

Advance of alpine glaciers during final retreat of the Cordilleran ice sheet in the Finlay River area, northern British Columbia, Canada

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Abstract

Sharp-crested moraines, up to 120 m high and 9 km beyond Little Ice Age glacier limits, record a late Pleistocene advance of alpine glaciers in the Finlay River area in northern British Columbia. The moraines are regional in extent and record climatic deterioration near the end of the last glaciation. Several lateral moraines are crosscut by meltwater channels that record downwasting of trunk valley ice of the northern Cordilleran ice sheet. Other lateral moraines merge with ice-stagnation deposits in trunk valleys. These relationships confirm the interaction of advancing alpine glaciers with the regionally decaying Cordilleran ice sheet and verify a late-glacial age for the moraines. Sediment cores were collected from eight lakes dammed by the moraines. Two tephras occur in basal sediments of five lakes, demonstrating that the moraines are the same age. Plant macrofossils from sediment cores provide a minimum limiting age of 10,550–10,250 cal yr BP (9230±50 ¹⁴C yr BP) for abandonment of the moraines. The advance that left the moraines may date to the Younger Dryas period. The Finlay moraines demonstrate that the timing and style of regional deglaciation was important in determining the magnitude of late-glacial glacier advances.

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Introduction

The Cordilleran ice sheet in northern British Columbia covered an area of nearly 500,000 km² during the last glaciation (Fraser glaciation, late Wisconsinan), but its history is poorly known (Clague, 1989; Jackson et al., 1991; Ryder and Maynard, 1991; Bobrowsky and Rutter, 1992; Clague and James, 2002). This part of the ice sheet probably began to decay about 19,000–18,000 cal yr BP, when glaciers sourced in the northern Coast Mountains retreated from the Queen Charlotte Islands (Blaise et al., 1990). By approximately 14,000 cal yr BP, the ice sheet had retreated from Hecate Strait and its outlet glaciers terminated along the mainland coast and in fiords (Clague, 1985). Similarly,

glaciers in southeast Alaska had retreated from the continental shelf by approximately 16,000 cal yr BP (Mann and Hamilton, 1995). The eastern margin of the ice sheet had withdrawn from summits in the front ranges of the northern Rocky Mountains by 14,800–13,200 ³⁶Cl yr ago (Bednarski and Smith, 2007), although alpine glaciers persisted in the northern Rocky Mountains after the ice sheet decayed (Bednarski, 1999, 2000, 2001; Bednarski and Smith, 2007). Deglaciation was probably complete in northern British Columbia before approximately 11,000 cal yr BP (Clague, 1981).

Other than the broad constraints on deglaciation provided above, the timing and pattern of ice sheet decay are unknown in northern British Columbia. Importantly, nothing is known about the effects of late-glacial climatic oscillations such as the Younger Dryas (12,800–11,500 cal yr BP; Alley, 2000) on remnants of the northern half of the Cordilleran ice sheet. A change in ocean circulation and cooling of climate in the north-east Pacific Ocean coincident with the Younger Dryas have been documented in north-coastal British Columbia and adjacent

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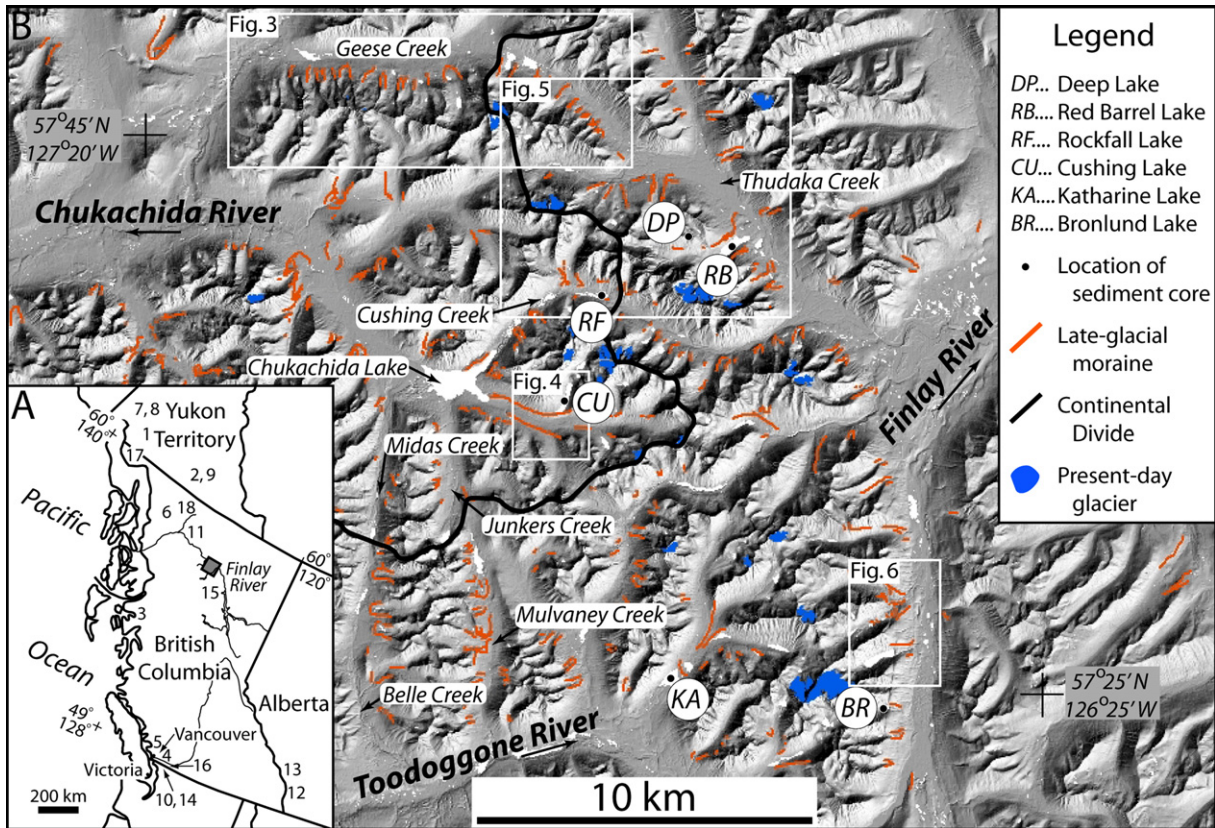


Figure 1. A) Location of study area (grey box). Numbers mark sites where late-glacial advances have been identified: 1—Bond (2004); 2—Bond and Kennedy (2005); 3—Clague (1985); 4—Clague et al. (1997); 5—Friele and Clague (2002a,b); 6—Hamilton (1981); 7—Hughes (1990); 8—Jackson et al. (1991); 9—Kennedy and Bond (2004); 10—Kovanen and Easterbrook (2001); 11—T. Lakeman, unpublished data; 12—Osborn and Gerloff (1997); 13—Reasoner et al. (1994); 14—Riedel et al. (2003); 15—Roots (1954); 16—Saunders et al. (1987); 17—Tallman (1975); 18—Watson and Mathews (1944). B) Hillshaded digital elevation model of the study area showing distribution of late-glacial moraines, locations of cored lakes, and locations of Figures 3–6. Hillshaded digital elevation model was produced with permission using digital elevation data from the Province of British Columbia, Base Mapping and Geomatic Services Branch.

Alaska (Engstrom et al., 1990; Mathewes, 1993; Mathewes et al., 1993; Patterson et al., 1995; Hansen and Engstrom, 1996; Hendy et al., 2002; Hetherington and Reid, 2003; Lacourse, 2005). Younger Dryas glacier advances have been identified in the central and southern Coast Mountains of British Columbia (Clague, 1985; Friele and Clague, 2002a), the southern Canadian and northern American Rocky Mountains (Reasoner et al., 1994; Osborn and Gerloff, 1997), the northern Cascade Range of Washington (Kovanen and Easterbrook, 2001; Riedel et al., 2003), and possibly southwestern Alaska (Briner et al., 2002).

