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Latest Pleistocene and Holocene glacier fluctuations in western Canada

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ABSTRACT

We summarize evidence of the latest Pleistocene and Holocene glacier fluctuations in the Canadian Cordillera. Our review focuses primarily on studies completed after 1988, when the first comprehensive review of such evidence was published. The Cordilleran ice sheet reached its maximum extent about 16 ka and then rapidly decayed. Some lobes of the ice sheet, valley glaciers, and cirque glaciers advanced one or more times between 15 and 11 ka. By 11 ka, or soon thereafter, glacier cover in the Cordillera was no more extensive than at the end of the 20th century. Glaciers were least extensive between 11 and 7 ka. A general expansion of glaciers began as early as 8.4 ka when glaciers overrode forests in the southern Coast Mountains; it culminated with the climactic advances of the Little Ice Age. Holocene glacier expansion was not continuous, but rather was punctuated by advances and retreats on a variety of timescales. Radiocarbon ages of wood collected from glacier forefields reveal six major periods of glacier advance: 8.59–8.18, 7.36–6.45, 4.40–3.97, 3.54–2.77, 1.71–1.30 ka, and the past millennium. Tree-ring and lichenometric dating shows that glaciers began their Little Ice Age advances as early as the 11th century and reached their maximum Holocene positions during the early 18th or mid-19th century. Our data confirm a previously suggested pattern of episodic but successively greater Holocene glacier expansion from the early Holocene to the climactic advances of the Little Ice Age, presumably driven by decreasing summer insolation throughout the Holocene. Proxy climate records indicate that glaciers advanced during the Little Ice Age in response to cold conditions that coincided with times of sunspot minima. Priority research required to further advance our understanding of late Pleistocene and Holocene glaciation in western Canada includes constraining the age of late Pleistocene moraines in northern British Columbia and Yukon Territory, expanding the use of cosmogenic surface exposure dating techniques, using multi-proxy paleoclimate approaches, and directing more of the research effort to the northern Canadian Cordillera.

1. Introduction

Mountain glaciers are sensitive indicators of environmental change. They respond to changes in climate by adjusting their width, length, and thickness. Because most valley glaciers are laterally constrained and ice deforms under its own weight, fluctuations in length are the most common response to long-term changes in climate. These adjustments, however, do not occur immediately; they depend on the response time of the glacier, which for mountain glaciers can range from several years to decades (Nye, 1960; Jóhannesson et al., 1989; Harrison et al., 2003). Nevertheless, the geologic evidence of glacier fluctuations can be used to document Holocene regional climate variability at time scales of decades to centuries.

In western Canada, as elsewhere in the Northern Hemisphere, glaciers expanded from minimum extents in the early Holocene to maximum extents some 150–300 years ago (Porter and Denton, 1967; Denton and Karlén, 1973; Osborn and Luckman, 1988). This long-term, progressive expansion has truncated the surface record of past glacier activity, because recent advances commonly overrode or destroyed lateral and terminal moraines that demarcate former glacier limits. However, fragmentary evidence of these events is preserved in some moraines and glacier forefields where subfossil wood in growth position, detrital wood, paleosols, or tephras are exposed. Glacier retreat in the 20th century and the first decade of the 21st century has provided many such exposures, which can be exploited through stratigraphic, sedimentologic, and geochronologic studies to improve understanding of Holocene

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glacier history (Osborn and Luckman, 1988; Smith and Desloges, 2000; Koch et al., 2007a,b; Osborn et al., 2007).

Sheared stumps in growth position in glacier forefields provide direct evidence of glacier fluctuations (Menounos et al., 2004). In contrast, detrital wood collected in glacier forefields may have been delivered onto the glacier by mass wasting, or it may have been reworked from older deposits. Ryder and Thomson (1986) provide an excellent discussion of the origin and interpretation of wood collected from glacier forefields.

Proglacial lake sediments afford an additional, indirect but continuous proxy of upvalley glacier fluctuations (Karlén, 1981; Leonard, 1986; Souch, 1994; Leonard and Reasoner, 1999; Menounos, 2002). Lengthy periods characterized by high sedimentation rates or high clastic content generally coincide with times of greater ice extent (Leonard, 1997). However, changes in sediment delivery from alpine glaciers, proglacial sources, or non-glacierized terrain may obscure the relation between proglacial lake sedimentation and glacier activity (Hallet et al., 1996; Leonard, 1997; Menounos, 2002). Lake sediment records in combination with more direct evidence from glacier forefields offer considerable scope for reconstructing alpine glacier activity during the Holocene. In some cases, the combined use of clastic records with detrital wood strengthens the case for regional glacier activity if both records accord (Menounos et al., 2004, 2008).

Twenty years have elapsed since the last major review of Holocene glacier fluctuations in western Canada (Osborn and Luckman, 1988). Over that time, many studies and discovery of new sites have contributed to an improved understanding of glacier activity in this region. Accelerated glacier retreat in the 1990s exposed new evidence at sites studied previously, and many new proglacial lake sediment records have become available. In this paper we summarize and interpret the evidence of the latest Pleistocene and Holocene glacier fluctuations in western Canada that has appeared in the past 20 years. Further, we consider some of the forcing factors responsible for decadal to millennial changes in glacier cover. Although the focus of the review is on western Canada, we refer to sites in the United States close to the International Boundary. We limit our review to evidence that is most directly associated with glacier fluctuations: lateral and end moraines, detrital and *in situ* (growth position) wood in glacier forefields, and proglacial lake sediment records. We present the evidence from the oldest to youngest periods and from maritime to continental localities. Ages are reported in both radiocarbon years (^{14}C yr BP) and calibrated radiocarbon years before AD1950 (ka) from 16 to 1.0 ka. We calibrated radiocarbon ages with the calibration program CALIB 5.02 (Stuiver et al., 2005) and report the 95% confidence limits of these calibrated ages. For the past millennium we use AD for true calendar ages and cal yr AD for calendar-equivalent radiocarbon years.

The most recent definition of the base of the Holocene is 11.7 ka (Walker et al., 2008). According to this definition Younger Dryas-equivalent and older advances described in this paper are latest Pleistocene in age, and events subsequent to the Younger Dryas interval are Holocene in age.

1.1. Geographical setting

The Canadian Cordillera is a broad mountainous region that extends from the International Boundary to the Arctic Ocean (Fig. 1). The region includes many rugged mountain ranges with local relief exceeding 600 m. The tallest mountain in the Cordillera is Mount Logan in southwest Yukon (5959 m asl), but the highest peaks in most mountain ranges are 2000–3500 m asl. The Cordillera can be subdivided into Western, Interior, and Eastern systems (Bostock, 1949). The Western System includes the Insular, Cascade, Coast, and St. Elias mountains. The Interior System includes the Purcell, Selkirk, Cariboo, and Monashee mountains in the south and the Hazelton, Skeena, Cassiar, Omineca, and Ogilvie mountains in the north (Fig. 1). These interior mountain ranges border extensive areas of low relief, referred to as the Interior Plateaus in British Columbia and the Yukon Plateaus in Yukon Territory. The Eastern System includes the Rocky Mountains in the south and the Selwyn, Mackenzie, and Richardson mountains in the north (Fig. 1). The Canadian Rocky Mountains are bordered by the Rocky Mountain Trench to the west and the Rocky Mountain Foothills to the east. The Tintina Trench, which is the northern extension of the Rocky Mountain Trench, borders the Selwyn Mountains.

Glaciers today cover about 30,000 km^2, or 3%, of the landmass of British Columbia and 12,500 km^2, or 2.6%, of Yukon (Schiefer et al., 2007; Moore et al., in press). The Coast and St. Elias Mountains contain the largest glaciers and are most heavily glacierized mountain ranges in the Canadian Cordillera. Glaciation levels rise eastward, largely due to the diminished influence of maritime air masses (Østrem, 1966). Median elevations of glaciers range from 1540 m asl in the Insular Mountains to 2550 m asl in the southern Rocky Mountains (Schiefer et al., 2008).

2. Latest Pleistocene glacier fluctuations (16,000–11,000 ka)

2.1. Decay of the Cordilleran ice sheet

The last 5000 years of the Pleistocene was a time of "flickering" climate, characterized by rapid switching between glacial and interglacial states. In northwest North America, the late-glacial period was marked by decay of the Cordilleran ice sheet, the large body of confluent glaciers that covered nearly all of British Columbia, southern Yukon, southern Alaska, and the northwestern conterminous United States (Fig. 2; Clague, 1989). This ice sheet achieved its maximum size about 16.5 ka, but had disappeared by the beginning of the Holocene (Fulton, 1971; Clague, 1989).

Deglaciation was characterized by complex frontal retreat at the periphery of the ice sheet and by down-wasting through much of the interior (Fulton, 1967, 1991). The western periphery of the ice sheet became unstable and rapidly retreated shortly after 16 ka due to climate warming and eustatic sea-level rise. Frontal retreat also occurred at the same time or shortly thereafter in southern Yukon and northernmost Washington, Idaho, and Montana. In areas of low and moderate relief nearer the center of the ice sheet, deglaciation proceeded mainly by down-wasting, stagnation, and complex retreat of active ice. Fulton (1967, 1991) proposes four stages in deglaciation of these areas: (1) an active ice phase during which regional flow continued but diminished as ice thinned; (2) a transitional upland phase with the highest areas becoming ice-free, but regional flow continuing in major valleys; (3) a stagnant ice phase during which ice was confined to valleys but was still thick enough to flow; and (4) a dead ice phase when dead ice occupied valleys.

Active glaciers persisted longest in some mountain valleys. Many of these glaciers coexisted with tongues of decaying and dead ice in trunk valleys in the Cordilleran interior. In detail, the pattern of retreat was complex, due largely to the considerable relief and topographic complexity of the Cordillera.

Ice sheet decay was interrupted by local, or perhaps regional, advances. Some advances were triggered by abrupt cooling, but others may have been the result of non-climatic causes, including sea-level change and topographic effects. Alpine glaciers advanced near the end of the Pleistocene, by which time the Cordilleran ice sheet had wasted so much that it was no longer responsive to climate cooling. In some areas, resurgent alpine glaciers came into contact with stagnant tongues of the ice sheet. In other areas, small glaciers reformed in cirques that had earlier been completely ice-free. Below, we summarize the evidence for these events.

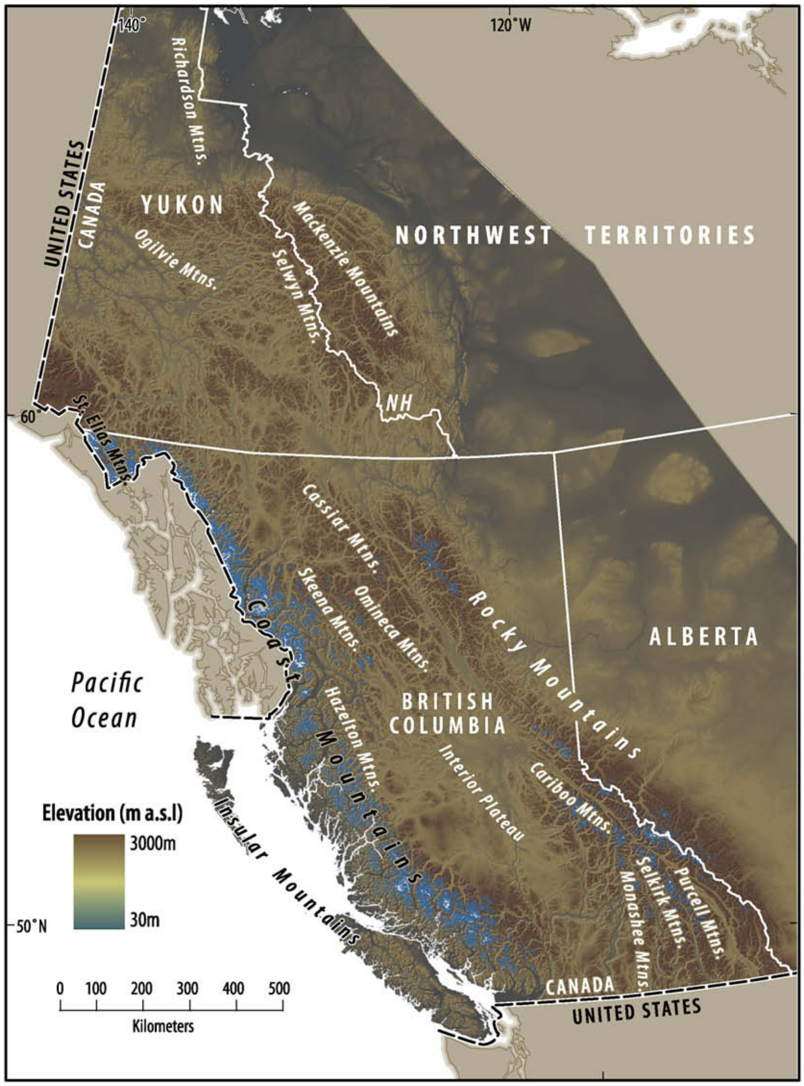

Fig. 1. Relief map of western Canada, showing major mountain systems and ranges. Blue shading shows present-day ice cover.

2.2. Sumas advances

The Puget lobe, at the southwest margin of the Cordilleran ice sheet, reached what is now Olympia, Washington, its southernmost limit, about 17.0–16.6 ka (Porter and Swanson, 1998). It remained there less than 500 years and then rapidly retreated northward, first in contact with glacial lakes that developed on the glacio-isostatically depressed crust, and later in contact with the sea, which invaded Juan de Fuca Strait and Puget Sound (Thorson, 1980, 1989). By 15.5 ka, Puget Sound was free of ice and a large calving embayment had developed in the southernmost Strait of Georgia near the International Boundary. Within perhaps a century or two, the remnant piedmont lobe in the Strait of Georgia had collapsed, in much the same way as the piedmont glacier in Glacier Bay, Alaska, disappeared in the 19th and 20th centuries. Glaciers on Vancouver Island became isolated from the decaying ice sheet on the British Columbia mainland and no longer terminated in the sea. On the mainland coast, active ice became restricted to fjords and to the central and eastern Fraser Lowland, where a large lobe of ice fed from the Coast Mountains and the British Columbia interior continued to flow into the sea (Fig. 3). The early stage of Pleistocene deglaciation was thus characterized by catastrophic, uninterrupted retreat and thinning of the southwest margin of the ice sheet.

Retreat slowed about 14.5 ka, followed by local advances of the margin of the now-depleted ice sheet, referred to by early researchers as the "Sumas advance" (Armstrong et al., 1965). A till overlying non-glacial sediments in the central Fraser Lowland records an advance of the Sumas piedmont lobe at this time (Clague et al., 1997). The lobe subsequently retreated and forest became established on the emergent lowlands to the west. The Sumas lobe advanced again at ca 13.5 ka (Clague et al., 1997; Kovanen, 2002; Kovanen and Easterbrook, 2002), overriding the lowland forest and reaching into northernmost Washington. This advance, the most significant of the Sumas events, brought the piedmont glacier in the Fraser Lowland 15–25 km farther west than its previous position and caused it to move several kilometers into lower Chilliwack Valley, which had earlier become deglaciated (Saunders et al., 1987). Shortly after 13.2 ka, the piedmont lobe resumed its easterly retreat, but it once again advanced sometime between 13.0 and 11.5 ka, during the Younger Dryas Chronozone, as recorded by an end moraine just south of the International Boundary at Sumas, Washington (Kovanen, 2002; Kovanen and Easterbrook, 2002), and by a till and associated end moraine in Chilliwack Valley (Saunders et al., 1987). In summary, the piedmont glacier in the eastern Fraser Lowland advanced at least three times during the last 3000 years of the Pleistocene.

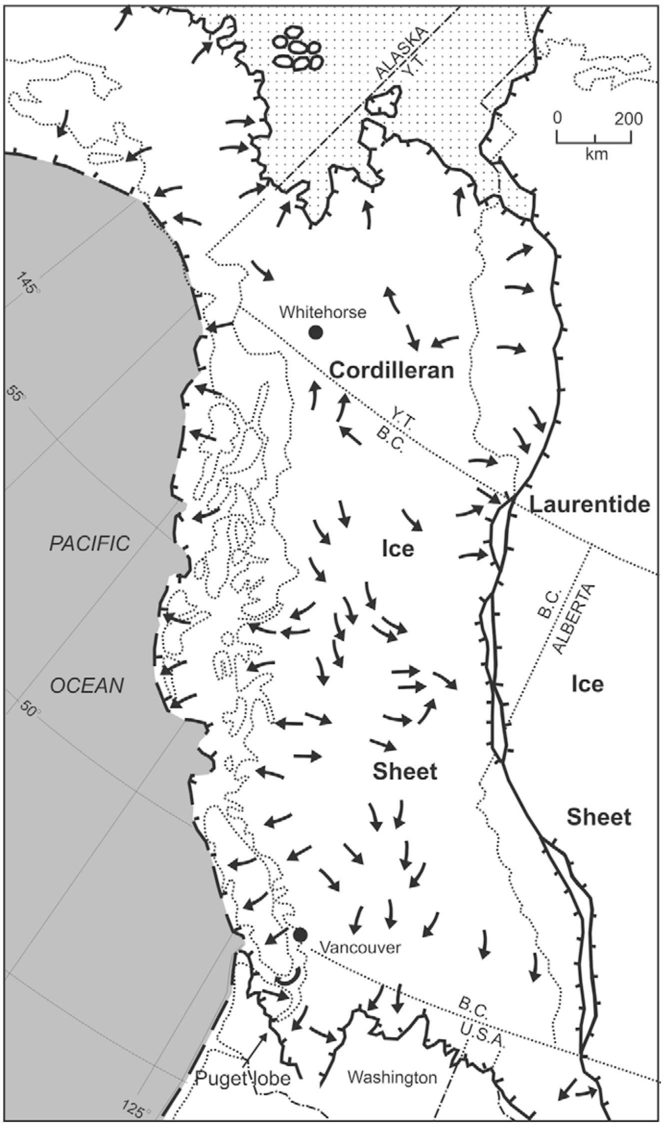

Fig. 2. Extent of the Cordilleran ice sheet and contiguous semi-independent glaciers ca 17.0 ka, at the time of maximum glacierization (modified from Clague, 1989, Fig. 1.12).

