Latest Pleistocene and Holocene behaviour of Franklin Glacier, Mt. Waddington Area, British Columbia Coast Mountains, Canada

by

Bryan Joel Mood
B.Sc., Mount Allison University, 2013

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Requirements for the Degree of

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in the Department of Geography

Supervisory Committee

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Abstract

Holocene climate variability in the British Columbia Coast Mountains has resulted in repeated intervals of glacier expansion and retreat. Since reaching their late Holocene maximum positions in the late 20th century, glaciers in the region have experienced significant volumetric loss. The subsequent downwasting and frontal retreat has revealed forests buried by glacier advances throughout the Holocene, enabling description of significant intervals of ice expansion using dendroglaciology. This thesis characterizes dendroglaciological evidence as it relates to climate at two scales: (1) at Franklin Glacier in the Mt. Waddington area, and; (2) throughout the Coast Mountains.

Dendroglaciological evidence from glacier forefields and lateral moraines in the Coast Mountains provides evidence for at least 11 intervals of glacier activity during the Holocene. The earliest recorded glacier activity is documented in the Pacific Ranges from 8.5 to 8.2 ka, after which glaciers in this region retreated during the early Holocene warm and dry interval. Following this period, glacier activity occurred from 7.3-5.3 ka in response to attendant cool and moist conditions in the Pacific Ranges. After 5.3 ka, glaciers in the Pacific Ranges exhibit a near-continuous record of activity from 4.8-2.5 ka and in the Boundary Ranges at 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.3 during intervals characterized wet conditions resulting from an intense, eastwardly positioned Aleutian Low pressure centre. Glaciers were again expanding downvalley in the Pacific and Boundary ranges from 1.4-1.2 and 1.7-1.1 ka, respectively before contemporaneous activity from 0.8-0.4 ka during the Little Ice Age. Common intervals of glacier activity throughout the Coast Mountains occurred at 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.5, 1.4-1.2, and 0.8-0.4 ka.

Franklin Glacier is an 18-km long valley glacier that originates below the west face of Mt. Waddington. Radiocarbon-dated wood samples from the proximal faces of lateral moraines flanking the glacier show that it expanded at least nine times since 13 ka. A probable Younger Dryas advance of Franklin Glacier at 12.8 ka followed the late glacial retreat and downwasting of the Cordilleran Ice Sheet from ca. 16.0 to 12.9 ka. During the succeeding early Holocene warm period, Franklin Glacier appears to have retreated significantly, leaving no record of glacial expansion until the mid-Holocene when it repeatedly advanced at 6.3, 5.4, and 4.6 ka in response to cool summer temperatures and generally moist conditions. Downwasting of the glacier surface after 4.6 ka was followed by intervals of expansion at 4.1, 3.1, and 2.4 ka contemporaneous with a period of increased precipitation. Following ice expansion at 2.4 ka into trees over

224 years in age, there is no record of the glacier activity until 1.5 ka when Franklin Glacier thickened and advanced into young subalpine fir trees, reflecting attendant cool and wet environmental conditions. During the Little Ice Age, advances at 0.8 and 0.6 ka preceded a mid-19th to early-20th century advance that saw Franklin Glacier attain its maximum Holocene extent in response to an extended interval of cold temperatures.

The dendroglaciological record at Franklin Glacier is among the most comprehensive recovered from the British Columbia Coast Mountains and showcases the complexity of latest Pleistocene and Holocene glacier behaviour in the region.

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Dedication

This thesis is dedicated to my brother, Dallas, and Kaitlin.

Chapter One: Introduction

1.1 Introduction

Glaciers in the British Columbia Coast Mountains have experienced substantial volumetric losses since the end of the Little Ice Age (LIA) in response to warming temperatures (Larocque and Smith, 2003, 2005a; Koch et al., 2007a; Schiefer et al., 2007; VanLooy and Forster, 2008; Wood and Smith, 2013; Tennant et al., 2012). The resultant downwasting and frontal retreat has exposed the remains of overridden forests entrained in sediment deposits that provide an opportunity to describe glacier behaviour using dendroglaciological techniques. Menounos et al. (2009) summarize the outcome prior dendroglaciological research and report that there were expansion events at 8.6-8.2, 7.4-6.4, 4.4-4.0, 3.5-2.8, 1.7-1.3 ka, followed by LIA terminal fluctuations linked to climatic changes at regional- and global-scales.

The purpose of this thesis is to enhance our understanding of Holocene glacier behaviour by systematically reviewing the existent literature and completing field investigation at a little studied glacier in the Mt. Waddington region. This thesis has three specific objectives:

1. To interpret Holocene activity at Franklin Glacier using dendroglaciological and radiocarbon dating techniques.

2. To compare the findings at Franklin Glacier to existing records of Holocene glacier fluctuations in the Coast Mountains.

3. To explore the underlying climates that influenced activity at both Franklin Glacier and other Coast Mountains glaciers during the Holocene.

1.2 Thesis Outline

This thesis is organized into four chapters. This chapter is followed by Chapter Two which summarizes published and unpublished research on the behaviour of glaciers in the Coast Mountains over the last 11,700 years. Chapter Three describes the outcome of dendroglaciological investigations at Franklin Glacier that were used to detail it's latest Pleistocene and Holocene behaviour. Chapter Four summarizes the findings of the thesis and offers directions for future research.

Chapter Two: A review of Holocene glacier expansion in the British Columbia Coast Mountains, Canada

2.1 Introduction

The Holocene glacial history of the British Columbia Coast Mountains, Canada, is characterized by repeated episodes of ice expansion and retreat. Menounos et al. (2009) describe six intervals of glacier expansion in western Canada between 8.6 and 1.3 ka, followed in the past millennia by the Little Ice Age (LIA) during which most glaciers in this region attained their maximum Holocene extent. Recent investigations augment our understanding of the regional character of these events and provide evidence of additional glacier activity in the mid- to late Holocene (Harvey et al., 2012; Coulthard et al., 2013; Craig and Smith, 2013; Hoffman and Smith, 2013; Osborn et al., 2013; Mood and Smith, 2015). Collectively, these reports serve to emphasize the complex and sensitive mass balance response of glaciers in this region to large-scale climate fluctuations (Bitz and Battisti, 1999; Yarnal, 1984; Wood et al., 2011).

In this paper, previously reported and unreported dendroglaciological and radiocarbon evidence are integrated to provide a regional assessment of Holocene glacial activity in the Coast Mountains. The summary results provide detailed insights into the mixed response of glaciers to past and current natural climate variability, and thereby supplement both the previous review by Menounos et al. (2009) and the more recent global review of Holocene glacier activity by Solomina et al. (2015). In addition, the findings offer a paleoglaciological record useful for verifying and standardizing the extended paleoclimate and sediment deposition records emerging from related research in

British Columbia (e.g., Dallimore et al., 2005; Gavin et al., 2011; Galloway et al., 2010, 2013; Babalola et al., 2013; Steinman et al., 2014).

