Dendrogeomorphic reconstruction of Little Ice Age paraglacial activity in the vicinity of the Homathko Icefield, British Columbia Coast Mountains, Canada

Sarah J. Hart a, John J. Clague b, Dan J. Smith a,⁎

a University of Victoria Tree-Ring Laboratory, Department of Geography, University of Victoria, Victoria, British Columbia, Canada V8W 3R4
b Department of Earth Sciences, Simon Fraser University, Burnaby, British Columbia, Canada V5A 1S6

1. Introduction

Glaciers throughout the British Columbia Coast Mountains are downwasting and receding (Schiefer et al., 2007; Vanlooy and Forster, 2008). This deglacialization has been accompanied by an increase in the number and severity of floods and other catastrophic events in many mountain valleys (Evans and Clague, 1994). In the southern British Columbia Coast Mountains, failures of moraine and glacier dams during the 1980s and 1990s caused severe downstream flooding along the Homathko River and its tributaries (Fig. 1; Blown and Church, 1985; Clague and Evans, 2000; Kershaw et al., 2005). The largest of the floods, which occurred on August 12, 1997, was triggered by the collapse of the lower part of Diadem Glacier into proglacial Queen Bess Lake (Clague and Evans, 2000; Kershaw et al., 2005). Waves induced by the massive ice avalanché overtopped and trenched the moraine dam, producing a flood that eroded Holocene-age sediments along the west fork of Nostetuko River. A series of sheets of silt and sand recording floodplain aggradation, separated by organic horizons containing tree stems and stumps in growth position, were exposed by the flood (Wilkie, 2006; Wilkie and Clague, 2009).

Streams can aggrade or incise their floodplains in response to changes in discharge, sediment load, and valley form (Leopold and Bull, 1979). In glaciated watersheds, changes in glacier cover can trigger floodplain aggradation or incision (Church and Ryder, 1972; Church and Slaymaker, 1989). Retreat of glaciers exposes a landscape that is susceptible to rapid change, due principally to transfers of large amounts of sediment from valley slopes to rivers. Church and Ryder (1972) coined the term “paraglacial” to define such a period of landscape adjustment and sediment transfer. Thirty years later, the paraglacial concept was extended by Ballantyne (2002a,b), who framed it in a context of sediment sources and sinks. He theorized that sediment transfers occur on different timescales (10 5 to 10 6 years), but at exponentially decreasing rates through time.

In glaciated basins, dendrogeomorphological research may provide insights into the activity of upstream glaciers on decadal to centennial timescales. Most prior dendroglaciological reconstructions in the Coast Mountains have relied on dating trees directly damaged or killed by glaciers (Smith and Laroque, 1996; Allen and Smith, 2007; Jackson et al., 2008). This approach, however, provides an incomplete record of past glacier activity because glacial advances of the past several hundred years destroyed or buried much of the evidence of earlier events (e.g. Koch et al., 2007). Dendrochronological dating of
paraglacial sedimentary sequences outside glacier forefields offers an opportunity to supplement and extend the direct glacial record — it provides a more complete, albeit less direct, record of climate and glacial activity (Wilkie, 2006).

In this paper, we identify sources and sinks of paraglacial sediment along the west fork of the Nostetuko River valley and develop an annually resolved chronology of late Holocene paraglacial aggradation events. These events are compared to independently documented periods of glacier activity in the northwestern North America Cordillera. Based on this comparison, we argue that aggradational events in the west fork of Nostetuko valley coincided with periods of glacial activity.

2. Study area

Our study area is the middle reach of the west fork of Nostetuko River valley (51° 17′ N., 124° 30′ W., ca. 1400 m asl) in the southern Coast Mountains of British Columbia (Fig. 1). The west fork of Nostetuko River originates at Diadem Glacier, an outlet glacier of the Homathko Icefield, and joins the main stem of Nostetuko River 8 km below Queen Bess Lake. Nostetuko River is a tributary of the Homathko River, which flows through the southern Coast Mountains to tidewater at the head of Bute Inlet (Fig. 1).