2.3. Squamish Valley advances

Friele and Clague (2002a,b) describe a similar sequence of late-glacial events in upper Howe Sound and Squamish Valley, north of Vancouver. When the tidewater glacier in the Strait of Georgia collapsed, ice flowing down Squamish Valley stabilized within the narrows of upper Howe Sound. A large end moraine was constructed at Porteau, where Howe Sound narrows to the north (Fig. 4), probably during an advance documented by Friele and Clague (2002a) in the upper Mamquam Valley, east of Squamish. Although not as securely dated as the Sumas advances, moraine construction probably occurred around 13.5 ka. The glacier that built the moraine subsequently retreated to a position north of Squamish and then, at 12.8 ka, advanced again during the Younger Dryas interval and built another moraine at the head of Howe Sound. Stratigraphic evidence for this second advance includes remnants of a forest buried in outwash (Friele and Clague, 2002b).

Both the Sumas and Squamish Valley advances involved repeated resurgence of the southern margin of the remnant Cordilleran ice sheet. A different story emerges from work recently completed in the Finlay River watershed of northern British Columbia.

2.4. Finlay moraines

Late-glacial moraines situated several kilometers downvalley of Little Ice Age glacier limits have been identified in several mountain ranges in western Canada (Menounos et al., 2005; Grubb, 2006; Lakeman et al., 2008). The moraines are larger and have much greater downvalley extents than both Little Ice Age moraines and Crowfoot moraines (see below); they delineate glaciers that were five to ten times greater in area than Little Ice Age glaciers (Fig. 5). They are found in relatively dry parts of western Canada, including the Finlay River area in the Omineca Mountains, the headwaters of Homathko River on the east flank of the Coast Mountains, the Skeena Mountains east of Dease Lake, the Nahanni River watershed in Northwest Territories, and on Level Mountain northwest of Dease Lake. The moraines in the Finlay River area have been attributed to the “Finlay Advance” by Lakeman et al. (2008), and we here informally refer to the whole family of such moraines in British Columbia and Yukon as ‘Finlay moraines’.

In the Finlay River area the moraines are sharp-crested, up to 120 m high, and as far as 9 km beyond Little Ice Age limits (Lakeman et al., 2008). Several lateral moraines are cross-cut by

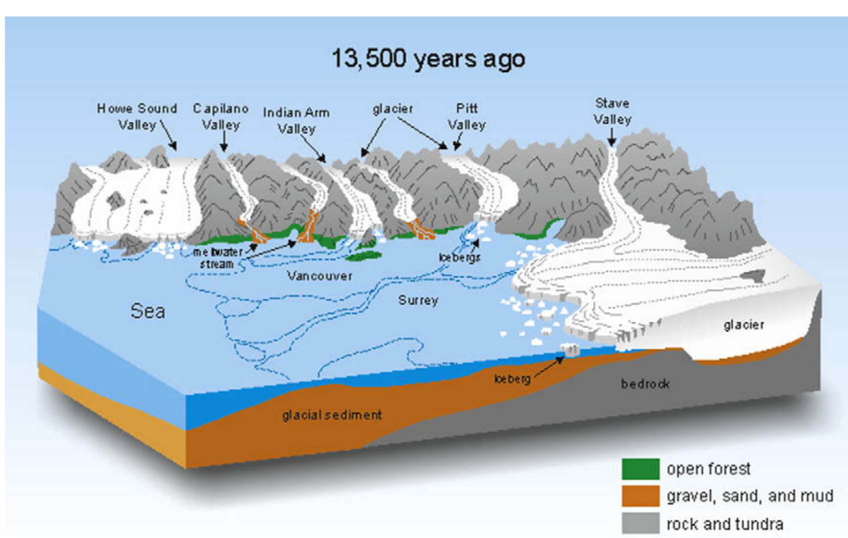

Fig. 3. Schematic diagram showing tidewater glaciers calving into the proto-Strait of Georgia about 13.5 ka (from Clague and Turner, 2003).

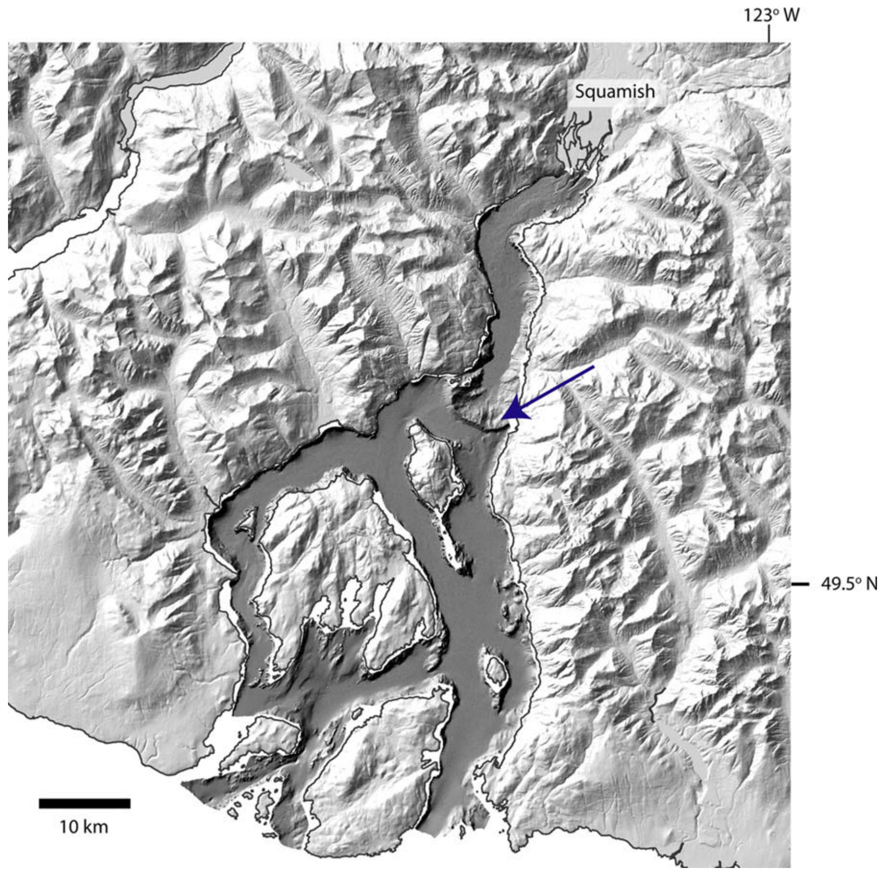

Fig. 4. Multi-beam image of the floor of Howe Sound, showing the submerged end moraine (arrow) at Porteau. The moraine was constructed about 13.5 ka, before the beginning of the Younger Dryas interval, by a glacier flowing down Squamish Valley from the Coast Mountains to the north. Image courtesy of the Geological Survey of Canada.

meltwater channels that record down-wasting of trunk valley ice of the northern Cordilleran ice sheet, whereas other moraines merge with ice-stagnation deposits in trunk valleys. These relationships demonstrate that advancing alpine glaciers interacted with the decaying Cordilleran ice sheet. Sediment cores collected from lakes dammed by the moraines have yielded plant macrofossils as old as 9180 ± 80 ^{14}C yr BP [10.56–10.22 ka], which is a minimum age for the Finlay advance and subsequent glacier retreat (Table 1).

Numerous lateral and terminal moraines extend up to 13 km beyond Little Ice Age moraines in the eastern Coast Mountains near Homathko River. Cyperaceae fragments recovered from basal sediments in a lake impounded by one of these moraines yielded a radiocarbon age of 9390 ± 40 ^{14}C yr BP [10.71–10.51 ka] (Grubb, 2006), which is a minimum age for glacier retreat from the moraine. The moraines record an equilibrium-line-altitude (ELA) depression of 275 m relative to the modern ELA. In contrast the average ELA depression in this area during the Little Ice Age was 85 m (Grubb, 2006).

Finlay moraines also occur in the headwaters of Nahanni River in the Ragged Range, where today only small cirque glaciers remain. The moraines are up to 80 m high and extend two to 10 km downstream from cirque headwalls. The moraines in the valley of Hole in the Wall Creek were assigned by Ford (1976) to his Hole in the Wall Glaciation, which he correlated with the classical Wisconsinan advance of Cordilleran ice. Ford (1976) had no chronologic control for this or other glacial deposits in the Nahanni River watershed, and he offered no compelling reason why his earlier, more extensive Flat River Glaciation could not be the classical Wisconsinan event.

The Finlay moraines are regional in extent and record climatic deterioration near the end of Pleistocene, but they are only loosely bracketed in age between the demise of the Cordilleran ice sheet and 10.50 ka. They could have been constructed during the Younger Dryas climatic reversal and, indeed, Lakeman et al. (2008) infer such a correlation, but it is also possible that they record a pre-Younger Dryas event such as that documented in the Canadian Rockies and the Puget and Georgia Strait lowlands.

2.5. Pre-Younger Dryas moraines and Crowfoot Advance

The southern Rocky Mountains in Canada supported valley glaciers that were semi-independent of the Cordilleran ice sheet during the Late Wisconsinan (Clague, 1989). The larger valley glaciers flowed east into the Rocky Mountain Foothills and onto the easternmost Interior Plains, where they briefly coalesced with ice flowing from the Keewatin sector of the Laurentide ice sheet (Jackson et al., 1997). As climate warmed, and these glaciers were starved of moisture by the Cordilleran ice sheet to the west, and retreated back to cirques. By 14.0 ka, most alpine glaciers were probably no larger than they were at the end of the 20th century. But evidence exists for two minor advances near the end of the Pleistocene – one older, and the other coeval with the Younger Dryas cooling event (Luckman and Osborn, 1979; Reasoner et al., 1994; Osborn and Gerloff, 1997).

Osborn and Gerloff (1997) describe lateral moraines in Jasper, Banff, and Yoho national parks that project to termini up to 6 km downstream of both Little Ice Age and Crowfoot end moraines (see below). Apart from the minimum age imposed by the upvalley Crowfoot moraines, these lateral moraines are undated. However, Osborn and Gerloff (1997) describe a potentially correlative moraine in the Mission Mountains (47.626°N, 113.934°W), which are part of the Montana Rocky Mountains. That moraine is overlain by Glacier Peak G tephra, which has been dated at ca 13.0 ka

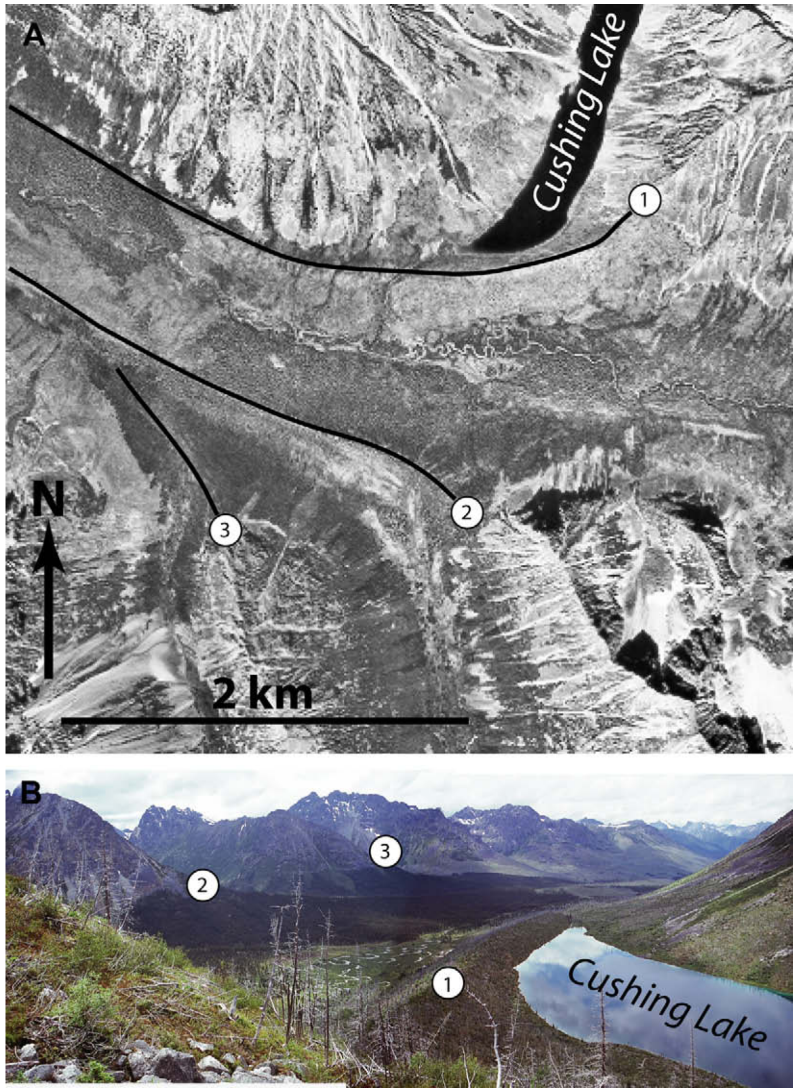

Fig. 5. Late-glacial moraines (1, 2, 3) in the Finlay River watershed, northern British Columbia. (A) Aerial photograph of Cushing Lake, which is dammed by moraine 1. (B) Moraines viewed downvalley (west) from Cushing Lake (from Lakeman et al., 2008).

(Mehringer et al., 1984). The cirque containing the moraine has neither Little Ice Age nor Crowfoot deposits, thus no comparison with those events can be made, but nevertheless the moraine must have been built prior to the Crowfoot Advance.

Pleistocene cirque glaciers formed or advanced near the end of the Pleistocene during the Crowfoot Advance (Luckman and Osborn, 1979; Osborn and Gerloff, 1997). Reasoner et al. (1994) correlated the type moraine of this advance, in Banff National Park, to sediment in adjacent Crowfoot Lake, which they dated to 13.00–11.50 ka. The moraine thus was built during the Younger Dryas chronozone. Correlative moraines, overlain by Mazama tephra and located short distances beyond, or partly covered by, Little Ice Age moraines are found in Banff and Jasper national parks (Luckman and Osborn, 1979), Waterton Lakes National Park and adjacent Glacier National Park, Montana (Osborn, 1985; Carrara, 1987), upper Elk Valley in the British Columbia Rockies (Ferguson, 1978), Mt. Assiniboine Provincial Park (Osborn, unpublished data), and the Purcell and Selkirk mountains (Osborn, unpublished data). In the southern Coast Mountains, what appear to be equivalent moraines have been found just outside Little Ice Age moraines in the Birkenhead Lake and Duffey Lake watersheds (Minkus, 2006) and adjacent to Lillooet Glacier (Walker, 2003). Basal lake sediments 0.5 km downstream of one of the Duffey Lake moraines yielded a radiocarbon age of 9680 ± 40 ^{14}C yr BP [11.21–10.80 ka], which is a minimum for the age of the advance that constructed the moraine (Minkus, 2006). Crowfoot-type moraines may also be present in the Washington Cascades, although the evidence is debated. The Brisingamen moraine in the Enchantment Lakes basin (Waitt et al., 1982) has the same relation to Little Ice Age moraines as Crowfoot moraines in Canada. Evidence from lake sediments indicates that the advance that formed the Brisingamen moraine ended shortly before 11.30 ka, suggesting temporal equivalence with the Younger Dryas (Bilderback and Clark, 2003).

In areas where Crowfoot moraines are common, such as Banff National Park in Alberta and Glacier National Park in Montana, the spatial relation between Crowfoot and Little Ice Age moraines is variable. For example, in Glacier National Park some Crowfoot moraines are hundreds of meters downvalley from Little Ice Age moraines, whereas others are partly buried by them. However, Crowfoot moraines are absent in most cirques in the park, suggesting that Crowfoot ice was limited in extent or absent and that Crowfoot moraines were overrun by Little Ice Age glaciers (Osborn, 1985). Mazama ash occurs below Little Ice Age till and above till of probable Crowfoot age in the north lateral moraine of Stutfield Glacier in Jasper National Park (Osborn et al., 2001), indicating the Crowfoot Advance there was less extensive than the maximum

Table 1
Radiocarbon ages pertinent to latest Pleistocene glacier activity in western Canada.