2.2 Study Area and Research Background

The Coast Mountains are situated in western British Columbia, adjacent to the Pacific Ocean (Figure 2.1). They are comprised of granitic igneous and metamorphic rocks, and their form is largely the result of Cenozoic tectonic uplift, fluvial and glacial erosion, and mass wasting (Mathews, 1991). The Coast Mountains are subdivided into the southern Pacific Ranges, the central Kitimat Ranges and the northern Boundary Ranges (Mathews, 1986; Church and Ryder, 2010). The Boundary Ranges extend from the Nass River, near the southern end of the Alaska Panhandle, northward into the Yukon Territory. Glacier history surveys have been conducted at the Juneau, Andrei, Frank Mackie, Todd, and Cambria icefields (Figure 2.1). The Kitimat Ranges are located between the Nass River valley on the north and the Bella Coola River valley on the south (Figure 2.1). The Pacific Ranges extend from the Bella Coola River valley on the north to the Fraser River valley on the south. Dendroglaciological surveys have been completed in the Monarch, Ha-Iltzuk, Homathko, and Lillooet icefields, as well as the Mt. Waddington area and Garibaldi Provincial Park (Figure 2.1). The Coast Mountains were repeatedly glaciated during the Pleistocene (James and Clague, 2002; Collins and Montgomery, 2011; Roed et al., 2013). The latest period of glaciation, referred to as the Fraser Glaciation, was underway by 30 ka, and by 25 ka eastward-flowing trunk valley glaciers coalesced to form the Cordilleran Ice Sheet (CIS) on the Interior Plateau (James and Clague, 2002).

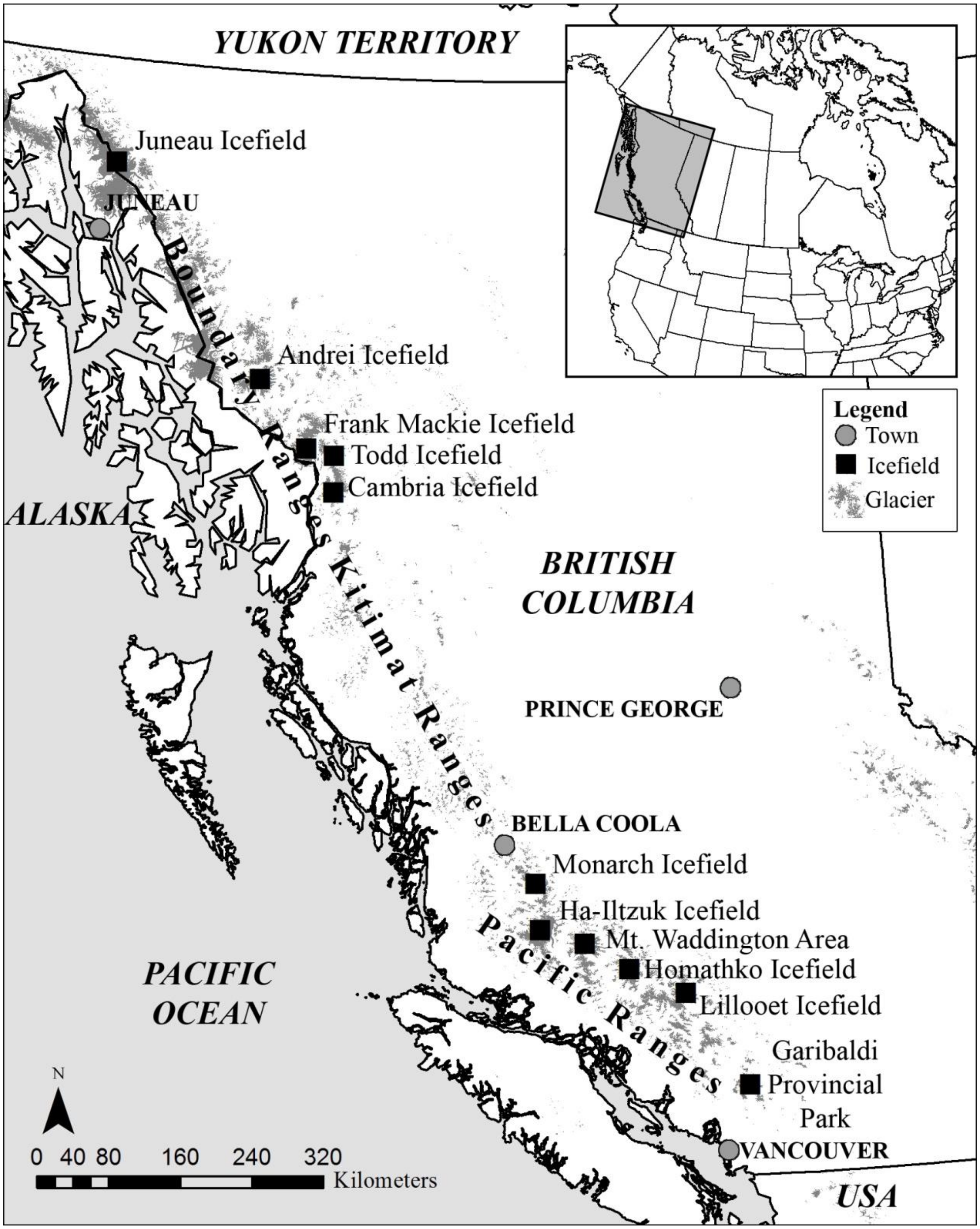

Figure 2.1 – Coast Mountains study area highlighting glacier and icefield locations.

By 16-17 ka the CIS reached its maximum thickness and extent, with a reversal in flow direction seeing westward flowing ice overwhelming most Coast Mountain summits (Mathews, 1986; Stumpf et al., 2000; Menounos et al., 2009). The CIS then began to downwaste and glaciers were increasingly restricted to trunk valleys, leaving most summits ice-free by 13.0 ka (Friele and Clague, 2002a; James and Clague, 2002; Margold et al., 2013). During subsequent Younger Dryas cooling at 12.9 ka, tributary glaciers began to thicken and readvance down valley, before many disappeared entirely during the warm and dry conditions of the early Holocene (Friele and Clague, 2002a; Menounos et al., 2009).

Holocene climate changes led to repeated intervals of cirque and valley glacier expansion that highlight the persistent, but variable, influence of Pacific ocean-atmosphere teleconnections on glacier activity in this region (Yarnal, 1984; Bitz and Battisti, 1999). Prominent expansion events recorded at 8.6–8.2, 7.4–6.4, 4.4–4.0, 3.5–2.8, 1.7–1.3 and 0.7-0.1 ka are represented at many locations in the Coast Mountains (Menounos et al., 2009). These events broadly coincide with intervals of reduced summer insolation in the northern Hemisphere and large volcanic eruptions (Renssen et al., 2006; Sigl et al., 2013; Solomina et al., 2015).

