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Multi-proxy record of Holocene glacial history of the Spearhead and Fitzsimmons ranges, southern Coast Mountains, British Columbia

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Abstract

Evidence from glacier forefields and lakes is used to reconstruct Holocene glacier fluctuations in the Spearhead and Fitzsimmons ranges in southwest British Columbia. Radiocarbon ages on detrital wood and trees killed by advancing ice and changes in sediment delivery to downstream proglacial lakes indicate that glaciers expanded from minimum extents in the early Holocene to their maximum extents about two to three centuries ago during the Little Ice Age. The data indicate that glaciers advanced 8630–8020, 6950–6750, 3580–2990, and probably 4530–4090 cal yr BP, and repeatedly during the past millennium. Little Ice Age moraines dated using dendrochronology and lichenometry date to early in the 18th century and in the 1830s and 1890s. Limitations inherent in lacustrine and terrestrial-based methods of documenting Holocene glacier fluctuations are minimized by using the two records together.

1. Introduction

Glaciers are valuable indicators of environmental change. The general expansion of alpine glaciers in the Northern Hemisphere from the early Holocene to the present, however, has limited our ability to reconstruct the full record of advances and retreats. As glaciers advance, they override and commonly destroy direct evidence delimiting former ice margins, such as terminal and lateral moraines. These larger advances may also remove evidence used to constrain the ages of the advances, such as fossil trees and tephras. Moraines and overrun trees in glacier forefields are the most direct evidence for prehistoric glacier fluctuations, but the chronologies they afford are fragmentary and incomplete.

Lakes draining glacierized terrain provide opportunities to infer upvalley glacier activity (Karlén, 1981; Leonard, 1986; Souch, 1994; Leonard and Reasoner, 1999; Menounos, 2002). Increased clastic sedimentation in these lakes is commonly interpreted to reflect enhanced sediment production beneath an expanded body of ice. However, changes in sediment availability and storage beneath alpine glaciers or changes in sediment from non-glacial sources can complicate the relation between sediment yield and glacier extent (Hallet et al., 1996; Leonard, 1997).

Both terrestrial (e.g., Luckman et al., 1993; Osborn et al., 2001) and lacustrine (references above) evidence have been used in North America and elsewhere to reconstruct Holocene glacier fluctuations. Surprisingly, few studies use *both* approaches to infer past glacier fluctuations, although there are notable exceptions (Leonard, 1997; Luckman, 2000; Menounos et al., 2004; Bakke et al., 2005).

Retreat of glaciers in the southern Coast Mountains of British Columbia during the 20th century exposed rooted stumps and detrital wood in glacier forefields (Mathews, 1951), providing a rich and largely unexploited record of glacier fluctuations. Meltwater streams issuing from many of the glaciers flow into lakes, in which sediments spanning thousands of years afford a continuous, albeit indirect, proxy of glacier activity. This paper reconstructs Holocene

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glacier fluctuations in Spearhead and Fitzsimmons ranges of the southern Coast Mountains of British Columbia by combining both lake and terrestrial evidence. We show that the record provided by study of these complementary lines of evidence is richer than that obtained using either method alone.

2. Physical setting

Garibaldi Provincial Park in southwest British Columbia is a spectacular mountain landscape, geologically notable for the interaction between glaciers and Quaternary volcanic eruptions (Mathews, 1958). The Spearhead and Fitzsimmons ranges, which are within the park, are underlain by Cretaceous quartz diorite and granodiorite, and by older, intruded metamorphic rocks (Monger and Journeay, 1994). This part of Garibaldi Park ranges in elevation from 915 m at Cheakamus Lake to 2892 m at the summit of Wedge Mountain (Fig. 1). Over 630 km^2 of Garibaldi Park was covered by glaciers in the early 18th century, but ice cover today is 390 km^2, a loss of 38% in less than 300 years (Koch et al., 2003).

The climate of Garibaldi Park is controlled by Pacific air masses. There are strong gradients in precipitation, runoff, and winter snowpack from southwest to northeast. For example, average annual precipitation at Squamish, just outside the southwest corner of the park, is 2370 mm, but it declines to 1230 mm at Whistler, 60 km to the north. Precipitation increases with elevation due to orographic effects.

Green Lake (2 km^2), which is considered in this paper, has a watershed of 180 km^2 in the Fitzsimmons and Spearhead ranges (Fig. 1). Important sediment sources to Green Lake include contemporary glaciers (7% of catchment area), glacier forefields, and steep hillslopes adjacent to Fitzsimmons Creek. The creek has incised into a thick sequence of Pleistocene sediments. These incised deposits are an important source of sediment to the lake, although only during infrequent landslides (Menounos et al., 2006). Green Lake has two main basins. The proximal basin, the larger of the two, is 40 m deep; the distal basin is 30 m deep. Acoustic (3.5 kHz) surveys of the lake indicate that latest Pleistocene and Holocene sediments are 10 to 20 m thick.

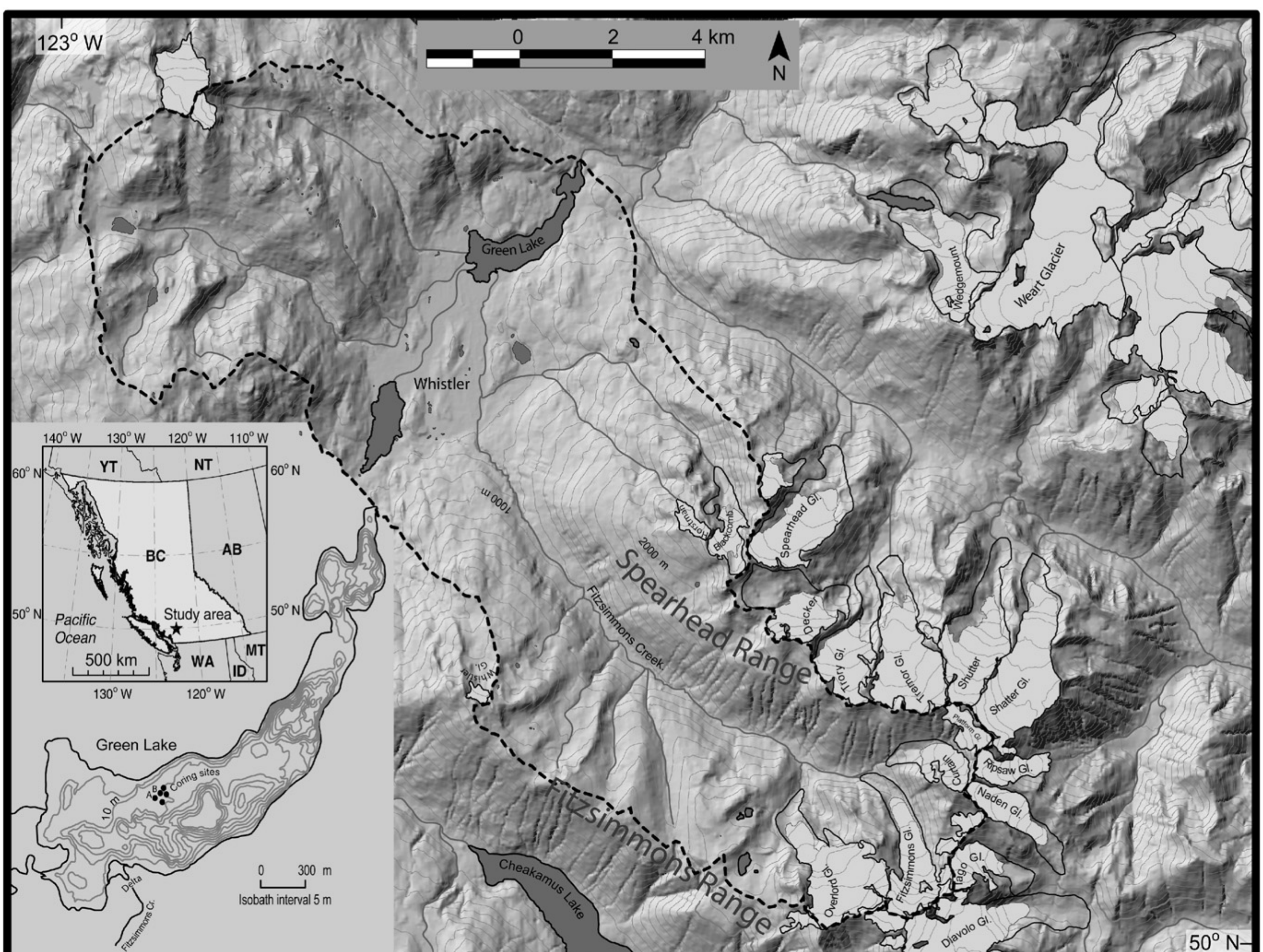

Fig. 1. Study area. Dashed line delineates the Green Lake watershed. Inset maps show the location of the study area in British Columbia and the bathymetry of Green Lake. Squamish, site of the closest weather station, is 25 km south–southwest of Cheakamus Lake. Lake bathymetry of Green Lake provided by E. Schiefer.