The Younger Dryas, however, was not the only cold interval during the late-glacial period. Advances several hundred years to more than 1000 yr earlier than the Younger Dryas have been identified in southwest British Columbia (Saunders et al., 1987; Clague et al., 1997; Friele and Clague, 2002b) and in the southern Canadian and northern American Rocky Mountains (Osborn and Gerloff, 1997). Their regional climatic significance, however, remains unknown.

The Finlay River area of north-central British Columbia (Fig. 1) contains large, sharp-crested lateral and terminal moraines that record a resurgence of alpine glaciers during late-glacial time (i.e. younger than 18,000 cal yr BP; Clague, 1981).

This paper reports on these moraines and related deglacial sediments and landforms. Our objectives are to: 1) describe the late-glacial moraines, 2) infer the pattern of deglaciation in the area, 3) present information on the age of the advance that constructed the moraines, 4) discuss the paleoclimatic significance of the moraines, and 5) compare the inferred late-glacial climate of the Finlay River area to that of other areas in the Canadian Cordillera where late-glacial moraines have been identified. We also consider the relation of glacier resurgence in northern British Columbia and late-glacial advances elsewhere in the Canadian Cordillera.

Study area

The Finlay River watershed is located in north-central British Columbia, in the rain shadow of the northern Coast Mountains (Fig. 2). The study area comprises approximately 3200 km² of mountainous terrain straddling the northern Omineca and southern Cassiar mountains (Figs. 1 and 2). The Continental Divide runs through the centre of the study area, separating Pacific from Arctic drainages (Fig. 1). Finlay River flows north and then east into the northern Rocky Mountain Trench, which borders the study area to the east (Fig. 1). Chukachida River

drains the western part of the study area and flows into Stikine River on Spatsizi Plateau west of the map area (Fig. 1).

Climate in the study area is continental. For the period AD 1971–2000, mean annual temperature and precipitation at Fort St. James, approximately 350 km southeast of the study area, are 3.1 °C and 487.1 mm, respectively (Environment Canada, 2002). Muncho Lake, approximately 150 km northeast of the study area, has a mean annual temperature of −0.2 °C and an average annual precipitation of 495.8 mm over the same period. Dease Lake, approximately 200 km northwest of the study area has a mean annual temperature of −0.8 °C and an average annual precipitation of 425.5 mm. White and black spruce (*Picea glauca*, *Picea mariana*), subalpine fir (*Picea engelmannii*), lodgepole pine (*Pinus contorta*), and aspen (*Populus tremuloides*) dominate vegetation at low elevations in the study area (Meidinger and Pojar, 1991).

The highest mountains are about 2500 m above sea level (asl), and local relief is up to 1400 m. Small cirque glaciers exist above 1800 m asl on the flanks of a few of the highest summits (Fig. 1). The study area was extensively glaciated during the Pleistocene. Cirques, broad U-shaped valleys, and arêtes formed during repeated, lengthy periods of alpine glaciation.

The Cordilleran ice sheet also covered the area repeatedly during the Pleistocene, most recently during the Fraser glaciation.

Methods

Glacial landforms and deposits were mapped on Province of British Columbia aerial photographs flown in 1986 and 1999. The photographs have nominal scales of 1:60,000 and 1:35,000, respectively. Mapped features were digitized, entered into a geographic information system (GIS), and checked during an aircraft-supported field survey in 2005. Identification of large moraine complexes; meltwater channels; ice-contact deltas, fans, and a landslide; eskers; and glaciolacustrine sediments enabled a detailed reconstruction of latest Pleistocene glacier activity and the regional pattern of deglaciation. We estimated equilibrium line altitudes (ELAs) of former glaciers to determine their paleoclimatic significance. ELAs were determined using the accumulation area ratio (AAR) method with an AAR of 0.6 (Porter, 1975).

We collected sediment cores from six lakes in the study area to constrain the absolute age of the moraines and the time of deglaciation (Fig. 1). Two additional cores were recovered from

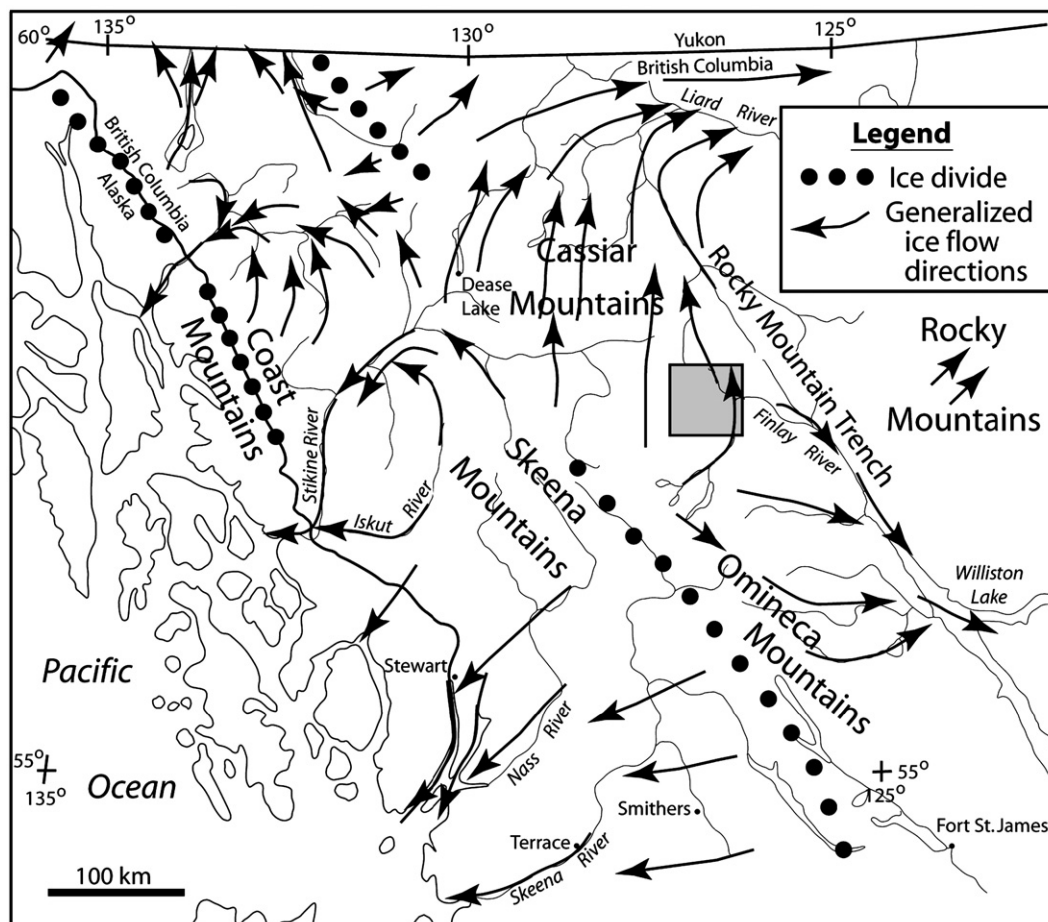


Figure 2. Generalized ice flow directions in northern British Columbia at the maximum of the Fraser glaciation. The grey box delineates the study area. Modified with permission from Ryder and Maynard (1991), with data from Bednarski (1999), Stumpf et al. (2000), and McCuaig and Roberts (2002).

