suggests they may have been deposited in close proximity to ice. Thinly-bedded silt and sand deposited near the top of this unit indicate a lower energy environment, implying retreat of the ice margin. Unit 3 contains sediments related to ice-proximal and distal conditions and occurs between two diamicts interpreted as basal till, thus it was likely deposited during an oscillation of the ice margin. An alternate hypothesis is that this unit was deposited subglacially.

Unit 4 occurs in sections DS-01/02/03/04/05 and in most surface and shallow exposures throughout the study area. It is a massive compact diamict and supported by a matrix of sandy silt, and typically shows good fissility and moderate jointing. Most of this unit is dark grey, except where it has weathered to light brown. Exposed thicknesses range from 5 to 25 m. In the stratigraphic exposures, subangular to subrounded clasts of mixed-lithologies comprise 10 – 20% of the material, although clast contents of up to 30% were observed. Clast a-axis fabrics were measured from this unit in various exposures throughout the study area and generally show a strong unimodal distribution parallel or sub-parallel to the local orientation of streamlines macroforms. Clast of unit 4 in sections DS-02 and DS-03 become slightly more rounded and diverse in lithology higher in the exposures. Sorted lenses of sand were observed in unit 4 at section DS-03. The lower contacts of unit 4 are commonly erosive (DS-02 and DS-04), but also occur as gradational (DS-03) or interbedded (DS-01). Unit 4 is interpreted as basal till based on compaction, sorting, clast morphology, the occurrence of fissility and jointing, the strong

unimodal clast a-axis fabric, and regional extent. This unit occurs at surface throughout most of the study area, and therefore, is interpreted as Fraser glaciation till.

Unit 5 was observed remotely from the opposite side of the river overlying unit 4 at section DS-01; only a limited description is possible. It is a crudely stratified diamict ranging in thickness from 3 to 6 m. It is dominantly matrix-supported with some clast-supported areas. The clast content appears to range from about 20 to 60%. The lower contact is defined by a thin, apparently continuous, large cobbly-gravel lag that is roughly parallel to the surface. This unit is interpreted as ablation till based on the sorting, crude stratification and stratigraphic location over basal till (unit 4). The clast lag observed between the units may represent basal sediment that was washed by subglacial meltwater prior to the deposition of en- and supra-glacially transported material.

Unit 6 is divided into two sub-units and was observed at section DS-03 (unit 6a and b) and DS-05 (unit 6b). Unit 6a is about 6 m thick and fines upwards from well rounded pebble gravel into stratified, cross-bedded medium sand that is sharply overlain by another layer of pebble gravel. The lower contact is gradational with unit 4 over 5 cm. Throughout the unit are discontinuous, 25 – 75-cm-thick, silty sandy matrix-supported diamict beds with sharp upper and lower contacts.

Unit 6b consists of laminated and finely bedded silt and sand. At section DS-05, this unit consists of 2.1 m of laminated silt and fine sand with 1 – 3 cm thick clay laminations at 10 – 20 cm intervals. At section DS-03, unit 6b consists of 1 m of laminated and finely bedded silt and fine sand with dropstones. The lower contact of unit 6b is conformable with unit 6a at section DS-03 and with unit 4 at section DS-05. Unit 6a is interpreted as an ice-proximal subaqueous fan deposit because of the interbedded sand and gravel, the discontinuous lenses of diamict that are consistent with subaqueous debris flows, and its stratigraphic location beneath more distal glacial

282 lacustrine sediments (i.e. Unit 6b). The alternation between sand and gravel suggests different
283 flow regimes, likely due to different meltwater discharges. The fine-grained laminations of Unit
284 6b indicate farther proximity from the ice where distal underflow and suspension settling occurs
285 in glacial lakes (Shaw and Archer 1979). Unit 6 is interpreted as deglacial glaciolacustrine
286 sediment because it stratigraphically overlies basal till.

287 Unit 7 is observed at sections DS-02/03/05/06 and in numerous shallow exposures throughout
288 the area. It is composed of moderately sorted to well sorted, rounded to well-rounded, pebble- to
289 cobble-sized gravel that generally fines upwards. Deposits are typically clast-supported with a
290 coarse sand matrix. All observed lower contacts are erosive. In stratigraphic exposures, these
291 deposits range from <1 to 5 m thick, but may be thicker in large outwash corridors. At section
292 DS-06, unit 7 is composed of moderately rounded to well rounded, moderately sorted gravel that
293 coarsens and becomes less sorted upwards. At section DS-02, discontinuous beds of unit 7 occur
294 perched on narrow terraces in the diamict at 20, 29 and 35 m above present-day river level. Unit
295 7 is interpreted as glaciofluvial sediment and, based on its stratigraphic position over till, is
296 interpreted to have been deposited during ice retreat.

297 Unit 8 was only observed at section DS-06. It is as a compact, dark grey, silty sand, matrix-
298 supported diamict with less defined jointing and fissility than units 2 and 4. It has a 20% clast
299 content, which is of mixed lithology, with stones that are moderately rounded; some are striated
300 and faceted. The lower contact is gradational over 10 cm with the underlying gravel of unit 7.
301 This unit is interpreted as basal till based on compaction, sorting, clast morphology, and the
302 occurrence of fissility and jointing. It is slightly less consolidated and thinner than till unit 4.
303 Unit 8 is underlain by unit 7 (retreat glaciofluvial) and unit 4 (basal till) indicating it was

deposited during a re-advance. The re-advance was likely a temporary oscillation of the ice margin during the retreat phase.

Unit 9 does not occur at any of the described stratigraphic sections, and therefore, is not depicted in the stratigraphic logs. It consists of well-sorted silt to fine sand with cross-bedding apparent in some exposures, and was observed overlying or adjacent to glaciofluvial and glaciolacustrine deposits. Unit 9 is interpreted as aeolian material that was sourced from unvegetated glaciofluvial and glaciolacustrine deposits. Most aeolian landforms are discontinuous mantles or poorly-formed dunes generally less than two metres thick with some dunes reaching several metres in height. The thin-nature of the deposits, generally poorly developed dune structures, and lack of identified buried soil horizons suggest that aeolian activity was short-lived with no indication of reactivation (Sacco 2012).

Chronology

The chronology of deglaciation in the study area is only understood in a broad context. Regional radiocarbon ages suggest that the study area was ice free before $\sim 10\,200\text{ }^{14}\text{C yr}$ ($\sim 12\,000\text{ cal yr}$) BP (Table 2). Refinement of this hypothesis using additional radiocarbon dating is restricted by the lack of datable organic material from the earliest part of the post-glacial period. For example, a piece of wood collected from the base of a peat deposit in an esker system yielded an age of $8\,775 \pm 30\text{ }^{14}\text{C yr BP}$ (UCAIMS 83989; Table 2; Figure 1). This age provides a minimum age for ice retreat beyond the esker system, but more accurately records the timing of initial post-glacial peat development, which significantly post-dates deglaciation. Therefore, we use optical dating to determine when post-glacial aeolian material was deposited as a proxy for ice retreat. The use and accuracy of this proxy is based on the assumption that the most recent aeolian activity

occurred soon after ice retreat. The true age for deglaciation is the optical age plus the duration of aeolian activity before final burial of the sampled sediment.

Four aeolian sediment samples were processed and analysed (samples MLD1, CLR5, OCDF4 and OCDF5; Figure 2). Samples MLD1, CLR5, and OCDF5 were collected from aeolian mantles about 50 cm above the underlying glaciofluvial material. Sample OCDF4 was collected near the base of a well-formed dune with a crest-height of about 2 m. The underlying contact could not be exposed.

Sample OCDF5 yielded optical ages of 8.79 ± 0.51 ka for quartz and 9.94 ± 0.77 ka for fine-grained K-feldspar (Table 3); the quartz age is considered minimum age for this sample as it failed a dose-recovery test (see Appendix A). Sample CLR5 yielded a fine-grained K-feldspar age of 9.35 ± 0.64 ka, and sample OCDF4 yielded a fine-grained K-feldspar age of 13.3 ± 1.0 ka. Sample MLD1 was abandoned; see discussion in Appendix A. All K-feldspar ages were corrected for anomalous fading using the method of Huntley and Lamothe (2001). Figure 5 summarizes the experimental methods.