At present, the majority of large glaciers and icefields in the Coast Mountains are located in the Boundary and Pacific ranges (Figure 2.1). Most glaciers in the Pacific Ranges are between 1 to 50 km^2 in area, whereas in the Boundary Ranges 40% of the glaciers are greater than 100 km^2 in area (Bolch et al., 2010). The southern Boundary Ranges support the Juneau, Andrei, Frank Mackie, Todd, and Cambria icefields (Figure

2.1), and in the Pacific Ranges the Monarch, Ha-Iltzuk, Whitemantle, Homathko, and Lillooet icefields (Figure 2.1). In both areas, large outlet glaciers flow from icefields into deeply entrenched valleys where multiple Holocene expansion and retreat events have left prominent terminal and lateral moraine complexes (e.g., Jackson et al., 2008; Lakeman et al., 2008; Craig and Smith, 2013).

In recent decades many Coast Mountain glaciers experienced significant retreat and downwasting (VanLooy and Forster, 2008), with the estimated loss in glacier area exceeding 2100 km^2 from 1985 to 2005 (Bolch et al., 2010). The greatest areal losses have been in the Kitimat Ranges (21.8%), followed by the Pacific Ranges (10.3%) and the Boundary Ranges (7.7%) (Bolch et al., 2010). In the Pacific Ranges many cirque and valley glaciers are now smaller in extent than during the mid-Holocene (Harvey et al., 2012; Koch et al., 2014; Mood and Smith, 2015).

Low-frequency climate oscillations originating in the Pacific Ocean strongly influence Coast Mountain hydroclimates (Kiffney et al., 2002; Barnett et al., 2005; Stahl et al., 2006; Bonfils et al., 2008; Starheim et al., 2013). Since the last glacial maximum, fluctuations in the intensity and position of the Aleutian Low pressure centre, described by the Pacific Decadal Oscillation (PDO), have modulated the glacial climate of this region (Moore and Demuth, 2001; Mann et al., 2005; Wood et al., 2011; Galloway et al., 2013). Prolonged intervals of wet, cool conditions leading to glacier expansion tend to be associated with increases in the intensity of the Aleutian Low and an eastward shift of the pressure centre during periods of decreased solar activity (Christoforou and Hameed, 1997; Spooner et al., 2003; Mann et al., 2005; Shen et al., 2006). Glacier retreat and downwasting generally accompany the persistence of a warm and/or dry climate regime established following northwestward shifts of the Aleutian Low pressure centre (Heusser et al., 1985; Babalola et al., 2013).

2.3 Methodologies

Evidence of Holocene glacier activity in the Coast Mountains comes from several sources. Trees killed and buried during periods of glacier expansion provide a direct indication of activity (Coulthard and Smith, 2013a). While corroborative insights into the absolute and relative timing of glacial events in this region have been derived from sediment cores (e.g., Menounos et al., 2004, 2008; Arsenault et al., 2007; Galloway et al., 2010, 2011, 2013; Gavin et al., 2011; Steinman et al., 2012, 2014; Roe et al., 2013) and lichenometric studies (e.g., Smith and Desloges, 2000; Larocque and Smith, 2003; Koch 2009; Harvey and Smith, 2012), attention is focused here on the findings of dendroglaciological analyses completed within glacial and ice-proximal deposits (e.g., Ryder and Thomson, 1986). In the following section I outline dendroglaciological sampling techniques and potential sources of error. More comprehensive descriptions are provided by Ryder and Thomson (1986), and Coulthard and Smith (2013a).

Advancing glaciers often override and kill standing forests growing on valley floors and side walls, with the tree remains buried and preserved beneath till or proglacial stream deposits. Following deglaciation, these remains are exposed within the proximal faces of collapsing lateral moraines (e.g., Reyes and Clague, 2004; Hoffman and Smith, 2013; Osborn et al., 2013) or are revealed by stream and gully erosion of valley-bottom deposits (e.g., Allen and Smith, 2007; Koch et al., 2007; Koehler and Smith, 2011). In these settings dendroglaciological studies of rooted, glacially-sheared, stumps or detrital boles found pressed into buried paleosols are used to document intervals of glacier expansion (Smith and Koehler, 2011). In situ stumps provide the most reliable dating control, as these remains are still in their original growth position (Coulthard and Smith, 2013a). Re-transported detrital wood that is not found in growth position is less reliable for dating glacier advances, as it is out of context and may have been entrained from older deposits (Ryder and Thomson, 1986; Menounos et al., 2008) or represent the remains of avalanche-transported trees overridden by subsequent ice expansion but may represent a limiting maximum age on glacier advance (Jackson et al., 2008). Multiple units of subfossil wood samples within a stratigraphic sequence may be used to determine a chronology of glacier activity over time (e.g., Reyes and Clague, 2004; Jackson et al., 2008). Only in situ sheared stumps or wood mats/boles pressed into paleosols may provide a kill date for glacier activity although they do not account for glacier response times to climate perturbations and may post-date these events. Other samples may be the result of mass wasting events. The type of dendroglaciological sample provides possible sources of error but others still exist. Bark and perimeter wood should be dated in order to provide an accurate kill date but this is not always possible (Coulthard and Smith, 2013a). The presence of neither increases the error of dendroglaciological samples.

Kill dates are assigned to dendroglaciological samples by tree-ring cross dating or radiocarbon analyses (Luckman, 1998; Coulthard and Smith, 2013a). If the annual rings in a dendroglaciological sample can be cross-dated to a dated tree-ring chronology, an absolute kill-date can be established using standard dendrochronologic techniques (Fritts, 1976; Coulthard and Smith, 2013a, b). Radiocarbon dating is used to assign an approximate age to subfossil wood samples where tree-ring cross-dating fails. If more than one sample is collected from a study location, the radiocarbon-dated wood remains may be cross-dated to additional undated samples to develop a floating chronology (Coulthard et al., 2013; Hoffman and Smith, 2013). In this paper, the radiocarbon ages assigned to wood samples were calibrated using INTCAL13 (Calib v. 7.02; Stuiver and Reimer, 1993; Reimer et al., 2013) and are reported as thousands of years before present (ka), where present is 1950. The reported ages in the text are median calibrated dates rounded to the nearest 0.1 ka. This is a large source of potential error in the following analysis because many calibration ranges exceed 100-year intervals and the true date of the materials may fall outside of this period. Ranges (±2σ) are reported in calibrated years before present in Table 2.1.

Dendroglaciological surveys in the Coast Mountains between 49°30’ to 57°30’ N provide insight into the regional behaviour of over 50 glaciers (Table 2.1). Sample descriptions (see Table 2.1) have been used to determine the probability of subfossil wood evidence as glacially killed: a) In situ sheared stumps and wood mats with underlying paleosols were considered high probability; b)Wood mats in lateral moraines with no underlying paleosol or in situ boles with pressed into paleosols were considered medium probability; or, c) Samples described as detrital boles, forest litter, or branches were considered to have a low probability of being glacially overridden. In the following summary, I identify all samples that have previously been described as being glacially killed. In the discussion and summaries of activity that follow, the focus is narrowed to high probability samples used to describe regional activity. Glaciers in the Boundary Ranges provide a framework of activity from the mid- to late Holocene, whereas studies completed within the Pacific Ranges offer greater insights into Holocene glacier behaviour because these glaciers demonstrate greater frontal retreat and downwasting. By comparison, knowledge of the behaviour and timing of glacier activity in the Kitimat Ranges remains limited. I have chosen not to employ the regional glacier event terminology introduced by Ryder and Thomson (1986), agreeing with Clague et al. (2009) and Osborn et al. (2013) that discrete events such as the Garibaldi Phase (7.4-6.4 ka) and the Tiedemann Advance (3.5-2.8 ka) have less relevance today as research now describes a near-continuous record of Holocene glacier expansion and retreat (Menounos et al., 2009).