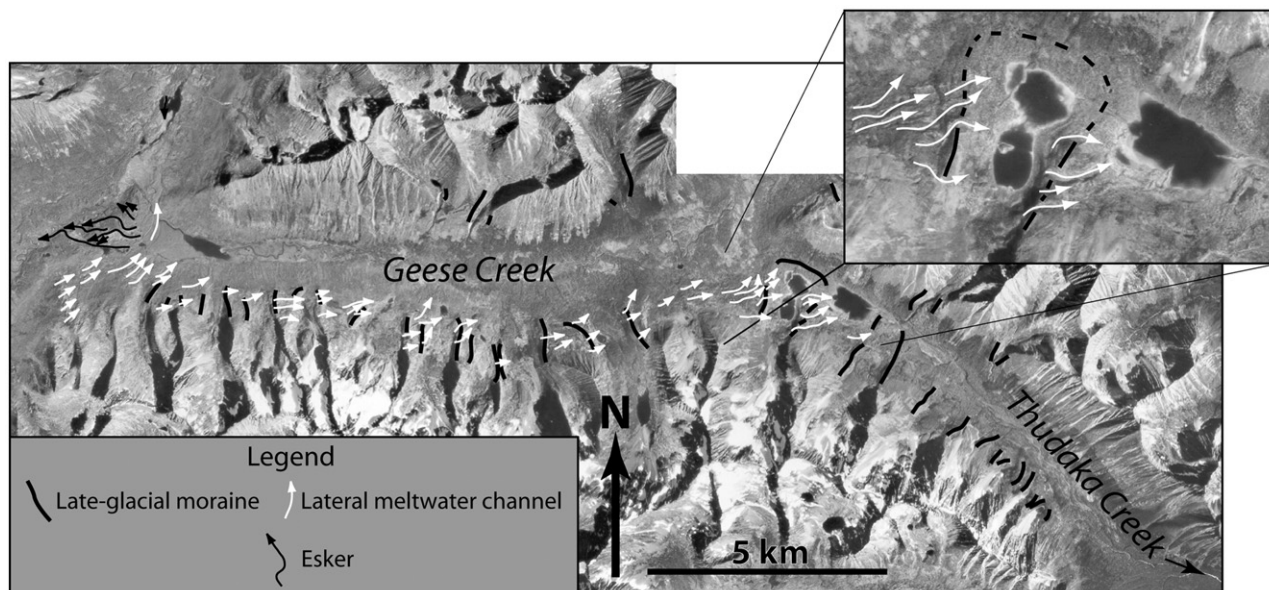


Figure 3. Aerial photograph mosaic of late-glacial landforms in Geese Creek valley. Ice-marginal meltwater channels demarcate the margin of a trunk glacier. The channels crosscut lateral moraines constructed by resurgent alpine glaciers. Inset shows the approximate terminus of the trunk glacier in Geese Creek valley when it was in contact with resurgent alpine glaciers; note deep lateral meltwater channels cut into the lateral moraines. Aerial photographs 15BC86082-002, 003, and 004; reproduced with permission of British Columbia Ministry of Agriculture and Lands, Base Mapping and Geomatic Services Branch.

Black and Sandwich lakes, which are, respectively, 10 km and 30 km south of the study area. We collected the cores from lake ice in January 2005 and from the floats of a De Havilland DHC-2 Beaver in July 2005, using a percussion coring system (Reasoner, 1993). They were transported to the University of Northern British Columbia, where they were split, logged, photographed, analyzed for bulk physical properties (grain size, magnetic susceptibility, and organic content), and sampled for terrestrial macrofossils and tephtras. Terrestrial macrofossils were radiocarbon dated at IsoTrace Laboratory (University of Toronto) and Beta Analytic Inc. by accelerator mass spectrometry. Calibrated ages were determined using the program OxCal v4.0, using the IntCal04 calibration curve (Reimer et al., 2004). All calibrated ages are reported at the 95% confidence interval (± 2 sigma). Glass shards were isolated from the tephtras using a heavy liquid separation procedure. The major element compositions of the glass shards were determined using electron microprobes at the Department of Geosciences, University of Calgary and the Department of Earth and Atmospheric Sciences, University of Alberta (Lakeman, 2006).

The Fraser glaciation in the Finlay River area

The northern half of the Cordilleran ice sheet during the Fraser glaciation was fed from accumulation centres located over the northern Coast, Skeena, and northernmost Cassiar mountains (Fig. 2; Mathews 1980; Ryder and Maynard, 1991; Stumpf et al., 2000; McCuaig and Roberts, 2002). Radial flow from the northern Skeena Mountains directed ice north into the Cassiar Mountains and Liard Plateau, east into the northern Rocky Mountains and foothills, and south into the Omineca Mountains and Nechako Plateau (Fig. 2; Mathews, 1980; Ryder

and Maynard, 1991; Bobrowsky and Rutter, 1992; Bednarski, 1999, 2000, 2001; Stumpf et al., 2000; McCuaig and Roberts, 2002). The Skeena Mountains also fed glaciers flowing west along major valleys through the northern Coast Mountains (Fig. 2; Ryder and Maynard, 1991). Smaller ice divides over the northern Coast Mountains and northernmost Cassiar Mountains directed ice westward onto the continental shelf and north into the Yukon, respectively (Fig. 2; Ryder and Maynard, 1991).

The Finlay River area was covered by ice flowing from the northern Skeena Mountains (Fig. 2; Mathews, 1980; Ryder and Maynard, 1991). Ice flow through the study area at the last glacial maximum may have been relatively unobstructed by topography. This condition, however, was probably short-lived, and for much of the Fraser glaciation, mountain ranges and valleys in the study area controlled the pattern of ice flow. Numerous meltwater channels parallel valley sides and others cross alpine ridges, recording drawdown of the ice sheet during initial deglaciation. At that time, glaciers continued to flow north and east from the northern Skeena Mountains along major valleys through the study area.

Geomorphic observations

Late-glacial moraines

Large, regionally extensive lateral and terminal moraines provide evidence for an advance of alpine glaciers in the Finlay River area during late-glacial time, herein referred to as the Finlay Advance (Fig. 1). The moraines are bulky, sharp-crested, up to 120 m high (Figs. 3 and 4), and extend up to 9 km beyond Little Ice Age terminal moraines (Fig. 5). Glaciers that constructed the Finlay moraines terminated between 1100 and 1200 m asl. Two or three nested moraines are present in most

areas, indicating multiple episodes of moraine construction. The moraines are uniformly distributed throughout the study area (Fig. 1). Morphologically similar moraines occur north and west of the study area in the Cassiar Mountains, but they were not investigated in this study (Lakeman, 2006).

Lateral moraines are common in tributary valleys below high cirques (Fig. 1). Associated terminal moraines occur at the mouths of some tributary valleys, where they join trunk valleys. These moraines are typically broad and have undulating to flat surfaces (Fig. 5). More commonly, however, terminal moraines are weakly developed or absent altogether (Fig. 3). Lateral moraines with no associated terminal moraine commonly merge with ice-stagnation deposits, such as ice-contact alluvial fans,

kames, eskers, and kettles that cover trunk valley floors (Fig. 6). Some lateral moraines curve at the mouths of tributary valleys, suggesting that tributary and trunk glaciers coalesced (Fig. 6). Lateral meltwater channels, formed during downwasting and retreat of trunk glaciers, crosscut many of the lateral moraines (Fig. 3).

Deglacial landforms

Several of the main valleys in the study area contain landforms that record thinning and retreat of trunk glaciers after construction of the late-glacial moraines. Lateral meltwater channels in the Finlay River valley crosscut late-glacial moraines 200–300 m above the valley floor, and concordant lateral