Ice-flow history

Streamlined macroforms were identified during mapping and are interpreted to represent ice flow at Fraser Glaciation maximum and possibly during early deglaciation (Figure 6). These landforms indicate that northeastward ice-flow was dominant with minor variation to the north and the east (Figure 6). The northward deviations are probably the result of local topographic control on a thinning ice sheet. East-west oriented landforms are only observed in the western part of the study area.

Microforms (striations, grooves and rat tails) were measured at 41 sites. Specific flow directions were determined at nine locations (Figure 6). From these data, three broad trends and relative ages are identified. Striations of the oldest group were measured at eight locations in the northern part of the study area with orientations from $320^{\circ} - 140^{\circ}$ to $280^{\circ} - 100^{\circ}$ (Figure 6; green symbols). They are nearly perpendicular to the streamlined features in some areas and are interpreted to record early ice flow from a source area in the Omineca Mountains. The second oldest group of striations has orientations from $190^{\circ} - 10^{\circ}$ to $260^{\circ} - 80^{\circ}$ (Figure 6; black symbols). Although the range of orientations in this group is high, they are in agreement with the trend of the local streamlined features; thus they likely developed contemporaneously during glacial maximum and the early stages of deglaciation. Striations of the youngest group were measured at eight locations only in the western portion of the study area and range from an orientation of $250^{\circ} - 70^{\circ}$ to a direction of 95° (Figure 6; red symbols). They are interpreted to relate to ice flow during later stages of deglaciation. At one key site, three ages of ice flow were determined: the oldest flow direction is 60° , the second trends at $80^{\circ} - 260^{\circ}$ and the final is 95° . The oldest striation set in the series is aligned with local streamlined macroforms. The younger two striation sets represent the transition to eastward ice-flow during deglaciation.

Till clast a-axis fabrics were measured at 13 locations in the study area (Figure 7; Table 4). At four locations (sites 6 – 9), two fabrics were measured at different depths in vertically aligned positions to identify potential changes in ice-flow direction. Most fabrics are thought to be within Unit 4 and are in general agreement with the orientations of the local streamlined macroforms, with a few exceptions. Fabric 09-01, with a major trend at $350^{\circ} - 170^{\circ}$ and a weaker one at $225^{\circ} - 45^{\circ}$, is interpreted to have developed primarily during the southeastward ice-advance and was

partly realigned during the north-northeastward flow indicated by the orientation of streamlined landforms.

The ice-flow direction inferred from fabric 09-03 is about 350° . However, the macroforms in the area are oriented in a more northeastward direction, at about 35° . Similarly, fabric 09-05 has a minor northward trending component. The primary trend in fabric 09-05 is about $85^\circ - 265^\circ$ and reflects ice flow during the late-glacial phase. The origin of these northward components has not been resolved.

The principal eigenvector of fabric 09-06 indicates ice flow to the northeast, which is consistent with the macroforms. The clast measurements, however, are spread along the outer edge of the southwest and northeast quadrants suggesting that more than one ice-flow direction may have affected the fabric; possibly during the transition to more eastward flow during the deglacial phase. Otherwise, the dispersion may simply be due to till-forming processes (e.g. local changes in the strain direction, clast collisions).

Two of the vertical fabric series show differences in dominant trend. In the west-central part of the study area, the lower till fabric (10-01) trends to the northeast (45°), parallel to the fluting from which it was collected. The upper fabric (10-02) trends more eastward (at 85°). This is interpreted to reflect a shift in ice-flow direction to the east during deglaciation, which is corroborated by sequenced striations to the north and south (Figure 6). In the northeast part of the study area, the lower till fabric (11-02) is in agreement with the macroforms, trending about 15° . The upper fabric (11-01) trends more to the northeast at about 50° , similar to that of a local striation measurement (Figure 6). The eastward deviations of the fabric trend and striations from

the macroforms are interpreted to reflect an eastward shift in ice-flow during deglaciation due to topographic influence (see discussion).

Three stages of ice-flow were determined from the orientations of macroforms, microforms and clast fabrics. During the first stage, ice flowed southeast into at least the northern part of the study area; it is unknown how much of the area was affected by this event. The second stage was to the northeast, which was the direction ice was flowing during the glacial maximum. During deglaciation, ice-flow direction transitioned back to the east. The absence of late-phase, eastward oriented indicators in the northeast suggest that the area was deglaciated when the southern and western sector were under the influence of eastward flowing ice.

Discussion

Local glacial history

At the onset of the last glaciation, ice advanced toward the study area from source areas in the Skeena, Omineca, Coast and Cariboo mountains. Initially, southeast flowing ice from the Omineca Mountains impeded drainage to the north through the ancestral Parsnip River north of McLeod Lake, forming a glacial lake of unknown extent in the northeast corner of the study area. This ice was likely the first to reach the study area. Unit 1a (moderately well sorted sand, section DS-04) was likely deposited in the early phases of the glacial lake. As the lake level increased, unit 1b (glaciolacustrine clay, silt and sand, section DS-04) was deposited. When ice was near section DS-01, a subaqueous outwash fan (unit 1c) formed. Ice then advanced, most likely toward the southeast as evidenced by the orientation of the early striations (Figure 6, green symbols) and the southward oriented till fabric (Figure 7, fabric 09-01), depositing unit 2. The extent of this early advance is unknown, but no evidence of southward flow was observed south

of $\sim 54.7^\circ$ N, thus the southern portion of the study area may not have been covered by ice at this time. Ice then retreated back to at least the location of section DS-01 when more subaqueous outwash (unit 4) was deposited over the early advance till.

As glaciation intensified, ice from the Coast, Omineca and Cariboo mountains coalesced producing a north eastward flow that continued throughout glacial maximum. The streamlined macroforms, striations (Figure 6, black symbols) and many of the till fabrics throughout the area record this northeast ice-flow.

As the climate ameliorated near the end of the Late Wisconsinan, the CIS began to retreat. Initially, the ice front thinned along the western flanks of the Rocky Mountains and then retreated to the southwest, toward the study area. Northward oriented macroforms in the northeast part of the study area (Figure 6) indicate that the northward oriented Rocky Mountain Trench caused local variations in ice flow to the north. Stacked till fabrics measured south of the trench (fabrics 11-02 and 11-01; Figure 7) indicate a transition in ice-flow direction from north, back to northeast, likely when the ice retreated enough that flow was not affected by the Rocky Mountain Trench. At this time, subglacial meltwater flowed north to the northeast through several systems into the Peace River Basin (Figure 8a). The longest meltwater system can be traced from Salmon to McDougall valleys. It transitions between eskers and subglacial channels as it crosses the drainage divide of Mount Mackinnon and continues over the uplands and beyond the study area. Two meltwater systems developed off the arms of Carp Lake and extended northeast into the northern McLeod Lake valley. Shallow meltwater channels and a series of eskers indicate that, at some point, meltwater from the east arm system was diverted north into McLeod River valley (Figure 8a).

Ice retreated from the northeast part of the study area along a mostly straight, northwest-southeast-trending margin as indicated by the orientation of lateral meltwater channels (Figure 8b). Meltwater continued to be discharged into the McLeod Lake valley, but was blocked just north of the study area, forming a large glacial lake, informally named glacial Lake McLeod. Silt and sand (unit 6b, section DS-05) were deposited in this lake about 20 m above the level of present-day McLeod Lake. Meltwater moving through the McDougall River valley was diverted southeast along the retreating ice margin, cutting a series of channels leading to the McLeod River; the youngest channel is the present course of the McDougall River. Meltwater, partly supplied by subglacial drainage flowing over the north end of Mount Mackinnon, continued to be directed into the uplands and through the McDougall River until the ice retreated to lower elevations (Figure 8b). Ice continued to retreat and a sediment or ice dam caused local ponding of a glacial lake in McDougall River valley. Cross-bedded sands were deposited by underflow currents, while failures from the unstable valley walls or ice front deposited discontinuous diamict beds (unit 6a; section DS-03). It is unlikely that both glacial lakes were connected due to the ~100 m difference in elevation between glaciolacustrine units in the McDougall River and McLeod Lake valleys. A 100 m deep lake extending from McLeod lake to the location of this section would submerge a significant area, from which no glaciolacustrine sediments or landforms were mapped. Eventually the blockage was removed and cobble gravel was deposited on the valley floor (unit 7; sections DS-02 and DS-03).