3. Methods

3.1. Forefield wood

We searched the forefields of Horstman, Spearhead, Decker, Trorey, Overlord, and Whistler glaciers for rooted stumps and detrital wood (Fig. 1). We used air photos and binocular-aided field traverses to limit the search to locations where detrital wood could not have been delivered to the glacier by snow avalanches or deadfall from living trees above the glacier (Ryder and Thomson, 1986). Proximal flanks of lateral moraines were also examined for wood and paleosols, but none was found. A GPS receiver was used to record the location of wood samples collected for radiocarbon dating. Discs of larger detrital logs were collected for tree-ring analysis.

3.2. Dendrochronology

Tree rings were used to date moraines at Overlord Glacier (Fig. 1). We sampled the oldest living trees on each moraine to determine a minimum age for moraine formation (Lawrence, 1946; Sigafoos and Hendricks, 1969; Luckman, 1998). Ecesis, the time required for trees to colonize a new substrate, was estimated from successive aerial photographs and the difference in age between the kill date of a glacier advance and the oldest trees on the moraine constructed during that advance. Most trees were cored near ground level to reduce age–height errors (McCarthy et al., 1991). Nineteen small trees were cut at ground level to estimate seedling growth rates and thus correct for any age–height bias.

Trees killed by a glacier advance can be crossdated into living chronologies to determine the date of death and presumably glacier overriding (Luckman, 1995). We cut discs from trees overridden by glaciers in the forefield of Overlord Glacier and measured ring width (± 0.001 mm) along four radii for each disk.

A master tree-ring chronology was built from trees growing on the Overlord moraines. The series was checked and verified using the International Tree-Ring Data Bank (ITRDB) software program COFECHA (Holmes, 1983). Floating chronologies were developed from a killed tree, which included bark, on the second outermost moraine, and from several logs in the glacier forefield. The floating series were crossdated against the living chronology using COFECHA.

3.3. Lichenometry

We used lichenometry to estimate the ages of the three inner moraines of Overlord Glacier. This technique could not be used on the two outermost moraines, as tree cover on these moraines is too extensive. We assume that the largest *Rhizocarpon geographicum* lichen growing on a suitable substrate provides a minimum age for the surface (Innes, 1985). We were unable to construct a local *Rhizocarpon* growth curve and therefore used three calibration curves from nearby areas: (1) Vancouver Island, ca. 200 km to the west (Lewis and Smith, 2004); (2) the Bella Coola-Mt. Waddington area, 200–350 km to the northwest (Smith and Desloges, 2000; Larocque and Smith, 2004); and (3) the Cascade Range of Washington State, ca. 100–350 km to the south (O'Neal and Schoenenberger, 2003). Sixty *Rhizocarpon* thalli were measured on each of the three inner moraines.

3.4. Documentary evidence

We mapped 20th century changes in ice cover from ground and aerial photographs. Ground-level photographs of parts of the Fitzsimmons Range date back to 1928 and 1929, and national and provincial aerial photographs have been taken since 1931. We merged mapped former glacier margins with a digital elevation model (DEM) in a GIS to determine former glacier areas and recession rates.

3.5. Lake sediments

We recovered percussion and vibracores from Green Lake to assess Holocene changes in sedimentation. Coring sites are located on a wide shallow shelf northwest of the Fitzsimmons Creek delta (Fig. 1). The Holocene record of sedimentation at the shelf was constructed from two vibracores about 50 m apart [00-Grn(A), 00-Grn(B), Fig. 1]. Over-penetration resulted in the loss of the uppermost 20 cm of sediments from core 00-Grn(A). A continuous record of bulk physical properties was obtained by combining measurements taken from the two vibracores. Bulk physical properties of the uppermost 1000 cm of core 00-Grn(B) were combined with measurements from the lowermost 200 cm of core 00-Grn(A). A unique, 2-cm graded bed was used to link the two cores. The percussion cores were primarily used to refine the varve chronology of the upper, Little Ice Age part of the sediment sequence. This varve chronology is discussed elsewhere (Menounos, 2006).

We split, photographed, and sampled the cores for bulk physical properties (water, dry density, organic content, particle size) using standard procedures. The loss-on-ignition (LOI) method (2 h at 550 °C) was used to estimate organic matter content. Changes in LOI in proglacial sediments commonly reflect changes in glacigenic sediment supply to the lake (Karlén, 1981; Souch, 1994; Leonard and Reasoner, 1999; Nesje et al., 2001). Samples were treated with 35% H_2O_2 and dispersed in a sodium metaphosphate solution prior to particle size analysis. Particle size was determined with a sedigraph. Laminated sediments were impregnated with low-viscosity resin (Lamoureux, 2001) and then slabbed and thin-sectioned for laminae counting and measurement (± 0.05 mm) under a dissecting microscope. Laminae were also identified and counted in partially dried sediment cores.

3.6. Radiocarbon dating

Wood samples collected from glacier forefields were submitted for conventional (beta) radiocarbon dating. We collected macrofossils from the Green Lake sediments by wet-sieving fresh sediment from the two cores. Macrofossils were oven-dried at 70 °C and placed in sealed glass vials prior to submission for accelerator mass spectrometry (AMS) dating. Radiocarbon ages of terrestrial and lacustrine macrofossils (Table 1) were converted to calendric ages using the calibration program CALIB 5.02 (Stuiver and Reimer, 1993; Stuiver et al., 2005). Ages of wood and macrofossil samples and age ranges of clastic events are reported in both radiocarbon (^{14}C yr) and calendar (cal yr) years BP (before AD 1950); ages within the past millennium are reported in calendar years AD.

4. Results

4.1. Fossil wood in glacier forefields

Overlord Glacier is located in the headwaters of Fitzsimmons Creek and is 2.75 km long (Figs. 1 and 2). Three meltwater streams issue from the glacier terminus (Fig. 3). We found 12 wood fragments along the middle stream. The fragments were about 5–30 cm long, splintered, and partly abraded. They were scattered along a 50 m reach of the stream, some as close as 40 m to the glacier terminus. The fragments either washed out of the glacier or melted out from the receding ice margin. One of the wood fragments yielded a radiocarbon age of 6170 ± 70 ^{14}C yr BP [7250–6900 cal yr BP] (Table 1).

The western outlet stream of Overlord Glacier has incised into a sheet of till at least 25 m thick. Several pieces of wood were found on the floor of the incised gully about 100 m downstream of the glacier snout. The largest sample is a 3.5 m long, 30–70 cm diameter log with a preserved rootstock (Fig. 4). The log is splintered, abraded, and in places weathered smooth. Branches and fragments of wood were also exposed in the west wall of the gully, at about the same elevation and 40 m downstream of the log on the gully floor. It is likely that the log was originally embedded in the till but was exhumed by fluvial erosion. Flakes of the outermost rings from the log and a sawed-off piece of the branch sticking out of the gully wall yielded radiocarbon ages of 5890 ± 70 and 5980 ± 70 ^{14}C yr BP, respectively [6890–6510 and 6990–6660 cal yr BP] (Table 1). The similarity of the three Overlord radiocarbon ages suggests that all three trees died at about same time.

Decker Glacier is 1 km long (Fig. 1) and presently confined to an upper basin. During the Little Ice Age, however, it flowed from the upper basin down a steep, 100 m-high bedrock slope to a lower basin. The glacier extended about 1250 m farther downstream at the Little Ice Age maximum than today and coalesced with the terminus of adjacent Trorey Glacier.