The west fork of Nostetuko River occupies a U-shaped valley in an area with more than 2000 m of relief. The river flows through two rock canyons and across several alluvial reaches between Queen Bess Lake and the main stem of Nostetuko River. The gradient of the west fork averages 3.8°, but ranges from 14° in the rock canyons to 0.5° in broad alluvial reaches (Kershaw, 2002). Lateral migration of the river is constrained by the steep valley slopes, rock canyons, and colluvial aprons and fans (Kershaw, 2002; Wilkie and Clague, 2009).

The study area has cold wet winters and warm dry summers. Climate normals (1961–1990) indicate that the valley has an average annual temperature of about 0.7 °C; mean monthly temperatures range from −8.8 to 9.7 °C (Wang et al., 2006). Average annual precipitation is about 1180 mm/year, of which approximately 70–75% falls as snow from October through May.

The valley lies within the Engelmann Spruce–Subalpine Fir (ESSF) bioclimatic zone (Meidinger and Pojar, 1991) and contains mixed stands of Engelmann spruce (Picea engelmannii), subalpine fir (Abies lasiocarpa), and whitebark pine (Pinus albicaulis). Stands of mature trees are restricted to the valley floor and to the surface of several large alluvial fans and colluvial aprons. Tree growth on steep valley walls is limited by episodic snow avalanches and debris flows. The upper limit of tree growth in the valley is ca. 1750 m asl.

3. Previous research

Kershaw (2002) and Wilkie (2006) described late Holocene valley-bottom sediments eroded by floodwaters from Queen Bess Lake in 1997. Their investigations provide insights into the paraglacial history archived within the valley-fill sediments, upon which our research builds.

The west fork of Nostetuko River emerges from a rock canyon 1.25 km below Queen Bess Lake and truncates a large terminal moraine constructed by the south lobe of Bess Glacier (unofficial name; Fig. 2). The glacier probably advanced many times during the Holocene to this position, creating this moraine. A branch collected from the base of the moraine yielded a radiocarbon age of 2790±60 14C year BP (Table 1; Wilkie and Clague, 2009), coincident with the regional Tiedemann Glacier advance reported by Ryder and Thomson (1986). A tree stem, exposed about 10 m below the surface of the moraine returned a radiocarbon age of 150±50 14C year BP (TO-8932; Table 1).

River bank exposures downvalley of the moraine contain in situ stumps and boles rooted in a series of buried soil horizons and peat layers up to about 4 m above present river level. Many peat layers are not continuous or differ in thickness, but they record periods of floodplain stability and soil development that are separated by units of valley-wide silt and sand up to 1 m thick deposited during aggradational episodes (Wilkie and Clague, 2009).

![Fig. 1. Locations of study sites within the west fork of the Nostetuko River valley (modified from Clague and Evans, 2000).](image-url)
Radiocarbon dating of the exhumed stumps suggests that valley-floor forests were repeatedly buried over the past 6000 years. Wilkie and Clague (2009) interpreted these burial events as responses to up-valley glacier advances associated with the Garibaldi phase (mid-Holocene), Tiedemann advance (ca. 3000 years ago), the First Millennium CE advance (ca. 1500 years ago), and the Little Ice Age of the past 1000 years.

4. Methods

Kershaw (2002) mapped landforms in the west fork Nostetuko Valley using large-scale (1:5000 and 1:10,000), aerial photographs. We used the distribution of these and other landforms to assess –25°). Alluvial fans were discriminated from more-common sources and sinks of sediment in the west fork Nostetuko Valley. We established a living tree-ring chronology from 40 increment core samples (20 trees, 2 cores each) collected from a mature stand of subalpine fir trees (>300 years in age) growing adjacent to the river on a valley-bottom alluvial fan. The cores were transported to the University of Victoria Tree-Ring Lab (UVTRL), where they were air-dried and then glued into slotted mounting boards. The cores were subsequently sanded to a 600-grit polish to enhance annual rings. We used a chain saw to collect cross-sections of subfossil trees within 95 km of the study site on the relatively dry leeward side of the Coast Mountains. We constructed a chronology by cross-dating the series using common marker years (Stokes and Smiley, 1964). The International Tree Ring Database (ITRDB) computer program COFECHA was subsequently used to verify block correlations between each tree-ring series and the master chronology to ensure absolute dating accuracy (Holmes et al., 1986). We removed those cores from the data set that exhibited aberrant growth patterns and were not significantly correlated to the group.