2.4 Boundary Ranges

Juneau Icefield

The eastern side of the Juneau Icefield is located in northwestern British Columbia. Covering an area of approximately 3900 km^2, the icefield contains numerous large valley glaciers including Llewllyn and Tulsequah (Figure 2.2a). From 1988-2011, Juneau Icefield had a cumulative surface mass balance of -26.6 m water equivalent and a constantly increasing transient snow line (Pelto et al., 2013).

Llewelyn Glacier is a 30-km valley glacier that flows east-northeast from peaks at 2300 m asl to a terminus at 730 m asl (Figure 2.2a). Dendroglaciological and stratigraphic analysis revealed that it advanced between 1.7 and 1.5 ka and reached its late Holocene maximum position between 1.0 and 0.8 ka (Clague et al., 2010).

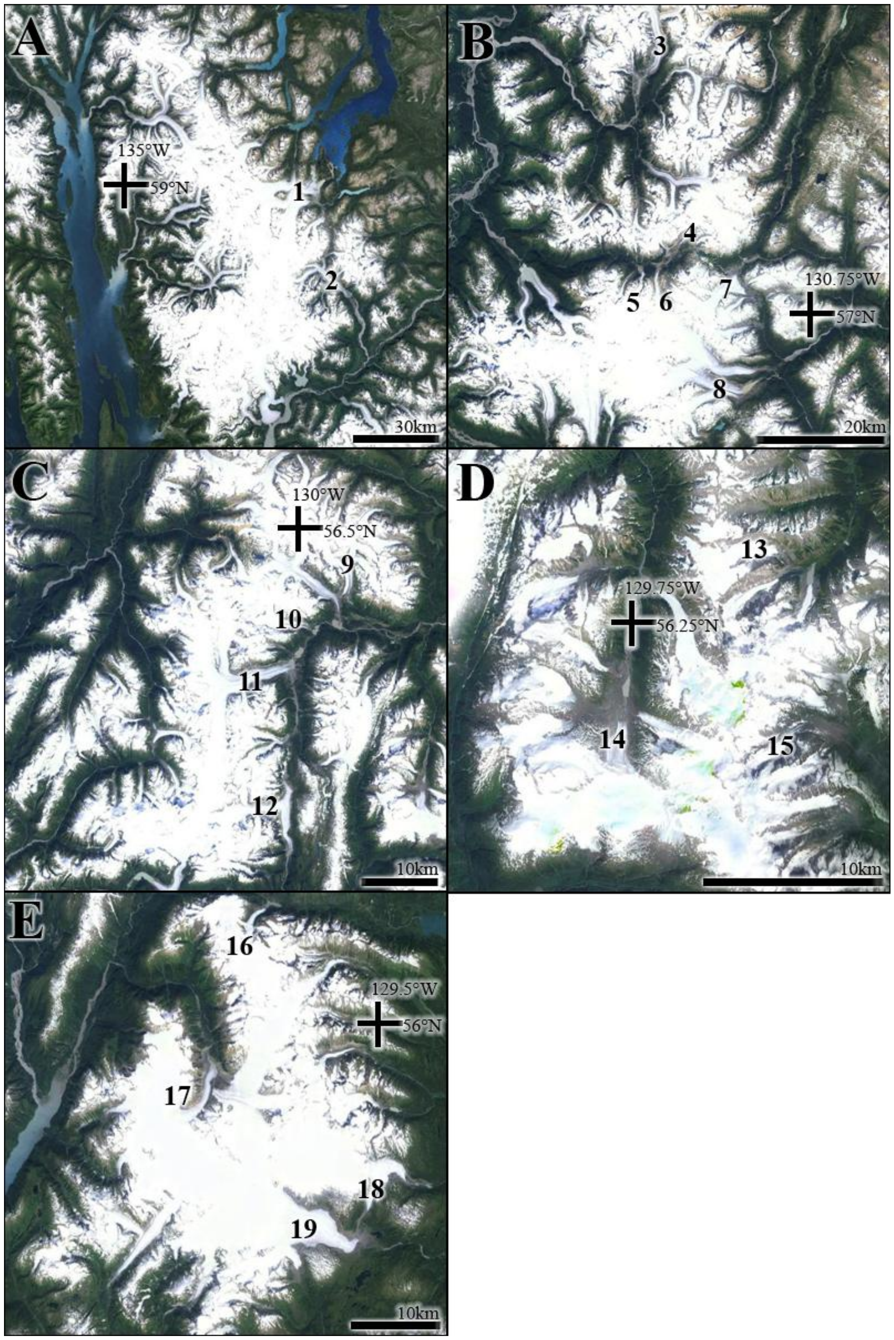

Figure 2.2 – Aerial view of study locations within the Boundary Ranges. Icefield names: (A) Juneau ([1] Llewllyn; [2] Tulsequah); (B) Andrei ([3] Scud; [4] Sphaler; [5] Glacier B; [6] Meringue; [7] South More, and; [8] Forrest Kerr glaciers); (C) Frank Mackie ([9] Charlie; [10] Canoe; [11] Frank Mackie, and; [12] Salmon glaciers); (D) Todd ([13] Horsefly; [14]Todd and Sage, and; [15]Surprise glaciers), and; (E) Cambria ([16] Bear River; [17] Bromley; [18] White, and; [19] South Flat glaciers). Images from Google Earth.

Tulsequah Glacier is a 30-km long valley glacier that flows east-southeast from 2100 m asl before terminating at 170 m asl in Tulsequah River Valley (Figure 2.2a). Radiocarbon dated subfossil tree stems reveal that it advanced at 1.1-1.2 ka (Clague et al., 2010).

Andrei Icefield area

The Andrei Icefield area is located east of the Stikine River and north of the Iskut River (Figure 2.2b). This rugged mountain landscape is characterized by numerous glaciers and snowfields with summit elevations ranging from 1600 to 3000 m asl. Field studies were initiated by Kerr (1934, 1936) who noted that following an interval of warm climate, glaciers began to advance approximately 1500 years ago. Subsequent investigations by Ryder (1987) describing LIA age activity at several glaciers in the Stikine-Iskut area provided the only additional insights, until recent investigations at Forest Kerr, Meringue, More, and Sphaler glaciers (Smith, *unpublished data*).