The ice mass that impeded northward drainage forming glacial lake McLeod is roughly delineated by hummocky gravels, a large esker complex and a large kettle lake (presently Tudyah Lake) (Figure 9a). The southern limit of the lake was the retreating ice margin and the lake expanded as this margin retreated up the valley. The highest water level was at least 750 m

a.s.l. as indicated by deltaic sediments, a spillway draining east into the Parsnip River valley, and a south-dipping ice-marginal channel cut into the east valley wall that connected glacial lakes McLeod and Fraser (Figure 9a). At this water level, the lake occupied about a 100 km length of the valley between the blocking ice mass on the north and the ice margin on the south (Figure 9a).

Continued ice retreat exposed the southern part of the Crooked River valley and widened the connection between glacial lakes McLeod and Fraser. The water elevation was likely controlled by the ice mass in the north. A second lake elevation of 690 m a.s.l. is inferred from the elevation of topsets on two deltas at the mouth of the McLeod River (Figure 9b). At this elevation, the lake could be contained by an ice mass the size of Tudyah Lake. Many of the topographic highs in the southern part of the valley were exposed. Glacial Lake Fraser was likely still draining north into glacial Lake McLeod as its water spilled over the drainage divide and through the channel at an elevation of 715 m a.s.l. in the south end of the valley.

As the ice continued to retreat, meltwater was diverted from the McDougall River and flowed southeast along the ice margin to Carp Lake (Figure 8c). A chaotic system of meltwater channels developed in which subglacial channels carried water northeast towards the ice margin. As the ice margin retreated, new lateral channels flowing to the southeast progressively formed at lower elevations, following the ice front position and cross-cutting the subglacial channels. Outwash plains that end abruptly at the limits of Carp Lake suggest that they formed against stagnant ice blocks that occupied the lake basin. A proglacial lake developed around Weadon Lake to a minimum elevation of 780 m a.s.l. Northern drainage was impeded by an ice mass in the Crooked River valley (Figure 8c, 9a). A series of eskers indicates some drainage below the ice

479 mass, and a channel cut into the west wall of the Crooked River valley suggests that drainage
480 also occurred adjacent to the ice mass (Figure 9a).

481 When the northern ice margin began retreating down the uplands in the northwest part of the
482 study area, several ice-marginal lakes were impounded and perched glaciolacustrine sediments
483 were deposited (Figure 8c). Spillways indicate a minimum elevation of about 1020 m a.s.l. for
484 the northern lake and 1100 m a.s.l. for the southern lake. The lakes existed along the retreating
485 ice margin until the water was able to drain southeast, south of Mount Mackinnon.

486 When the ice margin was south of the drainage divide separating the Arctic and Pacific
487 watersheds, all proglacial meltwater began flowing southeast along the ice margin (Figure 8d).
488 The ice front configuration, as reconstructed from the position of the lateral meltwater channels,
489 was generally straight and retreating to the southwest. During the retreat, there was a reduction
490 of influence from ice in the south resulting in a transition to eastward ice flow and a curved ice
491 margin (Figure 8e). The transition to eastward ice-flow is illustrated in the southwest by
492 macroforms (Figure 6), till fabrics 10-01, 10-02, and 09-05 (Figure 7), and striations northwest
493 of Great Beaver Lake (Figure 6, black and red symbols). Striations on the west side of NTS
494 093J12 (Figure 6, red symbols) and till fabric 09-06 (Figure 7) also indicate eastward ice flow.
495 At least one oscillation of the ice front was associated with the changing ice dynamics. This is
496 recorded at section DS-06 where a thin layer of till (unit 8) overlies outwash gravel (unit 7).

497 A glacial lake formed around what is now Great Beaver Lake to a minimum elevation of 795 m
498 a.s.l. (Figure 8e). Ice-contact deposits suggest that the two possible outlets at 780 m a.s.l. were
499 blocked by ice. The lake drained through the lower Salmon River valley.

Ice retreated south across the southern adjacent map area (Clague 1988) and west through the western adjacent map area (Plouffe 2000), implying that the ice margin continued to diverge beyond the study area. The western ice lobe retreated across the Interior Plateau, eventually breaking up into ice tongues that occupied depressions such as the Babine, Francois and Takla Lake valleys (Levson et al. 1998; Stumpf et al. 2000). Meltwater from these glaciers drained into glacial Lake Fraser, which is thought to have occupied much of the Fraser Basin including parts of the Nechako and Stuart river valleys (Tipper 1971a; Clague 1988, 1989; Plouffe 1997, 2000). The exact configuration and morphology of glacial Lake Fraser is unknown. It has been suggested that it existed as a series of lakes in multiple basins (Tipper 1971a), and conversely, as a single lake reaching a surface elevation of 800 m a.s.l. (Clague 1987). The elevation of the spillway into the Peace River basin limited the elevation of the northeastern extent of the lake to 715 m a.s.l., although early phases of the lake may have reached an elevation of 750 m a.s.l. when glaciers still covered the spillway. Elevations of mapped glaciolacustrine sediments attributed to glacial Lake Fraser increase to the southwest, and to a greater extent to the west (e.g. 760 – 780 m a.s.l. around Stuart lake (Plouffe 2000; this study); 760 m a.s.l. in the Prince George area (Tipper 1971b); 800 m a.s.l. west of Prince George (Blais-Stevens and Clague 2007); and 875 m a.s.l. southwest of Vanderhoof (Sacco et al. 2014); Figure 10). However, there is a decrease in elevation of the glacial lake sediments west of Vanderhoof (Plouffe 2000). The inconsistency between these measurements suggest that glacial Lake Fraser was a time-transgressive complex, likely influenced by the retreating ice margin and differential isostatic depression.

The glacial lake that occupied the Stuart Lake basin was likely isolated from the early phases of glacial Lake Fraser by a curved section of the retreating ice margin. The ice margin may have

523 crossed the Stuart River valley where gullied glaciolacustrine sediments transition into a more
524 defined meltwater channel (figure 11). The large subglacial channel and inset esker complex
525 extending southeast could have formed during the draining of the lake. As the ice thinned,
526 hydrostatic pressure forced the lake to drain subglacially and cut the large channel. As flow
527 slowed, the large esker system was deposited within the channel. Wood collected at the base of a
528 peat deposit from an excavation pit between two esker ridges was submitted for radiocarbon
529 analysis in an attempt to date the timing of deglaciation. The sample yielded an age of 8775 ± 30
530 ^{14}C yr BP (UCAIMS 83989; Table 2; Figure 1 and 11). It provides a minimum age for ice retreat
531 beyond the esker system, but most likely indicates the timing of initial post-glacial peat
532 development, which likely significantly post-dates deglaciation.

533 Reported radiocarbon ages from west of the study area suggest that ice retreated from the region
534 before $\sim 12\,000$ cal yr BP (Table 2). Based on radiocarbon chronology, we suspect that the
535 feldspar ages from samples OCDF5 (9.94 ± 0.77 ka) and CLR5 (9.35 ± 0.64 ka) are younger
536 than deglaciation. It is possible that aeolian activity continued for over 1000 years. However, this
537 explanation is not in agreement with the morphology of the landforms; if activity persisted for
538 that length of time, the dune landforms should be better developed and some migration from
539 their source areas would be expected (Sacco 2012). Alternatively, there may have been a brief
540 period of aeolian reactivation on a regional scale. Between 8000 and 10 000 yr BP, the climate in
541 the southern Interior was warmer and dryer than it is today (Hebda 1995). If this climate trend
542 extended north to the study area, some of the aeolian deposits could have reactivated, resulting in
543 the younger optical ages. Sample OCDF4 (13.3 ± 1.0 ka) yielded a fine-grained K-feldspar age
544 that is in better agreement with the reported regional radiocarbon ages, but is significantly older
545 than samples OCDF5 and CLR5. If this age is accurate, the younger ages (samples OCDF5 and

CLR5) suggest there was a period(s) of reactivation that only affected some landforms. Alternatively, sample OCD4 contains some grains that were not exposed to sufficient sunlight. More chronologic data is needed to better resolve the timing of deglaciation in the region.