Several snags and stumps are rooted on a bedrock cliff adjacent to Decker Glacier on the north side of the upper basin (Fig. 5). The snags appear to have been broken off and sheared in an easterly (i.e., down-glacier) direction, indicating they were probably killed by an advancing glacier. The fossil trees were found 40–75 m higher than the modern glacier surface. Three sampled stumps were identified as *Abies* (fir) from their cell patterns (Bond and Hamner, 1995). The three stumps returned radiocarbon

Table 1
Radiocarbon ages from lake sediments and glacier forefields

Laboratory no[a]	Field no.	Material	^{14}C age (yr BP)	Calendar age (cal yr BP)[b]	Elevation (m)
Lake sediment					
AA-38704	00-Grn(B); 241 cm	Conifer needles	1300 ± 50	1300–1090[c]	—
AA-38705	00-Grn(B); 342 cm	Conifer needles	1860 ± 50	1920–1630[d]	—
AA-38706	00-Grn(B); 619 cm	Conifer needles	3230 ± 50	3570–3360[e]	—
AA-38707	00-Grn(B); 800 cm	Conifer needles	5040 ± 50	5900–5660	—
AA-38708	00-Grn(B); 962 cm	Wood (twig)	7940 ± 50	8990–8630	—
Glacier forefield					
Beta-157268	Spearhead-2	18-cm-long wood fragment	3900 ± 80	4530–4090	1995
Beta-168423	Spearhead-3	14-cm-long wood fragment	3900 ± 60	4510–4150	1995
Beta-170665	OV-1	27-cm-long wood fragment	6170 ± 70	7250–6900	1640
Beta-170660	OV-5	Outermost rings of log	5890 ± 70	6510–6890	1625
Beta-170667	OV-8	8-cm-diameter branch	5980 ± 70	6990–6660	1610
Beta-157262	Decker-2	Outermost rings of detrital log	2960 ± 50	3320–2970	2010
Beta-157263	Decker-3	Outermost rings of rooted snag	2960 ± 40	3320–2990	2010
Beta-157264	Decker-4	Outermost rings of rooted stump	2920 ± 50	3240–2900	2030
Beta-157265	Decker-5	Outermost rings of rooted stump	3200 ± 70	3580–3260	2040

[a]Radiocarbon laboratory: AA-University of Arizona; Beta = Beta Analytic Inc.
[b]Calendar ages ($\pm 2\sigma$) determined using Calib 5.02 (Stuiver and Reimer, 1993; Stuiver et al., 2005).
[c]Varve age (1030 ± 20 cal yr BP) without insertion of bioturbated varves (~40).
[d]Varve age (1585 ± 20 cal yr BP).
[e]Varve age (3150 ± 60 cal yr BP).

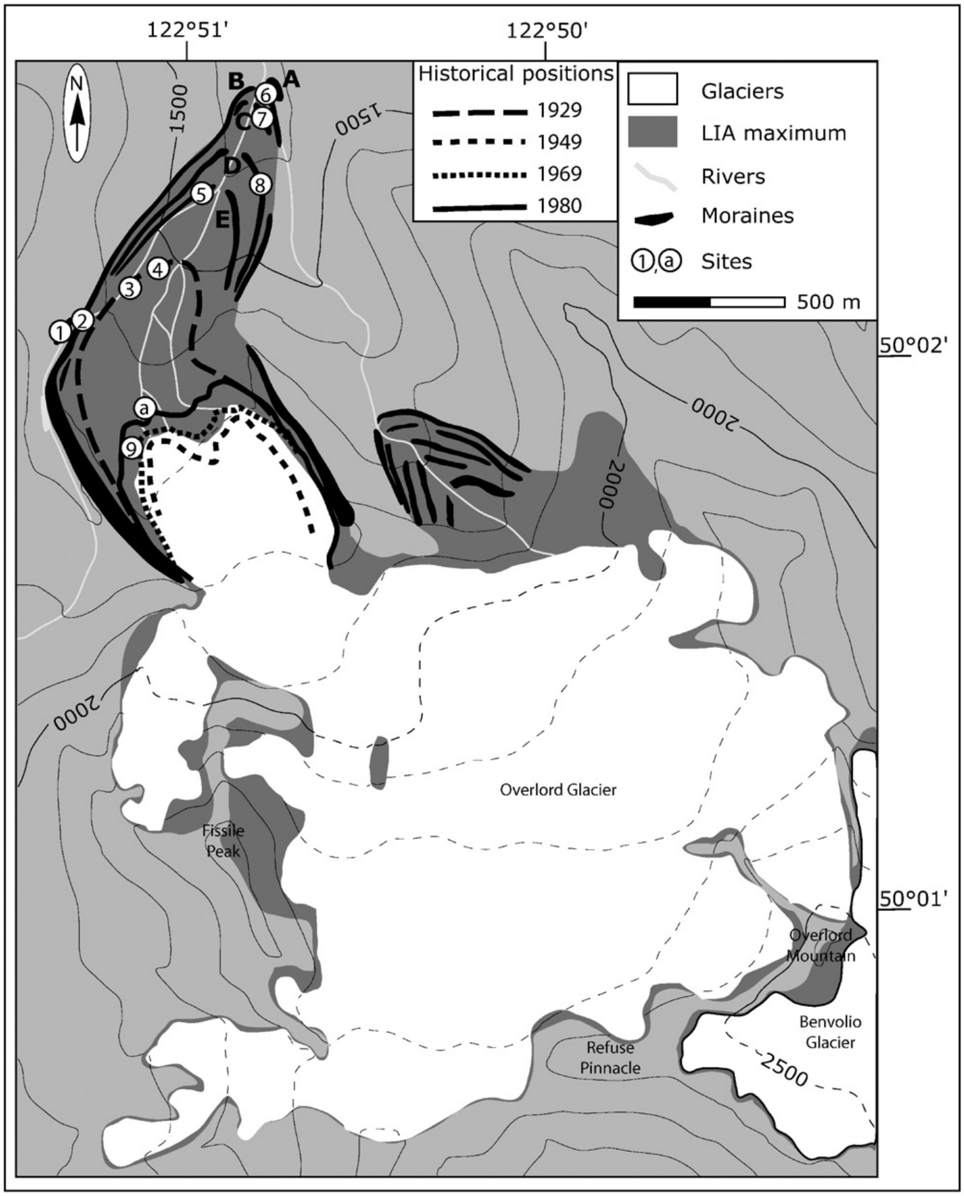

Fig. 2. Geomorphic map of Overlord Glacier showing the ice margin in 1996, dated moraines (A–E), and sampling sites (1–9). Sites 1, 2, 5, 6, 7, and 8 are where trees were sampled to determine moraines ages. Age/height and ecesis determinations were made at sites 3, 4, and 9. The map also shows positions of the snout of Overlord Glacier since 1929. The two streams that yielded detrital wood (“a”) appear in Fig. 3.

ages of 2960 ± 40 ^{14}C yr BP, 2920 ± 50, and 3200 ± 70 ^{14}C yr BP [3320–2990, 3240–2900, and 3580–3260 cal yr BP] (Table 1). A detrital log near the stump that was dated to 2960 ± 40 ^{14}C yr BP yielded a nearly equivalent age of 2960 ± 50 ^{14}C yr BP.

Spearhead Glacier is adjacent to Decker Glacier and flows 2.5 km northwestward from Blackcomb Peak (Fig. 1). Several small fragments of weathered, fractured wood were found over an elevation range of 10 m on a rock slope about 60 m north of the 1998 terminus and 15–25 m above the lateral ice margin. One conifer, 3.5 m tall in 2001, is growing nearby, but there is no source of modern wood upslope of the weathered fragments. The two largest fragments yielded radiocarbon ages of 3900 ± 80 and 3900 ± 60 ^{14}C yr BP [4530–4090 and 4440–4250 cal yr BP] (Table 1).

4.2. Dendrochronology

The master living tree-ring chronology spans the period from AD 1649–2001. Inter-series correlation is 0.53 ($p < 0.01$).

On the downstream side of the second-outermost moraine of Overlord Glacier, a rooted trunk is tilted away from the moraine that lies directly against it. The tree is interpreted to have been killed by glacier advance. A floating chronology from this tree was crossdated with the living chronology. The highest

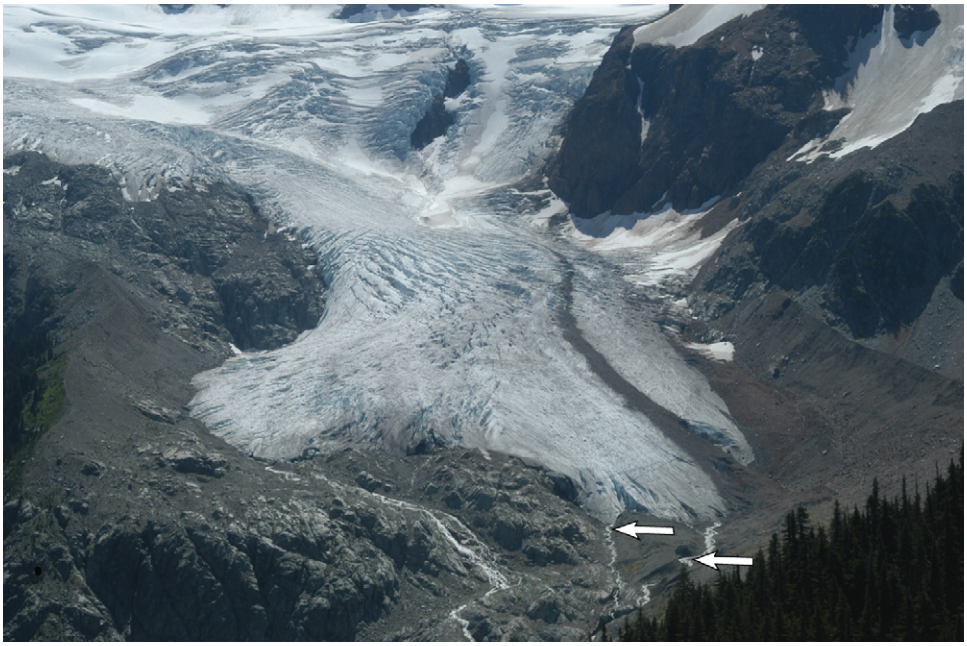

Fig. 3. Forefield of Overlord Glacier, showing sites where most detrital wood was found (arrows).