Scud Glacier is a large valley glacier located ca. 50 km north of the Andrei Icefield. It is formed from the confluence of many smaller tributary glaciers below its source area at ca. 1450 m asl (Figure 2.2b) and currently terminates 13 km down valley at ca. 590 m asl. The presence of a prominent trimline and lack of vegetation in the glacier forefield indicate that Scud Glacier downwasted during the last century. Detrital wood remains and buried stumps in growth position provide evidence for distinct intervals of glacier expansion at ca. 2.9, 1.5, 0.8, 0.6, 0.5, 0.2, and 0.1 ka (Ryder, 1987; Smith, *unpublished data*).

Sphaler Glacier is a large southwest-flowing valley glacier located north of the Andrei Icefield (Figure 2.2b). It originates at 1700 m asl and flows down valley 5.5 km to a terminus at ca. 1080 m asl. Partially-forested, prominent, valley-side lateral moraines illustrate significant downwasting of the glacier surface since the LIA maximum. Investigations of incised valley-floor deposits adjacent to the northern lateral moraine of Sphaler Glacier reveal evidence for Holocene expansion at ca. 8.9, 7.8, 3.2, and 1.4 ka (Smith, *unpublished data*).

Glacier B is a small northward-flowing outlet glacier of the Andrei Icefield (Figure 2.2b). Frontal retreat exposed an in situ sheared stump close to the presumed LIA terminal moraine dating to ca. 0.6 ka (Ryder, 1987).

Meringue Glacier is a small outlet glacier that originates at ca. 1660 m asl on the northern perimeter of the Andrei Icefield and flows 4.5 km northward to its terminus at ca. 820 m asl. Extensive frontal retreat and downwasting have revealed stratigraphic evidence within the eastward-facing lateral moraine complex indicating that the glacier expanded down valley at ca. 1.4 and 0.6 ka (Smith, *unpublished data*).

More Glacier is a large valley glacier that originates at ca. 1600 m asl and flows 10 km northeastward from the Andrei Icefield (Figure 2.2b). Downwasting, frontal retreat and stream erosion of a valley-bottom till have exposed subfossil boles and stumps in gullies indicating glacier expansion at ca. 1.4 ka (Smith *unpublished data*).

Forest Kerr Glacier is a large outlet glacier that flows 11 km eastward from ca. 1500 m asl at the southern edge of the Andrei Icefield to a calving terminus in a large proglacial lake at ca. 780 m asl. Field reconnaissance of eroded valley-bottom till units following drainage of an ice-marginal lake revealed stratigraphic evidence for late Holocene advances at ca. 2.6, 1.7-1.4, 0.9 ka (Smith, *unpublished data*).

Frank Mackie Icefield

Frank Mackie Icefield is located between 1300 and 1400 m asl in the headwaters of the Bowser River (56°18'N, 130°14'W; Figure 2.2c). East- and south-flowing outlet glaciers drain into the Bowser and Salmon river valleys. Investigations have focused on Berendon, Canoe, Frank Mackie, Charlie and Salmon glaciers (Clague and Mathews, 1992; Harvey et al., 2012; St-Hilaire, 2014, Smith, *unpublished data*), as well as on lake-bottom deposits associated with ice-dammed Tide and Summit lakes (Clague and Mathews, 1996).

Charlie Glacier is located in the northern sector of the Frank Mackie Icefield (Figure 2.2c). Originating at ca. 2000 m asl, the glacier flows 7 km southward down a broad valley to a terminus at ca. 970 m asl. Recent retreat and downwasting have exposed overridden wood remains within eroded lateral moraines. These remains provide evidence for glacier expansion at ca. 3.5, 3.4, 2.6, 2.5, and 0.5 ka (St-Hilaire, 2014).

Canoe Glacier is a 7-km long, eastward-flowing outlet glacier that terminates in Tippy Lake at 510 m asl (Figure 2.2c; Harvey et al., 2012). Originating at 1900 m asl, Canoe Glacier extended across and dammed the Bowser River valley (Clague and Mathews, 1992). Recent downwasting along the north-facing lateral moraine at 660 m asl has exposed the remains of glacially overridden forests killed by glacier expansion at ca. 5.2 and 3.6 ka (Harvey et al., 2012).

Frank Mackie Glacier is a 10-km long outlet glacier originating in the Frank Mackie Icefield at 1320 m asl. The glaciers flows eastward into the Bowser River valley where it calves into a proglacial lake at ca. 630 m asl (Figure 2.2c). Prior to the 1930s, Frank Mackie Glacier dammed Tide Lake into the headwaters of the Bowser River valley. The terminal moraine complex was breached by overflow from Tide Lake in 1934, allowing the lake to drain. Incised bedded Tide Lake sediments indicate glacier expansion at ca. 2.9, 2.8, 1.5, 1.4, and 1.3 ka (Clague and Mathews, 1992). Recent stratigraphic investigation and dating of extensive wood mats buried between till units in the flanking lateral moraine complex show that Frank Mackie Glacier was expanding into the Bowser River valley at ca. 2.6, 2.4, 1.7, 0.8, 0.5, and 0.2 ka (St-Hilaire, 2014; Smith, *unpublished data*).

Salmon Glacier is a 10-km long valley glacier originating at ca. 1400 m asl in the Frank Mackie Icefield (Figure 2.2c). It flows eastward into the Salmon River valley, where it bifurcates into north- and south-flowing lobes that terminate at ca. 820 and 430 m asl, respectively. Salmon Glacier previously drained northward through Tide Lake and into the Bower River Watershed but now drains southward under the glacier and into the Salmon River Valley (Mathews and Clague, 1993). Detrital wood remains encased within till at ca. 850 m asl near the LIA lateral moraine possibly indicate a period of glacier expansion at 5.6 ka (St-Hilaire and Smith, 2014).

Todd Icefield

Todd Icefield is located at 1400-1500 m asl within the headwaters of Todd Creek (Figure 2.2d). Meltwater from Todd Icefield flows northward into the Bowser River

Valley. Todd and Surprise glaciers were the focus of dendroglaciological investigations by Jackson et al. (2008), and Horsefly Glacier has been studied by D.J. Smith (*unpublished data*).

Horsefly Glacier is a small northeast-facing valley glacier that flows 4 km from 1700 m to 1250 m asl (Figure 2.2d). Downwasting of the ice surface along a south-facing lateral moraine revealed detrital wood remains dating to ca. 3.5 ka (Smith, *unpublished data*).

Todd Glacier is the largest of five outlet glaciers of Todd Icefield (Figure 2.2d). Until the 1950s, five tributary glaciers formed Todd Glacier, and ice flowed down valley to a terminus at 950 m asl (Jackson et al., 2008). Historic frontal retreat to a terminus at ca. 1000 m asl has been accompanied by extensive erosion of lateral moraines on the valley walls below Todd Glacier. Detrital wood remains and rooted stumps provide evidence for expansion of Todd Glacier at ca. 2.2, 1.6, 1.4, 0.6, and 0.1 ka (Jackson et al., 2008).

Surprise Glacier originates at ca. 1900 m asl from icefalls below Mt. Pattullo and spills eastward into a narrow, steep and deeply incised bedrock valley (Figure 2.2d; Jackson et al., 2008). A prominent 150-m-high, north-facing lateral moraine close to the present-day terminus at 910 m asl contains six distinct sediment units separated by detrital wood mats. The stratigraphy provides evidence for glacier expansion at ca. 3.0, 1.7, 1.4, 0.6 ka, as well as late LIA advances in 1764 and 1848 AD (Jackson et al., 2008).