Regional ice-flow model for the northern Interior Plateau

The ice-flow model is described below in 3 phases: 1) ice-expansion; 2) glacial maximum; and 3) late glacial (Figure 12). The ice-expansion phase probably lasted late into the last glaciation (Ryder and Maynard 1991; Bobrowsky and Rutter 1992; Stumpf et al. 2000). During this time, ice accumulated and then expanded into the Northern Interior of British Columbia from the north, west, southwest and southeast (Clague 1988, 1989; Plouffe 1997, 2000; Stumpf et al. 2000; Plouffe and Levson 2001a). The first glaciers that developed in the Omineca and Skeena mountains flowed southeast down the northern Rocky Mountain Trench and the Takla and Babine lake valleys (Ryder and Maynard 1991; Plouffe 1997; Stumpf et al. 2000; Levson 2001; green arrows, Figure 12a). Multiple striation sites and a till fabric from this study suggest that the glaciers also flowed southeast across the central part of the northern Interior Plateau and at least into the northern uplands of the study area.

Ice from the Kitimat Ranges flowed eastward across the northern Interior Plateau (Plouffe 1997, 2000; Stumpf et al. 2000; Plouffe and Levson 2001a). These glaciers interacted with the glaciers sourced in the Skeena and Omineca mountains producing an eastward transition in their flow direction (yellow arrows, Figure 12a). Eventually, ice from the Cariboo Mountains and Pacific Ranges reached the northern part of central British Columbia (Clague 1988, 1989; Plouffe 1997, 2000), which caused a deflection in flow to the northeast (blue arrows, Figure 12a). The exact locations of the ice confluences are unknown. Based on the proximity of the Omineca and

568 Skeena mountains to the north, the Kitimat Ranges and Hazelton Mountains to the west and the
569 Cariboo Mountains to the southeast, it is likely that ice masses from these source areas
570 coalesced, near the study area prior to the arrival of ice from the Pacific Ranges.

571 In the northern Interior of British Columbia the glacial maximum is thought to be marked by an
572 eastward shift of the ice divide from the Coast and Skeena mountains over Babine Lake Valley,
573 extending south to the Francois and Ootsa lake areas (Levson et al. 1998; Stumpf et al. 2000).
574 The east-west oriented macroforms in the adjacent map-area, NTS 093K (Plouffe 2000) reflect
575 the influence of this ice divide. In the study area, the dominance of northeastern ice-flow
576 indicators suggest that flow was controlled by the interaction between this ice divide and an ice
577 divide in the south through the maximum stage (Figure 12b).

578 During the late-glacial phase, the ice sheet thinned and retreated from the Rocky Mountains, and
579 ice flow at the margin became more influenced by the local topography. Initially, northern
580 deviations occurred as the ice surface lowered through the valleys north of the study area (orange
581 arrows, Figure 12c). As the ice retreated past the influence of these valleys, it began to flow more
582 to the east. The transition was likely affected by the reduced influence from ice flowing from the
583 south, due to downwasting and stagnation (Fulton, 1967, 1991; Plouffe et al. 2011). This allowed
584 an increased influence from the ice divide to the west, causing the final regional eastward ice-
585 flow (red arrows, Figure 12c).

586 *Conceptual model for the deterioration of the CIS*

587 Fulton's (1967, 1991) deglacial model for the Southern Interior describes active ice, transitional
588 upland, stagnant ice and dead ice phases. Although there are physiographic similarities between
589 northern and southern parts of the Interior, the absence of landforms associated with large-scale

dead and stagnant ice indicate Fulton's model is not entirely applicable in the north (*cf.* Plouffe 2000). This is likely influenced by: 1) the ice divide in the Northern Interior that largely remained east of the Coast Mountains during the early phases of deglaciation and connected with valley glaciers when it shifted back to the mountains (*cf.* Levson et al. 1998; Stumpf et al. 2000), 2) the proximity to local accumulation centres (Plouffe 2000), and 3) the limited occurrence of deep topographic lows that served as loci for stagnation in the south.

Plouffe (2000) applied Fulton's (1967) model of deglaciation with minor modifications to map-areas NTS 093N/K, west of the study area. He illustrates that ice tongues remained active and connected to their sources throughout deglaciation. Levson et al. (1998) and Stumpf et al. (2000) also conclude ice movement and connectivity late into deglaciation over the Nechako Plateau. In this study area, the lack of regional ice stagnation deposits, relative deglacial ice flow chronologies and the orientation of ice marginal meltwater channels illustrates localized stagnation, and large-scale modifications to ice-flow and the shape of the ice margin as local source areas became more influential during deglaciation. Here, Fulton's (1967, 1991) model of deglaciation is refined with these findings to emphasize the reorganization of the ice sheet, continued ice movement, and local rather than regional ice stagnation. The refined model is specific to the northern Interior Plateau of British Columbia and consists of four phases: 1) confluent ice sheet; 2) ice sheet reorganization; 3) valley; and 4) alpine.

The confluent ice sheet phase is analogous with Fulton's (1967, 1991) active ice phase. During this phase the northwest-southeast-oriented ice front retreated to the southwest from the Rocky Mountains into the northern Interior Plateau. The general ice-flow direction was similar to that of the glacial maximum, but large topographic features caused local ice-flow variations at the margin. Locally, ice masses detached from the ice front as it retreated across topographic

613 irregularities. Examples of local ice stagnation are most common on the east side of topographic
614 highs, such as the ice dam that impounded glacial Lake McLeod and ice stagnation topography
615 on the northeast side of the uplands in NTS 093J13. During the ice sheet reorganization phase,
616 ice dynamics shifted to reflect major source areas. Specifically, the southern part of the margin
617 began to trend more east-west and northern part more north-south, and ice-flow near the northern
618 margin shifted east. Closer to the major source areas, ice was thinning and ice-flow became
619 increasingly influenced by topography (Stumpf et al. 2000). Ice-flow continued through this
620 phase and the ice front continued to retreat west through the Nechako Plateau (Plouffe 2000) and
621 south through the Fraser Basin (Clague 1988). The valley phase (analogous with Fulton's (1967,
622 1991) transitional upland phase) occurred when ice was limited to large valleys such as the
623 Babine and Ootsa lake valleys. Topographically controlled ice-flow continued in valley glaciers
624 that were still attached to source areas in the northern mountain ranges (Levson et al. 1998;
625 Plouffe 2000; Stumpf et al. 2000). Local ice stagnation presumably occurred during this phase as
626 ice blocks detached from retreating ice margins. During the final alpine phase, active ice was
627 limited to alpine regions. Climate fluctuations such as the Younger Dryas may have caused some
628 readvances/reactivation that later stagnated in alpine areas (e.g. Lakeman et al. 2008) within the
629 Coast Mountains.

630 Margold et al. (2013) present a model of the CIS retreat in central British Columbia
631 reconstructed from regional-scale investigations of meltwater landforms interpreted from
632 satellite imagery (Google Maps, SPOT) and digital elevation data (GeoBase© 2009). Based on
633 ice surface profiles delineated through inversion modeling, they conclude that the pattern of
634 retreat was similar to that of the Fennoscandian ice sheet (*cf.* Kleman et al. 1997), in which ice
635 retreated back into the source area along an irregular ice front with minimal stagnation.

Specifically, the CIS retreated westward and southwestward back into the Coast Mountains. The findings of the present study are in general agreement with this proposed style of frontal retreat and the suggestion of major ice reorganization. However, the ice front locations during deglaciation as depicted by Margold et al. (2013) are simplified because of the limited details captured from the interpretation of small scale data. Our study, and others in the region (e.g. Giles and Levson, 1995; Clague 1998*a, b*; Stumpf et al., 2000; Plouffe and Williams, 2001) demonstrate that detailed mapping of landforms and ice-flow indicators at all scales provide a more accurate view on ice from positions during deglaciation.