Laboratory no.[a]	Reference	Material and context	^{14}C age (yr BP)	Calibrated age[b] (ka)	Latitude	Longitude	Elevation (m asl)	Comment
GSC-3340[c]	Clague et al., 1982	Plant detritus above till	16,000 ± 570	20.40–17.93	53°41.7′	131°52.8′	0	Minimum age for ice-sheet deglaciation, Graham Island
QL-1891	Porter and Swanson, 1998	Basal peat in bog	13,610 ± 80	16.64–15.80	47°34.8′	122°10.7′	–	Minimum age for ice-sheet deglaciation, Puget Lowland
GSC-5871	Clague et al., 1997	Stump in peat	11,900 ± 100	13.98–13.50	49°01.4′	122°24.8′	90	Minimum for oldest late-glacial advance, Fraser Lowland
TO-4056	Clague et al., 1997	Conifer needles in peat	11,300 ± 80	13.33–13.03	49°01.3′	122°25.5′	90	Maximum age of younger late-glacial advance, Fraser Lowland
GSC-4097	Saunders et al., 1987	Wood fragment in silt	11,600 ± 90	13.67–13.27	49°04.9′	121°50.6′	240	Maximum age of older late-glacial advance, Chilliwack valley
GSC-4036	Saunders et al., 1987	Branch in paleosol	11,300 ± 100	13.34–12.97	49°04.9′	121°50.6′	300	Minimum age of older late-glacial advance, Chilliwack valley
GSC-4037	Saunders et al., 1987	Branch in paleosol	11,200 ± 90	13.25–12.93	49°04.9′	121°50.6′	300	Maximum age of younger late-glacial advance, Chilliwack valley
GSC-6140	Friele and Clague,,2002b	Outer rings of log in silt	11,900 ± 100	13.98–13.50	49°41.1′	122°54.8′	600	Minimum age for ice-sheet deglaciation, Mamquam valley
TO-6838	Friele and Clague, 2002b	Stick in silt	11,550 ± 110	13.66–13.21	49°41.1′	122°54.8′	600	Maximum age for late-glacial advance, Mamquam valley
GSC-185	Armstrong, 1981	Wood fragment in silt	10,690 ± 180	12.96–12.08	49°35′	123°13′	–	Minimum age for retreat of glacier from Porteau end moraine
Beta-43865	Brooks and Friele, 1992	Root of stump in sand	10,650 ± 70	12.84–12.40	49°43.9′	123°05.3′	90	Minimum age for late-glacial advance, Squamish valley
CAMS 3065	Reasoner et al., 1994	Twig in lake sediment	11,330 ± 220	13.64-12.87	51°39.1′	116°25.4′	1950	Maximum age for Crowfoot Advance, Crowfoot Lake
CAMS 3177	Reasoner et al., 1994	Seed in lake sediment	10,070 ± 420	12.84–10.52	51°39.1′	116°25.4′	1950	Minimum age for Crowfoot Advance, Crowfoot Lake
GSC-1012	Lowdon and Blake, 1976	Wood in silt	10,100 ± 150	12.34–11.23	49°52.2′	118°05.2′	440	Minimum age for ice-sheet deglaciation, Columbia Valley
Beta-192241	Minkus, 2006	Plant matter in silt	9680 ± 40	11.21–10.80	50°16′	122°04′	1790	Minimum age for late-glacial advance, southern Coast Mountains
Beta-192954	Grubb, 2006	Plant matter in silt	9390 ± 40	10.71–10.51	51°43.3′	124°34.9′	1930	Minimum age for late-glacial advance, Chilcotin region
TO-12475	Lakeman et al., 2008	Plant matter in silt	9180 ± 80	10.56–10.22	57°35.6′	126°54.5′	1400	Minimum age for late-glacial advance, Finlay River

[a] Radiocarbon laboratory: Beta = Beta Analytic Inc; GSC = Geological Survey of Canada; CAMS = Center for Accelerator Mass Spectrometry; TO = Isotrace Laboratory; QL = Quaternary Isotope Laboratory.
[b] Calendar ages (±2σ) determined using CALIB 5.02 (Stuiver et al., 2005).
[c] Analytical uncertainty is 2σ for GSC radiocarbon ages and 1σ for all other ages.

advance of the Little Ice Age. The general conclusion that can be drawn from available observations is that the Crowfoot Advance and the maximum Little Ice Age advance were of similar magnitude, the latter being slightly more extensive in most cases.

2.6. Relations between late-glacial moraine systems

The widespread distribution of late-glacial moraines of Younger Dryas age points to regional climatic deterioration as the probable cause for the advance that produced them. Paleoecological records from British Columbia and Alaska (Mathewes, 1993; Mathewes et al., 1993; Patterson et al., 1995; Hansen and Engstrom, 1996; Lacourse, 2005) indicate that the most severe cold interval during late-glacial time coincided with the Younger Dryas chronozone. However, the Younger Dryas-age advances differ in extent, perhaps due to the differences in glacier size or response time. The substantial Sumas and Squamish Valley advances were associated with large valley and piedmont glaciers, whereas the minor Crowfoot Advance involved small high-elevation glaciers at drier, more continental sites. Cooling probably did not induce substantial advances in the eastern Cordillera because most of the large glaciers had disappeared prior to the Younger Dryas. However, differences in the styles of deglaciation, regional temperature and precipitation variations, and differences in the rates and magnitudes of late-glacial climate change may also be possible influences. The stochastic nature of the rapid changes in the atmosphere and oceans at the end of the last glaciation also may contribute to the sporadic occurrence and different scales of terminal Pleistocene glacier advances in western North America.

An important, and as yet unanswered, question is the relation of Finlay moraines to Crowfoot moraines. The two families of moraines have been reported from different regions, the Crowfoot moraines extending across southwest Alberta and southern British Columbia into the southernmost Coast Mountains, and the Finlay moraines occurring farther north in British Columbia. The lack of known overlap may mean that the Finlay moraines are also of Younger Dryas age. If so, the response of glaciers to Younger Dryas cooling was greater at more northerly latitudes, or perhaps in high areas at more northerly latitudes, than in alpine areas of southern British Columbia and southern Alberta. Perhaps the difference in the moraines stems from earlier deglaciation of southern locations – in the south cirque glaciers had to reform during the Younger Dryas, whereas farther north large ice masses were reactivated. Alternatively, the Finlay moraines may predate Crowfoot moraines. If that is the case, either Crowfoot moraines exist to the north but have not yet been found, or any Crowfoot moraines deposited were overrun and destroyed or buried during Little Ice Age advances. The latter scenario has an analog in the Sierra Nevada in California (Clark and Gillespie, 1997). No evidence of a Younger Dryas advance has been found there, but a pre-Younger Dryas event more extensive than the maximum Little Ice Age advance is now well established.

3. Early Holocene glacier fluctuations (11.0–7.50 ka)

The times, magnitudes, and even existence of early Holocene glacier advances in western North America have been debated for over 40 years. In the 1960s and 1970s, some workers proposed early

Holocene advances in the Canadian Rockies that were more extensive than Little Ice Age advances (see Osborn and Luckman, 1988). However, Luckman et al. (1978), Luckman and Osborn (1979), Davis and Osborn (1987), and Osborn and Gerloff (1997), discuss the poor dating control for these events. More recent studies in the southern Coast Mountains (Menounos et al., 2004; Koch et al., 2007a; Osborn et al., 2007a) show that early Holocene glaciers were restricted relative to their late Holocene counterparts. The evidence includes generally clastic-poor early Holocene sediments in proglacial lakes and exposure of early Holocene wood in glacier forefields by ongoing glacier recession. The latter evidence demonstrates that early Holocene forests existed at sites that were recently ice-covered. Detrital wood washed from the snout of Athabasca Glacier in the Canadian Rockies yielded ages from ca 8230 to 7550 ^{14}C yr BP [9.20–8.37 ka], indicating that the glacier was less extensive during this period than in 1993 when the wood was collected (Luckman, 1988; Luckman et al., 1993). Palynological evidence and the presence of subfossil tree remains above the present treeline in the Rockies suggest relatively warm conditions between 8100 and 5900 ^{14}C yr BP [9.10–6.70 ka] (Luckman and Kearney, 1986), as do trees recovered above treeline near Castle Peak in British Columbia (Clague and Mathewes, 1989).

Although early Holocene glacier ice was relatively restricted in western Canada, lake sediments and radiocarbon-dated detrital wood from glacier forefields provide evidence for one or more minor advances during this time. A minor glacier advance in the southern Coast Mountains is suggested by a silty interval, dated to between 10.15 and 7.51 ka in a core collected from lower Joffre Lake, and to between 8.59 and 7.51 ka in cores collected from Green Lake (Menounos et al., 2004). Detrital wood found in recently deglaciated forefields of Sphinx and Sentinel glaciers yielded radiocarbon ages ranging from 8.63 to 8.02 ka, leading Menounos et al. (2004) to suggest that an advance correlative with the 8200-year cold event (Alley and Ágústsdóttir, 2005) was responsible for both the wood and the silty lake sediment horizons (see discussion by Kovanen and Beget, 2005; and reply by Menounos et al., 2005).

It is not clear whether any of the wood washed out of the terminus of the Athabasca Glacier (Luckman et al., 1993; Table 2) is associated with a glacier advance. The detrital sample with an age of 7550 ± 100 ^{14}C yr BP [8.55–8.17 ka] overlaps the age range assigned to the 8200-year advance in the southern Coast Mountains and could relate to a minor advance of Athabasca Glacier during the early Holocene. The origin of the wood with ages of 8230 and 8000 ^{14}C yr BP remains uncertain; there is no supporting evidence for Holocene advances that predate 8.5 ka.

Significant early Holocene glacier advances have been proposed in the Cascade Range of Washington State. Begét (1983) described one such advance on Glacier Peak that was more extensive than the climactic advance of the Little Ice Age. Davis and Osborn (1987) criticized Begét's interpretation, which was later defended by Kovanen and Begét (2005) and further discussed by Menounos et al. (2005). Claims of extensive early Holocene advances have also been made by Heine (1998) at Mt. Rainier and by Thomas et al. (2000) and Kovanen and Slaymaker (2005) at Mount Baker. Reasoner et al. (2001) questioned Heine's (1998) conclusion on the basis of its incongruence with other paleoenvironmental records from the region and suggested that inaccurate bulk-sediment radiocarbon ages might be the problem. Heine's results have not as yet been proven wrong, but the purported early Holocene advance on Mount Baker has been challenged on the basis of new evidence (Davis et al., 2005; Riedel et al., 2006).

4. Neoglaciation

Porter and Denton (1967, p. 205) define "Neoglacial" as "the climatic episode characterized by rebirth and/or growth of glaciers following maximum shrinkage during the Hypsithermal interval." Because the Hypsithermal interval encompasses zones V through VIII of the Danish pollen sequence (Deevey and Flint, 1957) with temporal boundaries of ca 9000 and 2500 ^{14}C yr BP (Mangerud et al., 1974), the original definition of Neoglacial in a strict sense referred to the last 2500 years. But Porter (2000) notes that the Hypsithermal is a *time-stratigraphic unit*, whereas Neoglaciation was defined as a *geologic-climate unit* (American Commission on Stratigraphic Nomenclature, 1961), with the consequence that Neoglaciation and the Hypsithermal interval may overlap, depending on the region. Both Porter (e.g., Porter, 2000) and Denton (e.g., Denton and Karlen, 1973) and most other authors have abandoned the original strict definition and generally refer to Neoglaciation, or the Neoglacial, as the interval when glaciers in the Northern Hemisphere expanded from their minimal extents of the early to mid-Holocene. We use this informal definition in this paper. Matthes (1939) coined the phrase "Little Ice Age" for this interval, but this term is now restricted to fluctuations of the past millennium (e.g. Osborn and Luckman, 1988; Luckman, 2000, 2004; Matthews and Briffa, 2005).

The chronology of glacier fluctuations during Neoglaciation is based on tephras and dated plant fossils from lateral and end moraines, stumps and detrital wood in glacier forefields, and changes in the clastic content of lake sediments. The following survey is subdivided into early, early–middle, middle–late, and late Neoglacial intervals for the purpose of organization; no formal nomenclatural scheme is intended.

4.1. *Early Neoglacial (7.50–5.00 ka)*

4.1.1. *Coast Mountains and Yukon*

Much of the evidence for glacier activity during the early Neoglacial interval comes from the southern Coast Mountains (Fig. 6). In the past 30–50 years, glacier retreat exposed stumps in growth position in forefields of several glaciers in this region (Mathews, 1951). Radiocarbon ages from the stumps and from detrital wood were used by Ryder and Thomson (1986) to define a period of mid-Holocene glacier expansion that they referred to as the "Garibaldi Phase" (6000–5000 ^{14}C yr BP; 6.95–5.62 ka). They used the term "phase" rather than "advance" because they found no evidence to suggest that glaciers subsequently retreated.

Several studies over the past 20 years have demonstrated the complexity and regional nature of the Garibaldi Phase. Outermost rings of *in situ* stumps within several hundred meters of the contemporary margins of Warren and Lava glaciers in Garibaldi Provincial Park (Figs. 6 and 7, Table 2) yielded radiocarbon ages with calibrated age ranges of 7.47–5.89 ka (Koch et al., 2007a; Osborn et al., 2007a). Two of the ages from the Warren Glacier forefield (6370 ± 70 and 6360 ± 80 ^{14}C yr BP) fall outside the radiocarbon age range of 6000–5000 ^{14}C yr BP originally used to define the Garibaldi Phase (Ryder and Thomson, 1986). However, given their proximity to dated stumps originally used to define the Garibaldi Phase, we extend the age range of the Garibaldi Phase by several hundred years to include them. Similar radiocarbon ages have been obtained from detrital wood in till or washed from the termini of Overlord, Sphinx, Warren, Lava, and Stave glaciers (Fig. 6, Table 2). These data must be interpreted with caution, however, because the wood may have been carried to the glacier by non-glacial processes (Ryder and Thomson, 1986).

In situ and detrital wood from the forefields of Bridge and Tchaikazan glaciers, 100 km north of Garibaldi Provincial Park, yielded ages of 6000–5000 ^{14}C yr BP [6.95–5.62 ka] (Smith, 2003). Like the sites in Garibaldi Park, the *in situ* stumps are well inside Little Ice Age limits, but several hundred meters from contemporary glacier margins. Detrital wood of similar age has also been found at Tiedemann Glacier, 215 km northwest of Garibaldi Park

Table 2
Radiocarbon ages pertinent to Holocene glacier activity in western Canada.

Laboratory no.[a]	Reference	Material and context	^{14}C age (yr BP)	Calibrated age[b] (ka)	Latitude	Longitude	Elevation (m asl)
Coast Mountains							
Spearhead Glacier					50°05′N	122°50′W	
Beta-157268	Osborn et al., 2007a	Detrital wood (glacier forefield)	3900 ± 80	4.53–4.09			1995
Beta-168423	Osborn et al., 2007a	Detrital wood (glacier forefield)	3900 ± 60	4.51–4.15			1995
Wedgemount Glacier					50°09′N	122°48′W	
Beta-170671	Koch et al., 2007a	Branch 90 m from snout	8650 ± 60	9.88–9.52			
Helm Glacier					49°57′N	122°59′W	
Beta-168430	Koch et al., 2007a	Log 140 m from glacier snout (1997)	8900 ± 60	10.20–9.78			
Beta-186523	Koch et al., 2007a	Detrital stump at glacier snout (2003)	4080 ± 40	4.81–4.44			
Sphinx Glacier					49°55′N	122°58′W	
Beta-186509	Koch et al., 2007a	Snag in till 300 m from glacier snout	5830 ± 60	6.78–6.49			
Beta-208685	Koch et al., 2007a	Log in till 500 m from glacier snout	4280 ± 70	5.04–4.58			
Beta-186510	Koch et al., 2007a	Snag in till 800 m from glacier snout	3560 ± 70	4.08–3.64			
Beta-186511	Koch et al., 2007a	Snag in glacier forefield 600 m from snout	1570 ± 40	1.54–1.38			
GSC-6770	Menounos et al., 2004	Branch 100 m from glacier snout	7720 ± 80	8.42–8.56			
Sentinel Glacier					49°54′N	122°59′W	
Beta-148787	Menounos et al., 2004	Branch near glacier snout	7720 ± 70	8.63–8.39			
Beta-157267	Menounos et al., 2004	Branch near glacier snout	7470 ± 80	8.42–8.06			
Beta-148786	Menounos et al., 2004	Branch near glacier snout	7380 ± 80	8.36–8.02			
Beta-186508	Koch et al., 2007a	Snag in till 300 m from glacier snout	6040 ± 60	7.16–6.73			
Warren Glacier					49°52′N	123°00′W	
Beta-168425	Koch et al., 2007a	Stick 250 m from glacier snout (2000)	8050 ± 60	9.12–8.66			
Beta-148789	Koch et al., 2007a	*In situ* stump 95 m from glacier snout (2000)	6370 ± 70	7.42–7.17			
Beta-148790	Koch et al., 2007a	*In situ* stump 100 m from glacier snout (2000)	6360 ± 80	7.43–7.03			
Beta-148788	Koch et al., 2007a	*In situ* stump 600 m from glacier snout (2000)	5780 ± 70	6.74–6.41			
Beta-168424	Koch et al., 2007a	Log in till	5700 ± 50	6.64–6.36			
Lava Glacier					49°49′N	122°57′W	
Beta-168426	Koch et al., 2007a	*In situ* stump	6170 ± 60	7.25–6.91			
Beta-168427	Koch et al., 2007a	*In situ* stump	6050 ± 50	7.15–6.75			
Beta-186521	Koch et al., 2007a	*In situ* stump	5760 ± 60	6.71–6.41			
Beta-186520	Koch et al., 2007a	Detrital stump in forefield	5130 ± 40	5.99–5.75			
Beta-186517	Koch et al., 2007a	Log from wood mat in lateral moraine	3190 ± 40	3.55–3.34			
Stave Glacier					49°46′N	122°32′W	
Beta-170668	Koch et al., 2007a	Detrital stump in till	6250 ± 70	7.32–6.97			
Stave Glacier (west tongue)					49°45′N	122°32′W	
Beta-171096	Koch et al., 2007a	Log in lateral moraine	1080 ± 60	1.17–0.83			
Overlord Glacier					50°01′N	122°50′W	
Beta-170665	Osborn et al., 2007a	Detrital wood (glacier forefield)	6170 ± 70	7.25–6.90			1640
Beta-170660	Osborn et al., 2007a	Detrital wood (glacier forefield)	5890 ± 70	6.89–6.51			1625
Beta-170667	Osborn et al., 2007a	Detrital wood (glacier forefield)	5980 ± 70	6.99–6.66			1610
Decker Glacier					50°04′N	122°50′W	
Beta-157265	Osborn et al., 2007a	*In situ* stump rooted within LIA limit	3200 ± 70	3.58–3.26			2040
Beta-157262	Osborn et al., 2007a	Detrital wood (glacier forefield)	2960 ± 50	3.32–2.97			2010
Beta-157263	Osborn et al., 2007a	*In situ* snag rooted within LIA limit	2960 ± 40	3.32–2.99			2010
Beta-157264	Osborn et al., 2007a	*In situ* stump rooted within LIA limit	2920 ± 50	3.24–2.90			2030
Lillooet Glacier					50°45′N	122°44′W	
Wk-12311	Reyes and Clague, 2004	Detrital wood in paleosol below till	3030 ± 40	3.36–3.08			
GSC-6746[c]	Reyes and Clague, 2004	Detrital wood in paleosol below till	2960 ± 60[c]	3.24–3.00			
GSC-6756	Reyes and Clague, 2004	Detrital log in lateral moraine	2490 ± 60	2.72–2.37			
Wk-12307	Reyes and Clague, 2004	Wood fragment from paleosol below till	2440 ± 40	2.70–2.36			
Wk-12313	Reyes and Clague, 2004	Charcoal from paleosol below till	2090 ± 50	2.30–1.93			
Wk-12306	Reyes and Clague, 2004	Charcoal in paleosol between two tills	1720 ± 40	1.71–1.54			
GSC-6767	Reyes and Clague, 2004	Detrital log on top of paleosol below till	1700 ± 80	1.70–1.53			
TO-9754	Reyes and Clague, 2004	Branch in paleosol between two tills	1600 ± 70	1.69–1.35			
Wk-12309	Reyes and Clague, 2004	Wood in paleosol between two tills	1550 ± 50	1.54–1.34			
Wk-12310	Reyes and Clague, 2004	Twig in paleosol between two tills	1530 ± 40	1.52–1.34			
GSC-6760	Reyes and Clague, 2004	Log on top of paleosol between two tills	1390 ± 50	1.34–1.28			
Wk-12308	Reyes and Clague, 2004	Wood from paleosol below till	1090 ± 50	1.17–0.92			
GSC-6606	Reyes and Clague, 2004	Branch on top of peat below till	1090 ± 50	1.06–0.94			
Bridge Glacier					50°50′N	123°30′W	
Beta-197976	Allen and Smith, 2007	Log in gully within LIA limit	2980 ± 60	3.34–2.98			
Beta-181155	Allen and Smith, 2007	Log in paleosol	1930 ± 70	2.04–1.71			
Beta-171549	Allen and Smith, 2007	*In situ* stump	1500 ± 50	1.52–1.30			
Beta-171546	Allen and Smith, 2007	Log within LIA limit	1190 ± 60	1.26–0.98			
Beta-181856	Allen and Smith, 2007	Log within LIA limit	1040 ± 50	1.06–0.80			
Goddard Glacier					51°06′N	124°10′W	
GSC-6046	Menounos et al., 2008	Detrital wood in forefield	4120 ± 60	4.82–4.53			
Jacobsen Glacier					52°03′N	126°04′W	
GSC-4155	Desloges and Ryder, 1990	Wood mat in lateral moraine between two tills	2470 ± 50	2.71–2.36			1370
Tiedemann Glacier					51°19′N	124°59′W	
Beta-220941	Clague, unpublished	Log in lowest section of moraine	5010 ± 40	5.89–5.65			
UCIAMS-40663	Clague, unpublished	*In situ* stump	3865 ± 20	4.41–4.18			
UCIAMS-40660	Clague, unpublished	*In situ* stump	3820 ± 20	4.29–4.10			