Cambria Icefield

The Cambria Icefield encompasses an area 715 km^2 and feeds several large outlet glaciers (Figure 2.1, 2.2d). Bear River, Bromley, South Flat and White glaciers have been surveyed (Jackson et al., 2008; Johnson and Smith, 2012; Hoffman and Smith, 2013; Osborn et al., 2013).

Bear River Glacier flows northward from its source in the Cambria Icefield at ca. 2130 m asl to a calving terminus in Strohn Lake at ca. 440 m asl (Figure 2.2e; Osborn et al., 2013). Stratigraphic investigations of the terminal moraine above Strohn Lake indicate that Bear River Glacier expanded into valley-side forests in ca. 4.1-4.0, 3.8, 3.6, and 1.0 ka (Jackson et al., 2008; Osborn et al., 2013).

Bromley Glacier originates at ca. 1650 m asl in the Cambria Icefield and flows northward into the Bitter Creek valley to a terminus at ca. 820 m asl (Figure 2.2e). Stratigraphic investigations of the proximal face of a west-facing lateral moraine at about 1200 m asl show that Bromley Glacier advanced down valley at ca. 2.6-2.4, 1.8, 1.4, and 0.7 ka (Hoffman and Smith, 2013).

South Flat and White glaciers originate at 1500 m asl in the Cambria Icefield and flow eastward into White River valley (Figure 2.2e). The two glaciers were confluent until the early 20th century but have since separated. Historic retreat and draining of a series of ice-dammed lakes have exposed valley-bottom and valley-side wood remains indicating that the glaciers expanded at ca. 1.5-1.4, 1.3, 1.2, 1.0, 0.9, 0.7, and 0.6 ka (Johnson and Smith, 2012).

Summary

The earliest record of glacier expansion in the Boundary Ranges comes from Sphaler Glacier, where fir needles within sediments below till suggest that the glacier likely expanded at 8.9 and 7.8 ka. Canoe and Salmon glaciers provide evidence for mid-Holocene advances at 5.6 and 5.2 ka. Taken together, these records suggest glaciers in the Boundary Ranges were expanding down valley at 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.3, 1.7-1.1, and 0.8-0.4 ka.

2.5 Kitimat Ranges

Lichenometric surveys of nested lateral moraines at eastward-flowing Pattullo Glacier at the eastern periphery of the Kitimat Ranges in the vicinity of Eustuk Lake indicate glacier expansion prior to the late thirteenth century and in 1550–1610, 1680–1710 and 1850 AD (Harvey and Smith, 2013). Fifty-five kilometres to the south, Tzeetsaytsul Glacier flows from a small icefield at ca. 1850 m asl. Lichenometric surveys and cross-dating of detrital tree remains in till to a living tree-ring chronology provides evidence for ice expansion in the 17th and mid-18th centuries (Smith and Desloges, 2000).

2.6 Pacific Ranges

Monarch Icefield

The Monarch Icefield has an area of approximately 620 km^2 and is located at the northern end of the Pacific Ranges, about 65 km southeast of Bella Coola (Figure 2.1, 2.3a). Over the past century large outlet glaciers flowing from the icefield have retreated at rates ranging from 11 to 47 m yr^{-1} (VanLooy and Forster, 2008). Dendroglaciological

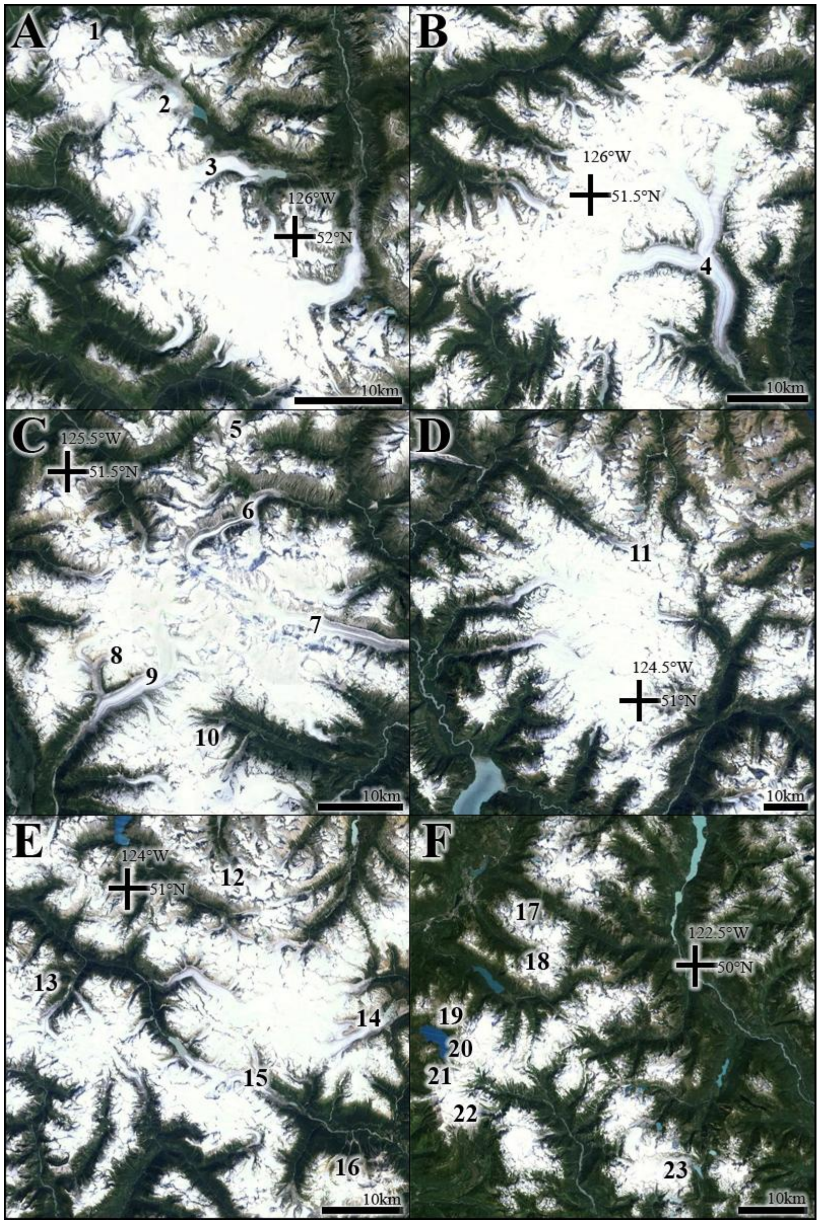

Figure 2.3 – Aerial view of study locations within the Pacific Ranges. Icefield names: (A) Monarch ([1] Purgatory; [2] Fyles, and; [3] Jacobson glaciers); (B) Ha-Itzuk ([4] Klinaklini Glacier); (C) Mt. Waddington area ([5] Whitesaddle; [6] Scimitar; [7] Tiedemann; [8] Confederation; [9] Franklin, and [10] Jambeau glaciers); (D) Homathko Icefield ([11] Queen Bess Glacier); (E) Lillooet Icefield ([12] Tchalkazan; [13] Gilbert; [14] Bridge; [15] Lillooet; [16] Manatee, Orca, and Beluga glaciers), and; (F) Garibaldi Provincial Park ([17] Spearhead and Decker; [18] Overlord; [19] Helm; [20] Sentinel and Sphinx; [21] Warren; [22] Lava, and; [23] Stave glaciers). Images from Google Earth.

surveys have been completed at Purgatory, Fyles, and Jacobsen glaciers at the eastern margin of the icefield (Desloges and Ryder, 1990; Harvey et al., 2012).