We strengthened and extended back in time the living tree-ring chronology by supplementing it with six cross-dated subalpine fir chronologies previously collected by UVTRL researchers in the region (Table 3). The regional chronology incorporates only tree-ring data collected within 95 km of the study site on the relatively dry leeward slopes of the Coast Mountains. We constructed the regional chronology using all series from the pooled tree-ring data and removing only those series that were not significantly correlated to the group.

4.1. Living chronology

We established a living tree-ring chronology from 40 increment core samples (20 trees, 2 cores each) collected from a mature stand of subalpine fir trees (>300 years in age) growing adjacent to the river on a valley-bottom alluvial fan. The cores were transported to the University of Victoria Tree-Ring Lab (UVTRL), where they were air-dried and then glued into slotted mounting boards. The cores were subsequently sanded to a 600-grit polish to enhance annual rings (Stokes and Smiley, 1964). The ring widths of each sample were measured to the nearest 0.01 mm along a single radius using WinDendro software and a high resolution flatbed scanner (Guay et al., 1992). A Velmex stage equipped with a microscope and video display was used with the computer program MeasureJ2X (VoorTech Consulting, 2008) to measure widths of exceptionally narrow rings.

We conducted a chronology by cross-dating the series using common marker years (Stokes and Smiley, 1964). The International Tree Ring Database (ITRDB) computer program COFECHA was subsequently used to verify block correlations between each tree-ring series and the master chronology to ensure absolute dating accuracy (Holmes et al., 1986). We removed those cores from the data set that exhibited aberrant growth patterns and were not significantly correlated to the group.

We strengthened and extended back in time the living tree-ring chronology by supplementing it with six cross-dated subalpine fir chronologies previously collected by UVTRL researchers in the region (Table 3). The regional chronology incorporates only tree-ring data collected within 95 km of the study site on the relatively dry leeward slopes of the Coast Mountains. We constructed the regional chronology using all series from the pooled tree-ring data and removing only those series that were not significantly correlated to the group.
4.2. Subfossil dendrogeomorphology

We constructed floating tree-ring series from the subfossil cross-sectional samples by measuring ring widths along at least two radii. Care was taken to use paths with the maximum number of rings and to avoid areas with reaction wood. Companion series from each subfossil sample were cross-dated using CDendro 7.1 (Larsson, 2003).

We built species-specific floating chronologies from subfossil samples rooted in the same organic layer. The species of each sample was determined using a simple visual identification key and a light microscope (Hoadley, 1990). We calculated correlation coefficients for the floating chronologies as an average linear correlation between each series and every other series. We used this method because it provides conservative estimates of the year-to-year agreement between members of the chronologies.

Attempts were made to cross-date individual floating chronologies to the local and regional living chronologies using CDendro 7.1 (Larsson, 2003). Cross-dating was deemed successful based on high correlation coefficients, t-statistics, and visual similarity between the floating chronology and a dated chronology. Where successful, cross-dating provided absolute death dates. Where this approach was unsuccessful, radiocarbon ages of samples included in the floating chronology or located on the same organic horizon provided approximate age constraints for the burial event (Koch et al., 2007). We assume an ecesis interval of 5 years (Wilkie, 2006).

When flooded, conifers characteristically undergo changes in ring width (Sigafouos, 1964; SigafosFriedman et al., 2005; Mizugaki et al., 2006) or produce adventitious roots (Kozlowski et al., 1991). The absence of adventitious roots and suppression of growth at the end of the trees’ lives indicate that most aggradation events were fatal, a less common but still documented response (Strunk, 1988). We interpret periods of abnormal outer ring growth to represent the beginning of the aggradation period.

5. Results

5.1. Sediment sources and sinks

Primary sediment sources in the west fork Nostetuko valley include rockwalls and glacigenic sediments (Fig. 3). Pleistocene and modern glaciers have steepened rock slopes, which have increased tensile stresses within the slopes and facilitated slope failure and downslope transfers of sediment. The glacier forefields are areas of abundant unconsolidated glacigenic sediment that have only incipient soil and scanty vegetation. During periods when glaciers are advancing or retreating, some of this sediment is mobilized by a variety of processes – debris flows, snow avalanches, fluvial erosion, and periglacial processes – and carried downvalley by the west fork of Nostetuko River. The forefields of glaciers, including Diadem and Bess glaciers, are important sources of sediment to the fluvial system. In particular, the Nostetuko west fork flows past the Little Ice Age terminal moraines of Diadem and Bess glaciers, eroding both at times of high discharge (Fig. 3; Kershaw, 2002).