Conclusions

The purpose of this study is to establish a glacial history for the McLeod Lake map area based on surficial geology mapping and ground observations. Our interpretation is placed in a regional context that is founded upon similar research and together used to improve the understanding of the events of the Fraser Glaciation in central British Columbia.

Previous studies indicate that the initial ice advance from the Omenica and Skeena mountains flowed southeast down the major peripheral valleys. Data from this study suggest the ice also occupied the northernmost parts of the Interior Plateau between the valleys, and that an ice-front fluctuation occurred prior to the arrival of east flowing ice from the Coast Mountains. During glacial maximum, ice masses from the north, west, southwest and southeast coalesced over the Interior Plateau, and the ice divide shifted east from the Coast Mountains into the western part of the northern Interior Plateau. Ice flow reversed on the west side of the divide, and continued in an eastward direction, east of the divide. The eastward flow was diverted to the northeast by glaciers sourced in the south. It is unknown when this diversion occurred in relation to the

658 shifting ice divide. During initial deglaciation, decreased influence from the south resulted in a
659 transition of ice flow back to the east over much of the region.

660 Evidence of glacial lake formation during the advance and retreat phases of the Fraser Glaciation
661 is present in the lithostratigraphic record and surface landforms. At present, little is understood
662 about the configurations of advance-phase glacial lakes as evidence is sparse. The evolution of
663 glacial lakes during deglaciation, however, suggests the configuration of the lakes and drainage
664 networks was heavily influenced by the retreating ice margin. Glacial Lake Fraser occupied a
665 large area in the northern Interior Plateau during deglaciation. Differences in elevation of glacial
666 lake sediments and landforms across the region suggest glacial Lake Fraser existed as a time-
667 transgressive complex of smaller lakes. A better understanding of the glacial Lake Fraser
668 complex may be developed with increased temporal and spatial resolution of the ice margin
669 configuration using higher resolution surficial geology mapping and considering the effect of
670 isostasy.

671 A conceptual model of deglaciation specific to the northern Interior Plateau is proposed here. It
672 is refined from Fulton's (1967, 1991) conceptual model, and consists of four phases: 1) a
673 confluent ice sheet phase; 2) ice sheet reorganization phase; 3) valley glaciation phase; and 4)
674 alpine glaciation phase. The refinements consider the effects of interacting ice masses from
675 multiple source regions on the morphology and ice-flow of the CIS during deglaciation.

676 Deglacial chronological control is generally lacking over much of the interior of British
677 Columbia due to a lack of organic material for radiocarbon dating. Optical dating of post-glacial
678 aeolian sediments is an alternative means of constraining the time of deglaciation (*cf.* Lian and
679 Huntley 1999; Hickin et al. 2016). Our optical dating experiments demonstrate that K-feldspar is

a better geochronometer than quartz in this region. Fine silt K-feldspar post-glacial aeolian landforms yielded optical ages of 13.3 ± 1.0 , 9.94 ± 0.77 and 9.35 ± 0.64 ka. There are two possible interpretations of the optical ages: 1) deglaciation of the study area occurred around 13 ka and there were period(s) of reactivation in some localities, perhaps as a result of changes in climate; or 2) at two sigma, the optical ages are in general agreement with the regional radiocarbon ages associated with deglaciation, that range between 12.0 and 10.4 ka cal yrs BP (Table 2), and can be used together as minimum ages for deglaciation. More optical dating of stabilized aeolian landforms resting directly on glacial sediments would refine our knowledge of the timing of deglaciation in the region.

Acknowledgments

Generous funding for this project was provided by Geoscience BC. We would like to thank all of the field assistants, and the staff at the Luminescence Dating Laboratory at the University of the Fraser Valley. Additional support was provided through Natural Sciences and Engineering Research Council (NSERC) of Canada Discovery grants to B.C. Ward and O.B. Lian, and through an NSERC Research Tools and Instruments Grant to O.B. Lian. We would also like to thank Alain Plouffe and an anonymous reviewer for their thoughtful comments and suggestions that improved the quality of the manuscript.

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911 **Table 1** Summary of stratigraphic units.

Unit	Genesis	Geologic event
9	Aeolian	Post-Fraser Glaciation
8	Glacial (Basal)	Fraser Glaciation (re-advance)
7	Glaciofluvial	Fraser Glaciation (retreat phase)
6	Glaciolacustrine	Fraser Glaciation (retreat phase)
5	Glacial (ablation?)	Fraser Glaciation
4	Glacial (basal)	Fraser Glaciation
3	Glaciolacustrine	Fraser Glaciation (advance phase)
2	Glacial (basal)	Fraser Glaciation (early advance (?))
1	Glaciolacustrine or glaciofluvial sand	Fraser Glaciation (advance phase)

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914 **Table 2** Details for radiocarbon ages discussed in text. See Figure 1 for sample locations.

Reference number	Radiocarbon age (yrs BP)	Calibrated age (yrs BP) ^a	Locality	Material	Location (UTM)	UTM Zone	Sample Number	Reference
RC1	8775 ± 30	9910 – 9640	Hart, BC	peat from esker system	515094, 59779763	10	UCAIMS-83989	This study
RC2	9440 ± 110	11 140 – 10 400	Skunk Lake	peat overlying glacial sediment	430319, 6141989	10	GSC-5869	Plouffe 2000
RC3	9830 ± 130	11 760 – 10 780	Nautley River	peat overlying glacial sediment	396118, 6000514	10	GSC-5891	Plouffe 2000
RC4	10000 ± 140	12 040 – 11 200	Omineca Valley	peat overlying glacial sediment	368628, 6184920	10	GSC-2036-2	Blake 1986
RC5	10100 ± 90	12 040 – 11 320	Omineca Valley	marl overlying glacial sediment	368628, 6184920	10	GSC-2036	Alley and Young 1978

915 ^aCalibration range is 2s. Calibration was done using OxCal 4.2 (Brock-Ramsey 2009) and the
916 IntCal13 data set (Reimer et al. 2013).

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918 **Table 3** Dose rates, equivalent doses and optical ages. See Figure 2 for sample locations.

Sample	$\dot{D}_c$ (Gy/ka) ^a	$\dot{D}_T$ (Gy/ka) ^b	D_e (Gy) ^c	Optical age (ka) ^d
OCDF5 (feldspar)	0.154 ± 0.0077	2.17 ± 0.15	17.6 ± 0.2	9.94 ± 0.77
OCDF5 (quartz)	0.154 ± 0.0077	1.79 ± 0.11	15.7 ± 0.7	8.79 ± 0.51
OCDF4 (feldspar)	0.161 ± 0.008	2.21 ± 0.14	23.8 ± 0.5	13.3 ± 1.0
CLR5 (feldspar)	0.163 ± 0.0082	3.27 ± 0.2	26.4 ± 0.3	9.35 ± 0.64

919 ^a $\dot{D}_c$: dose rate due to cosmic rays, calculated using present day burial depths and the
 920 relationship of Prescott and Hutton (1994).

921 ^b $\dot{D}_T$: total dose rates due to cosmic rays and α , β and γ radiation.

922 ^c D_e : equivalent dose found using the central age model.

923 ^d Apparent optical ages (feldspar ages corrected for AF) and their analytical uncertainties at $\pm 1\sigma$.

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Table 4 Till clast fabric details. Refer to figure 7 for locations.

Fabric number	Depth from surface (m)	Clasts measured	Site Number	Eigenvector(°)	Mean plunge	Apparent ice flow direction (°) ^a	Eigenvalue 1
09-01	3	50	1	172	1	352	0.689
09-02	1.75	50	2	205	7	25	0.822
09-03	2	50	3	170	14	350	0.788
09-05	2	50	4	261	0	81	0.720
09-06	1.75	50	5	229	1	49	0.683
10-01	2.1	25	6	224	9	44	0.780
10-02	1.2	25	6	268	4	88	0.809
11-01	2	30	7	61	0	241	0.819
11-02	3.5	30	7	202	8	22	0.722
11-03	4	50	8	204	15	24	0.846
11-04	2	50	8	211	11	31	0.880
11-05	2.75	30	9	240	5	60	0.745
11-06	1.5	35	9	61	8	241	0.465

^a Apparent ice flow directions are calculated 180° from the principal eigenvector and may differ from inferred ice flow direction (see text).