Purgatory Glacier flows northward 1.5 km from the Monarch Icefield at ca. 1820 m asl to its terminus at ca. 790 m asl in the Noeick River valley (Figure 2.3a). Fluvial incision of the valley bottom has exposed till units with detrital wood remains at unit contacts. The stratigraphy reveals glacier expansion at ca. 1.0, 0.7, 0.6 and 0.5 ka (Desloges and Ryder, 1990).

Fyles Glacier spills northward from the Monarch Icefield to formerly ice-dammed Ape Lake (Figure 2.3a; Harvey et al., 2012). The snout of Fyles Glacier has receded at rates ranging from 15 to 30 m yr^{-1} over the past few decades (VanLooy and Forster, 2008), triggering significant outburst floods in 1984 and 1986 (Gilbert and Desloges, 1987; Desloges and Gilbert, 1992) that exposed a stratigraphy interpreted as indicating glacier expansion at ca. 6.8, 6.4, 6.3, 6.0, 5.8, and 5.6 ka (Harvey et al., 2012).

Jacobsen Glacier is a large outlet glacier that flows northeastward from the Monarch Icefield to the Talchako River valley (Figure 2.3a; Harvey et al., 2012). Since AD 1915 the glacier snout has retreated over 6 km (Desloges, 1987). Erosion and collapse of the proximal faces of lateral moraines along the northern and southern flanks of the glacier have exposed in situ stumps and detrital wood indicating glacier expansion at ca. 6.6-6.4, 2.6, and 1.0 ka (Desloges and Ryder, 1990; Harvey et al., 2012).

Ha-Iltzuk Icefield

The Ha-Iltzuk Icefield has an area of approximately 854 km^2 and is located west of Mt. Waddington and south of Bella Coola (Figure 2.1, 2.3b). Klinaklini Glacier flows southward from the icefield and has retreated at rates ranging from 33 to 69 m yr^{-1} (1974-2001). The glacier has downwasted significantly since the LIA, losing over 20 km^3 of water equivalent from 1949 to 2006 (VanLooy and Forster, 2008; Tennant et al., 2012). Exposed detrital wood remains indicate Klinaklini Glacier was advancing down valley at ca. 0.8 and 0.4 ka (Ryder and Thomson, 1986).

Mt. Waddington area

Mt. Waddington (4019 m asl) is the highest peak in the Coast Mountains and is surrounded by valley glaciers flowing from high elevation ice and snowfields (Figure 2.1, 3c). Ryder and Thomson (1986) surveyed the Tiedemann and Franklin glaciers in the Mt. Waddington area, and since then research has been conducted at Whitesaddle, Scimitar, Tiedemann, Confederation, Franklin and Jambeau glaciers (Larocque and Smith, 2003; Menounos et al., 2009; Coulthard et al., 2013; Craig and Smith, 2013; Menounos et al., 2013; Mood and Smith, 2015).

Whitesaddle Glacier, located 40 km northeast of Mt. Waddington, flows eastward from a cirque at 2200 m asl to a terminal position at 1910 m asl (Figure 2.3c). Detrital boles and in situ stumps date glacial expansion events at ca. 0.7 and 0.6 ka (Larocque and Smith, 2003).

Scimitar Glacier originates at ca. 3000 m asl in the Mt. Waddington-Combatant col below the northeast face of Mt. Waddington (Figure 2.3c). The glacier flows 12 km northeast to a terminus at Scimitar Lake at ca. 980 m asl. Downwasting of the ice surface has exposed the remains of trees buried in the flanking lateral moraines, which indicate the glacier was advancing at 3.0, 2.8, 1.5, and 0.2 ka (Craig and Smith, 2013).

Tiedemann Glacier originates at 2870 m asl on the east face of Mt. Waddington and flows 18 km eastward to its present terminus at 860 m asl (Figure 2.3c). The glacier has retreated over 4.5 km from a set of LIA terminal moraines at 450 m asl since the early 20th century (Larocque and Smith 2005), exposing wood mats at unit contacts in the south-facing lateral moraine. Tiedemann Glacier advanced down valley at ca. 5.8, 4.4, 4.3-4.0, 3.6, 3.1, 2.9, 2.8, 2.4, 2.3, 1.2, 0.4, and 0.1 ka (Ryder and Thomson, 1986; Menounos et al., 2008, 2009, 2013).

Confederation Glacier is located 15 km southwest of Mt. Waddington (Figure 2.3c). Its source is at ca. 2000 m asl below Mt. Bell. Until the 1960s Confederation Glacier was confluent with Franklin Glacier, but it has since retreated up valley 4.5 km to its present terminus at 1440 m asl. In situ stumps and detrital wood indicate that Confederation Glacier advanced down valley at ca. 5.6, 3.8, 3.5, 3.4, and 0.6 ka (Coulthard et al., 2013).

Franklin Glacier is an 18-km long valley glacier that originates in an icefield at 2150 m asl below the west face of Mt. Waddington (Figure 2.3c). It advanced at 6.3, 5.4, 5.3, 4.6, 4.1, 3.1, 2.4, 1.5, 0.8, and 0.6 ka (Ryder and Thomson, 1986; Coulthard et al., 2013; Mood and Smith, 2015).

Jambeau Glacier, located 16 km south of Mt. Waddington, flows northeast from its source at ca. 1650 m asl in the Whitemantle Icefield (Figure 2.3c). Since the 1920s, Jambeau Glacier has downwasted and retreated over 1 km, exposing remnants of forest dating to ca. 3.2, 2.9 and 0.2 ka (Coulthard et al., 2013).

Homathko Icefield

The Homathko Icefield is located 50 km southeast of Mt. Waddington and covers an area of 570 km^2 (Figure 2.1, 2.3d; VanLooy and Forster, 2008). Field surveys at Bess and Queen Bess glaciers indicate outlet glaciers from the Homathko Icefield were advancing down valley at ca. 2.8, 1.4, 0.8-0.7, and 0.2 ka (Wilkie and Clague, 2009; Hart et al., 2010; Mood and Smith, 2015).

Lillooet Icefield area

The Lillooet Icefield has an area of 603 km^2 and is the source of Toba, Lillooet, Bishop, and Lord rivers (Figure 2.1, 2.3e). Field investigations have focused at Gilbert, Bridge, Tchaikazan, Icemaker, Lillooet, Manatee, Orca, Beluga and Goddard glaciers (Ryder and Thomson, 1986; Reyes and Clague, 2004; Allen and Smith, 2007; Koehler and Smith, 2011; Harvey et al., 2012.)