The major sediment sinks in the west fork of the Nostetuko River valley are debris cones, composite alluvial fans, and the alluvial fill on the valley floor (Fig. 3). Wilkie and Clogue (2009) documented aggradation units over 1 m thick covering the valley floor. Other important sediment sink are five composite fans and one debris cone on the valley sides within the study area. The largest fans are on the western, more glacierized side of the valley. These features are largely depositional, but may also represent important secondary sources of sediment for later paraglacial events (Fig. 3).

5.2. Tree-ring chronologies

The local living subalpine fir chronology comprises 29 series from 18 trees and spans 279 years (CE 1729 to 2007; Table 4). The mean series intercorrelation value, calculated using 50-year segments, lagged successively by 25 years, is significant at the 95% confidence level ($r = 0.444$, Table 4).

The composite regional chronology consists of over 250 series and extends the tree-ring record back to CE 1572 (Table 4). COFECHA revealed a high degree of agreement between series (Table 4). The chronology, however, loses significant sample depth before CE 1725 (Fig. 4). Statistics calculated on the regional and site chronologies indicate similar sensitivities and intercorrelation values (Table 4). A strong correlation ($r = 0.53$) between the two chronologies was found using CDendro after detrending both average raw ring width series with a negative exponential curve.

5.3. Subfossil samples

We sampled 135 subfossil trees on the Nostetuko valley floor. Most of the samples were found in growth position, rooted in organic layers abruptly overlain by thick layers of silt and sand. The samples were identified as Engelmann spruce, subalpine fir, or whitebark pine. None of the 135 trees sampled showed evidence of adventitious roots. Periods of suppressed radial growth, traumatic resin canals, and reaction wood were observed in some samples, but they were restricted to the end of the ring series. We interpret the absence of adventitious roots and suppression of growth at the end of the trees’ lives to mean that catastrophic aggradation caused the death of the trees.

Of the 135 samples, only 57 were suitable for dendrochronological analysis (17 whitebark pine, 32 subalpine fir, and 8 Engelmann spruce samples). Many of the samples were deemed unsuitable for analysis because they contained too few rings. Other samples could not be used because the tree-ring series were too complacent or contained too much reaction wood.

Site A is the fluvially eroded face of the large moraine fan below Bess Glacier (Fig. 2; Table 2). A laterally continuous mat of stumps, boles, and branch fragments is exposed ca. 10 m below the surface the fan. The downvalley orientation and character of the detrital wood indicate that the trees were killed and buried during an advance of

<table>
<thead>
<tr>
<th>Site</th>
<th>Interval</th>
<th>No. series</th>
<th>Average series intercorrelation</th>
<th>Average mean sensitivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Local</td>
<td>1792–2007</td>
<td>29</td>
<td>0.444</td>
<td>0.214</td>
</tr>
<tr>
<td>Regional</td>
<td>1572–2007</td>
<td>257</td>
<td>0.513</td>
<td>0.214</td>
</tr>
</tbody>
</table>

*Calculated using computer program COFECHA; default setting of 50-year segments with a 25-year overlap.*
Bess Glacier into the valley. The perimeter rings of a bole recovered from the site in 2004 yielded a radiocarbon age of $150 \pm 60$ 14C year BP (TO-8932; Table 1). When visited in 2008, further erosion and collapse of the face had exposed additional detrital wood.

Cross-sections were collected from nine subfossil subalpine fir samples found within the moraine face and on the talus below. The oldest sample contained 97 annual growth rings. No bark was present on the samples, and some perimeter rings were assumed missing because of weathering. Three of the nine samples cross-date to form a 76-year-long floating chronology (float VII; Table 5). This chronology cross-dates to the regional chronology and suggests that in ca. CE 1807 Bess Glacier was advancing over a moraine surface likely first colonized by trees about CE 1731 (Fig. 4). Assuming that the local ecesis interval is five years and that the oldest tree sampled was the first colonizer of a recently deglaciated surface, we infer that Bess Glacier had retreated some distance upvalley by the mid-1720s.