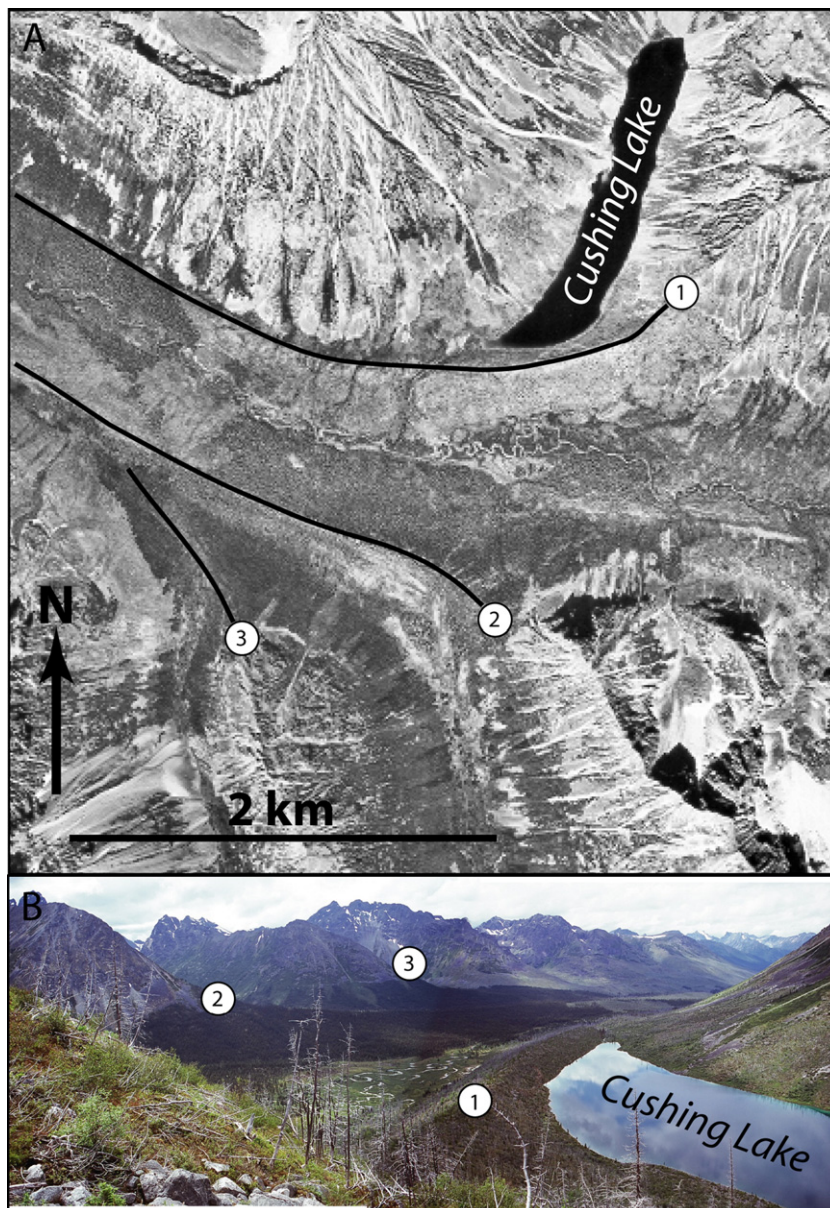


Figure 4. A) Aerial photograph of Cushing Lake, which is dammed by a late-glacial moraine, labeled 1. B) Photograph of late-glacial lateral and medial moraines looking west downvalley towards Chukachida Lake (Fig. 1). Moraines 2 and 3 are labeled on both photographs. Aerial photograph 15BCB99017-131; reproduced with permission of British Columbia Ministry of Agriculture and Lands, Base Mapping and Geomatic Services Branch.

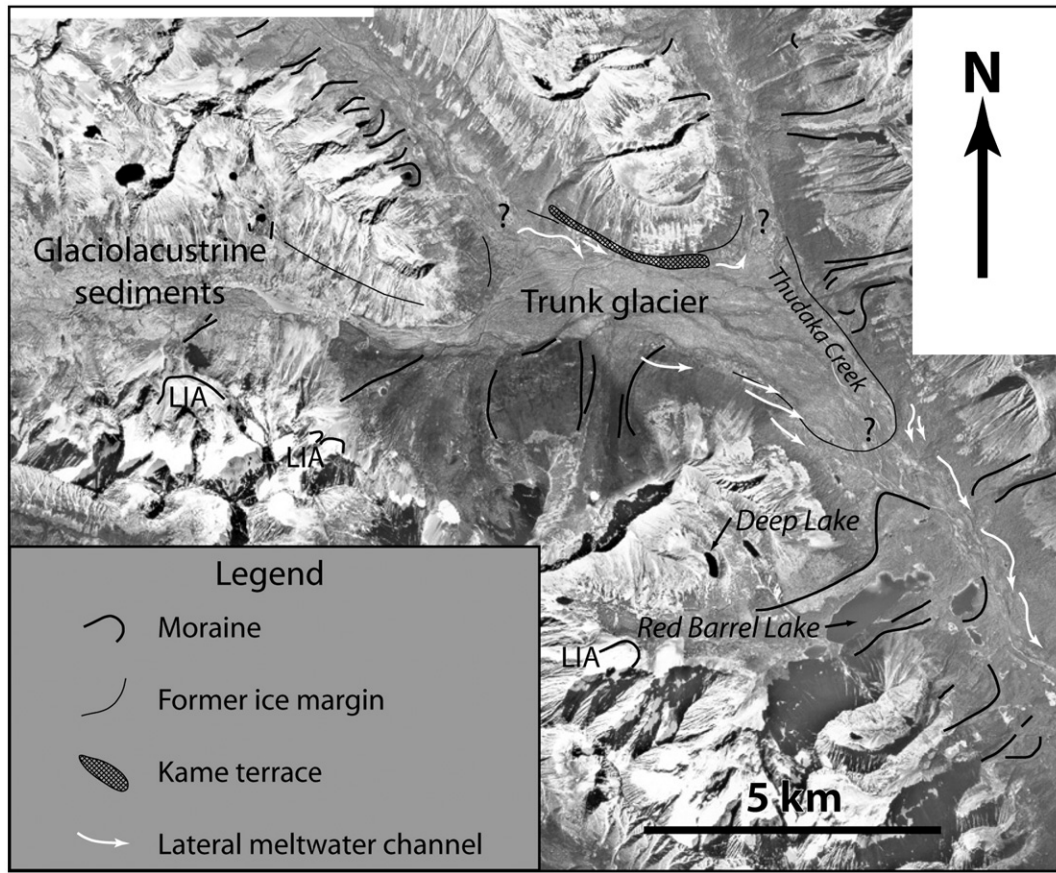


Figure 5. Aerial photograph mosaic of late-glacial landforms in Thudaka Creek valley. LIA—Little Ice Age moraine. Aerial photographs 15BCB99018-097, 164, 165, 166, and 167; reproduced with permission of British Columbia Ministry of Agriculture and Lands, Base Mapping and Geomatic Services Branch.

meltwater channels descend in elevation to the north. Hummocky terrain and large, subglacial meltwater channels occur at the confluence of Finlay River and the Rocky Mountain Trench, and large kettles, eskers, and subglacial meltwater channels are present farther up-valley. Prominent kame terraces are located 60–80 m above the Finlay River valley floor, below late-glacial moraines complexes in two east-trending tributary valleys (Fig. 6).

Resurgent alpine glaciers constructed lateral moraines in Geese Creek valley, but terminal moraines are absent (Fig. 3). Lateral meltwater channels crosscut the moraines up to 200 m above the valley bottom. The channels descend in elevation to the east and terminate against several late-glacial terminal moraines in the northwest tributary of Thudaka Creek (Fig. 3). Eskers and ice-contact alluvial fans occur below lateral moraines in adjacent cirques and tributary valleys (Fig. 3).

Lateral meltwater channels in Cushing Creek valley are approximately 200 m above the valley floor, descend eastward, and terminate near Thudaka Creek (Fig. 1). A large rock avalanche dams a lake in Cushing Creek valley. The rock avalanche debris consists mainly of large, angular, blocks of local bedrock, but it also contains exotic, rounded, granite cobbles and boulders. The debris lobe extends about 2 km to the east down Cushing Creek valley. The western margin of the debris, bordering the lake, has a steep linear front. Numerous eskers occur farther west in the valley, beyond the rock avalanche deposit.

Belle, Midas, Mulvaney, Junkers, and Jack Lee Creek valleys contain several late-glacial moraines that were constructed by resurgent alpine glaciers in small tributary valleys (Fig. 1). Lateral meltwater channels, up to 250 m above the trunk valley bottoms, crosscut the moraines. Late-glacial terminal moraines built by alpine glaciers are well preserved on valley floors. Lateral and subglacial meltwater channels abut several ice-contact deltas, and large ice-contact alluvial fans occur at the mouths of Junkers and Midas Creek valleys.

The pattern of moraines and lateral meltwater channels in Thudaka Creek valley indicates that a trunk glacier occupied the valley during late-glacial time. Lateral meltwater channels on the sides of Thudaka Creek valley descend to the southeast and terminate 1–2 km up-valley of the terminal moraine that impounds Red Barrel Lake (Fig. 5). The Red Barrel moraines have no crosscutting features indicative of interaction with the trunk glacier in Thudaka Creek valley, but there are several meltwater channels on the distal side of the Red Barrel terminal moraine (Fig. 5). Glaciolacustrine sediments occur in the westernmost tributary of Thudaka Creek and indicate that a small glacial lake formed there during deglaciation (Fig. 5).