Fig. 4. Log in the forefield of Overlord Glacier (the site is denoted by the right-hand arrow in Fig. 3).

correlation ($r = 0.49$; $p < 0.01$) with the latter was achieved for the period AD 1649–1702, suggesting that the tree was killed in AD 1702 when the second-outermost moraine was constructed.

A second floating chronology was developed for four detrital wood samples from Overlord Glacier, including the radiocarbon-dated log. The ring series spans 138 years, and the inter-series correlation is 0.56 ($p < 0.01$). Crossdating showed that all four trees died within 6 years of one another.

Overlord Glacier was more extensive during the Little Ice Age than today (Figs. 2 and 6). Trees on moraine E, sampled in 2002, were as old as 52 years. Maximum tree ages for moraines D, C, and B are, respectively, 84, 146, and 294 years (2002 datum). Adding age-height corrections and ecesis values to the ring counts gives minimum limiting ages of AD 1920s, 1890s, 1830s, and 1700s for moraines E, D, C, and B, respectively. The inferred age of moraine B compares favorably to the date of AD 1702 for the death of a tree killed by an advance of Overlord Glacier.

4.3. Lichenometry

The three regional lichen growth curves (Fig. 7) yielded different ages for stabilization of the three inner moraines of Overlord Glacier (moraines C, D, and E; Fig. 2) (Table 2). The Bella Coola—Mt. Waddington curve gave stabilization dates of AD 1600s, 1730s, and 1870s; the Cascades curve gave dates of AD 1870s, 1900s, and 1920s; and the curve from Vancouver Island gave dates of AD 1830s, 1890s, and 1920s. The Vancouver Island curve is supported by our dendrochronological results and thus is considered most appropriate for Garibaldi Park.

Fig. 5. Rooted stumps on the bedrock cliff above Decker Glacier. The stumps indicate the level of the glacier at approximately 3000 ^{14}C yr BP.

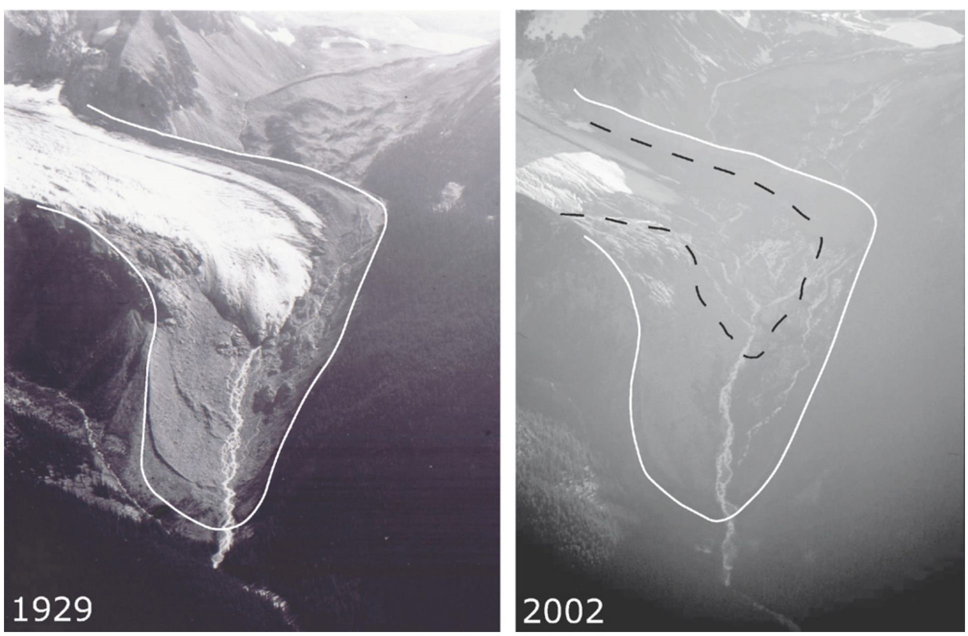

Fig. 6. Overlord Glacier in 1928–1929 (British Columbia Archives I-67251) and 2002. White lines indicate Little Ice Age maximum position; black dashed line on the 2002 photo is the 1928–1929 position.

4.4. Green lake sediment record

Recovered sediments from Green Lake are mainly laminated silty clay and clayey silt. The sediments have little or no sand ($<$ 1% by weight), except at the base of the sequence. In general, density, organic matter, and mineral flux (g yr^{-1}) increase upcore (Fig. 8).

The lowest 10 cm of the sequence is inorganic, dense sandy silt with isolated pebbles. The basal sediments are overlain by 100 cm of inorganic, rhythmically laminated clayey silt, comprising 79 couplets. Each couplet consists of a silt lamina conformably overlain by a clay lamina. Silt–clay couplets thin up the core in a non-linear fashion. Silt and clay laminae in the lowest 20 couplets are about equal in thickness, whereas silty laminae exceed 70% of the total couplet thickness of the upper 59 couplets.

The inorganic sediments are overlain by less dense, laminated, organic-rich silty clay (Fig. 8). A radiocarbon age of 7940 $\pm$ 50 ^{14}C yr BP [8630–8990 cal yr BP] was obtained from a small twig at 962 cm depth (Table 1).

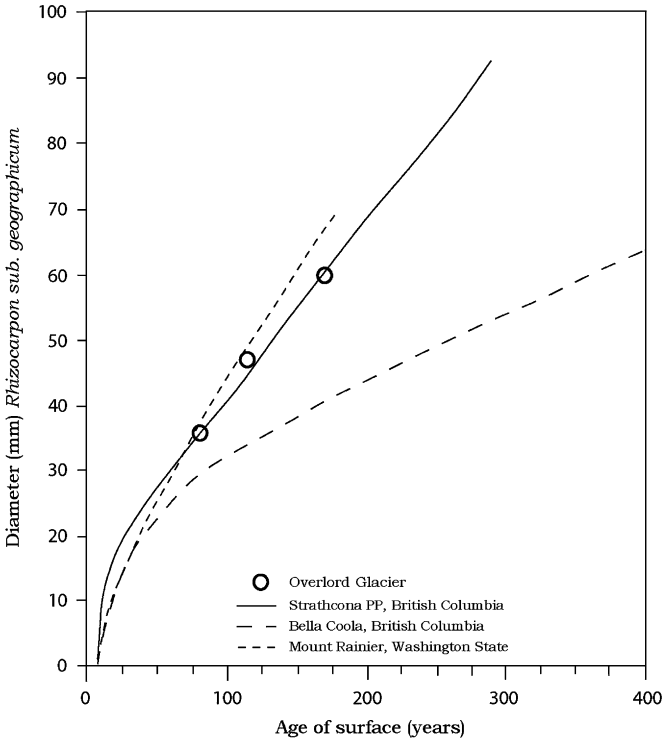

Fig. 7. Three lichen growth curves considered in this study, and data points from the Overlord Glacier forefield. The curve from Strathcona Provincial Park on Vancouver Island best fits the data points. Strathcona Provincial Park—Lewis (2001), Lewis and Smith (2004); Bella Coola—Smith and Desloges (2000), Larocque and Smith (2004); Mount Rainier—Porter (1981), O'Neal and Schoenenberger (2003).

Table 2
Diameters of single largest and five largest (mean) lichens on Little Ice Age moraines at Overlord Glacier

Moraine	Lichen diameter (mm)		Date AD using growth curve from		
	Single	Five	Vancouver Island[a]	Cascades[b]	Central Coast Mts.[c]
A	n/a	n/a	n/a	n/a	n/a
B	n/a	n/a	n/a	n/a	n/a
C	59.7	58.4	**1830s**	1870s	1600s
D	47.5	46.8	**1890s**	1900s	1730s
E	35.6	34.9	**1920s**	1920s	1870s

Note: Dates derived from the Vancouver Island curve are considered most reliable and appear in bold.
[a]Lewis (2001); Lewis and Smith (2004).
[b]Porter (1981); O'Neal and Schoenenberger (2003).
[c]Smith and Desloges (2000); Larocque and Smith (2004).