Site B is the eroded face of a fluvial terrace on the east side of the river across from site A (Fig. 1; Table 2). In situ roots and stumps are associated with a buried organic layer located about 3 m below the terrace surface. The outer rings of an in situ root yield a radiocarbon age of $370 \pm 50$ 14C year BP (TO-8932; Table 1). We examined another stump, collected in 2004 by Wilkie (2006). Accounting for ecesis, the minimum period of stability recorded by the organic layer at site B is 107 years.

Site C is located on the west side of the valley, about 295 m north of site A and adjacent to a small meltwater stream that eroded by the 1997 outburst flood scouring an alluvial fan. Erosion during the 1997 outburst flood scoured an elongate depression, exposing a thick sequence of alternating facies of stratified silt, sand, and peat.

Site D is located at the SE edge of the depression close to the forest. Site E is located at the NW edge of the abandoned floodway. The lowest exposed organic horizon at site D was radiocarbon-dated to $940 \pm 50$ 14C year BP (Beta-200728; Table 1). The outer rings of a stump rooted in another peat less than 1 m higher in the sequence yielded a radiocarbon age of $520 \pm 50$ 14C year BP (Beta-200727; Table 1). This finding and the presence of several other buried forest layers higher in the sequence suggest that trees repeatedly colonized overbank sediments on the valley floor at this site.

Forty-two cross-sections of stumps were collected at sites D and E. Two in situ stumps of subalpine fir in the lowest organic horizon cross-date and form a 160-year-long floating chronology (float III; Table 5). The trees were killed and buried by silt and sand in 940 $\pm 50$ 14C year BP. It is inferred that this event was followed by a lengthy interval without significant aggradation, as a subalpine fir stump with 152 rings, slightly higher in the sequence, dates to $520 \pm 50$ 14C year BP (Table 1).

Site F is located 100 m downhill from sites D and E at the north side of a forested bedrock knob (Fig. 1). Sand separates two laterally extensive organic horizons, each of which contains subalpine fir stumps in growth position. The lower organic horizon is at river level, about 3 m below the surface. Two cross-sections of these stumps cross-date to form a 180-year-long floating chronology (float V; Table 5) with an outer perimeter wood age of 600 $\pm 60$ 14C year BP (Table 1).

The upper organic horizon is buried by ca. 1 m of fine sand (Fig. 5). Two cross-sections of stumps from this layer cross-date to form a 202-year-long floating chronology (float VI; Table 5). This floating chronology cross-dates with the sample from site B; thus a perimeter date of $370 \pm 50$ 14C year BP was assigned to both (Table 5).

Site G is an exposure of silt and sand eroded by the 1997 outburst flood and located 315 m downhill from site F (Fig. 1). A buried peat layer $\approx 2$ m below the surface contains in situ stumps. Cross-sections of two stumps radiocarbon-dated by Wilkie (2006; KW15 and KW18)
cross-date with samples from site H (float IV; Table 5). This result indicates that the trees at site G were killed in ca. 620 ± 50 14C year BP (Table 1), after a period of floodplain stability lasting at least 165 years.

Site H is a riverbank exposure 50 m downstream from site G (Fig. 1; Table 2), along a side channel. Exhumed stumps are exposed in the side of the riverbank (Fig. 6). Two cross-sections of in situ stumps collected from the lowest organic horizon ca. 3.5 m below the surface cross-date to form a floating chronology. They also cross-date with samples from site I to form a 142-year-long floating chronology (float II, Table 5) with a perimeter radiocarbon age of 1030 ± 50 14C year BP (Beta-200731; Table 1). These results suggest that the trees growing at site H were killed and buried in 1030 ± 50 14C year BP following a period of floodplain stability that lasted at least 142 years (Table 5).

Six samples were collected from stumps rooted in a second woody layer about 2.5 m below the surface. Three samples (two subalpine fir and one Engelmann spruce) cross-date with samples from site G (float IV, Table 5). This finding suggests that these trees were killed 620 ± 50 14C year BP after a period of floodplain stability lasting at least 245 years (Table 5).

A third organic layer is about 1 m below the present-day ground surface. It contains snags that extend above ground level. One of the four stumps sampled from this horizon cross-dates with the regional subalpine fir chronology and has a death date of CE 1657. This chronology extends back 152 years, indicating the woodplain was stable for at least 126 years, after which an episode of sediment aggradation buried and killed the trees at sites J and K shortly after 1160 ± 50 14C year BP.