Figure 1 a) Location map showing the study area (grey box) within the Interior Plateau physiographic region (dashed line; after Holland (1976)) of British Columbia with locations of radiocarbon ages (green symbol) and place names referred to in the text. b) Physiography and local place names of the study area which comprises NTS map areas 093J05/06/11/12/13/14. Major forest service roads are shown by thin black lines. Shaded relief derived from Canadian digital elevation data (Geobase[®] 2009).

Figure 2 a) Simplified surficial geology for the study area based on terrain mapping. Locations of logged stratigraphic sections are indicated by red stars and aeolian samples for optical dating by yellow stars. The original maps can be found at <http://www.geosciencebc.com/s/Report2013-10.asp>. Examples of landforms used to determine the glacial history: b) streamlined macroforms; c) subglacial meltwater complex; d/e) ice-contact glaciofluvial deposits; f/g). meltwater channels. Orthophotography base processed from imagery collected in 1996 and 1997.

Figure 3 Representative stratigraphic logs from the study area. See Figure 2 for locations.

Figure 4 a) Lower part of section DS-01 exposing subaqueous outwash fan sediments (unit 1c) below an early advance till (unit 2). Scale is approximate. Insets show: b) undulating, interstratified sand, gravel and diamict beds in unit 1c with approximate scale; and c) dewatering structures in the upper sand beds of unit 1c (dashed line) with scale increments of 1 cm.

Figure 5 Modified SAR methods used to date feldspar (a) and quartz (b) fractions of samples. L_i is the luminescence measured after administration of a regenerative dose, D_i . Regenerative doses ($i = 1$ to 5) for feldspar were 6.55, 13.1, 26.2, 52.4 and 0 Gy and quartz 4.98, 9.97, 19.93, 39.87 and 0 Gy; the 0 Gy data point is used to check for thermal transfer resulting from the preheat, which lasted for 10 s at 180° C for feldspar and 240° C for quartz. T_i is the luminescence measured after a test dose of 0.87 Gy at a cut-heat of 160° C and 14.95 Gy at a cut-heat of 220° C for feldspar and quartz respectively. A higher D_t was used for quartz due to the limited luminescence produced from the grains. The 13.1 Gy (feldspar) and 9.97 Gy (quartz) doses were repeated for each aliquot to determine the effectiveness of the test dose for correcting for sensitivity change. Typical examples of luminescence decay curves for aliquots from OCDF4 (c; feldspar) and OCDF5 (d; quartz). Inset graphs show the dose-response data for these aliquots, determined from the sensitivity corrected luminescence (L_i/T_i) measured over the first 0.4 s of stimulation. The data are fitted with a saturating exponential curve, onto which the natural luminescence is interpolated in order to determine the equivalent dose (D_e).

Figure 6 Locations and orientations of macroforms and striations discussed in text. Some macroforms were removed when plotting where symbol concentrations were high. Striation data indicate early southeastward ice flow (green symbols), general northeast ice flow at glacial maximum (black symbols) and eastward flow in the late glacial period (red symbols). Orange circle refers to striation sequence discussed in text.

Figure 7 Locations and measurements of till clast fabrics. Stereonet plots use a Schmidt equal area projection in the lower hemisphere. Refer to Table 4 for fabric details.

Figure 8 Glacier and drainage networks during deglaciation. Successive ice margins are hypothesized based on the trend of successive lateral meltwater channels. a) Near complete ice cover; subglacial meltwater flowing northeast. b) Linear ice margin retreating into study area; lateral meltwater channels forming; proglacial meltwater flowing northeast into glacial Lake McLeod. c) Linear ice margin continues to retreat; ice marginal lakes develop against uplands in north and around Weldon Lake in south. d) Ice margin retreats past drainage divide; proglacial meltwater drains southeast. e) Ice margin trending more north-south in the west and east-west in the south; proglacial meltwater continues to drain southeast; glacial lakes develop around Great Beaver Lake and in the southeast.

Figure 9 Inferred configurations of glacial Lake McLeod. a) Minimum water-level elevation for the maximum extent is determined from a spillway draining east into the Parsnip River valley, and a south-dipping ice-marginal channel cut into the east valley wall at ~750 m a.s.l. b) Second extent at a minimum water-level elevation of ~690 m a.s.l. inferred from the elevation of delta topsets at the mouth of the McLeod River valley. Glacial Lake Fraser drained into glacial Lake McLeod at this stage.

Figure 10 Glacial Lake Fraser delineated at a water level of 715 m a.s.l that was controlled by a spillway into the Peace River watershed. Glaciolacustrine sediments could not have been deposited contemporaneously at the depicted extents (light blue lines from Plouffe (2000); brown from this study; purple from Sacco et al. (2014); red from Clague (1998a); and dark blue from Tipper (1971b)). Therefore, this map suggests that a glacial lake complex consisting of non-

contemporaneous lakes occupied a large part of the region. The configuration of the complex was likely time-transgressive and the extent and elevation of the lakes were in part controlled by retreating glaciers that obstructed the drainage.

Figure 11 Hypothetical ice configuration used to explain glacial Lake Stuart at a water level of 780 m a.s.l., and the formation of the Prince George esker system. Glacial Lake Fraser is depicted at a water level of 715 m a.s.l., which was controlled by the outlet draining north in to the Peace River watershed. Location of organics analysed by radiocarbon dating indicated by black symbol. See text for details.

Figure 12 Regional ice-flow model for the northern Interior with data from Tipper (1971*a*); Plouffe (2000); Stumpf et al. (2000); Levson (2001); Plouffe and Levson (2001*b*) and this study. Physiographic regions after Holland (1976). Field area of this study is outlined in red. a) Glaciers first flowed southeast from the Omineca Mountains (green arrows) until more dominant ice from the west, south and southeast deflected the flow direction to the northeast (yellow and blue arrows). b) During the maximum phase, the ice divide shifted east from the Coast Mountains. Ice flow maintained a dominantly northeastern flow direction over most of the northern Interior Plateau, while reversing direction west of the divide (black arrows). c) During the late glacial phase, ice flow was initially modified by topography on the foothills of the Rocky Mountains until it retreated (orange arrows). The influence from the south diminished and the glaciers sourced in the west became dominant causing an eastward shift in ice flow (red arrows).

1018 **Appendix A:** Optical dating sample collection and preparation procedures

1019 Samples were collected by inserting an opaque ABS tube (~15 cm long and ~8 cm in diameter)
1020 into a freshly exposed vertical section. Zones with evidence of bioturbation or other post-
1021 depositional disturbances were avoided as these processes can introduce grains with different
1022 light exposure histories to the sample. All sample tubes were completely filled with material to
1023 keep the contents immobile and the slip caps were reinforced with duct tape to provide an air-
1024 tight seal to retain the as-collected water contents.

1025 Under filtered light in the laboratory, all samples were sub-sampled to determine water content.
1026 The as-collected, saturated and dry weights were used to calculate the as-collected and saturated
1027 water contents. For each sample, about 20 g of sediment was removed, dried and milled, and sent
1028 to a commercial laboratory where U, Th, and K concentrations were determined, which are
1029 needed to calculate environmental dose rates. Material from the centre of the tube was extracted
1030 for optical dating.

1031 Coarse-grained quartz (180-250 μm) and fine-grained feldspar (4-11 μm) were isolated from all
1032 processed samples for dating. Quartz was chosen because it is highly resistant to weathering and
1033 requires less exposure to sunlight to empty electron traps and effectively reset the 'luminescence
1034 clock'. However, in some cases quartz may not emit measurable luminescence (Lian 2013).
1035 Feldspar produces a higher intensity luminescence signal, but it suffers from anomalous fading
1036 (discussed below). To mitigate this effect, the fine-grained size fraction of feldspar was used, as
1037 it is thought to be less susceptible to anomalous fading (Lian and Roberts 2006).