The organic-rich sediments contain several intervals of light-colored, clastic-rich sediment (Fig. 8). The most prominent of the light-colored intervals is centered near 900 cm, 10 cm below a 1 cm thick, silty tephra. The tephra has physical properties similar to those reported for Mazama tephra which include a fine texture, absence of biotite-rich phenocrysts, and thin bubble-wall glass shards (Reasoner and Healey, 1986). Mazama tephra dates to 6730 ± 40 ^{14}C yr BP [7670–7510 cal yr BP] (Hallett et al., 1997). Mineral flux reaches minimum values for the core between 962 and 800 cm depth (Fig. 8).

Sediments between 800 and 600 cm depth are weakly laminated, inorganic silty clay (Fig. 8). Laminae in this interval are diffuse, strongly bioturbated, and less dense than overlying sediments. Terrestrial macrofossils at 800 cm depth yielded an age of 5040 ± 50 ^{14}C yr BP [5900–5660 cal yr BP], and conifer needles at 619 cm depth gave an age of 3230 ± 50 ^{14}C yr BP [3630–3360 cal yr BP] (Fig. 8 and Table 1).

Sediments above 600 cm depth comprise inorganic, rhythmically laminated, clayey silt couplets (Fig. 8). Chironomid head capsules and trace fossils are common in bioturbated sections above 600 cm depth. Conifer needles from 342 and 241 cm yielded radiocarbon ages of 1860 ± 50 ^{14}C yr BP [1920–1630 cal yr BP] and 1300 ± 40 ^{14}C yr BP [1300–1080 cal yr BP]), respectively (Table 1).

Couplets above 600 cm are interpreted to be varves. Average couplet thickness is 1.96 ± 1.00 mm yr^{-1}, which is in agreement with the average sedimentation rate of the uppermost 619 cm of sediment based on radiocarbon ages (1.71–1.84 mm yr^{-1}). A varve interpretation is also supported by the cesium activity of the sediments and the formation of new couplets at sites cored over periods of several years (Menounos et al., 2005a). The possible error in varve counts was determined by repeat counting and by calculating the number of missing and extra varves between marker beds. It is 1.7%, or approximately 2 varves/century.

We produced a varve chronology by averaging varve thicknesses among the contributing cores. Standardization was not required because varve thickness for a given year did not differ significantly among cores. Thickness measurements could not be completed on a 5 cm interval of bioturbated sediment, which was estimated from varve thickness above and below the interval to represent 40 years of sedimentation. We added 40 varves of random thickness to approximate the number and the thickness of missing varves. The chronology is based on one core prior to 1000 cal yr BP, two cores for the period 1000–300 cal yr BP, four cores from 300–150 cal yr BP, and five cores after 150 cal yr BP. Mean inter-series correlation is 0.73. Varve thickness at the sill explains 80% of annual lake-wide sedimentation for the period AD 1931–1998 based on an extensive network ($n = 120$) of surface sediment cores (Menounos et al., 2006).

The thickest varves date to about 2700–2450, 1700–1550, 1300–1200, and 700–600 cal yr BP. These ages may be in error by up to 100 years because of the lower accuracy of counts below 360 cm depth. Thick varves were also deposited ca. AD 1620–1665, 1670–1755, 1770–1785, and 1920–1950 (Fig. 9). The thickest varves in the 50-year smoothed record date to AD 1920–1945.

4.5. Photographic record

Aerial and ground photographs show that Overlord Glacier receded through most of the 20th century (Fig. 2).

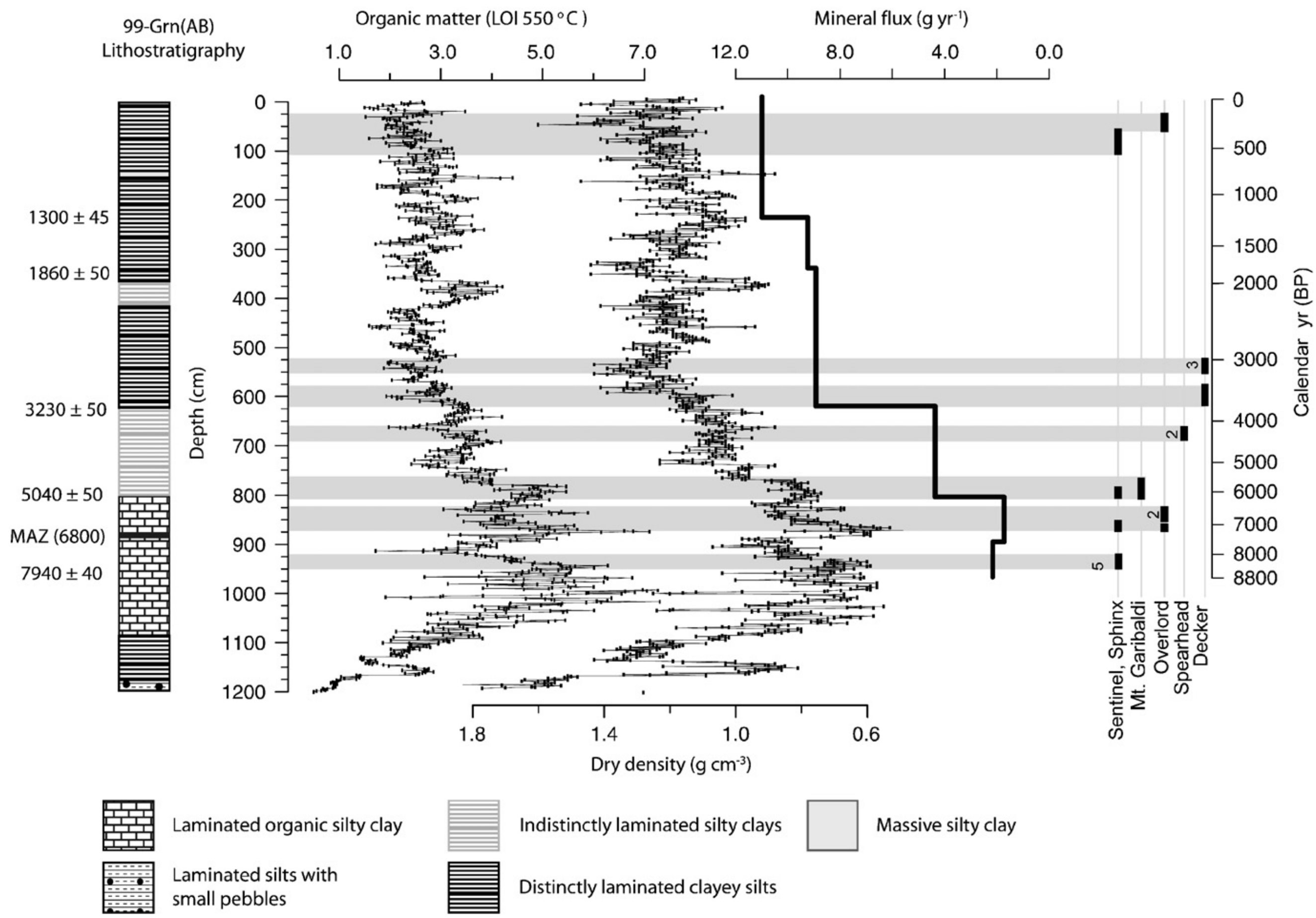

Fig. 8. Green Lake core 00-GRN(AB) lithostratigraphy, bulk physical properties, and mineral flux records. The age scale is based on calibrated radiocarbon and tephra ages. Black boxes and associated gray rectangles indicate the 95% confidence limits of calibrated age ranges for detrital wood samples in the forefields of Sentinel, Sphinx, Mt. Garibaldi (neve), Overlord, Spearhead, and Decker glaciers. Numbers to the left of the boxes denote the number of contributing radiocarbon samples if the number is more than one. Mineral flux represents the product of: (a) sedimentation rates calculated for intervals between bracketing calibrated radiocarbon or tephra ages (cm yr^{-1}); (b) the area of the core (cm^2); and (c) the average, organic-free dry density (g cm^{-3}) of the sediments within the interval.

In 1929 the glacier had retreated about 700 m from its Little Ice Age limit and about 250 m from its 1920s terminal position. The corresponding average annual recession rates are 3 and 28 m yr^{-1}. Between 1929 and 1949, the rate of retreat increased to 33 m yr^{-1}. The glacier reached its farthest upvalley position sometime between 1949 and 1969, and advanced about 120 m at an average rate of 6 m yr^{-1} between 1969 and 1980. Overlord Glacier retreated about 120 m at an average of 7.5 m yr^{-1} between 1980 and 1996. Recession rates increased to 10 m yr^{-1} between 1996 and 2002. All glaciers in the Fitzsimmons Creek watershed show a similar pattern of fluctuation over the last 70 years.