Age of the Finlay Advance

Crosscutting lateral meltwater channels (Fig. 3) and the spatial association of the moraines with sediments deposited

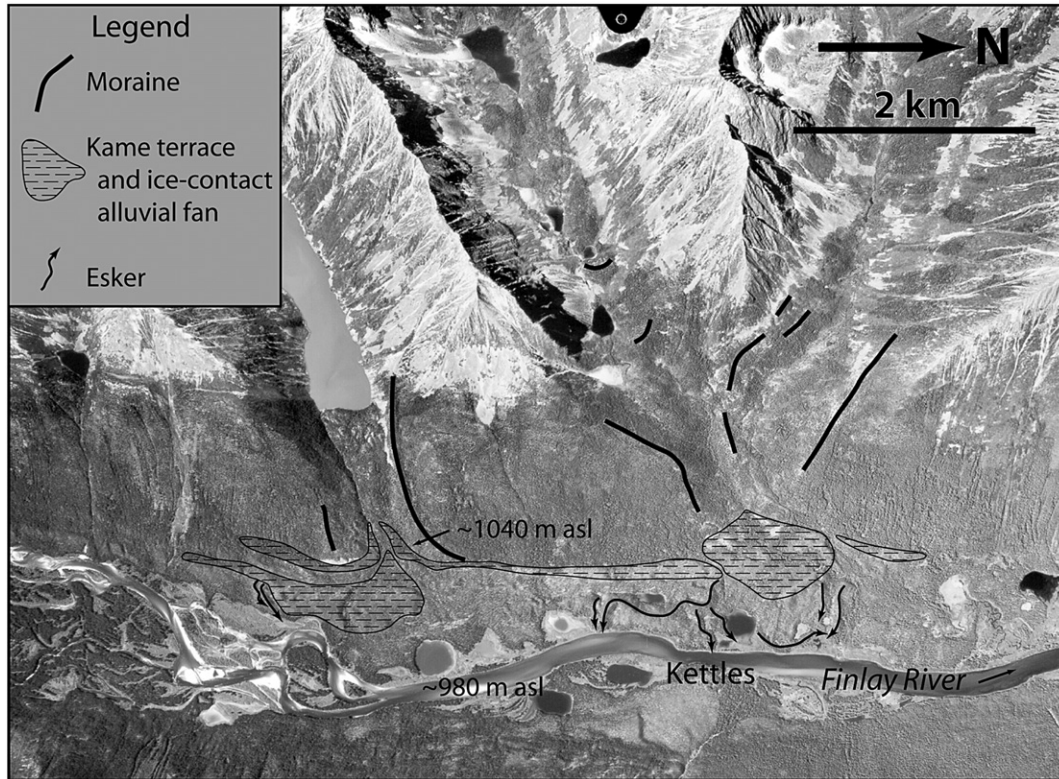


Figure 6. Aerial photograph of late-glacial landforms in the Finlay River valley and its tributaries. When alpine glaciers abandoned their lateral moraines, they built kame terraces against trunk ice in the Finlay River valley. Aerial photograph 15BCB99016-137; reproduced with permission of British Columbia Ministry of Agriculture and Lands, Base Mapping and Geomatic Services Branch.

during final wastage of the Cordilleran ice sheet (Figs. 3 and 6) demonstrate that the moraines were built during final deglaciation of northern British Columbia and are thus late Pleistocene in age. The age of the moraines is further constrained by radio-

carbon ages from macrofossils in basal organic sediments recovered from several lakes in the study area.

The lakes sampled in this study formed at the end of the Finlay Advance, thus basal ages on postglacial sediment cores

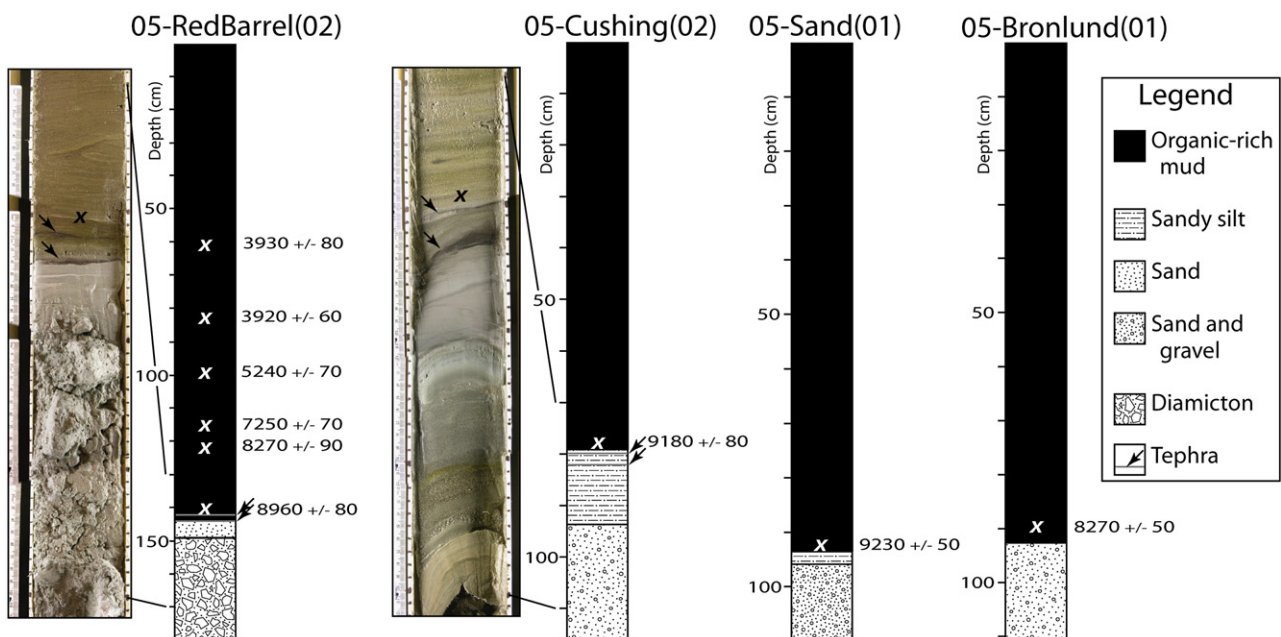


Figure 7. Stratigraphic logs and radiocarbon ages from Red Barrel, Cushing, Sandwich, and Bronlund lakes. Tephrae are thin black laminae delineated by arrows in the core photos. White and black Xs indicate locations of radiocarbon ages in ¹⁴C yr BP.

Table 1
Radiocarbon ages from the Finlay River area

Location (latitude, longitude)	^{14}C age ^a (yr BP)	Calibrated age range ^b (yr BP)	Laboratory number ^c	Dated material	Depth in core (cm)
Red Barrel Lake (57°40.623'N, 126°44.029'W)	3930±80	4580–4100	TO-12469	Conifer needle	60.5
	3920±60	4520–4160	TO-12470	Twig	82.5
	5240±70	6210–5890	TO-12471	Spruce terminal bud	98.5
	7250±70	8190–7950	TO-12472	Twig	115.0
	8270±90	9460–9030	TO-12473	Conifer Needle	121.5
	8960±80	10,250–9780	TO-12474	Wood	139.5
Cushing Lake (57°35.607'N, 126°54.450'W)	9780±80	10,560–10,220	TO-12475	Terrestrial plant matter	77.5
Bronlund Lake (57°25.463'N, 126°35.460'W)	8270±50	9430–9090	Beta-202608	Conifer needles	89.5
Sandwich Lake (57°01.565'N, 126°29.747'W)	9230±50	10,550–10,250	Beta-202609	Wood	93.5

^a Laboratory-reported errors are 1-sigma.

^b Calibrated ages determined using the program OxCal v4.0 Beta, which uses the IntCal04 calibration curve (Reimer et al., 2004). The range represents the 95% confidence interval (± 2 sigma).