(*continued on next page*)

Table 2 (*continued*)

Laboratory no.[a]	Reference	Material and context	^{14}C age (yr BP)	Calibrated age[b] (ka)	Latitude	Longitude	Elevation (m asl)
Beta-220940	Clague, unpublished	*In situ* stump	3760 ± 60	4.38–3.93			
Beta-220936	Clague, unpublished	*In situ* stump	3690 ± 50	4.22–3.89			
UCIAMS-40661	Clague, unpublished	*In situ* stump	2940 ± 20	3.21–3.00			
UCIAMS-40662	Clague, unpublished	*In situ* stump	2820 ± 20	2.97–2.86			
Beta-220939	Clague, unpublished	*In situ* stump	2710 ± 40	2.92–2.75			
Beta-220937	Clague, unpublished	*In situ* stump	2670 ± 50	2.87–2.73			
Beta-220938	Clague, unpublished	*In situ* stump	2520 ± 50	2.75–2.37			
UCIAMS-40664	Clague, unpublished	Wood mat in uppermost section of moraine	365 ± 20	0.50–0.32			
Bear River Glacier					56°06′N	129°39′W	
Beta-181857	Jackson et al., 2008	Log in moraine	3680 ± 60	4.22–3.84			
Beta-185808	Jackson et al., 2008	Log eroded (?) from moraine	3340 ± 60	3.72–3.41			
Surprise Glacier					56°12′N	129°36′W	
Beta-181858	Jackson et al., 2008	Log in moraine 125 m below crest	2960 ± 70	3.34–2.93			
Beta-197984	Jackson et al., 2008	Log in moraine 75 m below crest	1690 ± 60	1.73–1.42			
Beta-197986	Jackson et al., 2008	Branch in moraine 55 m below crest	1440 ± 60	1.51–1.27			
Todd Glacier					56°12′N	129°46′W	
Beta-181859	Jackson et al., 2008	Log in till	2300 ± 60	2.49–2.15			
Beta-199708	Jackson et al., 2008	Log in till	1690 ± 60	1.73–1.42			
Beta-181560	Jackson et al., 2008	Log in till	1540 ± 60	1.54–1.31			
Tide Lake[d]					56°20′N	130°05′W	
GSC-1372	Clague and Mathews, 1992	Wood in glaciolacustrine sediment	2730 ± 170	3.08–2.72			612
TO-2205	Clague and Mathews, 1992	Conifer needles in glaciolacustrine sediment	2700 ± 60	2.94–2.74			603
TO-2204	Clague and Mathews, 1992	Conifer needles in glaciolacustrine sediment	2650 ± 60	2.92–2.54			604
GSC-5349	Clague and Mathews, 1992	Log in glaciolacustrine sediment	1640 ± 60	1.61–1.42			616
TO-2898	Clague and Mathews, 1992	Conifer needles in glaciolacustrine sediment	1600 ± 40	1.57–1.39			603–606
GSC-5386	Clague and Mathews, 1992	Wood in glaciolacustrine sediment	1520 ± 50	1.52–1.34			595
TO-2897	Clague and Mathews, 1992	Wood in glaciolacustrine sediment	1440 ± 40	1.40–1.29			603
Canadian Rockies							
Peyto Glacier					51°41′N	116°32′W	
SRC-3117	Luckman et al., 1993	Detrital wood	3220 ± 80	3.64–3.26			
GSC-5157	Luckman et al., 1993	Detrital wood	3140 ± 70	3.45–3.26			
SRC-2950	Luckman et al., 1993	Detrital wood	3150 ± 75	3.56–3.17			
SRC-2949	Luckman et al., 1993	Detrital wood	2990 ± 70	3.36–2.97			
GSC-4680	Luckman et al., 1993	Detrital wood	2800 ± 80	3.00–2.79			
SRC-3119	Luckman et al., 1993	Detrital wood	2980 ± 70	3.35–2.96			
Beta-39934	Luckman et al., 1993	*In situ* stump	2929 ± 50	3.25–2.93			
GSC-4658	Luckman et al., 1993	Detrital wood	2880 ± 60	3.14–2.89			
GSC-4665	Luckman et al., 1993	Detrital wood	2870 ± 50	3.08–2.89			
GSC-4936	Luckman et al., 1993	Detrital wood	2840 ± 60	3.06–2.86			
SRC-3107	Luckman et al., 1993	Detrital wood	2490 ± 70	2.74–2.36			
Beta-35391	Luckman et al., 1993	Detrital wood	3300 ± 70	3.69–3.38			
Beta-33011	Luckman et al., 1993	Detrital wood	2980 ± 60	3.34–2.98			
Beta-48499	Luckman et al., 1993	Detrital wood	1710 ± 60	1.81–1.42			
Beta-38678	Luckman et al., 1993	Detrital wood	1550 ± 60	1.55–1.32			
SRC-2990	Luckman et al., 1993	Detrital wood	1140 ± 75	1.26–0.93			
SRC-3112	Luckman et al., 1993	Detrital wood	1110 ± 60	1.17–0.93			
Beta-48499	Luckman, 2006	Detrital wood	1710 ± 60	1.81–1.42			
Beta-38678	Luckman, 2006	*In situ* rootstock	1550 ± 60	1.55–1.32			
Beta-62063	Luckman, 2006	Detrital wood	1960 ± 80	2.11–1.72			
Beta-62066	Luckman, 2006	Wood in till	1500 ± 80	1.55–1.28			
Saskatchewan Glacier					52°09′N	117°09′W	
BGS-1369	Luckman et al., 1993	Detrital wood	3180 ± 80	3.61–3.21			
Beta-29957	Luckman et al., 1993	Detrital wood	2940 ± 80	3.34–2.88			
Beta-135588	Wood and Smith, 2004	Detrital wood	2910 ± 60	3.24–2.88			
Beta-31359	Luckman et al., 1993	Detrital wood	2880 ± 80	3.25–2.80			
Beta-135587	Wood and Smith, 2004	*In situ* stump	2870 ± 70	3.21–2.80			
Beta-135586	Wood and Smith, 2004	*In situ* stump	2830 ± 60	3.14–2.78			
Boundary Glacier					52°12′N	117°11′W	
WAT-1182	Gardner and Jones, 1985	*In situ* stump	4050 ± 70	4.82–4.41			
Beta-160362	Wood and Smith, 2004	*In situ* stump	3880 ± 40	4.42–4.16			
WAT-1183	Gardner and Jones, 1985	Peat below till	3880 ± 60	4.51–4.10			
Cavell Glacier					52°38′N	118°02′W	
Beta-74549	Luckman, 2006	Detrital wood	1680 ± 60	1.71–1.42			
Beta-64223	Luckman, 2006	Large detrital log in stream	1910 ± 80	2.04–1.63			
Stutfield Glacier					52°15″N	11721′W	
Beta-62421	Osborn et al., 2001	Wood in till	2380 ± 70	2.72–2.21			
Dome Glacier					52°12′N	117°15′W	
Beta-33007	Luckman et al., 1993	Detrital wood	6380 ± 80	7.46–7.16			
Beta-33008	Luckman et al., 1993	Detrital wood	6120 ± 60	7.17–6.80			
Yoho Glacier					51°35′N	116°32′W	
GSC-5118	Luckman et al., 1993	Detrital wood	2830 ± 80	3.07–2.85			
Athabasca Glacier					52°12′N	117°14′W	
Beta-17373	Luckman et al., 1993	Detrital wood	8230 ± 80	9.42–9.02			
Beta-20047	Luckman et al., 1993	Detrital wood	8000 ± 90	9.11–8.60			

Table 2 (*continued*)

Laboratory no.[a]	Reference	Material and context	^{14}C age (yr BP)	Calibrated age[b] (ka)	Latitude	Longitude	Elevation (m asl)
Beta-29957	Luckman et al., 1993	Detrital wood	7550 ± 100	8.55–8.17			
Beta-85918	Luckman, unpublished	Detrital wood	8170 ± 80	9.42–8.81			
Robson Glacier					53°08′N	119°06′W	
Beta-38309	Luckman et al., 1993	Detrital wood	3360 ± 60	3.82–3.45			
Beta-28439	Luckman et al., 1993	*In situ* stump	3300 ± 70	3.69–3.38			
Beta-33012	Luckman et al., 1993	*In situ* stump	3230 ± 70	3.64–3.28			
Beta-35010	Luckman et al., 1993	Detrital wood	3130 ± 70	3.55–3.08			
Beta-200484	Luckman, unpublished	Detrital wood in lake	3350 ± 80	3.83–3.40			
Beta-200485	Luckman, unpublished	Detrital wood in lake	3730 ± 70	4.35–3.88			
Beta-200486	Luckman, unpublished	Detrital wood	3550 ± 60	4.06–3.65			
Beta-84817	Luckman et al., 1996	Detrital wood	3160 ± 70	3.56–3.22			
Beta-62064	Luckman, 1995	Detrital wood from lake	3500 ± 60	3.96–3.63			
Beta-65381	Luckman, 1995	Root in paleosol, glacier toe site	3710 ± 70	4.28–3.85			
Beta-65382	Luckman, 1995	Large snag in gravels, glacier toe site	3650 ± 60	4.15–3.83			
Beta-187090	Luckman et al., 2005	Detrial wood, Extinguisher paleosol site	4780 ± 60	5.61–5.32			
Interior Ranges							
Castle Glacier					53°03′N	120°26′W	
UCIAMS-40542	Menounos, unpublished	Detrital log in forefield	4720 ± 20	5.58–5.33			
GSC-6709	Menounos et al., 2008	*In situ* stump in forefield	4210 ± 80	4.85–4.62			
UCIAMS-40543	Menounos et al., 2008	*In situ* stump in forefield	3720 ± 20	4.15–3.98			
GSC-6700	Menounos et al., 2008	Detrital log in forefield	3710 ± 80	4.22–3.93			
UCIAMS-40544	Menounos et al., 2008	Detrital log in forefield	3690 ± 20	4.09–3.93			
Haworth Glacier					51°42′N	117°54′W	
GSC-6772	Menounos et al., 2008	*In situ* stump in forefield	3870 ± 60	4.41–4.16			

[a] Radiocarbon laboratory: Beta = Beta Analytic Inc; GSC = Geological Survey of Canada; UCIAMS = University of California; Wat = University of Waterloo.
[b] Calendar ages (±2σ) determined using CALIB 5.02 (Stuiver et al., 2005).
[c] Analytical uncertainty is 2σ for GSC radiocarbon ages and 1σ for all other ages.
[d] Ice-dammed Tide Lake existed when Frank Mackie Glacier was more extensive than at present.

(Fig. 8, Table 2). A piece of wood in a peat layer between two tills in the north lateral moraine returned a radiocarbon age of 5010 ± 40 ^{14}C yr BP [5.89–5.65 ka]. Detrital wood within sediments of a former subaqueous ice-contact delta at Fyles Glacier, 120 km northwest of Tiedemann Glacier, yielded radiocarbon ages between ca 5980 and 4860 ^{14}C yr BP [6.80–5.60 ka] (Laxton et al., 2003). The wood was interpreted to be the remains of a forest overrun by Fyles Glacier during the Garibaldi Phase.

In southwest Yukon, Farnell et al. (2004) radiocarbon-dated caribou dung and plant matter at three of many known Holocene

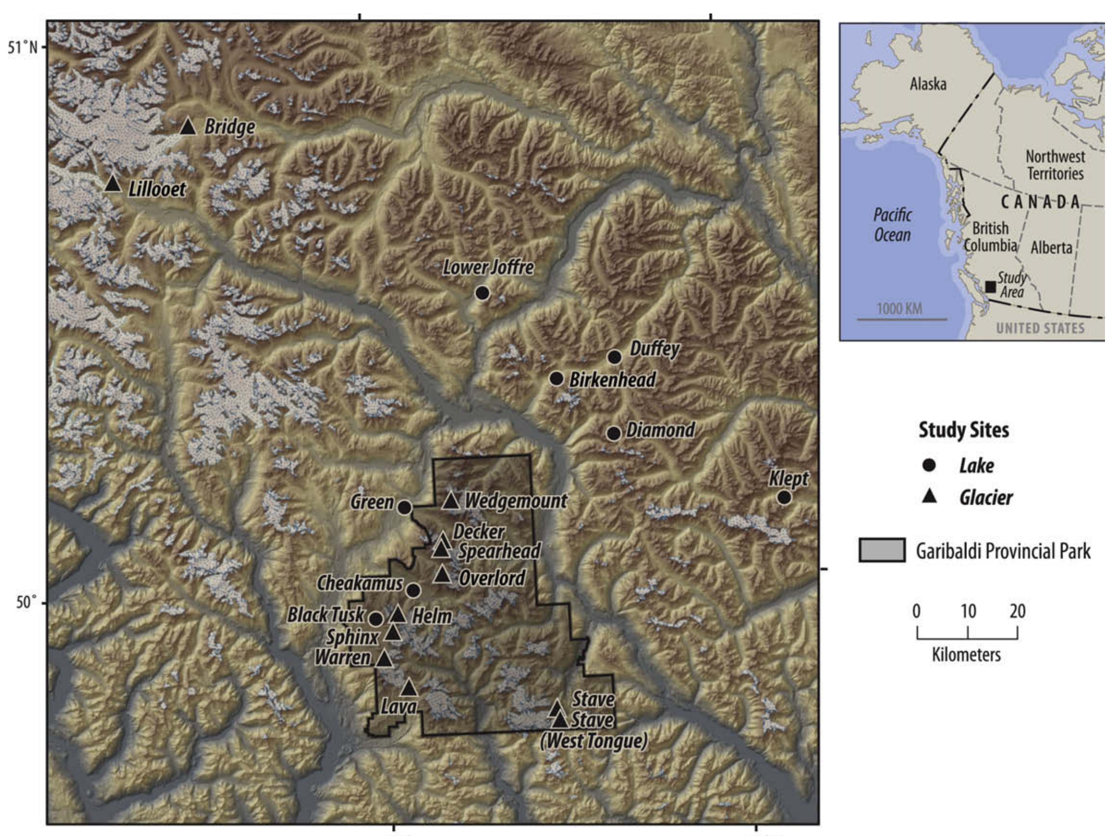

Fig. 6. Relief map of the southern Coast Mountains, showing localities mentioned in the text and current glacier extent.

Fig. 7. Rooted stump in the forefield of Warren Glacier, Garibaldi Provincial Park. The stump was overrun by the glacier 6360 ± 80 ^{14}C yr BP [7.47–7.03 ka]. Five other stumps were found near the dated stump.

"ice patches", masses of accumulated ice too small to flow. One patch formed between 8.3 and 8.0 ka, another at ca 7.4 ka, and the third at ca 4.5 ka. Ice patch formation was nearly continuous after ca 8 ka except for the interval 6.7–4.7 ka. The authors suggest that the apparent lack of net ice accumulation in that interval was caused by relatively high temperature or reduced precipitation or both, a conclusion at odds with the evidence for glacier expansion during part of that interval. But the authors note that the record may have been affected by melting, and concede that "mid-Holocene ice may yet be dated" (Farnell et al., 2004, p. 251).