5. Discussion

5.1. Interpretation of glacier forefield evidence

Sheared in situ tree stumps provide strong evidence for glacier fluctuations. In contrast, detrital wood in glacier forefields must be carefully interpreted. Detrital wood may have been delivered onto the glacier by snow avalanches (Ryder and Thomson, 1986), or it may have been reworked from older glacial deposits. Furthermore, even if a glacier is presently above treeline, trees may have been present on slopes above the ice earlier in the Holocene. Consideration of such possibilities must be done on a case-by-case basis, but the case for glacial killing of trees that produced the detrital wood is strengthened if the wood ages are consistent with regional evidence external to the study area and with other proxies for glacial activity, such as a dated interval of clastic-rich sediment in a proglacial lake.

Taken alone, the evidence from the Overlord detrital wood for a glacier advance ca. 6000 ^{14}C yr BP [ca. 6900–6700 cal yr BP] is equivocal, as treeline is presently above the lower part of the glacier and presumably was still higher during the early Holocene (Clague and Mathewes, 1989). However, avalanching of wood seems unlikely because germination of trees is inhibited by unsuitably rough and steep substrates in all positions that feed directly onto the glacier and its forefield. No obvious sites exist that, even in the past, could support trees as large as the trunk found in the Overlord forefield (Fig. 4). We cannot,

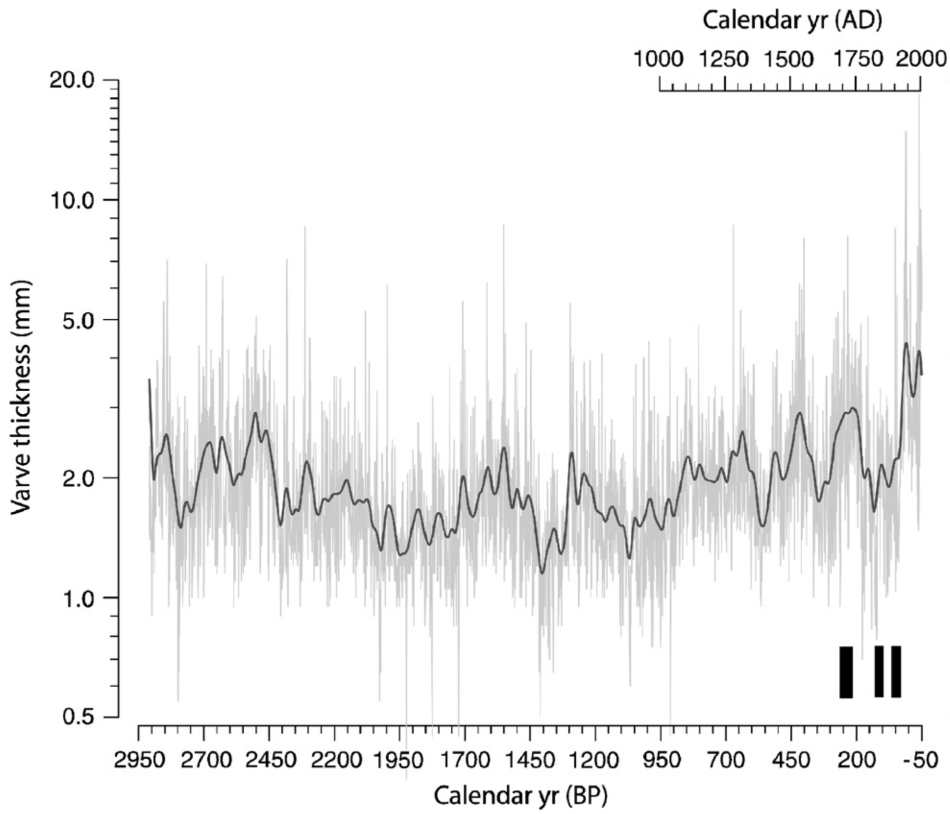

Fig. 9. Thickness of varves in the upper part of the Green Lake sediment sequence. Gray lines show varve thicknesses; the darker line is the smoothed series (50 year filter); black boxes show age ranges of moraines dated by dendrochronology and lichenometry.

however, rule out the possibility that trees grew above the glacier 6000 ^{14}C yr BP.

Decker Glacier was generally thickening and hence advancing between 3200 and 2920 ^{14}C BP [3580–2900 cal yr BP]. The oldest glacially sheared stump is the highest on the cliff, so ice probably thickened ca. 3200 ^{14}C yr BP to kill that tree, then retreated enough to allow new trees to grow on the cliff below, and finally readvanced to kill trees ca. 2960–2920 ^{14}C yr old. During the ~3200 ^{14}C yr BP advance, the glacier margin reached an elevation at least 75 m higher than that in 1998. No direct comparisons to Little Ice Age glacier thickness and length can be made, because the cliff containing the radiocarbon-dated stumps lacks a Little Ice Age moraine or trimline. However, assuming the glacier reached halfway up the bedrock cliff, it would have extended into the lower basin and nearly reached the Little Ice Age limit.

Interpretation of the wood samples from Spearhead Glacier that gave radiocarbon ages of 3900 ^{14}C yr BP [4530–4090 cal yr BP] is equivocal. The dated wood fragments may derive from a single tree and conceivably could have been moved by gravity from a higher position when treeline was higher. If, for the sake of argument, we assume glacial killing, the glacier margin would have been about at the level of the wood fragments 3900 ^{14}C yr BP; the glacier would have been 20 m thicker than today, with a terminus 500 m upvalley of the LIA limit.

5.2. Changes in Holocene clastic sedimentation

As is the case with detrital forefield wood, intervals of clastic-rich, proglacial lacustrine sediment, taken in isolation, constitute an equivocal proxy for glacial activity. For example, sediments can be delivered episodically to montane lake systems during non-glacial events such as large floods (Menounos et al., 2005a). But a glacial origin is suggested if the clastic interval coincides with tightly constrained ages on detrital forefield wood (Menounos et al., 2004), or is consistent with regional evidence.

The Green Lake sediment record reveals relatively low clastic sediment delivery to the lake during the early Holocene, followed by a steady increase to the present (Fig. 8). We attribute this trend to waning availability of glaciogenic sediment following late Pleistocene deglaciation, followed by episodic and progressively more extensive expansion of glaciers over the past 9000 calendar years. Similar changes in proglacial sedimentation have been documented in other proglacial lakes in the region (Menounos, 2002) and elsewhere in the Canadian Cordillera (Souch, 1994; Leonard, 1997; Leonard and Reasoner, 1999).

Sediments between 1070 and 800 cm depth [$>7940\pm40$ to 5040 ± 50 ^{14}C yr BP; >8990–8630 to 5900–5660 cal yr BP] contain less mineral and rock detritus than both underlying and overlying sediments (Fig. 8). The presence of diatoms and vivianite, a phosphorus replacement mineral, and the strongly bioturbated nature of the sediments indicate an increase in autochthonous sedimentation. Clastic sediment delivery increases abruptly at 800 and 600 cm [ca. 5040 ± 50 ^{14}C yr BP; 5900–5660 cal yr BP; and 3230 ± 50 ^{14}C yr BP; 3630–3360 cal yr BP] (Fig. 8). These transitions are superimposed on the general increase in clastic sediment delivery to the lake from early Holocene to present.

The clastic content of the lake sediments remains high after 3230 ± 50 ^{14}C yr BP [3630–3360 cal yr BP], but the record is characterized by substantial, century-scale variability, reflecting shorter-term changes in sediment delivery to the lake basin. The densest and most inorganic sediments date to ca. 3200, 2400, 1600, and 300 cal yr BP (Fig. 8). Centennial-scale variability in the clastic content of glacigenic sediments has been documented at other lakes in the Cordillera (e.g. Souch, 1994; Leonard and Reasoner, 1999) and elsewhere (Karlén, 1981; Bakke et al., 2005).

5.3. Formation of neoglacial clastic varves

Clastic varves form in proglacial lakes when sedimentation outpaces physical or biologic reworking and when sediment delivery is strongly seasonal. Differences in the type, quantity, and times of major inflow events provide the heterogeneity required for varve formation (Smith and Ashley, 1985). Green Lake sediments younger than ca. 3230 ± 50 ^{14}C yr BP [3630–3360 cal yr BP] are dominantly well preserved varves. Bioturbation is limited to zones in which LOI is above 3% and dry density is less than 1.1 g cm^{-3}. Mineral flux increases twofold from 4 g yr^{-1} between 800 and 600 cm depth to over 8 g yr^{-1} above 600 cm (Fig. 8). The most likely explanation for the onset of varve formation about 3230 ^{14}C yr BP is the increase in the flux of clastic sediment to the lake.