^c Beta—Beta Analytic Inc.; TO—IsoTrace Laboratory (University of Toronto).

recovered from the lakes serve as minimum limiting ages for the demise of the Cordilleran ice sheet and for retreat of alpine glaciers from the late-glacial moraines. Cushing and Red Barrel lakes are impounded behind late-glacial lateral and terminal moraines, respectively (Figs. 4 and 5), whereas Katharine Lake lies on the distal side of a late-glacial terminal moraine (Fig. 1). Bronlund Lake is situated in a basin 2 km up-valley from several late-glacial lateral moraines (Fig. 1). Deep Lake is located in a cirque that supported the glacier that constructed the terminal moraine impounding Red Barrel Lake (Fig. 5). Black Lake is located in a trunk valley and is dammed by a late-glacial alluvial fan. Sandwich Lake is situated in a late-glacial lateral meltwater channel. Rock avalanche debris in Cushing Creek valley dams Rockfall Lake (Fig. 1).

The sediment cores from these lakes contain thick organic-rich mud (gyttja) overlying inorganic silt, sand, gravel, and diamict (Fig. 7). Several of the cores, including those from Katharine, Red Barrel, and Cushing lakes, contain two black phonolitic tephra near the contact between inorganic and overlying organic-rich sediments (Figs. 1 and 7). Sediment cores from Deep and Black lakes contain only one phonolitic tephra at their base. No tephra were found in the cores from Bronlund and Rockfall lakes. The tephra comprise fine sand to silt-sized mineral grains and glass shards. They are coarser than the bounding sediments and lie within 2–4 cm of one another in the cores (Fig. 7).

An age of 10,250–9780 cal yr BP (8960±80 ^{14}C yr BP) was obtained on a piece of wood 2 cm above the upper tephra in a

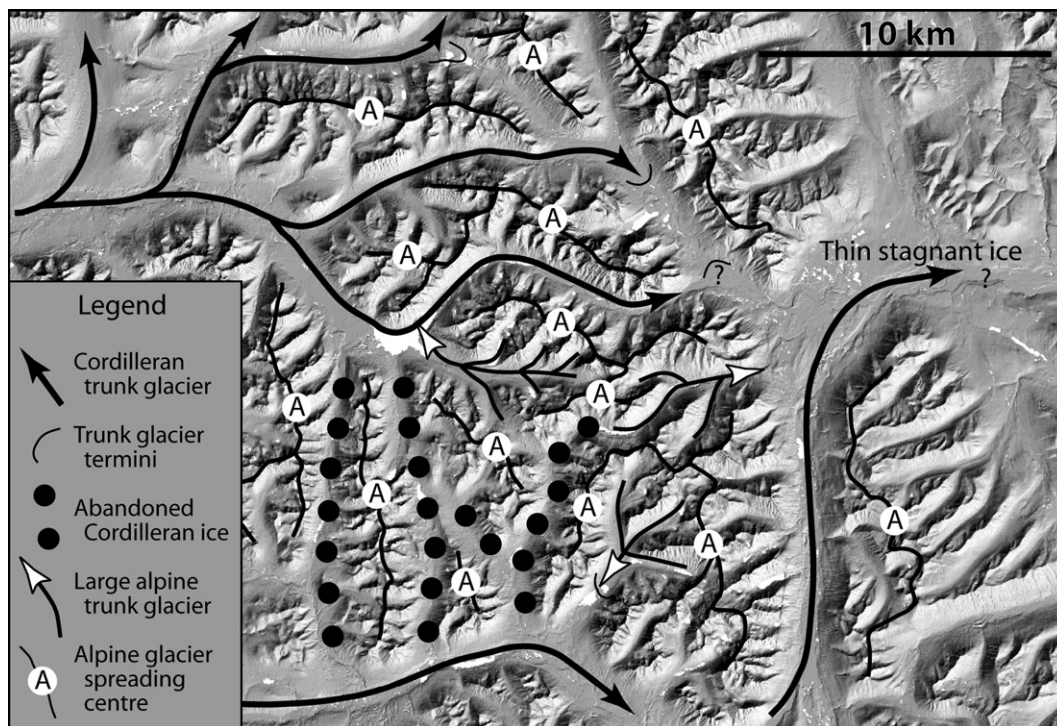


Figure 8. Hillshaded digital elevation model of the study area showing the pattern of ice flow during the late-glacial advance. Hillshaded digital elevation model was produced with permission using digital elevation data from the Province of British Columbia, Base Mapping and Geomatic Services Branch.

core from Red Barrel Lake (Fig. 7; Table 1). Unidentified plant material 1 cm above the upper tephra in Cushing Lake returned an age of 10,560–10,220 cal yr BP (9180 ± 80 ^{14}C yr BP; Fig. 7; Table 1). The occurrence of the phonolitic tephra in basal sediments from five lakes in the study area, of which three are inside late-glacial terminal and lateral moraines, indicates that deglaciation was complete by no later than 10,560–10,220 cal yr BP (9180 ± 80 ^{14}C yr BP). This age is also a minimum age for construction and abandonment of the late-glacial moraines.

Another limiting date for deglaciation is an age of 10,550–10,250 cal yr BP (9230 ± 50 ^{14}C yr BP) for wood, obtained from basal sediments in a core from Sandwich Lake (Fig. 7; Table 1). Sandwich Lake lies within a lateral meltwater channel at the same elevation as adjacent meltwater channels that crosscut late-glacial lateral moraines. This age is thus a minimum date for abandonment of the moraines.

Data from the Rocky Mountains to the east provide a maximum age for the Finlay Advance. Bednarski (2000, 2001) and Bednarski and Smith (2007) reported three cosmogenic chlorine-36 surface exposure ages on striated rock surfaces in the northern Rocky Mountains (Fig. 2). Bednarski and Smith (2007) concluded that the last ice sheet in northern British Columbia withdrew from summits in the front ranges of the northern Rocky Mountains as late as 14,800–13,200 ^{36}Cl yr ago. Ice cover in the Finlay River area at that time must have been greater than at the time of the Finlay Advance, because the exposure ages were obtained on high ridges that had been covered by ice flowing east from the Skeena ice divide 200–300 km to the west and 150 km west of the Finlay River area (Fig. 2; Mathews, 1980; Ryder and Maynard, 1991). Thus, the Finlay Advance is younger than 14,800–13,200 ^{36}Cl yr.

Discussion

Relation between trunk and tributary glaciers

The Finlay Advance occurred when valley glaciers were present at intermediate and high elevations in the northern Omineca and southern Cassiar mountains. At that time, remnant tongues of decaying Cordilleran ice sourced from the northern Skeena Mountains occupied many trunk valleys in the study area (Fig. 8).

Trunk glaciers in the Finlay, Toodoggone, and Chukachida River valleys flowed from the Skeena Mountains (Fig. 2). The large trunk glacier in the Chukachida River valley diverged into several smaller trunk glaciers in Geese Creek, Cushing Creek, and intervening valleys (Fig. 8). The trunk glacier in Finlay River valley was 200–300 m thick and thinned to the north. Its terminus was a lobe of thin stagnant ice that reached into the northern Rocky Mountain Trench. The two prominent kame terraces in Finlay River valley were deposited against the Finlay trunk glacier by two retreating alpine glaciers in adjacent tributary valleys (Fig. 6), indicating that the Finlay trunk glacier was about 60–80 m thick when alpine glaciers abandoned their late-glacial moraines. The abundant lateral moraines and the absence of terminal moraines in Geese Creek valley suggest that an active trunk glacier flowed to the east when the late-glacial

moraines were constructed (Fig. 3). Based on the location and elevation of lateral meltwater channels, this trunk glacier was about 200 m thick, thinned eastward, and terminated against resurgent alpine glaciers in the northwest tributary of Thudaka Creek (Fig. 3). A trunk glacier of comparable thickness also occupied Cushing Creek valley. It thinned to the east and likely terminated near Thudaka Creek. The rock avalanche in Cushing Creek valley is late-glacial in age based on the lithology and distribution of the debris. The steep, linear west margin of the debris indicates that the rock avalanche fell onto the terminus of the Cushing Creek trunk glacier as it retreated to the west. The exotic granite cobbles and boulders in the debris are likely derived from the Cushing Creek trunk glacier.