Clastic sedimentation increased in lakes in the Coast Mountains during the early Neoglacial interval. A clastic unit in sediment cores recovered from Green Lake in the southern Coast Mountains has an age range of 7.03–6.62 ka (Osborn et al., 2007a), coincident with evidence from nearby glacier forefields for an advance of glaciers in nearby Garibaldi Park (Table 2). A clastic-rich unit was deposited in nearby Black Tusk Lake between 5.90 and 4.90 ka (Cashman et al., 2002), and a prominent clastic unit dating to about 6.50 ka is present in cores taken from lower Joffre Lake (Filippelli et al., 2006). Souch (1994) found a peak in clastic sedimentation between 6.95 and 5.62 ka in cores from Klept Lake, ca 100 km northeast of Garibaldi Park (Fig. 6).

4.1.2. Interior ranges and Rocky Mountains

Evidence for early Neoglacial glacier activity in the ranges of the British Columbia interior and the Rocky Mountains are limited to

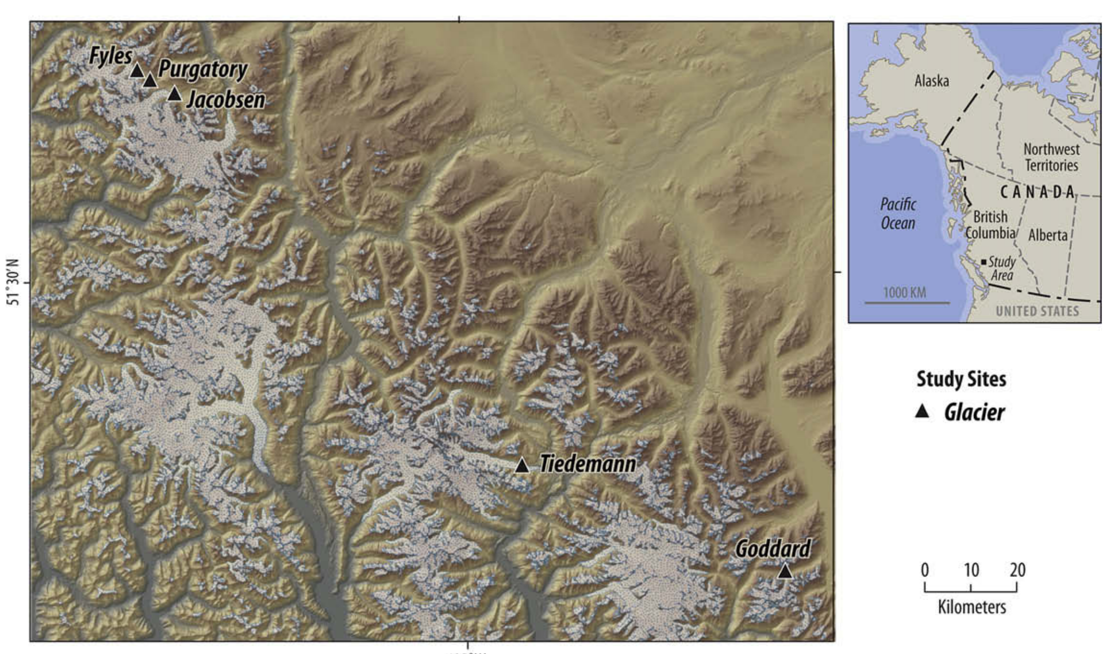

Fig. 8. Relief map of the Coast Mountains near Tiedemann Glacier, showing localities mentioned in text and current glacier extent.

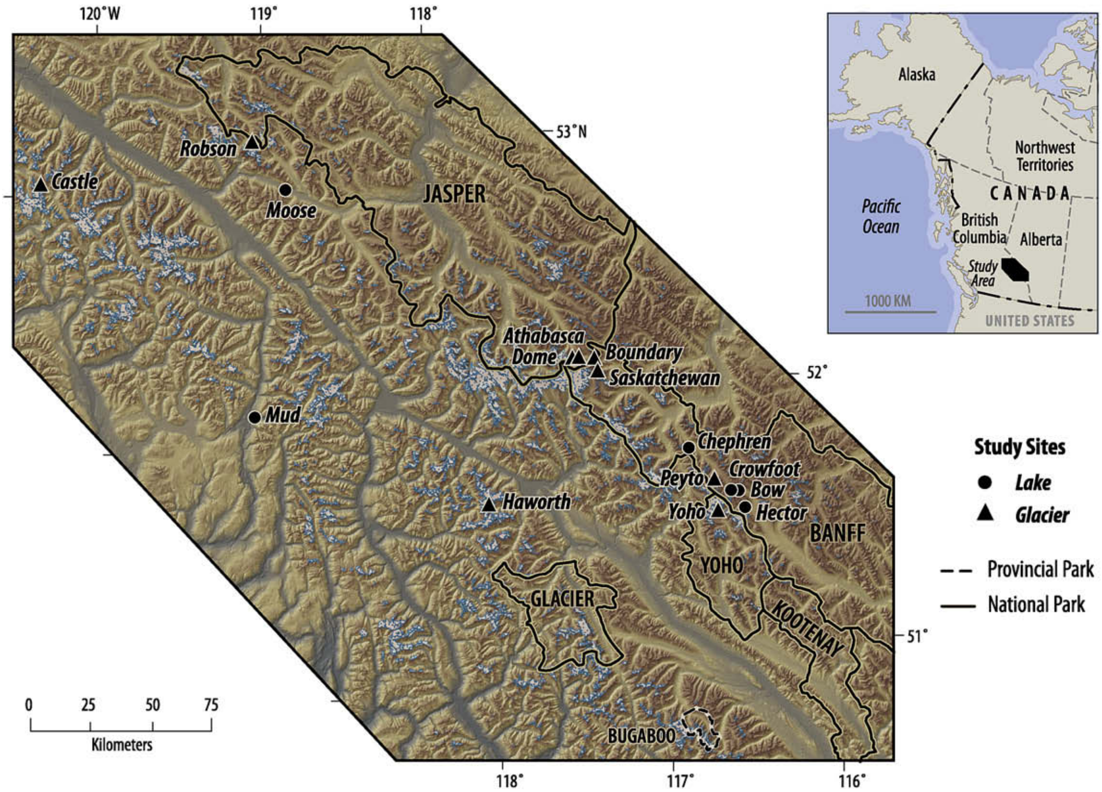

Fig. 9. Relief map of part of the Canadian Rockies and interior ranges, showing localities mentioned in text and current glacier extent.

detrital wood in glacier forefields (Fig. 9, Table 2). Radiocarbon ages of 6380 ± 80 and 6120 ± 60 ^{14}C yr BP [7.46–6.80 ka] were obtained on wood that washed out of the terminus of Dome Glacier (Luckman et al., 1993). These ages were originally interpreted as evidence of higher-than-present treeline, rather than trees that had been overridden by an advancing glacier, because no *in situ* wood of that age had been found in the Canadian Rockies (Luckman et al., 1993). However, the discovery of new sites with remnants of glacially overridden forests of this age suggests that it is more likely that the detrital wood at Dome Glacier also records glacier advance during the Garibaldi Phase.

A paleosol with *in situ* stumps as much as 500 years old and detrital logs up to 70 cm in diameter were found near the lateral margin of Robson Glacier in the Canadian Rockies (Luckman, 2007). One of the trees died about 4780 ± 60 ^{14}C yr BP [5.61–5.32 ka]. Detrital wood in till 10 m from the terminus of Castle Creek Glacier (unofficial name), located in the Cariboo Mountains about 90 km west of Robson Glacier, returned a radiocarbon age of 4720 ± 20 ^{14}C yr BP [5.58–5.33 ka]. These radiocarbon ages provide the first direct evidence of an early Neoglacial glacier advance in eastern British Columbia.

4.1.3. Summary

Evidence from sites throughout western Canada shows that alpine glaciers expanded between 6400 and 5000 ^{14}C yr BP [7.35–5.77 ka], a period loosely defined as the Garibaldi Phase. The lake sediment records and the range of radiocarbon ages suggest that this expansion was not slow and progressive, but rather episodic and separated by intervals of recession. The first Holocene advances were even earlier: Sphinx and Sentinel glaciers advanced ca 8.4 ka, Warren Glacier overrode a forest between 7.42 and 7.12 ka, and Dome Glacier advanced between 7.46 and 6.80 ka. Hence it appears that glaciers advanced as early as 8.4 ka in western Canada.

No evidence for Neoglacial activity before 3.50 ka has yet been found in the northern Coast and St. Elias mountains. Given the records from farther south, this absence of evidence is probably an indication of fewer sites visited rather than absence of glacier activity in these northerly locations prior to 3.50 ka.

4.2. Early–middle Neoglacial (5.00–3.50 ka)

Evidence for at least two glacier advances between 5.00 and 3.50 ka is provided by stumps in growth position, detrital branches and logs, peat layers below till near present glacier margins, and clastic intervals in proglacial lake sediments at several sites throughout western Canada.

4.2.1. Coast Mountains

Detrital wood near the 2003 snout of Helm Glacier (Fig. 6, Table 2) in Garibaldi Provincial Park (Koch et al., 2007a) returned a radiocarbon age of 4080 ± 40 ^{14}C yr BP [4.81–4.44 ka]. Detrital wood in till at nearby Sphinx Glacier yielded radiocarbon ages of 4280 ± 70 and 3560 ± 70 ^{14}C yr BP [5.04–4.59 and 4.08–3.64 ka], and two samples of detrital wood in the forefield of Spearhead Glacier, also in Garibaldi Park, gave ages of 3900 ± 60 and 3900 ± 80 ^{14}C yr BP [4.53–4.09 ka] (Osborn et al., 2007a). The stem of a glacially overridden tree in the forefield of Goddard Glacier, 150 km north–northwest of Spearhead Glacier (Menounos et al., 2008), provided an age of 4120 ± 60 ^{14}C yr BP [4.82–4.53 ka]. Stumps rooted in soils in outwash sand and silt in the north lateral moraine of Tiedemann Glacier (Fig. 10) returned radiocarbon ages of 3860 ± 20, 3820 ± 20, 3750 ± 60, and 3720 ± 50 ^{14}C yr BP [4.41–3.93 ka]. The outwash is overlain and underlain by till.

4.2.2. Interior ranges and Rocky Mountains

Stumps in growth position and detrial wood with preserved bark at Castle Glacier in the Cariboo Mountains (Table 2, Fig. 9) returned five radiocarbon ages between 4210 ± 80 and 3690 ± 20 ^{14}C yr BP [4.85–3.93 ka]. A rooted stump near the snout of Haworth Glacier in the Selkirk Mountains gave an age of 3870 ± 60 ^{14}C yr BP [4.41–

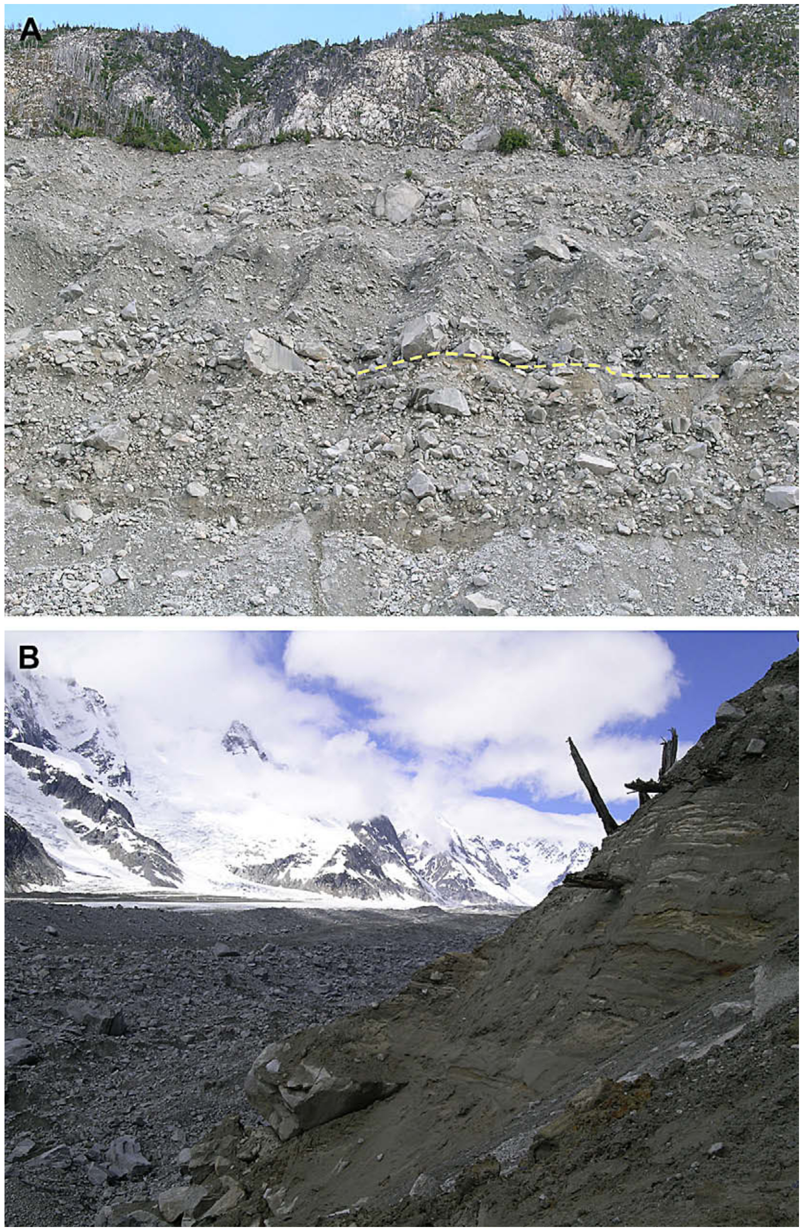

Fig. 10. (A) Multiple till and outwash units exposed in the north lateral moraine of Tiedemann Glacier. Dashed line denotes upper contact of the outwash unit containing Tiedemann-age [3.30–2.80 ka] stumps. (B) Stumps rooted in outwash sand and silt demarcated in A.

4.16 ka]. Two rooted stumps in the forefield of Boundary Glacier in the Canadian Rockies yielded ages of 4050 $\pm$ 70 and 3880 $\pm$ 40 ^{14}C yr BP [4.82–4.41 and 4.42–4.16 ka], and peat below till at the same site gave an age of 3880 $\pm$ 60 ^{14}C yr BP [4.51–4.10 ka] (Gardner and Jones, 1985; Wood and Smith, 2004).

4.2.3. Discussion

The 21 radiocarbon ages mentioned above cluster in two groups. Sixteen samples have a calibrated age range of 4.53–3.83 ka; eight of the 16 samples are outer rings of *in situ* stumps and one is a root (Table 1). Five samples have a calibrated age range of 5.04–4.44 ka; they too are derived from both detrital and *in situ* wood. The combination of rooted and detrital wood in the two groups implies that glaciers advanced at least twice during this time interval: an early advance at about 4.90 ka and another, more regional advance at about 4.2 ka (Menounos et al., 2008). Lake sediment records from the southern Coast Mountains include a clastic interval at 4.40–4.00 ka and a less prominent one at about 4.90 ka (Menounos et al., 2008).

4.2.4. Establishment of modern sedimentation rates

Cores recovered from several proglacial lakes in western Canada reveal an increase in clastic sedimentation during the period 5.00–3.00 ka. Sedimentation rates comparable to those of the past millennium were reached at Green Lake soon after 3200 ^{14}C yr BP [3.45–3.39 ka] and at nearby lower Joffre Lake by 3.50 ka (Fillipelli et al., 2006). Modern sedimentation rates were achieved at Mud Lake in the Monashee Mountains by 3.50 ka (Hodder et al., 2006) and at Moose Lake, 95 km to the north, between 4.10 and 3.10 ka (Desloges, 1999). It appears that glaciers reached positions comparable to those of the Little Ice Age by 3.50 ka. This period coincides with the establishment of modern pollen spectra throughout the Cordillera (Hebda, 1995) and likely reflects establishment of climatic conditions favorable for significant and widespread glacier expansion.

A similar transition is apparent in sediment cores from lakes in Banff National Park (Leonard and Reasoner, 1999), but it is earlier than in proglacial lakes in the Coast Mountains and interior ranges.

In Hector, Bow, and Crowfoot lakes the transition from low-clastic to high-clastic sediments occurred before ca 4.18, 4.99, and 4.40 ka, respectively. Reasons that could account for the time-transgressive nature of this facies change remain uncertain, but may include errors in dating or climatic differences between the coastal and interior ranges of the Canadian Cordillera.

4.3. Middle–late Neoglacial (3.50–1.00 ka)

In situ remains of forest floors in glacier forefields and within lateral moraines, and clastic sediment intervals in alpine lakes, indicate many glaciers achieved downvalley positions between 3.50 and 1.00 ka that were slightly less extensive than during the Little Ice Age. The evidence suggests that there were at least three advances during this period – an early advance centered at 3.00 ka, a second advance that probably culminated at 2.30 ka, and a third advance that occurred between 1.60 and 1.30 ka. The first two events occur within the Tiedemann Advance of Ryder and Thomson (1986) in the Coast Mountains and within the Peyto Advance of Luckman et al. (1993) in the Rockies. The third advance, during the period 1.60–1.30 ka, has been named the "First Millennium Advance" by Reyes et al. (2006a).

4.3.1. Coast and St. Elias Mountains

Glaciers in the southern Coast Mountains were more extensive during the period 3.50–1.00 ka than today. Evidence for advances at this time comes from Lava and Decker glaciers in Garibaldi Park (Koch et al., 2007a; Osborn et al., 2007a). A log in a wood mat in the west lateral moraine of Lava Glacier yielded a radiocarbon age of 3190 ± 40 ^{14}C yr BP [3.55–3.34 ka] (Koch et al., 2007a). Several snags and stumps are rooted on a bedrock cliff inside the Little Ice Age limit of Decker Glacier and 40–75 m above the 1998 ice margin (Osborn et al., 2007a). Outer rings from the stumps provided radiocarbon ages of 3200 ± 70, 2960 ± 40, and 2920 ± 50 ^{14}C yr BP [3.58–3.26, 3.32–2.97, and 3.22–2.99 ka], and a detrital log yielded an age of 2960 ± 50 ^{14}C yr BP [3.32–2.99 ka]. The snags are sheared in a down-glacier direction, suggesting that they were killed by an advancing glacier.