1038 Material extracted for dating was first treated with a 10% HCl solution to dissolve any
1039 carbonates and then was wet-sieved using deionized water to isolate the 180-250 μm size

fraction; the fine-grained fraction was set aside for the feldspar experiments. Quartz and feldspar were separated from heavier minerals using a sodium polytungstate solution with a density of 2.72 g ml⁻¹. Using a centrifuge, quartz and feldspar were then separated from each other in a sodium polytungstate solution with a density of 2.6 g ml⁻¹. The quartz grains were subsequently exposed to a 48% HF solution for 40 minutes to etch away alpha-affected surfaces and to dissolve any remaining feldspar grains.

The retained fine-grained polymineral fractions were treated with 10% H₂O₂ to remove any organic matter that might have been present. The sediment was then deflocculated by adding a 1 g L⁻¹ solution of sodium hexametaphosphate and mechanically agitating the slurry on a shaker table. The 4-11 µm size fraction was isolated using the Stokes settling technique that is based on particle settling rates in a column of water. The >11 µm size fraction was removed by allowing the sediment to settle in a 20 cm column of water for 30 minutes and discarding the any settled (>11 µm) material. The <4 µm size fraction was removed by allowing the sediment to settle for 4 hours and discarding the material still in suspension (<4 µm).

Individual aliquots from each coarse-grained subsample were made by sprinkling about 100 grains of quartz on 9.8-mm diameter aluminium discs using silicone oil as an adhesive. Aliquots of fine-grained sediments were made by suspending the 4-11 µm size fraction in acetone and pipetting 1 ml of the solution into vials with aluminum discs placed at their bottoms and allowing the acetone to evaporate.

Environmental dose rate determination

Environmental radiation absorbed by the samples after burial comes from alpha, beta and gamma radiation produced by the decay of U, Th, their daughter products, and 40K in the sediment

grains and in the surrounding environment. There is also a contribution from cosmic rays which decreases with depth beneath the ground surface. Subsamples extracted for dosimetry were sent to Becquerel Laboratories (Mississauga, Ontario) for neutron-activation analysis to determine the element contents (Table A1). Environmental dose rates were calculated using standard formulae (Aitken 1985; Lian et al. 1995) and the dose rate conversion factors of Adameic and Aitken (1998). Pore water within the sediment attenuates or absorbs radiation differently from mineral matter and therefore must be accounted for in dose rate calculations. For each sample the as-collected water contents were used because it is unlikely that the well-drained sand-rich aeolian sediment was ever water-saturated (Table A1). A 10% uncertainty was included with each water content value to account for periodic wet or dry conditions. The contribution from cosmic rays was determined using the current burial depths and the formulae given by Prescott and Hutton (1994) (Table A2).

Equivalent dose determination

Equivalent dose (D_e) determinations were carried out with a Risø TL/OSL-DA-20 reader equipped with a $^{90}\text{Sr}/^{90}\text{Y}$ beta source that delivered 5.98 ± 0.12 Gy/min to coarse-grain samples and 5.24 ± 0.12 Gy/min to fine-grain samples. Quartz grains were stimulated using light-emitting diodes that delivered 45 mW cm⁻² of blue (470 nm) light to the sample aliquot; the luminescence measured was ultraviolet (~350 nm) detected by an Electron Tubes Ltd. 9235QB photomultiplier tube placed behind a single 7.5 mm-thick Hoya U-340 optical filter. Potassium feldspar on the polymineral fine-grained aliquots was stimulated preferentially using light-emitting diodes that delivered 135 mW cm⁻² of infrared (870 nm) light; the luminescence characteristic of K-feldspars (~400 nm) was isolated using a Kopp 7-59 optical filter. A Schott BG-39 filter was

1084 used to absorb scattered infrared light. The photomultiplier tube was the same as that used for
1085 quartz.

1086 A series of pilot D_e determination experiments were first conducted on 5-10 aliquots from each
1087 sample to determine the best measurement protocols. The pilot experiments indicated that quartz
1088 from samples OCDF4 and CLR5 is not suitable for dating because the measured luminescence
1089 was too dim or the samples behaved poorly (e.g. aliquots frequently failed recycling tests and/or
1090 their luminescence signal did not appear to consist of the desired fast component; see below).
1091 Pilot tests using K-feldspar from sample MLD1 yielded a preliminary age value of ~ 75 ka,
1092 suggesting that mineral grains with different light exposure histories, possibly glaciofluvial, were
1093 inadvertently sampled, so no further analysis was done on quartz or feldspar from this sample.

1094 For all of the other samples, D_e values were determined using a standard single-aliquot
1095 regenerative-dose (SAR) procedure (Murray and Wintle 2000; see Figure A1 for procedure
1096 details). Two additions were made to the procedure for quartz: 1) grains were treated with an
1097 infrared wash prior to the measurement of luminescence (Figure A1b, step 3); and 2) after the
1098 measurement of luminescence from the test dose, aliquots were treated with an additional blue
1099 light stimulation at a higher temperature than the first preheat (Figure A1b, step 9). The infrared
1100 wash removes any signal from feldspar contamination (Wintle and Murray 2006), which was
1101 only minor when it occurred. The additional blue light stimulation is intended to reduce the
1102 effect of thermal transfer of electrons from optically insensitive traps to optically sensitive traps
1103 that occurred during the second preheat (Figure A1b, step 6; Murray and Wintle 2003). The SAR
1104 protocol used for the fine-grained feldspar samples was based on one found to be successful in
1105 dating the same mineral and particle sizes from aeolian samples of known age near Clinton in the
1106 Southern Interior of British Columbia (W.L. Moon, pers. comm. 2011; see Figure A1 for

1107 procedure details). For both minerals, a zero-dose point was included in construction of the dose-
1108 response curves to assess the severity of preheat-induced thermal transfer and one of the
1109 regenerative-dose points was repeated to assess the efficacy of the test dose to correct for
1110 sensitivity changes. The luminescence measured over the first 0.4 s of excitation was used to
1111 construct dose-response curves from which the D_e was determined by interpolation (Figure A1c,
1112 d). Aliquots were rejected for D_e determination if the luminescence measured at the 0 Gy step
1113 was above 5 % of the natural signal or if the recycling ratio (repeated dose) was greater than 10
1114 % of unity. D_e values for the accepted aliquots from each sample were plotted on radial plots,
1115 and the central age model (Galbraith et al. 1999) was used to determine the weighted mean
1116 equivalent doses used to calculate the optical ages (Figure A2; Table A2).

1117 Dose recovery experiments were performed on some of the samples to test the adequacy of the
1118 chosen SAR protocol (Murray and Wintle 2003; Wintle and Murray 2006). The dose recovery
1119 procedure involved exposing a number of aliquots to sunlight to empty relevant electron traps. A
1120 known laboratory dose, similar to the sample's D_e , was then given to the aliquots and the same
1121 SAR protocol used for the D_e determination was used to attempt to recover this dose. If the SAR
1122 protocol is working correctly and if the mineral is suitable for optical dating (i.e., it does not
1123 exhibit anomalous luminescence properties; cf. Demuro et al. 2008), the procedure should be
1124 able to recover the laboratory dose. For quartz, a successful dose recovery test may also indicate
1125 that the luminescence measured is dominated by that arising from the thermally-stable fast
1126 component of a sample's luminescence signal, and that there is not a significant contribution
1127 from the thermally unstable slow and medium components. Less is known about the structure of
1128 the luminescence emitted from infrared-stimulated K-feldspar, but it is likely that it arises from a
1129 continuum of traps that are thermally stable (Wintle 2008). Measurements on known-aged

1130 samples appear to confirm this hypothesis if anomalous fading is taken into account. Anomalous
1131 fading refers to the loss of electrons in thermally stable traps due to quantum mechanical
1132 tunnelling of electrons from the traps of interest to traps that are thermally unstable over
1133 geological time (Huntley and Lamothe 2001). If not corrected for, anomalous fading can cause
1134 erroneously low De determinations and result in calculated optical ages that are too young. Thus,
1135 any discrepancy in the dose recovery test for K-feldspar can be attributed to the choice of an
1136 inadequate SAR protocol, or to AF if there is a delay between irradiation and measurement. The
1137 effect of the latter is best observed if a significant delay is made between laboratory irradiation
1138 and measurement.