Trunk glaciers in Belle, Midas, Mulvaney, Junkers, and Jack Lee Creek valleys were up to 250 m thick. Given their north–south orientations, these glaciers were likely independent of the late-glacial Cordilleran ice sheet (Fig. 8). Several well preserved terminal moraines that cross the floors of these valleys indicate that the glaciers were either stagnant or had thinned significantly before alpine glaciers abandoned their late-glacial moraines. Several ice-contact deltas demonstrate that ice-dammed lakes formed in these valleys after alpine glaciers constructed their late-glacial moraines. The lakes were probably dammed by large trunk glaciers in Toodoggone River valley to the south and Chukachida River valley to the north or by several resurgent alpine glaciers that blocked local drainage (Fig. 8).

Some trunk valleys in the study area, such as lower Thudaka Creek valley, were ice-free prior to construction of the late-glacial moraines (Fig. 5). The moraines in lower Thudaka Creek valley have no crosscutting landforms indicative of interaction with Cordilleran trunk glaciers; thus the Thudaka trunk glacier must have retreated before the late-glacial moraines there were constructed. Lateral meltwater channels and a kame terrace in upper Thudaka Creek valley delineate the margin of the late-glacial Thudaka trunk glacier (Fig. 5). This glacier terminated approximately 1–2 km north of the terminal moraine that impounds Red Barrel Lake and coalesced with a Cordilleran trunk glacier in an adjacent valley to the west (Fig. 5). Meltwater channels on the distal side of the Red Barrel terminal moraine record meltwater drainage to the southeast into lower Thudaka Creek valley (Fig. 5).

Lateral meltwater channels indicate that trunk glaciers thinned and retreated rapidly after construction of the late-glacial moraines. Eskers and deeply incised subglacial meltwater channels in many of the major valleys record stagnation and final wastage of trunk glaciers. Alpine glaciers abandoned their moraines and retreated back into cirques. Some alpine glaciers constructed small moraines during retreat, but in general recessional moraines are uncommon.

Pattern and style of deglaciation

The Finlay Advance occurred during a period of deglaciation that Fulton (1967, 1991) termed the *transitional upland phase*, when downwasting of the ice sheet uncovered upland areas but regional ice flow continued in valleys. In the Finlay River area, drawdown rapidly transformed the ice sheet into a labyrinth of

downwasting valley glaciers (Fig. 8). Under these conditions, the regional equilibrium line was sufficiently low that alpine glaciers could persist at intermediate to high elevations. These alpine glaciers later advanced and constructed large moraines when the equilibrium line fell to 1650–1750 m asl. It is possible that alpine glaciers completely disappeared prior to the late-glacial advance, but this scenario seems unlikely given the size of the moraines and the glaciers that constructed them. Following the advance, a large and rapid rise in equilibrium line forced alpine glaciers to retreat and trunk glaciers to stagnate as the latter became detached from their source areas in the northern Skeena Mountains. By approximately 11,000 cal yr BP, glaciers in northern British Columbia probably were no more extensive than today (Clague, 1981).

The pattern of deglaciation in the Finlay River area, excluding final stagnation of trunk glaciers, appears to differ from that documented in southern British Columbia by Fulton (1967, 1991). Fulton concluded that upland areas in southern British Columbia appeared through the ice sheet first due to downwasting and complex frontal retreat. Cirques at intermediate and, locally, high elevations on the southern Interior Plateau were deglaciated before adjacent valleys during late-glacial time. In contrast, ice persisted in high mountain valleys in the northern Skeena, Omineca, and Cassiar mountains while the ice sheet downwasted. The alpine physiography of the study area and its proximity to major ice accumulation areas in the northern Skeena Mountains are probably the reasons alpine glaciers persisted at intermediate to high elevations during late-glacial time. A climate favourable for glacier nourishment during the latest Pleistocene may also partly account for widespread persistence of alpine glaciers in northern British Columbia.

Glacier resurgence during the Finlay Advance

No stratigraphic evidence was found to unequivocally demonstrate that alpine glaciers advanced during late-glacial time. However, some glaciers clearly constructed moraines in areas that had previously been deglaciated (Fig. 5), which is strong evidence for an advance of alpine glaciers during or after retreat of the Cordilleran ice sheet in northern British Columbia. Elsewhere, alpine glaciers constructed lateral moraines abutting active but decaying ice in trunk valleys (Figs. 3 and 6). The moraines record a discrete advance and not a stillstand during separation of tributary alpine glaciers from trunk glaciers, because the size and character of the moraines can only be explained by abundant remobilization of late-glacial sediments (e.g., the Red Barrel moraines; Fig. 5). Equilibrium line altitudes calculated for alpine glaciers that did not coalesce with remnant Cordilleran ice in trunk valleys at the time of moraine construction suggest that snowline was 1650–1750 m asl compared to approximately 1900–1950 m asl during the Little Ice Age.

A drop in equilibrium line altitude in the Finlay River area to 1650–1750 m asl allowed alpine glaciers to advance and construct lateral and terminal moraines in high tributary valleys (Fig. 3) and in areas where trunk glaciers were stagnant or absent (Figs. 5 and 6). Trunk glaciers were lower than the regional equilibrium line and thus did not subsequently override

and remove the Finlay moraines. The complex geometry of the trunk glacier network and the distance of these glaciers from major source areas in the northern Skeena Mountains attenuated the effects of positive mass balance that would have otherwise led to their thickening and advance. Ice in the northern Skeena Mountains probably thinned so much earlier during deglaciation that accumulation centres were unable to nourish the distal, complex valley glacier network when alpine glaciers advanced in the Finlay River area.

The large size of the lateral moraines may be due to the large amount of sediment available to the glaciers at the time of the Finlay Advance. As mentioned above, alpine glaciers at intermediate and high elevations survived initial downwasting and retreat of the Cordilleran ice sheet, with attendant transport and deposition of large amounts of supraglacial and subglacial debris. In particular, rockfall and landslides from freshly deglaciated cirque headwalls probably contributed large amounts of debris, which were transported down-glacier and deposited in the moraines.

Correlative landforms and deposits

Late-glacial landforms that may correlate with the Finlay moraines are Younger Dryas and pre-Younger Dryas in age. Younger Dryas glacier advances have been documented in many areas in western North America, including the central and southern Coast Mountains of British Columbia (Clague, 1985; Friele and Clague, 2002a,b), the southern Canadian Rocky Mountains (Reasoner et al., 1994), the northern Cascade Range of Washington State (Kovanen and Easterbrook, 2001; Riedel et al., 2003), and the Colorado Front Ranges (Menounos and Reasoner, 1997). Other advances of possible Younger Dryas age have been reported in several other areas of western North America (Zielinski and Davis, 1987; Gosse et al., 1995; Briner et al., 2002; Kovanen, 2002; Kovanen and Easterbrook, 2002; Owen et al., 2003; Licciardi et al., 2004).

Friele and Clague (2002a,b) documented a Younger Dryas advance of the Squamish Valley glacier in the southern Coast Mountains, 50 km north of Vancouver, British Columbia. East of the crest of the southernmost Coast Mountains, glaciers were restricted to cirques and high valleys with accumulation areas above about 2000 m asl (Souch, 1989; Friele and Clague, 2002a). Clague (1985) reported a Younger Dryas advance of a tidewater glacier in the central Coast Mountains near Terrace, British Columbia, but its climatic significance is uncertain. Crowfoot moraines record an advance of glaciers in the Canadian Rocky Mountains during the Younger Dryas (Reasoner et al., 1994). These moraines commonly lie immediately outside Little Ice Age moraines or have been incorporated into them, thus Crowfoot glaciers were comparable in extent to those of the Little Ice Age (Reasoner et al., 1994; Osborn and Gerloff, 1997; Osborn et al., 2001). The small size of Crowfoot glaciers suggests that they had decayed significantly from their maximum late Wisconsinan limits and in some cases may have formed in cirques that had been completely deglaciated prior to the onset of the Younger Dryas (Reasoner et al., 1994; Osborn and Gerloff, 1997; Osborn et al., 2001).