The oldest of the three stumps at Decker Glacier is 10 m upslope of the two stumps that gave ages of ca 2900 ^{14}C yr BP. The data suggest that the glacier thickened ca 3200 ^{14}C yr BP and killed the tree, then retreated enough to allow new trees to grow on the cliff below before advancing again ca 2900 ^{14}C yr BP. The stumps also indicate that Decker Glacier was at least 75 m thicker 3200 ^{14}C yr BP than in 1998, and almost as thick during a subsequent advance a few hundred years later.

In situ and detrital wood inside the Little Ice Age limits of Lillooet and Bridge glaciers yielded radiocarbon ages between 3000 and 1900 ^{14}C yr BP (Reyes and Clague, 2004; Allen and Smith, 2007). Tills within the north lateral moraine of Lillooet Glacier, about 60 m below the Little Ice Age moraine crest and 120 m above the present glacier surface, are separated by two paleosols (Reyes and Clague, 2004). Wood from the lower paleosol yielded ages of 3030 ± 40 and 2960 ± 60 ^{14}C yr BP [3.36–3.08 and 3.24–3.00 ka], and the outer rings of a log in the upper paleosol gave an age of 2490 ± 60 ^{14}C yr BP [2.72–2.37 ka]. A radiocarbon age from a subalpine fir stem collected from a nunatak below the Little Ice Age limit of nearby Bridge Glacier yielded an age of 2980 ± 60 ^{14}C yr BP [3.34–2.98 ka] (Allen and Smith, 2007). Four other stems from the nunatak were cross-dated to this sample using dendrochronological methods; all five stems died within a 20-year period. The data indicate that Bridge Glacier overrode a forest about 3.20 ka (Allen and Smith, 2007).

Ryder and Thomson (1986) document a major advance of Tiedemann Glacier beginning about 3.30 ka and culminating at 2.80 ka. They term this event the "Tiedemann Advance" and postulate that Tiedemann Glacier remained in an advanced position until 1.90 ka. New evidence, gathered over the past several years (Clague, unpublished data), suggests a more dynamic response of Tiedemann Glacier to mid-Neoglacial climate change than previously recognized. Dating of stratigraphically higher soils in this moraine indicate that the Tiedemann Glacier was 120–130 m thicker than today at a site 7 km upvalley of the present-day terminus between 3720 ± 50 and 2660 ± 50 ^{14}C yr BP [4.38–2.73 ka], and thickened even more, and thus advanced farther downvalley, shortly after 2520 ± 50 ^{14}C yr BP [2.75–2.37 ka]. The time of retreat from this maximum position is uncertain.

Sites in the central and northern Coast Mountains (Figs. 8 and 11) that record glacier expansion during the Tiedemann interval include Jacobsen and Purgatory glaciers (Desloges and Ryder, 1990) and Frankmackie Glacier (Clague and Mathews, 1992). The south lateral moraine of Jacobson Glacier contains a wood mat about 10 m below the Little Ice Age moraine crest. A log with bark from this wood layer gave an age of 2470 ± 50 ^{14}C yr BP [2.71–2.36 ka]. The proximity of the wood mat to the moraine crest shows that Jacobsen Glacier was close to its Little Ice Age limit about 2.50 ka. Radiocarbon ages at Tide Lake, a former glacier-dammed lake in the northern Coast Mountains, indicate that Frankmackie Glacier was more extensive ca 2.75 ka than in the late 20th century. Several other glaciers in the northern Coast Mountains (Fig. 11) advanced ca 3.00 ka, including Berendon, Bear River, Todd, and Forrest Kerr glaciers (Clague and Mathews, 1992; Haspel et al., 2005; Jackson and Smith, 2005; Laxton, 2005; Lewis and Smith, 2005; Jackson et al., 2008).

On the basis of a study of lake sediments at White Pass in the northernmost Coast Mountains, Lamoureux and Cockburn (2005) conclude that Neoglaciation did not commence in this area until about 2.00 ka, much later than in the Coast Mountains farther south (studies listed above), the Cassiar Mountains to the east (Lakeman et al., 2008), and the St. Elias Mountains to the northwest (see below). The reason for the difference remains uncertain.

Valley glaciers in the St. Elias Mountains in eastern Alaska and southwest Yukon expanded to near Little Ice Age limits between 3.30 and 2.40 ka (Denton and Karlén, 1973). The advance of debris-covered glaciers was slightly more extensive than the subsequent Little Ice Age advance; the opposite was true in the case of clean glaciers. Denton and Karlén (1977) infer that glaciers in the vicinity of the White River valley, near the Alaska–Yukon border, expanded in the interval ca 3.0–2.1 ka, with a culmination at 2.9–2.7 ka. Glacier expansion in the St. Elias Range is coincident with the Tiedemann Advance in the southern Coast Mountains and the Peyto Advance in the Canadian Rockies. In their study of southwest Yukon ice patches, Farnell et al. (2004, p. 251) note that ice "accumulation over the last 2000 years is especially noticeable and may broadly correlate to the Neoglaciation."

Proglacial lake sediment records from the southern Coast Mountains confirm the extensive nature of glaciers during the Tiedemann Advance. They indicate that glaciers remained more extensive than today from 3.30 ka to AD1945. Clastic varves began to form in Green Lake ca 3230 ^{14}C yr BP [3.63–3.36 ka]; mineral flux to the lake more than doubled at that time and remained high thereafter (Osborn et al., 2007a). An abrupt increase in clastic sedimentation has also been documented ca 3.00 ka at Diamond Lake (informal name; Minkus, 2006), and sedimentation rates increased in Kokwaskey and Kwoiek lakes at 2.40 ka (Souch, 1994). Clastic sedimentation rates in Duffey Lake ca 2.40 ka were comparable to those of the Little Ice Age (Menounos, 2002).

Numerous intervals of clastic-rich sediment occur within the Tiedemann-age sedimentary sequence in lakes cores, confirming the complex behavior of glaciers during the Tiedemann Advance. Two clastic-rich intervals in the Green Lake sediment record coincide with the advances of Decker Glacier between 3.58 and 2.99 ka

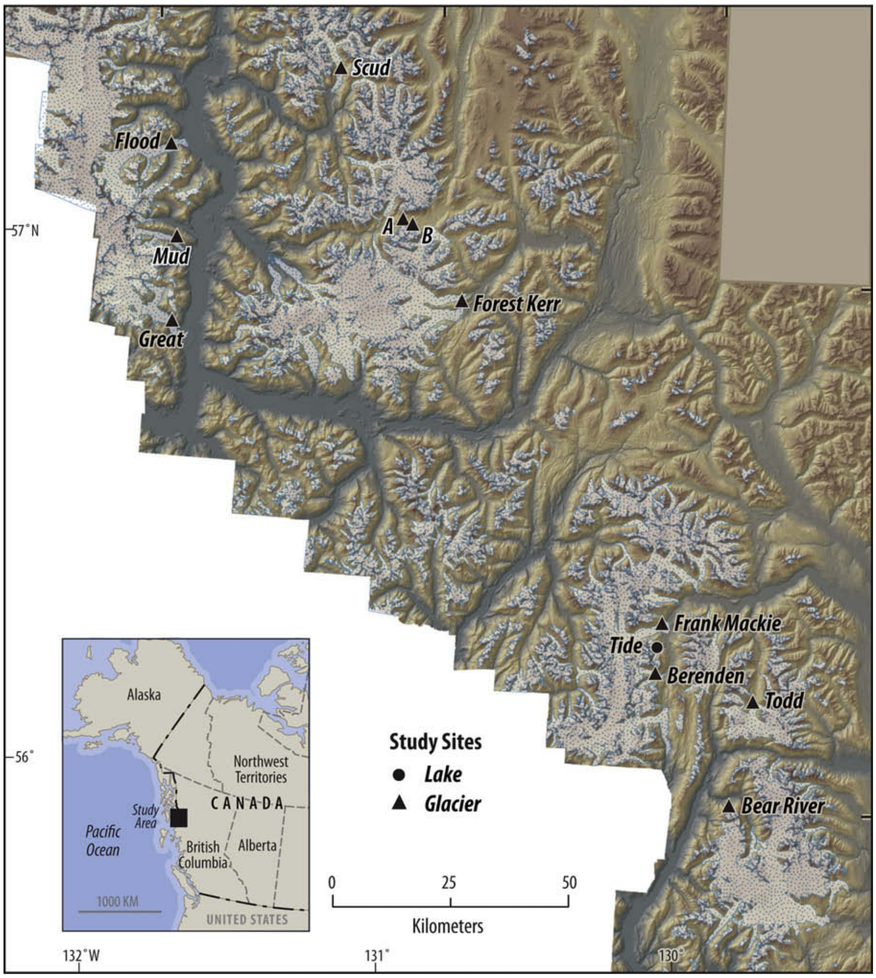

Fig. 11. Relief map of part of the northern Coast Mountains, showing localities mentioned in text and current glacier extent.

(Osborn et al., 2007a). An interval of clastic-rich sediment in Black Tusk Lake records a glacier advance at about 2.75 ka (Cashman et al., 2002), and a clastic-rich unit in nearby lower Joffre Lake dates to 3.20–3.00 ka.

Glaciers in coastal ranges of western Canada, Washington, and Alaska advanced between 1.60 and 1.30 ka during the First Millennium Advance (Reyes et al., 2006a). Evidence supporting an advance of this age includes detrital wood and stumps in growth position at Sphinx, Bridge, Deming, Lillooet, Miserable, Tiedemann, Frankmackie, Todd, Surprise, and Forrest Kerr glaciers (Reyes et al., 2006a; Osborn et al., 2007b). Glaciers began advancing downvalley as early as 1.80 ka, but most radiocarbon ages that delimit the advance fall between 1.60 and 1.30 ka (Reyes et al., 2006a). Slightly younger ages, 1.25–1.05 ka, have been reported for an advance of glaciers in the St. Elias Range (Denton and Karlén, 1973).

In their Yukon ice patch studies, Farnell et al. (2004) speculate that ice did not accumulate in the interval 1.44–1.03 ka (AD510–920), an interval they claim falls within the Medieval Warm Period. Such a claim is questionable, considering that the Medieval Warm Period is normally considered to be between the 9th and 14th centuries AD (Hughes and Diaz, 1994).

Lake sediment records indicate one or more glacier advances during the period 1.85–1.00 ka (Menounos, 2002; Fillipelli et al., 2006; Osborn et al., 2007a). Sedimentation rates in Green Lake were high between 1860 $\pm$ 50 [1.92–1.63 ka] and 1300 $\pm$ 45 ^{14}C yr BP [1.30–1.09 ka]. Especially thick varves date to 1.70–1.55 and 1.30–1.20 ka. A prominent clastic unit is centered at 1.50 ka in nearby lower Joffre and Duffey lakes.

4.3.2. *Interior ranges and Rocky mountains*

Bugaboo, Peyto, Robson, Saskatchewan, Stutfield, and Yoho glaciers advanced during the period 3.50–1.85 ka (Osborn, 1986; Luckman et al., 1993; Luckman, 1995, 2006; Osborn et al., 2001; Wood and Smith, 2004). Most of the relevant radiocarbon ages are older than 2800 ^{14}C yr BP [3.07–2.78 ka] and were obtained from detrital wood and stumps in growth position in the forefields of Peyto and Saskatchewan glaciers (Fig. 9). Thirteen ages from Peyto Glacier alone range from 3300 $\pm$ 70 to 2840 $\pm$ 70 ^{14}C yr BP [3.69–3.60 and 3.20–2.78 ka]. Two stumps in growth position near the late 20th century terminus of Saskatchewan Glacier returned ages of 2830 $\pm$ 60 and 2870 $\pm$ 60 ^{14}C yr BP [3.21–2.86 ka]. Wood and Smith (2004) cross-dated 36 stumps and detrital logs with the radiocarbon-dated stumps from this site and showed that all samples died between 2940 $\pm$ 60 and 2760 $\pm$ 60 ^{14}C yr BP [3.32–2.78 ka].

A log 1 km downvalley from the 1989 terminus of Yoho Glacier, 10 km south of Peyto Glacier, yielded a radiocarbon age of 2830 $\pm$ 80 ^{14}C yr BP [3.20–2.77 ka] (Luckman et al., 1993). A stump in growth position and two pieces of detrital wood within the Little Ice Age limit of Robson Glacier (Luckman et al., 1993) gave ages between 3300 $\pm$ 70 and 3130 $\pm$ 70 ^{14}C yr BP [3.83–3.09 ka]. A paleosol with *in situ* and detrital tree remains, exposed at two downvalley sites near the position of the 1993 glacier terminus, yielded ages ranging from 3730 $\pm$ 70 to 3350 $\pm$ 80 ^{14}C yr BP [4.35–3.88 and 3.83–3.40 ka].

Several sites in the interior and eastern ranges record glacier advances after 2.50 ka, but none of these sites has wood in growth position. A branch within outwash, about 80 m from the 1989 terminus position of Peyto Glacier, gave an age of 2490 ± 70 ^{14}C yr BP [2.74–2.36 ka] (Luckman et al., 1993). A log in till exposed in the north lateral moraine of Stutfield Glacier, 20 km northwest of Saskatchewan Glacier, yielded an age of 2380 ± 70 ^{14}C yr BP [2.72–2.21 ka]. The till is capped by a paleosol containing Bridge River tephra [2.46–2.35 ka]. Bugaboo Glacier generally advanced through the period 2500–1900 ^{14}C yr BP [2.71–1.83 ka] and then retreated (Osborn, 1986; Osborn and Karlstrom, 1989).

Lacustrine sediment records from the central and eastern ranges support the extensive nature of glaciers during the period 3.50–1.85 ka. Clastic sedimentation rates were high, although variable on a centennial time scale. Leonard and Reasoner (1999) identify two periods of increased clastic sedimentation in lakes in Banff National Park, one centered at 3.00 ka and another at 1.80 ka. Sedimentation in Hector Lake, however, was high throughout the period 3.00–1.80 ka (Leonard, 1997). The midpoints of intervals of notably thick varves at Hector Lake date to 3.35, 3.20, 2.90, 2.40, and 1.80 ka. Dirszowsky and Desloges (1997) associate rhythmite units dating to ca 3.46 and 2.33 ka in Chephren Lake sediment cores to times of maximum glacier extent. Thick varves in Moose Lake sediment cores date to ca 3.10, 2.50, and 2.40–2.30 ka (Desloges, 1999).

Some evidence has been found in the Rocky Mountains and interior ranges for the First Millennium Advance. A log and detrital stump associated with a paleosol in the lateral moraine of Peyto Glacier (Luckman, 2006) returned ages of 1710 ± 60 and 1550 ± 60 ^{14}C yr BP [1.81–1.42 and 1.55–1.32 ka]. The innermost 27 rings of a 180-year-old log washed out of Cavell Glacier in Jasper National Park yielded an age of 1910 ± 80 ^{14}C yr BP [2.04–1.63 ka], and an associated branch returned an age of 1660 ± 60 ^{14}C yr BP [1.70–1.41 ka] (Luckman et al., 1995; Luckman, 1999). Deposition of inorganic sediments in Chephren Lake, thought to immediately postdate an unnamed glacier advance, began about 1.47 ka (Dirszowsky and Desloges, 1997).

4.3.3. Discussion and summary

Glaciers in western Canada expanded during the period 3.50–1.90 ka. Ryder and Thomson (1986) argue that glaciers advanced several times during this period, each time reaching positions slightly less extensive than those of the climactic advances of the Little Ice Age. The evidence obtained since 1986 supports this assertion.

Numerous advances and retreats characterize the Tiedemann interval. Probably not all have left evidence at any one site, thus the records are not necessarily consistent in detail. For example, Bugaboo Glacier was retreating at 3070 ± 120 ^{14}C yr BP [3.56–2.95 ka], during the Tiedemann Advance (Osborn and Karlstom, 1989). Unconformities in lateral moraines at Jacobsen, Lillooet, and Stutfield glaciers that date to ca 2.50 ka suggest that these glaciers advanced at least twice during the period 3.50–1.90 ka. Lake sediment records, locations of dated stumps in growth position at Decker Glacier, and detrital wood below a till at Bridge Glacier provide additional support for multiple advances during the Tiedemann interval. Therefore, the use of the singular term ‘advance’ to describe a 1600-year interval when glaciers advanced repeatedly is problematic. The age range used for the Peyto Advance (3.50–2.90 and ca 2.55 ka) similarly highlights this problem. The limitations in the current use of nomenclature applied to glacier fluctuations in western Canada are discussed by Clague et al. (2009).

The First Millennium Advance, first documented by Reyes et al. (2006a), is strongly supported by abundant data from the coastal mountains of western North America. However, the large range of ages, between 1.80 and 1.30 ka, suggests that this advance may have complexity similar to that of the Tiedemann and Peyto advances.

4.4. Latest Neoglacial (1000 cal yr BP to present)

The history of glacier fluctuations in the past millennium has been the focus of considerable research in the past two decades and is known in considerable detail. The evidence for these fluctuations is of two types, morphological and stratigraphic. Moraines and related landforms have been dated from documentary sources or by dendrogeomorphic or lichenometric techniques, usually with accuracies of about 20 years within the past 200–300 years, in some cases with annual precision. In contrast, less extensive glacier advances prior to the last few centuries have been reconstructed mainly from moraine stratigraphy or from remnants of forests overridden and exposed during recent glacier recession. The fossil organic materials have been dated by radiocarbon techniques or, in some cases, by dendrochronology. Such evidence, however, only indicates times when glaciers were advancing or retreating. Furthermore, the dating control for the event is less precise because of two sources of uncertainty. First, the uncertainties of radiocarbon ages relative to their absolute age of the last millennium are high (Luckman et al., 1985). Several studies have shown incompatibility of up to several hundred years between radiocarbon ages and calendar ages determined by dendrochronology on the same material (Luckman et al., 1985; Koch et al., 2007b; Masiokas, 2008). Second, there are uncertainties associated with the position of the dated materials in time and space. Although radiocarbon and tree-ring ages indicate times of glacier advance, and in some cases the minimum length of an ice-free period prior to the advance, they demarcate the actual position of the glacier only if the dated materials are in growth position. This limitation is of greatest concern when trying to determine whether two deposits of similar age represent two separate glacier advances or different stages of the same event. It is particularly problematic in discriminating events of the past millennium, but also affects the interpretation of earlier Holocene glacier advances. Accordingly, we differentiate between minimum limiting ages of lateral or end moraines provided by lichen or trees that colonize the surface following abandonment of the moraine and maximum-limiting ages provided by sheared stumps, detrital wood, and tephra in paleosols.