1139 The dose recovery test for the quartz fraction of sample OCDF5 consisted of 20 aliquots exposed
1140 to sunlight for 30 minutes and subsequently given a beta dose of 19.7 ± 0.7 Gy (a value close to
1141 its pre-determined De). After 3 days, the aliquots were measured using the same SAR protocol
1142 originally used to date the sample. A dose of 13.9 ± 0.7 (~70%) was recovered, suggesting that
1143 the luminescence signal of the quartz is significantly influenced by the thermally unstable
1144 medium and/or slow components.

1145 Dose recovery tests were conducted on feldspar samples CLR5 and OCDF4. For CLR5, 20
1146 aliquots exposed to sunlight for 30 min and subsequently given a beta dose of 26.0 ± 0.5 Gy;
1147 after 44 days, a dose of 25.0 ± 0.2 Gy was recovered. For OCDF4, 38 aliquots were given a
1148 similar exposure to sunlight and a beta dose of 17.3 ± 0.4 Gy; after a delay of one day, 20 were
1149 measured and a dose of 16.1 ± 0.2 Gy was recovered. The remaining 18 aliquots were measured
1150 after a period of 425 days, and the difference in the recovered dose was negligible (within the
1151 uncertainties of the measurement). Both dose-recovery tests are in agreement, and suggest that a
1152 fading rate of no more than 2.0 ± 0.5 % per decade (cf. Huntley and Lamothe 2001) must be

applied to correct for AF in the fine-grained K-feldspar fraction of these samples. Based on the success of the experiments, using the same SAR protocol, on aeolian material in the lower Fraser River valley (unpublished results), it is unlikely that a significant part of the small difference in given and recovered dose was due to a minor inadequacy of the chosen SAR protocol. However, more fading tests should be done to confirm this assertion as the samples dated in this experiment and those from the lower Fraser River valley are about 300 km apart.

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- 1191

1192 **Table A1** Sample depths, K, U, Th and water contents.

Sample	d (cm) ^a	K (%) ^b	U (ppm) ^b	Th (ppm) ^b	Δ^w ^c
OCDF5 (feldspar)	180	1.33 ± 0.07	1.2 ± 0.13	2.6 ± 0.09	0.076 ± 0.008
OCDF5 (quartz)	180	1.33 ± 0.07	1.2 ± 0.13	2.6 ± 0.09	0.076 ± 0.008
OCDF4 (feldspar)	155	1.4 ± 0.1	0.9 ± 0.11	2.7 ± 0.12	0.047 ± 0.005
CLR5 (feldspar)	145	2.07 ± 0.08	1.3 ± 0.14	4.8 ± 0.16	0.046 ± 0.005

1193 ^a d : sample depth beneath the ground surface.

1194 ^b K, U and Th contents were determined using neutron-activation analysis.

1195 ^c Δ^w : water content = water mass/mineral mass. As collected water contents with a 10% error

1196 applied were used for calculations.

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1198 **Table A2** Dose rates, equivalent doses and optical ages.

Sample	$\dot{D}_c$ (Gy/ka) ^a	$\dot{D}_T$ (Gy/ka) ^b	D_e (Gy) ^c	Optical age (ka) ^d
OCDF5 (feldspar)	0.154 ± 0.0077	2.17 ± 0.15	17.6 ± 0.2	9.94 ± 0.77
OCDF5 (quartz)	0.154 ± 0.0077	1.79 ± 0.11	15.7 ± 0.7	8.79 ± 0.51
OCDF4 (feldspar)	0.161 ± 0.008	2.21 ± 0.14	23.8 ± 0.5	13.3 ± 1.0
CLR5 (feldspar)	0.163 ± 0.0082	3.27 ± 0.2	26.4 ± 0.3	9.35 ± 0.64

1199 ^a $\dot{D}_c$: dose rate due to cosmic rays, calculated using present day burial depths and the
1200 relationship of Prescott and Hutton (1994).
1201 ^b $\dot{D}_T$: total dose rates due to cosmic rays and α , β and γ radiation.
1202 ^c D_e : equivalent dose found using the central age model.
1203 ^d Apparent optical ages (feldspar ages corrected for AF) and their analytical uncertainties at $\pm 1\sigma$.

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Figure A1 Modified SAR methods used to date feldspar (a) and quartz (b) fractions of samples. Li is the luminescence measured after administration of a regenerative dose, D_i . Regenerative doses ($i = 1$ to 5) for feldspar were 6.55, 13.1, 26.2, 52.4 and 0 Gy and quartz 4.98, 9.97, 19.93, 39.87 and 0 Gy; the 0 Gy data point is used to check for thermal transfer resulting from the preheat, which lasted for 10 s at 180° C for feldspar and 240° C for quartz. Ti is the luminescence measured after a test dose of 0.87 Gy at a cut-heat of 160° C and 14.95 Gy at a cut-heat of 220° C for feldspar and quartz respectively. A higher D_t was used for quartz due to the limited luminescence produced from the grains. The 13.1 Gy (feldspar) and 9.97 Gy (quartz) doses were repeated for each aliquot to determine the effectiveness of the test dose for correcting for sensitivity change. Typical examples of luminescence decay curves for aliquots from OCDF4 (c; feldspar) and OCDF5 (d; quartz). Inset graphs show the dose-response data for these aliquots, determined from the sensitivity corrected luminescence (L_i/T_i) measured over the first 0.4 s of stimulation. The data are fitted with a saturating exponential curve, onto which the natural luminescence is interpolated in order to determine the equivalent dose (D_e).

Figure A2 Radial plots showing the distribution of equivalent doses from each sample. More precise data plot farther from the origin. The gray bands are centred on the weighted mean equivalent dose values provided by the central age model and encompass aliquots with equivalent doses consistent with the mean at a 95% confidence interval.

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Material type		Contact type	
Clay	Gravel	Erosional	
Silt	Diamict	Interstratified	
Sand	Volcanic bedrock	Sharp	
Covered	Clastic dyke	Gradational	
	Boulder lag		

Lithofacies codes

Dmm - diamict, massive, matrix supported
Dms - diamict, matrix supported, stratified
Dcs - diamict, clast supported, stratified
Gm - gravel, clast supported, massive
Gms - gravel, matrix supported, stratified
St - medium sand, trough cross-bedded

Sm - sand, massive
Sd - sand with deformed bedding
Fl - fine laminations of silt, clay and minor sand
(d) - dropstones
(w) - dewatering structures

Unit number and lithofacies

2 (*Dmm*)
Dropstone
Fault

124°W

Datum: NAD 83

Projection: UTM Zone 10

20 km

123°W

Striation (direction known; unknown)

Drumlin

Fluting

Crag and tail

- 980 m Glacial lake outlet and approximate elevation
- 980 m Maximum elevation of delta topsets
- Detailed study area boundary

-
- The map displays the Fraser River Watershed and its tributaries, including the Peace River, Stuart River, Nechako River, and Fraser River. It highlights the extent of glaciolacustrine sediments in shades of blue and cyan, with specific elevation markers (980 m, 760 m, 780 m, 715 m, 800 m, 875 m) indicating the maximum elevation of these sediments. A detailed study area boundary is outlined in red, and a drainage divide is shown as a black line. A scale bar at the bottom left indicates a distance of 40 km. The map is overlaid with a grid showing latitude (54°) and longitude (124°).
- Glacial lake outlet and elevation
 - Extent of glaciolacustrine sediment
 - Drainage divide
 - Detailed study area boundary
 - Maximum elevation of glacial lake sediments

124°

Glacial Lake (715 m)

980 m

Extent of glaciolacustrine sediment

Drainage divide

Detailed study area boundary

980 m

Maximum elevation of glacial lake sediments

Peace River Watershed

715 m

Stuart River

Fraser River Watershed

Vanderhoof

54°

Nechako River

800 m

875 m

Prince George

760 m

Fraser River

54°

0 40 km

124°

Datum: NAD 83
Projection: PC Albers

0 50 km

CLR5 (quartz)

OCDF4 (quartz)

OCDF5 (feldspar)

OCDF5 (quartz)