5.4. Comparison of Green Lake and glacier forefield records

We examined the relation between sediment delivery to Green Lake and glacier activity by comparing the Green Lake record to the radiocarbon-dated, in situ and detrital wood from the forefields of Spearhead, Decker, and Overlord glaciers, and to previously published evidence from Sphinx and Sentinel glaciers in Garibaldi Park south of the Fitzsimmons range. We consider a glacier advance to be the simplest and most economical explanation for correlative ages of the forefield wood and intervals of above-average, clastic sediment accumulation in Green Lake.

The clastic interval centered at 900 cm in the Green Lake core (see *Green Lake sediment record* above) is correlative with radiocarbon ages obtained on wood from the forefields of Sentinel and Sphinx glaciers (Menounos et al., 2004). The age range for the clastic event is 8400–7750 cal yr BP. This age range overlaps the calibrated range of Sentinel and Sphinx detrital wood samples [8630–8020 cal yr BP] (Fig. 8) and is suggestive of a minor glacial advance at a time of generally restricted ice.

Detrital wood from Overlord Glacier dates to 6170–5890 ^{14}C yr BP [7250–6510 cal yr BP] and coincides with a clastic interval in the Green Lake sediment record between 860 and 840 cm depth. This interval is bracketed by underlying Mazama tephra [7670–7510 cal yr BP] at 890 cm and radiocarbon-dated conifer needles [5900–5660 cal yr BP] at 800 cm (Fig. 8). Using minimum and maximum sedimentation rates determined from the radiocarbon ages, we calculate the average age range of the clastic interval to be 7030–6620 cal yr BP. Taken together, the Overlord detrital wood and the Green Lake clastic event is interpreted to record an advance of Overlord Glacier at ca. 6000 ^{14}C yr BP [6950–6750 cal yr BP]. The timing of the advance is consistent with ages of in situ stumps killed by advancing ice at nearby glaciers (see regional comparisons). The simplest explanation for the proximity of the detrital wood to the 2002 Overlord terminus and its position about 1 km upvalley of the Little Ice Age limit is that Overlord Glacier had a similar extent ca. 6000 ^{14}C yr BP as today.

Ages of Spearhead Glacier detrital wood (3900 ^{14}C yr BP; 4530–4090 cal yr BP) and a Green Lake clastic interval between 670 and 650 cm depth [4160–3900 cal yr BP] overlap and may record an advance of Spearhead Glacier about 3900 ^{14}C yr BP. The sparse wood recovery in the Spearhead forefield, however, makes this conclusion speculative.

In situ stumps in the Decker Glacier forefield indicate that the glacier was at least 75 m thicker than today when it overran trees ca. 3200 ^{14}C yr BP, and almost as thick during a readvance a few hundred years later. Two clastic intervals in the Green Lake sediment record coincide with the ages of the stumps and detrital wood samples (Fig. 8). The Decker Glacier advances also coincide with the formation of late Holocene clastic varves in Green Lake. After 3230 ^{14}C yr BP [3630–3360 cal yr BP], mineral flux to the lake basin more than doubled and remained high thereafter. This evidence suggests that glaciers reached and remained in downvalley positions comparable to those achieved during the Little Ice Age.

Age control provided by dendrochronology, lichenometry, and varve counting allows us to scrutinize the Little Ice Age glacial and lacustrine records in greater detail than is possible for older advances. Varves, on average, thicken from AD 1000 to 1945 (Fig. 9). Especially thick varves date to ca. AD 1300, 1500, 1700, and 1920–1945 (Fig. 9). Thick 18th-century varves coincide with construction of the second outermost moraine at Overlord Glacier.

Overlord Glacier built recessional moraines in AD 1830 and 1890, times when lake sedimentation was lower than average. Thin varves during most of the 19th century may reflect: (1) greater ice extent than at present but with minor frontal fluctuations; or (2) decreased sediment availability. The thickest varves in Green Lake date to AD 1920–1945; they were deposited after construction of a small recessional moraines in AD 1920.

Several prominent clastic events with no obvious glacier forefield counterparts are evident in the Green Lake sediment core (Fig. 8). Although their amplitude, duration, and visual characteristics are comparable to events discussed in this paper, their relation to glacier fluctuations remains uncertain. Given the fragmentary nature of the terrestrial record of glacier fluctuations, we suspect that many of these events may reflect century-scale periods of glacier advance. Additional chronologic control, in addition to new evidence from the regional record of glacier fluctuations, is required to substantiate this claim.

6. Regional comparisons

6.1. Holocene prior to the Little Ice Age

We interpret the increase in sediment flux observed in the Green Lake sediment record to an increase in glacier cover in the Fitzsimmons Creek watershed from early Holocene to present. The Green Lake sediment record suggests: (1) relatively small extent of glaciers between 7940 $\pm$ 40 and 5040 $\pm$ 50 ^{14}C yr BP [8990–5660 cal yr BP]; (2) growth of glaciers between 5040 $\pm$ 50 and 3230 $\pm$ 50 ^{14}C yr BP [5900–3360 cal yr BP], with some glaciers achieving extents comparable to maximum Little Ice Age positions by the end of this interval; and (3) repeated glacier advances and retreats after 3230 $\pm$ 50 ^{14}C yr BP, but with average ice cover during this interval more extensive than today.

The Green Lake clastic sediment trend mirrors the record of Holocene glacier fluctuations elsewhere in western North America (Davis, 1988; Osborn and Luckman, 1988). The latter indicates minimum glacier extent in the early Holocene, followed by episodic growth of alpine glaciers culminating with the climactic advances of the Little Ice Age. These long-term changes broadly coincide with reductions in Northern Hemisphere summer insolation (Berger, 1978) from early Holocene to present.

Evidence for retracted early Holocene glaciers is provided by fossil forest remnants close to or upstream of middle 20th century ice limits in Garibaldi Park (Stuiver et al., 1960; Lowdon and Blake, 1973, 1975; Koch et al., 2004; Menounos et al., 2004), at Mt. Breakenridge (Lowdon and Blake, 1968), Bridge Glacier (Blake, 1983), Dome Glacier (Luckman et al., 1993), and Athabasca Glacier (Luckman, 1988). Corroborating evidence for relatively high, early Holocene treeline comes from Castle Peak in the southeastern Coast Mountains (Clague and Mathewes, 1989).

Though generally restricted, glaciers probably advanced and retreated small amounts in the early Holocene, judging from small variations in clastic input to Green Lake. One of these minor clastic horizons, at 900 cm depth, overlaps in age the ages of detrital wood from Sphinx and Sentinel glaciers, and is interpreted above to record a minor glacier advance. Menounos et al. (2004, 2005b; discussion by Kovanen and Begét, 2005) related this minor advance to the well documented "8200 yr cold event" (also known as the "8k event"), first recognized in Greenland and the North Atlantic region (Alley and Ágústsdóttira, 2005).

Some authors have reported early Holocene advances of more significant magnitude, but the substance and/or timing of most of these advances has been questioned: (1) an advance on Mt. Baker dated to between 8400 and 7700 ^{14}C yr BP [9470–8430 cal yr BP] (Thomas et al., 2000); criticized by Reasoner et al. (2001) and Davis et al. (2005); (2) an advance on Glacier Peak ~8400 ^{14}C yr BP [~9470–9430 cal yr BP] (Begét, 1981); criticized by Davis and Osborn (1987); and (3) an advance on Mt. Rainier ~8900 ^{14}C yr BP [~10,170–9920 cal yr BP] (Heine, 1998); criticized by Reasoner et al. (2001). The point of contention in these debates is whether early Holocene advances were more extensive than those of the Little Ice Age. The early and middle Holocene clastic intervals in the Green Lake sediment record, which we ascribe to glacier advances, contain substantially less mineral matter than sediments younger than 3500 cal yr BP (Fig. 8). These data, together with the proximity of detrital wood samples to contemporary ice limits, suggests that early to middle Holocene advances were less extensive than advances of the late Holocene.

Glacier expansion in British Columbia ca. 6000–5000 ^{14}C yr BP [~7000–5600 cal yr BP] is referred to the "Garibaldi Phase" (Ryder and Thomson, 1986). This event was originally termed a "phase" rather than an "advance" on the basis of sheared stumps and detrital wood in glacier forefields in Garibaldi Park, on Mt. Breakenridge, and near Bridge Glacier. The term has since been applied elsewhere in Garibaldi Park (Koch et al., 2004) and at glaciers farther north in the Coast Mountains (Smith, 2003). It has not been used in the Canadian Rockies, but detrital wood of Garibaldi age occurs in the forefield of Dome Glacier (Luckman et al., 1993). The detrital wood in the Overlord Glacier forefield, ranging in age from 6170 to 5890 ^{14}C yr BP, and an equivalent clastic interval in the Green Lake sediment record indicate an advance of glaciers during the Garibaldi Phase. The lake sediment record suggests that glaciers retreated slightly after the Garibaldi Phase, although not to their early Holocene positions.