4.4.1. Definition of the Little Ice Age

The Little Ice Age (LIA) includes the glacier advances of the past millennium, more specifically those beginning during the 12th and 13th centuries, achieving maximum extents in the 17th–19th centuries, and ending near the beginning of the 20th century. The events are broadly synchronous across the Cordillera of the Americas from Alaska through Canada to Patagonia (Luckman and Villalba, 2000). The Canadian data are reviewed on a regional basis below.

4.4.2. Vancouver Island

Investigations have been carried out at three small glaciers on Vancouver Island (Smith and Laroque, 1996; Lewis and Smith, 2004a). Based on dendrochronological dating, Moving, Colonel Foster, and Septimus glaciers constructed their outermost moraines between AD1699 and 1718 and built moraines in the 19th century (Lewis and Smith, 2004b).

4.4.3. Northern Coast Mountains and the Yukon

Relatively little work has been carried out in northern British Columbia. Clague and Mathews (1992) dated the outer moraine of Berendon Glacier to the 1660s. Radiocarbon ages on wood recovered from lacustrine deposits of Tide Lake indicate that Frankmackie Glacier advanced and impounded the lake in the 1400s. Ryder (1987) used reports of early travelers and field observations to identify moraines dating to the early to mid-1700s and early to mid-1800s at both Flood and Mud glaciers. She dated moraines of similar age at Great Glacier using dendrochronology and obtained radiocarbon ages on overridden wood that indicated that Scud Glacier and Glacier "B" advanced in the mid-15th century.

Recent work at Todd Icefield in the northern Coast Mountains indicates a complex LIA history (Jackson et al., 2008). Three advances have been identified and dated to ca AD1200–1420, 1400–1630, and 1640–1950 based on radiocarbon dating of *in situ* stumps and log mats within lateral moraines.

Few studies have been conducted on LIA glacier advances in southwest Yukon. Denton and Karlen (1973) concluded that valleys glaciers on the northeast flank of the St. Elias Mountains were more restricted in the period AD900–1490 and expanded in the period AD1490–1920. Denton and Stuiver (1966) date construction of the LIA terminal moraine of the Donjek Glacier to ~290 ^{14}C yr BP (~ AD1580). Several beaches of Lake Alsek, which was dammed by advances of Lowell Glacier, have been dated by lichenometry and dendrochronology. The youngest lake dates to the mid-19th century (Clague and Rampton, 1982; Clague et al., 1991). Trees tilted and killed by Kaskawulsh Glacier as it approached its LIA limit have been dated by dendrochronology to between AD1671 and 1757; the latter is probably close to the time of the maximum LIA advance of this glacier (Reyes et al., 2006b). The advance of Kaskawulsh Glacier across the Slims-Kaskawulsh valley induced a rapid rise in the level of Kluane Lake in the 1650s and its subsequent overflow to the north, establishing the present outlet of the lake (Clague et al., 2006).

4.4.4. Central Coast Mountains

Dendrochronological dating of the outer moraines of 16 glaciers in the Bella Coola area (Desloges, 1987; Desloges and Ryder, 1990) suggests that the LIA maximum in this region occurred during the 19th century. Most of the moraines were abandoned between AD1860 and 1900, and only part of the lateral moraine of Talchako Glacier is older than AD1850. Smith and Desloges (2000) developed a local lichen growth curve for *Rhizocarpon geographicum* that spans the past 165 years and used the curve to date unvegetated outer moraines at four glaciers in this area. Lichenometric ages of AD1699, 1715, and 1761 (at two glaciers) indicate that the outer-most moraines are older than previously thought. Radiocarbon-dated wood recovered from lateral moraines suggest that Purgatory Glacier advanced in the 13th century and that Purgatory and Jacobsen glaciers advanced in the 15th century.

4.4.5. Southern Coast Mountains

Ryder and Thomson (1986) report lateral moraine stratigraphy at several glaciers in the southern Coast Mountains. The stratigraphy provides evidence for an 11th century advance of Klinaklini Glacier, 12th century advances of Franklin and Bridge glaciers, and mid-15th century advances of Klinaklini, Sphinx, and Bridge glaciers.

Koch (2006) and Koch et al. (2007b) summarize the LIA history of nine glaciers in Garibaldi Park. All glaciers have well developed LIA terminal moraines that date to between the 1690s and 1720s. Only one of the glaciers (Overlord) has an undated moraine farther downvalley. Several, less extensive, 19th and early 20th century moraines, which occur inside the terminal moraines at all sites, were dated by dendrochronology. Koch et al. (2007b) also report lichenometric data for moraines at four sites and estimated moraine abandonment ages using a previously published lichen growth curve from Vancouver Island (Lewis and Smith, 2004a).

Four glacier forefields in Garibaldi Park contain abundant subfossil wood that was overridden during the expansion of glaciers in the first half of the past millennium. Several tree-ring series were cross-dated and floating chronologies developed, but most of the chronologies could not be tied to a long living-tree chronology to provide true calendar ages. Therefore periods of glacier advance are reported and correlated based on radiocarbon ages, with appropriate uncertainties. The general picture is one of advances ca AD1030–1170 at Warren Glacier, AD1150–1270 at Lava, Stave, and Helm glaciers, and AD1510–1600 at Stave Glacier. Helm, Lava, Sphinx, and Garibaldi glaciers advanced into mature forest in the 14th and 15th centuries, in some cases twice. Advances in the 16th and 17th centuries culminated in the construction of moraines dated to about AD1690–1720. Reyes and Clague (2004) describe lateral moraine stratigraphy at Lillooet Glacier, with tills of at least four LIA advances postdating 470 ± 50, 290 ± 60, 170 ± 60, and 10 ± 50 ^{14}C yr BP [AD1320–1955]. Tills of the last three advances are separated by paleosols or wood mats.

A different picture emerges from lichenometric studies in the Mount Waddington area and from Bridge Glacier (Larocque and Smith, 2003, 2004; Allen and Smith, 2007). At those localities, moraines were dated by lichenometry back to the 6th century. The local lichen growth curve, however, has only one control point older than 165 years, which is based on a radiocarbon age, and all lichenometric ages older than 680 years are extrapolated (Larocque and Smith, 2004). Larocque and Smith (2003) identify moraine building periods at ca AD925–933, 1203–1236, 1443–1458, 1506–1524, 1562–1575, 1597–1621, 1657–1660, 1767–1784, 1821–1837, 1871–1900, 1915–1928, and 1942–1946 from 14 glaciers, nearly all of which are based on lichenometric dating. However, in an appendix, they present, but do not discuss, tree-ring-derived ages for 57 of the 181 moraine segments they dated by lichenometry. Of the 37 moraines with lichenometric ages older than AD1700, one has a tree-ring age of AD1635 and the other 36 have tree-ring ages ranging from AD1696 to 1948. Unfortunately, there is no supporting evidence to corroborate the lichenometric ages for older moraines at the Mount Waddington sites.

Allen and Smith (2007) provide lichenometric ages of AD1367, 1409, 1649, 1756, 1831, 1856, 1912, and 1949 for moraines at Bridge Glacier using the Waddington-area growth curve. Ryder (1991) had previously assigned ages of AD1856, 1899, 1908, 1921, and 1935 to the outer five moraines based on reconnaissance tree-ring dating. Ryder and Thomson (1986) report five radiocarbon ages ranging from 530 ± 65 and 685 ± 60 ^{14}C yr BP [AD1230–1470] from remnants of an overridden forest on a nunatak rising above Bridge Glacier about 7 km upvalley from the LIA terminus. The advance of Bridge Glacier also ponded a lake in a tributary valley close to the terminus. The uppermost peat underlying the lake deposits yielded an age of 1115 ± 40 ^{14}C yr BP [1.17–0.93 ka], indicating that the glacier must have been close to the LIA limit at the time of lake formation. These data show that the glacier advanced during the 13th and 14th centuries and possibly during the 10th or 11th century.

Dendrochronology affords accurate ages of moraines back to the 18th century, but older, lichen-based ages commonly require extrapolation beyond data points and thus are less reliable. Although there is abundant evidence throughout western North America for glacier advances in the first half of the past millennium, few are more extensive than the climactic advances of the 18th and 19th centuries. Independent evidence is required from the Waddington sites to confirm the older lichen ages for the moraines there. Furthermore, a concerted effort is needed to acquire more control points beyond 150 years in age to verify and calibrate the Waddington lichen growth curve. More secure control points might be developed by obtaining dendrochronological and lichenometric ages from the same moraines.

4.4.6. Canadian Rockies

Sufficient data exist within the Canadian Rockies to reconstruct the regional history of glaciers for most of the past millennium, although gaps remain. Following the initial work of Heusser (1956), new data and revisions have been provided by Luckman and

Osborn (1979), Luckman (1986, 1996a, 2000), Osborn and Luckman (1988), Smith et al. (1995), and Osborn et al. (2001). Luckman (2000) provides a data set for moraines at 66 glaciers in the Rockies and Premier Range (Cariboo Mountains). Glacier fluctuations during the 20th century have been dated using aerial and ground photographs. Earlier activity has been dated through dendroglaciology (e.g. Luckman, 1998; Smith and Lewis, 2007) and lichenometry (Luckman, 1977; McCarthy and Smith, 1994, 1995). The following account is based on Luckman (1996a, 2000), where the record and its limitations are discussed in more detail.

Luckman (2000) groups the moraines in his data set in 25-year age classes (Fig. 12). Because multiple moraines occur at many sites, he includes the oldest preserved moraines at each site as a separate population. The LIA maximum event for each glacier is defined as the farthest downvalley extent of the glacier during the past few centuries.

The oldest moraines at each site range widely in age (Fig. 12) – 15 predate AD1700; 27 date to the 18th century; 21 date to the 19th century; and one is from the 20th century (three glaciers have undated outer moraines). At several sites, however, the oldest moraine fragments are only preserved along small sections of the former glacier margin, typically in lateral moraines, and therefore the date of the oldest moraine is not necessarily the date of maximum ice extent at a specific locality. Regional moraine construction occurred at AD1700–1725 and 1825–1850 (Fig. 12). Most of the mid-19th century moraines were dated from trees that were damaged or killed by ice. Where both 18th and 19thcentury moraines occur, they are commonly separated by less than 200 m. At the most northerly sites (Mount Robson area and Premier Range), 13 of 14 glaciers sampled have outermost moraines dating to the 18th century. Farther south, the majority of the glaciers sampled reached their maximum extent in the 19th century (most clearly seen in the Kananaskis area), suggesting that the 19th century event increases in relative extent southward. Only 19th century moraines have been identified near the International Boundary in Waterton and Glacier national parks (Osborn, 1985; Carrara, 1989). Based on the moraine evidence and the narrow range of dendroglaciological moraine dates, Luckman (1996a, 2000) argues that regional glacier extent was probably greatest in the 1840–1850s. This conclusion is partially based on an assumption that investigators target sites with multiple moraine sequences, ignoring relatively simple glacier forefields with few moraines, which are thereby underrepresented in this sample.

Detrital or *in situ* wood killed during glacier advances has been recovered from six glacier forefields, and their tree-ring series have been cross-dated to provide calendar ages for the advances (Luckman, 2000). Detrital surface logs indicate Saskatchewan Glacier was advancing into a forest in the periods AD1540–1664, 1710–1739, and 1798–1862; the LIA maximum was in 1862 (Robinson, 1998). Dendrochronological dates from overridden stumps and detrital wood document maximum LIA advances of other glaciers: Athabasca Glacier was advancing in AD1838 and reached its a maximum extent in AD1843–1844; Peyto Glacier advanced between AD1721 and 1836, and reached its maximum position about AD1845; and Manitoba Glacier advanced after AD1474 and in the periods AD1690–1698 and 1796–1808, with a maximum about AD1837 (Robinson, 1998). Dating of moraines older than AD1700 is mainly based on extrapolated lichenometric ages on small moraine fragments (Luckman and Osborn, 1979), many of which may be minimum ages. These data indicate locally more extensive, earlier LIA advances, but generally along only a small part of the glacier margin.

Four sites provide unequivocal evidence for earlier glacier advances that are not seen in the moraine record. Sheared stumps and/or paleosols at Robson (Heusser, 1956; Luckman, 1986), Kiwa (Watson, 1986), Peyto (Luckman et al., 1993), and Stutfield (Osborn et al., 2001) glaciers have yielded radiocarbon ages of ca 1200–600 ^{14}C yr BP [1.27–0.53 ka]. Dendrochronological dating (Reynolds, 1992; Luckman, 1995, 1996b; Robinson, 1998) show that Robson, Peyto, and Stutfield glaciers advanced into forest during the periods AD1150–1350, ca AD1246–1375, and after AD1272, respectively. Robson Glacier advanced ca 550 m at 2–3 m/year (Luckman, 1995). The similar ages and wide distribution of these sites are strong evidence for a regional period of glacier advance between AD1150 and 1350, during which glaciers reached to within 500 m of their LIA maximum positions.

Most glacier forefields contain a nested series of moraines between Little Ice Age maximum positions and present glacier fronts. Commonly, two or more moraines date from the 1880s and the first two decades of the 20th century. However, as the differences in age between these features are comparable to the uncertainties associated with the age estimates, it is difficult to correlate events between forefields. Most photographs from the early 1900s show glacier fronts close to their LIA maximum positions. Generally, glacier recession was most rapid during the first half of the 20th century (Gardner, 1972). Many glaciers advanced short distances during the period AD1950–1970 (Luckman et al., 1987). Rapid recession resumed in the 1970s and has continued to the present. Measurements at Peyto Glacier showed 6 years with a positive annual mass balance between 1965 and 1976, but with a slightly negative mean net balance of −0.134 m water equivalent for the period. In contrast, there were only two slightly positive years between 1977 and 2004 (mean −0.75 m water equivalent per year) (Demuth and Keller, 2006).

Hilda rock glacier in Banff National Park has been advancing continuously for the past 430 years and at a rate of ca 1.6 cm/year since the AD1790s (Bachrach et al., 2004). Adjacent Hilda Creek rock glacier, at the front of Hilda Glacier, has been advancing at ca 1.2 cm/year since the mid-19th century (Carter et al., 1999). Neither rock glacier appears to be responding to climate warming over this period.

No comprehensive glacier inventory is available for the Canadian Rockies, but Harding (1985) reports on 120 glaciers in the Premier Range and estimated a 25% loss of total ice cover between the LIA maximum and 1970. Mount (1995) determined similar losses in a smaller inventory of glaciers adjacent to the Columbia Icefield. McCarthy and Smith (1994) estimated an average reduction of 34% in the length of glaciers in the Kananaskis Valley between 1916 and 1988. Average equilibrium-line changes over these periods, estimated as half the increase in toe elevation between the two dates, are 131, 68, and 69 m for the Premier, Columbia Icefield, and Kananaskis regions, respectively. Recent photographs indicate that Columbia Glacier, a western outlet of the Columbia Icefield, receded about 1.5 km between 1992 and 2006 by calving into a proglacial lake.

4.4.7. *Proglacial lake sedimentation during the past millennium*

Proglacial lake sediments throughout western Canada clearly record glacier expansion during the past 1000 years. Given the dating uncertainties of age-depth models developed from radiocarbon ages over the past millennium, which are typically 10–20%, we focus most of the following discussion on varved lake sediment records.

Holocene clastic sedimentation rates in lakes in the southern Coast Mountains peaked during the past 1000 years. The highest Holocene rates in Black Tusk and Green lakes occurred after 1.00 ka (Cashman et al., 2002; Osborn et al., 2007a), and the highest rates in lower Joffre Lake, Diamond Lake, and lakes of the Kwoiek Creek watershed date to the last 0.75 ka (Souch, 1994; Fillipelli et al., 2006; Minkus, 2006).