Site K is located 60 m downstream from site J adjacent to the main river channel and is being actively eroded by the river (Fig. 1). A prominent nonneroded peat layer which gave a radiocarbon age of 1160 ± 50 14C year BP (Table 1) occurs at water level. Five cross-sections were collected from stumps rooted in the peat layer. One sample cross-dates with a 94-year floating chronology at site J (float I; Table 5). We infer that the floodplain was stable for at least 126 years, after which an episode of sediment aggradation buried and killed the trees at sites J and K shortly after 1160 ± 50 14C year BP.

6. Synthesis

Significant geomorphic change occurred on the valley floor of the west fork of Nostetuko River during the Holocene. This change occurred within a context of the episodic advance and retreat of glaciers from upper valley positions to the edge of the trunk valley. Of greatest importance is the late Holocene delivery of sediment by Diadem and Bess glaciers. We identified and dated six aggradation events during the past 1200 years on the basis of overlapping radiocarbon ages and stratigraphic evidence (Fig. 4; Table 6). The oldest event (no. 1) occurred CE 718–990, based upon the radiocarbon age of 1160 ± 50 14C year BP and floating chronology I from an organic horizon within the valley-fill sedimentary sequence (Fig. 7). The chronology overlaps chronologies II and III in radiocarbon time, but none of the three chronologies cross-dates. The three organic layers

Table 6
Aggradation events.

<table>
<thead>
<tr>
<th>Event</th>
<th>Calibrated age range (CE)</th>
<th>Radiocarbon age (14C year BP)</th>
<th>Floating chronologies</th>
<th>Cross-dated chronologies</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>718–990</td>
<td>1160 ± 50</td>
<td>I</td>
<td>08A002</td>
</tr>
<tr>
<td>2</td>
<td>892–1209</td>
<td>940 ± 50</td>
<td>II</td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>1284–1424</td>
<td>1030 ± 50</td>
<td>III</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>1445–1637</td>
<td>600 ± 60</td>
<td>IV</td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>1657</td>
<td>620 ± 50</td>
<td>V</td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>1669–1954</td>
<td>110 ± 60</td>
<td>VI</td>
<td>08A108</td>
</tr>
<tr>
<td></td>
<td>1794</td>
<td></td>
<td></td>
<td>08A134</td>
</tr>
</tbody>
</table>
differ significantly in character, suggesting that they are not of the same age. This event appears to coincide in time with a well-documented advance of Bridge Glacier, 90 km to the SE, at 1190 ± 60 14C year BP (Allen and Smith, 2007).

Event 2 occurred at about CE 892–1153 (940 ± 50 14C year BP; Table 6). It is recorded at three sites in the valley (Fig. 7). An early Little Ice Age advance of this age has been identified at numerous sites in the southern Coast Mountains (Reyes and Clague, 2004; Allen and Smith, 2007; Koch et al., 2007) and in the northern Coast Mountains (Haspel et al., 2005; Spooner et al., 2005).

Event 3 occurred in the late fourteenth century (600 ± 60 14C year BP; Table 6) and is recorded by aggradation at valley-floor sites E, G, and H (Fig. 7). This time period is widely recognized as one when glaciers were advancing throughout the southern Coast Mountains (Ryder and Thomson, 1986; Larocque and Smith, 2003; Lewis and Smith, 2004; Allen and Smith, 2007; Koch et al., 2007).

Event 4 is a valley-wide aggradational episode that occurred about CE 1445–1637 (370 ± 50 14C year BP; Fig. 4, Table 6). Evidence for this event was found at sites B and F (Fig. 7). These sites are located within alluvial fan sediments spilling from the northern lobe Bess Glacier and within downstream floodplain sediments. Glaciers advances in the early to mid-1500s have been reported in the Mount Waddington area, 50 km to the NW (Larocque and Smith, 2003).

A sheet of sediment deposited during event 5 dates to CE 1657 at site H (Fig. 4; Table 6). Trees growing at site C are believed to have colonized a mineral surface ca. CE 1660 (sample 08A134; Fig. 6). This observation suggests that this event may have been short lived. Moraines elsewhere in the region stabilized shortly after this time, in the mid to late 1660s (Larocque and Smith, 2003; Lewis and Smith, 2004).