Briner et al. (2002) suggested that the late-glacial Mount Waskey advance in the Ahklun Mountains of southwestern Alaska culminated between 12,400 and 11,000 $^{10}\text{Be}/^{26}\text{Al}$ yr ago, based on cosmogenic surface exposure ages of boulders on moraines. During this advance, resurgent valley glaciers constructed large terminal and lateral moraines several kilometres downvalley of modern glacier termini. The Mount Waskey moraines are about 80 km inside the late Wisconsinan glacial limit, indicating that the advance occurred after substantial deglaciation during late-glacial time.

Late-glacial advances prior to the Younger Dryas have been documented at several localities in the Canadian Cordillera. Clague et al. (1997) presented evidence for two pre-Younger Dryas advances of a lobe of the Cordilleran ice sheet in southwest British Columbia. Saunders et al. (1987) reported evidence for the same advances in nearby Chilliwack River valley, British Columbia. A pre-Younger Dryas advance or stillstand of the trunk glacier in Squamish River valley is marked by an end moraine at Porteau Cove, 15 km south of Squamish, British Columbia (Friele and Clague, 2002b).

Poorly dated latest Pleistocene moraines have been reported elsewhere in the North American Cordillera. Some moraines in the Canadian and U.S. Rocky Mountains pre-date the Crowfoot advance and record local stillstands or advances during the late-glacial interval (Reasoner et al., 1994; Osborn and Gerloff, 1997). Undated moraines have been identified in at least four sites in northern British Columbia: the Tuya–Teslin area (Watson and Mathews, 1944), Level Mountain (Hamilton, 1981), the Atlin area (Tallman, 1975), and the Omineca Mountains (Roots, 1954). Late-glacial advances and stillstands of the northern Cordilleran ice sheet have been reported from several valleys in southern Yukon Territory (Hughes, 1990; Jackson et al., 1991; Bond, 2004; Kennedy and Bond, 2004; Bond and Kennedy, 2005).

Late-glacial moraines in the Cassiar Mountains near Dease Lake are similar in size and character to the Finlay moraines (Lakeman, 2006). Many of the valley glaciers that constructed the moraines coalesced with trunk glaciers that were part of the northern Cordilleran ice sheet. Sediment cores recovered from lakes inside the moraines in the Dease Lake area contain the same phonolitic tephra as the cores from the Finlay River area reported in this paper (Lakeman, 2006). The presence of the same tephra in the two areas some 250 km apart suggests that the moraine complexes are correlative and are the product of the same late-glacial climate event.

Late-glacial climate of northern British Columbia

Paleoecological data from north-coastal British Columbia provide evidence for regional climate change during deglaciation. The most pronounced and consistent change in the records is the Younger Dryas climate deterioration (Engstrom et al., 1990; Mathewes, 1993; Mathewes et al., 1993; Hetherington and Reid, 2003; Lacourse, 2005). Mathewes (1993) and Mathewes et al. (1993) documented a late-glacial shift from closed forest to open, herb-rich parkland at two sites near Cape Ball, Graham Island; the parkland persisted from about 12,860–12,670 cal yr BP

(10,730±60 ^{14}C yr BP) to 12,350–11,400 cal yr BP (10,200±100 ^{14}C yr BP). Lacourse (2005) presented similar evidence for cooling during the Younger Dryas on northern Vancouver Island. Mathewes et al. (1993) showed that waters on the central British Columbia continental shelf were cooler from 13,200–12,850 cal yr BP (11,070±120 ^{14}C yr BP) to 12,840–10,790 cal yr BP (10,170±360 ^{14}C yr BP) than either earlier or later, based on abundances of the cold-water benthic foraminifera *Cassidulina reniforme* in several marine sediment cores. Similarly, surface waters in Hecate Strait, east of the Queen Charlotte Islands, were cooler from 12,920–12,820 cal yr BP (10,890±50 ^{14}C yr BP) to 12,380–11,650 cal yr BP (10,250±80 ^{14}C yr BP) than at times immediately earlier or later based on the absence of most bivalves, except species tolerant of low sea-surface temperatures (Hetherington and Reid, 2003). Engstrom et al. (1990) reported that pine parkland on Pleasant Island, southeastern Alaska, was replaced by shrub- and herb-dominated tundra about 12,960–12,400 cal yr BP (10,770±120 ^{14}C yr BP).

These paleoecological studies document a return to cool and wet conditions in northern British Columbia and southeast Alaska during the Younger Dryas. The records provide no evidence of a comparable late-glacial climate oscillation either before or after the Younger Dryas, but these oscillations may not be resolvable in the existing paleoecological records. Nevertheless, the Younger Dryas was likely the most significant event of its kind in northern British Columbia during the late glacial period.

Regional synthesis

The disparity in the magnitude of glacier advances in the Finlay River area and elsewhere in the Canadian Cordillera indicate that glaciers responded differently to climate reversals during late Pleistocene deglaciation. The Finlay moraines imply that the timing and pattern of ice sheet retreat were important in determining glacier extents at the onset of late-glacial climate deterioration. Different styles and rates of deglaciation were probably primary mechanisms causing varied glacier responses to late-glacial climate change. Extensive advances in coastal and northern alpine areas near former accumulation centres or ice divides where ice persisted early during deglaciation (e.g., the Squamish and Finlay advances) support this assertion. Near-complete deglaciation of the southern Canadian Rocky Mountains by Younger Dryas time and perhaps relatively less precipitation may explain the comparatively small Crowfoot advances. Different temperature and precipitation gradients and the variable effects of remnant, large ice masses on local atmospheric circulation may have also contributed to the complexity in glacier fluctuations at the end of the Pleistocene in western North America. Without regional glaciological modeling, however, the relative importance of these factors remains uncertain.

Conclusions

Glaciers flowing from the northern Skeena Mountains covered the Finlay River area of northern British Columbia at the climax of the Fraser glaciation. Large alpine glaciers persisted at intermediate and high elevations early during deglaciation as

the Cordilleran ice sheet decayed by downwasting and frontal retreat.

Large terminal and lateral moraines were constructed in tributary valleys in the Finlay River area during a regional late-glacial advance of alpine glaciers following early deglaciation. The Finlay Advance coincided with final decay of a network of remnant, thin tongues of the northern Cordilleran ice sheet in trunk valleys, and it occurred when regional equilibrium line altitudes fell to 1650–1750 m asl. Trunk glaciers were detached from major accumulation centres and were well below the lowered regional equilibrium line altitude that allowed alpine glaciers to advance. The late-glacial moraines are well preserved and clearly record interaction between alpine glaciers and decaying trunk glaciers. The large size of the moraines is probably due the abundance of sediment available for transport during deglaciation.

Two phonolitic tephra slightly older than 10,560–10,220 cal yr BP (9180±80 ¹⁴C yr BP) occur in the study area and closely delimit the time of abandonment of the late-glacial moraines. An age of 10,550–10,250 cal yr BP (9230±50 ¹⁴C yr BP) from a meltwater channel cut during retreat of trunk glaciers from the area confirms a late-glacial age for the moraines. The advance occurred after the Cordilleran ice sheet had withdrawn from summits in the front ranges of the northern Rocky Mountains 14,800–13,200 ³⁶Cl yr ago.

The widespread distribution of the late-glacial moraines points to regional climatic deterioration as a probable trigger for the Finlay Advance. Paleoecological records from northern British Columbia indicate that the most severe cold interval during late-glacial time coincided with the Younger Dryas chronozone. The late-glacial moraines in the Finlay River area may thus record a regional advance of alpine glaciers at that time. This correlation, however, must be tempered by the fact that the absolute age of the Finlay Advance remains poorly constrained.

Younger Dryas and older glacier advances in the southern Coast Mountains of British Columbia, the southern Canadian Rocky Mountains, southwestern Alaska, the northern Cascade Range, and other areas in the western United States differ in size, suggesting variable glacier responses to late-glacial climate change. The different responses probably reflect different styles of deglaciation, different temperature and precipitation gradients, differing rates and magnitudes of late-glacial climate change, and variable effects of remnant ice masses on local atmospheric circulation. The stochastic nature of rapid changes in the atmosphere and oceans at the end of the last glaciation also may account for the sporadic occurrence and variable scale of terminal Pleistocene glacier advances in western North America.

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