Varved sediment records from Green and Cheakamus lakes reveal that, at the century time scale, maximum sedimentation coincides with times when glaciers were in expanded positions or

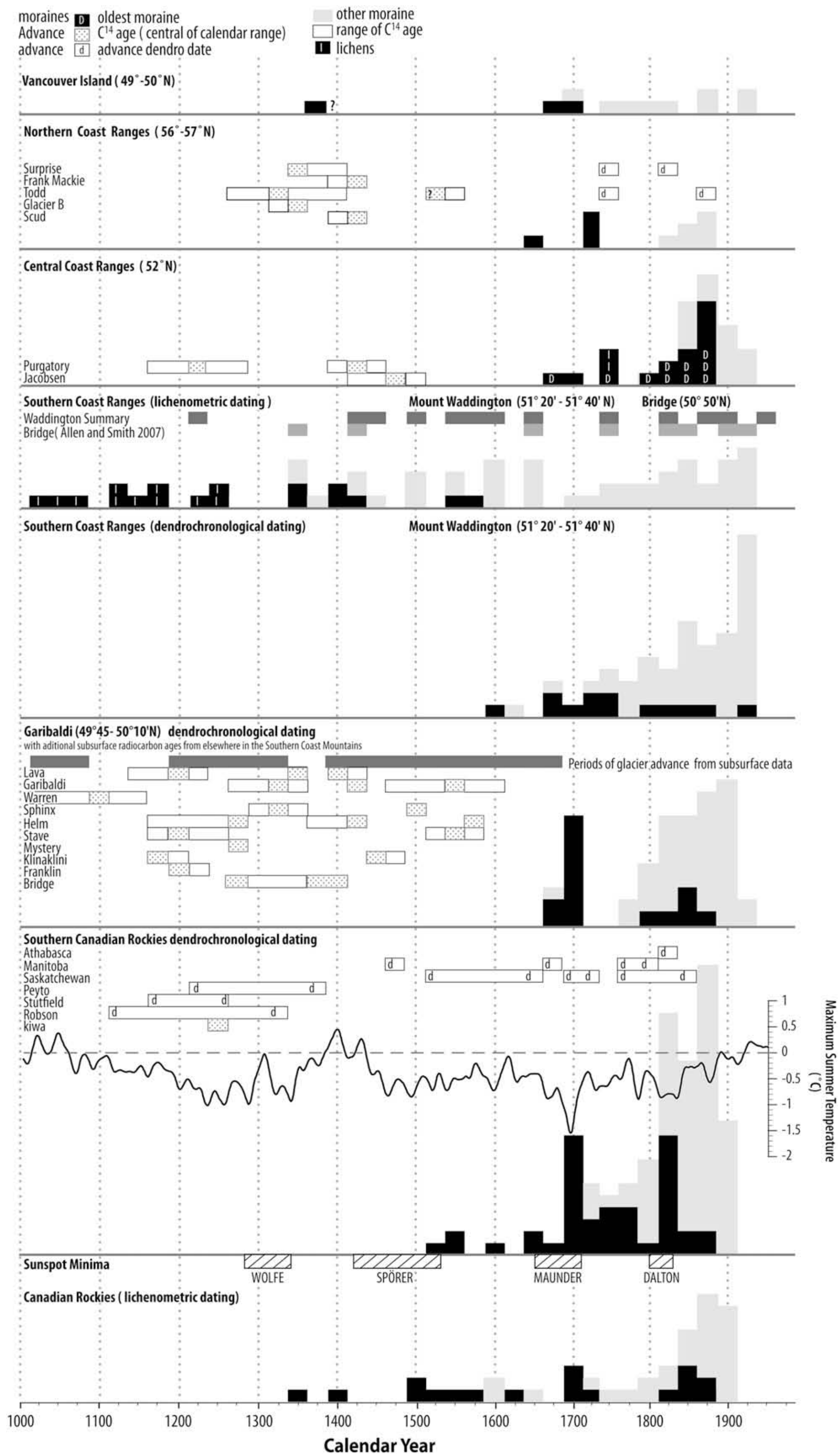

Fig. 12. Dated moraines and glacier advances during the past millennium in the Canadian Cordillera. Lichen and tree-ring ages on moraines are grouped in 25-year intervals. Black rectangles identify the oldest dated moraine at each site; gray rectangles represent moraines constructed during subsequent advances. Horizontal bars represent periods of glacier advance based on tree-ring or radiocarbon dating of subsurface material. The central age and range of ^{14}C ages are shown. The range of "kill" dates is indicated in instances in which a period of glacier advance is defined by multiple tree-ring "kill" dates. The May–August maximum temperature reconstruction at Athabasca Glacier (Luckman and Wilson, 2005), smoothed with a 20-year filter, is shown on the Canadian Rockies graph, as are the major periods of sunspot minima (Wiles et al., 2004). See text for data sources.

undergoing rapid retreat. Notably thick varves accumulated in both lakes during the early to middle 18th century and during the early to middle 20th century (Menounos, 2006; Menounos and Clague, 2008). The earlier period coincides with the climactic advances of the Little Ice Age in Garibaldi Park (Koch et al., 2007b), and the later period was a time of significant glacier retreat in the southern Coast Mountains and elsewhere in western Canada (Osborn and Luckman, 1988; Menounos et al., 2005).

At time scales shorter than a century, there is only moderate correspondence between the lake sediment and glacier records, and over periods of a decade or less, sediment delivery to lakes is poorly predicted by percent glacier cover. In Cheakamus Lake, the thickest varves date to the decades AD1090–1110, 1120–1170, 1210–1250, 1310–1330, 1390–1450, 1720–1780, 1860–1900, and 1920–1945 (Menounos and Clague, 2008). Thick varves accumulated in Green Lake during the periods AD1620–1665, 1670–1755, 1770–1785, and 1920–1945 (Osborn et al., 2007a). During the photographic period, the thickest varves coincide with times when glaciers were rapidly retreating.

Thick varves in Summit Lake in the northern Coast Mountains date to AD1450–1520 and 1680–1750 (Cockburn and Lamoureux, 2007). The latter interval coincides with the advance of Kaskawulsh Glacier to its maximum LIA extent (Reyes et al., 2006b) and to ages of end moraines of Flood, Mud, Scud, and Great glaciers to the south (Ryder, 1987).

Varved sediments from lakes in the Interior Ranges and Rocky Mountains show some correspondence with records of glacier activity, but as with varved sediments in the Coast Mountains, the agreement is poor at time scales shorter than 50–100 years (Leonard, 1997). Sedimentation rates are high in Hector Lake over the past 500 years, based on a 4450-year varve record from a single sediment core and a decadal-averaged composite developed from a number of shorter cores (Leonard, 1997). Thick varves accumulated in Hector Lake during the periods AD1200–1250, 1350–1400, 1550–1650, 1700–1750, 1830–1870, and 1920–1950.

Annually laminated sediments from Moose Lake reveal a complex pattern of sedimentation over the past millennium that differs substantially from that of Hector Lake (Desloges and Gilbert, 1995; Desloges, 1999). The catchment has only 3% glacier cover, thus the importance of glacigenic sediments to the lake is probably less than at Hector Lake, with 6% glacier cover in its watershed. Thick varves in cores from proximal locations in Moose Lake date to AD1450–1500, but this interval of thick varves is not apparent in sediment cores from the distal end of the lake. The discrepancy in varve thickness observed for distal and proximal sites of Moose Lake may originate from differences in the mechanisms by which sediment reaches the coring sites (Luckman, 2000). The century-scale variability in the composite Moose Lake record generally accords with the pattern of clastic sedimentation in Hector Lake (Luckman, 2000).

4.4.8. Summary

Geomorphic and stratigraphic evidence indicates that glaciers repeatedly advanced during the past millennium and that the advances were broadly synchronous throughout the southern Canadian Cordillera on a centennial time scale. Most glaciers were at or close to their Holocene limits during the middle to late 19th century; the maximum regional ice cover probably dates to between AD1830 and 1880. Glaciers advanced several times during the late 19th century and were still close to their maximum positions in the early 1900s. Ice cover decreased dramatically in the Rockies during the 20th century, mainly before 1960 and after 1975 (Luckman and Kavanagh, 2000; Moore et al., in press).

An earlier period of glacier advance that culminated in the early 1700s was locally more extensive, but regionally less extensive than the 19th century advance. Dating uncertainties make it difficult to evaluate synchroneity, but it appears that the advance occurred at about the same time across the region. The time of the maximum advance differs slightly from site to site depending on local factors. Although several moraines have been assigned dates prior to AD1700, the only convincing evidence for earlier regional glacier advances is from overridden forests in glacier forefields and paleosols and associated *in situ* and detrital wood in lateral moraines. There is limited evidence for a possible advance in the 11th century but evidence at many sites for a regional advance during the 13th and 14th centuries. Several sites also have evidence of glacier advances in the mid-15th, 16th, and 17th centuries, culminating in moraine construction in the late 1600s and early 1700s. At many sites these earlier advances reached downvalley to positions close to the LIA maximum. A similar pattern of glacier advances is also seen throughout Alaska and in Patagonia (Luckman and Villalba, 2000), suggesting common forcing of these events. Complementary paleoenvironmental reconstructions of proxy climate and mass balance data offer great promise for better elucidating the controls on glacier fluctuations during the past millennium and earlier in the Holocene.

5. Summary and discussion

5.1. Summary

Multi-proxy evidence of latest Pleistocene and Holocene glacier activity in western Canada reveals that glaciers responded dynamically to climate change over the past 15.0 ka. Several lobes of the Cordilleran ice sheet advanced about 13.5 and after 12.0 ka. Some regions record an advance of cirque or valley glaciers during the Younger Dryas; others record a relatively extensive advance that may correlate with the Younger Dryas or may be older. Both Younger Dryas and pre-Younger Dryas advances occurred in the Canadian Rockies.

Glaciers expanded, sometime episodically, through the Holocene and reached maximum downvalley positions during the Little Ice Age. Numerous glacier advances and retreats occurred on shorter time scales, often synchronously, throughout western Canada. The studies reported here show that glaciers advanced many times during the Garibaldi Phase and the Tiedemann and Peyto Advances. The continued use of these latter terms may impede an understanding of the complex, dynamic behavior of glaciers during these intervals (Clague et al., 2009).

5.2. Comparison of evidence from lake sediments and glacier forefields

In situ wood in glacier forefields provides the most direct and unequivocal evidence for glacier fluctuations, but the record is fragmentary. Carefully chosen detrital wood may date glacier advances, but other interpretations are possible. Proglacial lake sediments provide continuous records of upvalley glacier fluctuations, but these records may be complicated by other processes that affect sediment availability. The history of Holocene glacier fluctuations is best understood using a combination of the two approaches (Osborn et al., 2007a).

Comparison of terrestrial and lacustrine evidence of glacier fluctuations in western Canada reveals notable concordance between the records (Fig. 13). Peaks in the distribution of radiocarbon ages of wood from glacier forefields occur at 8.59–8.18, 7.36–6.45, 4.40–3.97, 3.54–2.77, 1.71–1.30, and 0.70 ka. These intervals are interpreted to be times when glaciers throughout western Canada were advancing into forests. The termination of an advance may not necessarily be constrained by the upper limit of the radiocarbon age range because glaciers may have remained in extended positions for centuries without killing additional trees. Peaks in the clastic content of the lake sediments typically occur one to several hundred years after peaks in the radiocarbon ages, although both begin to rise at about the time (Fig. 13).

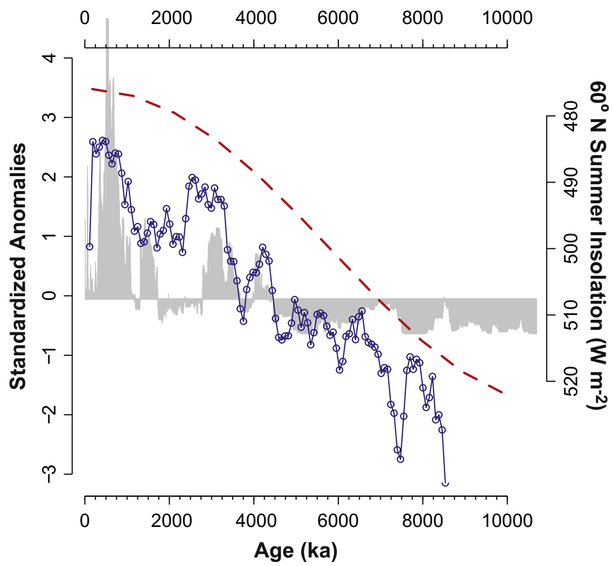

Fig. 13. Comparison of terrestrial and lacustrine evidence of Holocene glacier fluctuations in western Canada. The lacustrine record (blue line with open circles) represents the first principal component (PC1) of loss-on-ignition data, an index for the clastic content of lake sediments, for cores from Green, lower Joffre, Diamond, and Red Barrel lakes (Menounos et al., 2008). Control points for the linear age-depth models include AMS radiocarbon ages of terrestrial macrofossils and tephras. The curves were spline interpolated to a common sampling frequency of 75 years, which reflects the lowest temporal resolution of the sediment records. PC1 explains 68% of the variance common to the records. The filled gray curve represents the probabilities of 240 radiocarbon ages obtained from glacier forefields in western Canada (Table 2 and references in this paper). The dashed red line is the record of Holocene insolation at 60°N (Berger, 1978). The two datasets are standardized (zero mean and unit variance) to facilitate comparison of the records.

Significant differences between the lake and forefield records include the substantial lag between the distribution of radiocarbon ages in the early Holocene, the lack of a peak in radiocarbon ages at 5.00 ka, and high sedimentation rates at 2.50 ka without corresponding forefield wood ages. Reasons for the lag in the early Holocene may include dating imprecision and artificially high clastic influx into lakes due to reworking of Mazama tephra. We speculate that differences in the terrestrial and lacustrine records at 5.00 and 2.50 ka may be due to differences in the way the proxies record glacier activity. For example, one might find little wood dating to 6.00–5.00 ka if the rate of treeline depression at that time kept pace with the rate of glacier expansion. Glacier expansion, on the other hand, might be rapidly recorded via an increase in sedimentation in proglacial lakes. Similarly, if glaciers remained in extended positions at 2.50 ka, clastic sedimentation would be high but there might be little evidence in the form of overrun trees because most of the glacier expansion occurred earlier. With better dating control, these hypotheses may be tested in the future.

5.3. *Factors responsible for glacier fluctuations in western Canada*

The data presented in this paper demonstrate that glaciers in western Canada have fluctuated on a variety of time scales. This fact indicates that several different climate drivers have operated over the past 10.0 ka, each significant on a different time scale. The extents of glaciers at a given time can only be explained when all of these drivers are considered.

The progressive, episodic growth of alpine glaciers through the Holocene, previously noted in western Canada by Osborn and Luckman (1988) and elsewhere in western North America by Davis (1988), broadly coincides with the gradual reduction in summer insolation in the Northern Hemisphere through the Holocene (Berger, 1978). This concordance between long-term changes in glacier cover and summer insolation is revealed by comparing the clastic content of four proglacial lake sediment records reviewed in this paper and summer insolation at 60°N (Fig. 13).

Recent dendro-climatological research has yielded climate reconstructions for some of the primary variables that control glacier mass balance at time scales shorter than millennia. These proxy records provide a means of comparing the record of glacier fluctuations to climate variability and their external forcing mechanisms. Luckman and Wilson (2005) developed a maximum summer temperature reconstruction for the Columbia Icefield area for the period AD950–1994. Concordance is evident between sunspot minima, cool summer temperatures, and periods of glacier advance in the 12th and 13th centuries, the late 17th–early 18th century, and the mid-19th century (Fig. 12). One of the coldest intervals, in the early and mid-15th century, coincided with the Sporer sunspot minimum. This event is also reflected in sharply reduced growth in ring-width records for the major treeline species in the region (Luckman 1993, 1994), but there is little evidence for glacier expansion in the Rockies at this time. However, glaciers advanced in the northern Coast Mountains of British Columbia and in Alaska in the early and mid-15th century. Recent studies by Koch (2006), Koch et al. (2007b), and Wiles et al. (2008) also note the correspondence between glacier advances and solar minima during the past millennium. The correspondence suggests that variations in solar irradiance may play an equivalent role earlier in the Holocene, although a detailed examination of this possibility is beyond the scope of this paper.

Dendro-climatological records also afford a means to extend glacier mass balance records by several hundred years. Of sites where proxy mass balance records have been developed (e.g., Larocque and Smith, 2005; Watson et al., 2008), only Peyto Glacier incorporates a temperature proxy to estimate summer balance (Watson and Luckman, 2004). The Peyto record shows a strong temperature influence on 20th century glacier mass balance, and low-frequency links with sunspot minima.

Years of positive mass balance at Peyto Glacier also occur when winter precipitation is high. Winter balance of Peyto Glacier is correlated with the Pacific Decadal Oscillation (Mantua et al., 1997) and the El Niño-Southern Oscillation (ENSO), particularly when they are in phase (Watson et al., 2006). Continued development of such studies and longer reconstructions will provide additional insights into the climate drivers of glacier fluctuations in this region over the Holocene.

5.4. *Directions for further work*

Although much progress has been made since the last review of glacier fluctuations in western Canada 20 years ago, much important research remains to be done. Key tasks include: (1) obtaining better age control on the Finlay moraines; (2) applying cosmogenic surface exposure techniques to date moraines; (3) using multi-proxy methods in targeted catchments; and (4) greatly expanding research in the northern Canadian Cordillera.

The age of the Finlay moraines is important because, if they are of Younger Dryas age, they represent a different response to Younger Dryas cooling than the Crowfoot moraines; if they are not Younger Dryas in age, they record a significant, late Pleistocene climatic reversal that interrupted deglaciation. This issue can be addressed by radiocarbon dating basal sediments in lakes directly inside and outside the Finlay moraines and by cosmogenic surface exposure dating of boulders on the moraines.

Surface exposure dating has revolutionized studies of Pleistocene glaciation by dating the landforms produced during glacier advances. Until recently, exposure ages have had one-sigma uncertainties of ±20%, which effectively precluded the use of the methods for dating Holocene surfaces. However, improvements in sample preparation and detection limits of many isotopes yield exposure errors that are close to ±10% of the age range for late Holocene samples (Schaefer et al., 2007). This increase in precision may allow a refinement of the Little Ice Age moraine chronology where lichenometric ages are suspect or deviate significantly from minimum-limiting ages derived from trees that have colonized moraines.

This paper highlights the advantages of using a multi-proxy approach to understand the complex behavior or glaciers during the Holocene. Moraines, stratigraphy, and remains of forest overrun by glaciers provide the most direct and unequivocal evidence of glacier fluctuations prior to the 20th century. This evidence, however, is fragmentary and to some degree has biased our perception of glacier activity during the Holocene. Proglacial lake sediments offer the continuity that forefield evidence cannot, but sediment can be delivered to alpine lakes by non-glacial processes. Additional research should focus on using both types of records in tandem because each tells a different but complementary story about how glaciers fluctuated in the past.

Where possible, tree-ring chronologies exceeding 1000 years should be developed to provide a means of cross-dating wood collected in glacier forefields. As noted earlier, dendrochronology offers an opportunity to derive calendar year dates for the death of trees killed by an advancing glacier. Although the number of sites where this approach can be used is limited, the technique provides a means of narrowing age ranges of glacier advances that, prior to the past millennium, are based primarily on radiocarbon ages.

Our final proposition for future research is to increase the number of investigations in the northern Canadian Cordillera. Our understanding of Holocene glacier fluctuations is strongly biased to the south. Fruitful, future research should employ multi-proxy methods including surface exposure dating in glacierized catchments in the northern Rocky and Cassiar mountains and in the mountains of Yukon and Northwest Territories.

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