Aggradation during event 6 dates to ca. 1794 at site C (Fig. 4; Table 6). A radiocarbon-dated stump from site H may have been buried during event 6 (Table 6). This event is contemporaneous with advances of other glaciers in the southern Coast Mountains during the late Little Ice Age (Larocque and Smith, 2003; Menounos et al., 2009).

7. Discussion

All six aggradation events documented in this study correspond to periods of regional glacial activity. Dated samples were recovered from peat layers separating thick silt and sand units that record valley-wide aggradation. The general coincidence between aggradation events and periods of independently documented regional glacier activity is characteristic of a paraglacial system. We infer from this synchronous behavior that sediment from glacier forefields, moraines, and valley slopes was delivered at higher-than-normal rates during times when climate was cooler and wetter, and thus glaciers more extensive, than today.

We compared tree death dates determined in this study to local dates of glacier activity to determine at what point in the glacial cycle the west fork of the Nostetuko River aggraded its valley and killed valley-bottom forests. In the mid-1720s, Bess Glacier was some distance inside its maximum Little Ice Age limit, whereas in 1806 it was advancing toward this position (Fig. 2). The aggradation events closest in time to this dated activity are events 5 (CE 1657) and 6 (CE 1794). Given that small watersheds will respond to glacier fluctuations more rapidly than large watersheds (Harbor and Warburton, 1993), we infer that aggradation event 6 occurred in response to the advance in 1806. The temporal and spatial relations between the advance and aggradation suggest that the latter may occur prior to glaciers reaching their maximum extent during any given advance. Aggradation events appear to coincide with, rather than lag, periods of regional glacial advance and moraine stabilization. We caution, however, that climate and glaciers in British Columbia have fluctuated on a range of timescales (10^1 to 10^3 years; Clague et al., 2009). This variability and inherent limitations in our ability to resolve prehistoric glacier activity on timescales of years complicate
our comparison of aggradation events and glacier activity in the Nostetuko River watershed. More research at high temporal and spatial resolution is needed to understand these relationships.

8. Conclusions

We constructed a regional subalpine fir chronology for the period CE 1572–2007 from samples collected at seven sites in the southern Coast Mountains. This chronology was used to cross-date in situ, subfossilized trees bounded by sheets of silt and sand in the valley of the west fork of the Nostetuko River. Some of the exhumed forest horizons predate the regional chronology. In these cases, floating chronologies were pinned with radiocarbon ages on outer rings of subfossil trees in the valley. Using these two methods, we documented and dated six, spatially extensive episodes of aggradation in the Nostetuko River valley during the past 1200 years. The six sedimentation events coincide with independently dated periods of glacier expansion in the Coast Mountains and reflect increased sediment delivery to the west fork of the Nostetuko River by Bess, Queen Bess, and other glaciers within the upper part of the watershed.

Most previous dendrochronological reconstructions of Holocene glacial history in the Canadian Cordillera have focused on trees that have been directly impacted by glaciers. This approach only captures some of the most recent glacier fluctuations, because the climatic advances of the late Little Ice Age buried or destroyed much of the evidence of earlier advances. Sediment supply to rivers draining glaciated catchments may change in response to an increase or decrease in ice cover. In some cases, the response is manifested in aggradation or incision of floodplains. As these environments lie outside Little Ice Age glacier limits, they may yield a more complete record of Holocene glacier activity than the evidence from glacier forefields themselves.

Acknowledgements

Kirsten Brown, Bethany Coulthard, Kate Johnson, and Kyla Patterson helped in the field. Sandy Allen, Lynn Koehler, and Sonya Larocque allowed us to use their chronologies in building our regional chronology. Kenna Willie provided radiocarbon ages on subfossil wood. Mike King (White Saddle Air Services) provided helicopter access to the study area. We also thank Dennis Jelinski, Terri Lacourse, and Vic Levenson for their comments and scientific insight. This research was supported by the Natural Sciences and Engineering Research Council of Canada (NSERC) Discovery Grants to Clague and Smith and by a Canadian Foundation for Climate and Atmospheric Sciences (CFCAS) grant to the Western Canadian Cryospheric Network.

References


Author's personal copy