The most abrupt increase in the clastic content of Green Lake sediments is between 5040 [5900–5650 cal yr BP] and 3230 ^{14}C yr BP [3630–3360 cal yr BP] (Fig. 8). The regional glacier record provides substantial evidence for glacier expansion at this time. An age of 4360 $\pm$ 80 ^{14}C yr BP [5290–4820 cal yr BP] was obtained on detrital wood recovered in fluvial sediments in a moraine of Gilbert Glacier in the central Coast Mountains (Ryder and Thomson, 1986). Two rooted stumps in the forefield of Boundary Glacier in the Canadian Rockies yielded ages of 4050 $\pm$ 70 ^{14}C yr BP [4830–4410 cal yr BP] (Gardner and Jones, 1985) and 3880 $\pm$ 40 ^{14}C yr BP [4420–4160 cal yr BP] (Wood and Smith, 2004). Wood washed out of the snout of Helm Glacier in Garibaldi Park in 2003 gave an age of 4080 $\pm$ 40 ^{14}C yr BP [4810–4440 cal yr BP] (J. Koch, unpublished data). Detrital wood in the forefields of Goddard Glacier in the southern Coast Mountains and Haworth Glacier in the Selkirk Mountains yielded ages of 4120 ± 60 ^{14}C yr BP [4830–4450 cal yr BP] and 3870 ± 60 ^{14}C yr BP [4440–4100 cal yr BP], respectively (J.J. Clague, unpublished data). The wood from Spearhead Glacier discussed above dates to this interval, but its context does not allow it to be conclusively linked to a glacier advance. However, a clastic interval in the Green Lake record is the same age as wood from Spearhead Glacier, strengthening the case for a short-lived glacier advance at about 3900 ^{14}C yr BP [ca. 4420–4260 cal yr BP].

Decker Glacier overran trees at 3200 and 2900 ^{14}C yr BP [3580–3260, 3320–2900 cal yr BP], at which times the glacier was near its maximum Little Ice Age extent. These advances date to the time of inception of the Tiedemann Advance ($\sim$3300–1900 ^{14}C yr BP; $\sim$3560–1820 cal yr BP) of Ryder and Thomson (1986). Tiedemann Glacier, in the southern Coast Mountains, was more extensive between 3350 and 1300 ^{14}C yr BP [3630–1180 cal yr BP] than during the Little Ice Age, and it reached its maximum Holocene extent $\sim$2300 ^{14}C yr BP [$\sim$2350–2330 cal yr BP] (Arsenault, 2004). Other glaciers in the Coast Mountains show similar behavior. Gilbert Glacier advanced prior to 2200 ^{14}C yr BP [2300–2150 cal yr BP] and reached its Tiedemann maximum $\sim$1900 ^{14}C yr BP [$\sim$1880–1820 cal yr BP] (Ryder and Thomson, 1986). Frankmackie Glacier was relatively extensive ca. 2700 ^{14}C yr BP [2840–2760 cal yr BP] (Clague and Mathews, 1992). During the equivalent Peyto Advance in the Canadian Rockies, glaciers overrode a forest at 3300–2800 and $\sim$2500 ^{14}C yr BP [3560–2860 and 2710–2500 cal yr BP] (Luckman et al., 1993). The Peyto Advance of Stutfield Glacier peaked $\sim$2400 ^{14}C yr BP [$\sim$2460–2350 cal yr BP] (Osborn et al., 2001). Bugaboo Glacier advanced at least twice during the Peyto interval, once between $\sim$3400 and 3070 ^{14}C yr BP [$\sim$3690 and 3260 cal yr BP], and again between 3070 and 2000 ^{14}C yr BP [$\sim$3350 and 1900 cal yr BP] (Osborn and Karlstrom, 1989).

Glacier fluctuations inferred from the Green Lake sediment record are broadly synchronous with the Tiedemann and Peyto advances. The Green Lake record suggests glacier extent similar to that of the Little Ice Age beginning soon after 3200 ^{14}C yr BP [3450–3390 cal yr BP] and reaching a maximum at 2400 ^{14}C yr BP [2460–2350 cal yr BP]. The lake record implies that glaciers remained more extensive than today from Tiedemann time to AD 1945. It also shows that numerous small-scale advances and retreats were superimposed on that extension.

Evidence from 17 forefields in the Pacific Northwest indicates that glaciers expanded at about 1700 cal yr BP (Reyes et al., 2006). We found no evidence in glacier forefields in the Fitzsimmons and Spearhead ranges for glacier expansion at this time. However, the Green Lake clastic sediment record indicates intervals of increased sedimentation between 1860 ± 50 [1920–1630 cal yr BP] and 1300 ± 45 ^{14}C yr BP [1300–1090 cal yr BP]. Thick varves date to 1700–1550 and 1300–1200 cal yr BP.

6.2. Little Ice Age

The Little Ice Age history of Overlord Glacier is similar to that of most glaciers in the western Cordillera of the Americas (Luckman and Villalba, 2001), especially western North America (Heikkinen, 1984; Luckman, 2000; Larocque and Smith, 2003; Koch et al., 2004; Lewis and Smith, 2004). Moraines C (AD 1830s), D (AD 1890s), and E (AD 1920s) have their counterparts in the Cascade Range (AD 1820–1890 and 1920; Heikkinen, 1984), on Vancouver Island (AD 1840, 1890, and 1930; Lewis and Smith, 2004), in the central Coast Mountains (AD 1820–1840, 1870–1900, and 1915–1930; Larocque and Smith, 2003), and in the Canadian Rocky Mountains (AD 1825–1850 and 1850–1920; Luckman, 2000). Moraine B is the same age as moraines on Vancouver Island (AD 1690–1710; Lewis and Smith, 2004) and in the Canadian Rockies (AD 1700–1725; Luckman, 2000). Moraines of similar age are also found in the Cascades (AD 1740; Heikkinen, 1984) and the central Coast Mountains (AD 1660 and 1760–1785; Larocque and Smith, 2003). Moraine A is undated but is older than AD 1649, the age of the oldest tree on the moraine (Fig. 2, site 1). The moraine has much forest deadfall, thus the AD 1649 age is considered a significant underestimation. Consequently, moraine A could be early Little Ice Age or even older.

Leonard (1997) concluded that the thickest varves in Hector Lake in the Canadian Rocky Mountains were deposited during or immediately following moraine construction. Both the Green and Hector Lake chronologies indicate high sedimentation rates in the early 20th century related to increased sediment availability following retreat of glaciers from maximum Little Ice Age positions (Leonard, 1997; Menounos, 2002). The relation between rapid glacier recession and high sedimentation rates has been observed at other montane lakes in the southern Coast Mountains (Menounos et al., 2005a).

7. Conclusions

Glacial and proglacial lacustrine records are complementary data sets for inferring Holocene glacier fluctuations. Our lacustrine record provides context for interpreting detrital wood in the glacier forefields of this study. Overlapping age ranges of radiocarbon-dated detrital wood and clastic sediment intervals in Green Lake is evidence that the wood was probably killed by advancing glaciers. The evidence from the Spearhead and Fitzsimmons ranges in Garibaldi Park indicates that glaciers expanded episodically from minimum extents in the early Holocene. The main pre-Little Ice Age advances occurred 8630–8020, 7250–6510, and 3580–2900 cal yr BP. There may also have been an advance at 4530–4090 cal yr BP, but the evidence, from the Spearhead Glacier forefield, is tenuous. Glacier extent during the Garibaldi interval (7250–6510 cal yr BP) were comparable to those of the mid-20th century. By 3500 cal yr BP, glaciers were nearly as extensive as during the Little Ice Age. Maximum Holocene extents were reached two to three centuries ago during the final advances of the Little Ice Age. Overlord Glacier constructed moraines in the early AD 1700s, 1830s, 1890s, and 1920s.

The following general conclusions can be drawn from this study: (1) in situ wood in glacier forefields provides the most direct and unequivocal evidence for glacier fluctuations, but the record is fragmentary; (2) carefully chosen detrital wood may date glacier advances, but other interpretations are possible, thus supporting evidence is required; (3) proglacial lake sediments provide continuous records of upvalley glacier fluctuations, but changes in sediment availability and the lack of dating control can complicate interpretations; (4) long-term changes in the clastic content of proglacial lake sediments may reflect changes in glacier extent; and (5) the complex history of Holocene glacier fluctuations is best understood using a multi-proxy approach.

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