Gilbert Glacier originates at 3000 m asl below Mt. Gilbert and flows northeast to terminate down valley at 1350 m asl (Figure 2.3e). Stratigraphic units within a small canyon incised into the distal side of the northern lateral moraine indicate that the glacier was advancing at ca. 5.0, 3.7, 2.2, and 2.0 ka (Ryder and Thomson, 1986).

Bridge Glacier flows 18 km eastward from the Lillooet Icefield to a calving terminus in Bridge Lake (Ryder, 1991). The glacier is flanked by the eroded remains of large lateral moraines exposed during historical downwasting and retreat (Ryder and Thomson, 1986). Wood samples collected in the recently deglaciated forefield and between till units in the lateral moraines show that Bridge Glacier advanced at ca. 11.2, 7.5, 6.3, 2.1, 1.9, 1.4, 1.3, 1.1, 1.0 and 0.6-0.5 ka (Blake, 1983; Ryder and Thomson, 1986; Allen and Smith, 2007).

Tchaikazan Glacier is a 7-km long valley glacier located at the northern margin of the Lillooet Icefield. The glacier originates at 2480 m asl and flows northward to a terminus at 1900 m asl (Figure 2.3e; Harvey et al., 2012). Large detrital bole fragments washing out from beneath the snout were interpreted to indicate advances at ca. 6.2, 5.4, 5.1, and 4.1 ka (Harvey et al., 2012).

Icemaker Glacier is a small north-flowing glacier that originates at 2530 m asl within a cirque located on the north face of Icemaker Mountain (Figure 2.3e). Within the past century, the glacier has retreated >1.5 km and now terminates at 1770 m asl. The remains of in situ glacially-sheared stumps indicate that the glacier was expanding down valley at ca. 6.4 ka (Harvey et al., 2012).

Lillooet Glacier is a large valley glacier extending 9.5 km down valley from the Lillooet Icefield to a terminal position at ca. 1100 m asl (Figure 2.3e). Rooted sheared stumps in growth position and detrital boles on paleosols located within the flanking lateral moraine indicate that the glacier was advancing into forests at ca. 3.2, 3.1, 2.6, 2.5, 1.6, 1.5, 1.3, 1.0, and 0.5 ka (Reyes and Clague, 2004).

Manatee, Beluga, and Orca glaciers flow into a tributary of Lillooet River (Figure 2.3e). Since 1948, these glaciers have retreated 1.3-3.2 km up valley at rates ranging from 11 to 39 m yr^{-1} (Koehler and Smith, 2011). Subfossil stumps and detrital wood found within the proximal faces of lateral moraines and exposed by erosion indicate glacier expansion at ca. 4.8, 3.8, 3.7, 2.4, and 0.3 ka (Koehler and Smith, 2011).

Goddard Glacier is a 6-km long valley glacier that flows northeast from Mt. Cradock to a terminus at ca. 1570 m asl (Figure 2.3e). Subfossil wood remains in the glacier forefield indicate that the glacier advanced at ca. 6.3 and 4.7 ka (Blake, 1983; Menounos et al., 2008).

Garibaldi Provincial Park

Garibaldi Provincial Park is located 70 km north of Vancouver and contains some of the southernmost glaciers in the Coast Mountains (Koch et al., 2007) (Figure 2.1, 2.3f). Field investigations have targeted Spearhead, Decker, Overlord, Helm, Sphinx, Sentinel, Warren, and Lava glaciers in the northwestern part of the park and Stave Glacier at the southern extremity of the park (Preston et al., 1955; Barsendsen et al., 1957; Stuiver et al., 1960; Lowdon and Blake, 1973, 1975; Menounos et al., 2004; Koch et al., 2007; Osborn et al., 2007).

Spearhead Glacier flows 3.5 km from ca. 2300 m asl on Blackcomb Peak to a terminus at ca. 1900 m asl (Figure 2.3f). Small, weathered wood fragments found in the recently deglaciated forefield provide evidence of ice expansion at ca. 4.2 ka (Osborn et al., 2007).

Decker Glacier is a 1-km long cirque originating on the flank of Mount Trorey and Pattison at ca. 2460 m asl (Figure 2.3f). Glacier retreat has exposed rooted stumps and detrital boles on a bedrock cliff adjacent to the glacier that indicate expansion at ca. 3.4, 3.1, and 2.1 ka (Osborn et al., 2007).

Overlord Glacier is a 2.7-km long valley glacier located in the headwaters of Fitzsimmons Creek (Figure 2.3f). Wood fragments found within outwash channels provide evidence for ice expansion at ca. 7.1 and 6.7 ka (Osborn et al., 2007).

Helm Glacier originates at ca. 2030 m asl at Gentian Peak and flows 1.7 km northward to calve in a proglacial lake at ca. 1850 m asl (Figure 2.3f). Wood remains exposed in the recently deglaciated forefield indicate that the glacier advanced down valley at ca. 10.0 and 4.6 ka (Koch et al., 2007).

Sphinx Glacier is a small valley glacier originating at ca. 2020 m asl and terminating at ca. 1670 m asl (Figure 2.3f). Detrital wood fragments and in situ sheared stumps found in the glacier forefield indicate intervals of glacier expansion at ca. 8.5-8.4, 6.6, 4.8, 3.8, 1.5, and 0.5 ka (Barendsen et al., 1957; Lowdon and Blake, 1975; Menounos et al., 2004; Koch et al., 2007).

Sentinel Glacier flows 1 km northward from Deception Peak at ca. 2200 m asl to a terminus at ca. 1700 m asl near the shore of Garibaldi Lake (Figure 2.3f). In situ stumps located near several recessional moraines indicate periods of glacier expansion at ca. 8.5, 8.2, 7.0, 6.9, and 6.1 ka (Lowdon and Blake, 1973; Lowdon and Blake, 1975; Menounos et al., 2004; Koch et al., 2007).

Warren Glacier is a 1.6-km long glacier that originates at ca. 1900 m asl on the flank of Sharkfin Mountain and flows northwest to its present-day terminus at ca. 1600 m asl (Figure 2.3f). Fluvial erosion of proglacial sediments has provided evidence for glacier expansion at ca. 8.9, 7.3, 6.8, and 6.3 ka and reached its maximum position in the early 18th century (Koch et al., 2007a, 2007b).

Lava Glacier flows from the Garibaldi Neve east of Mount Garibaldi (Figure 2.3f; Koch et al., 2007). Sheared in situ stumps located on a nunatak close to the neve indicate that the glacier expanded at ca. 7.1, 6.9, 6.7, 6.6, 5.9, and 3.4 ka (Koch et al., 2007).

Stave Glacier originates at ca. 1870 m asl and flows eastward to its present terminus at ca. 1240 m asl (Figure 2.3f). Detrital stumps in till indicate the glacier was advancing at ca. 7.2 and 1.0 ka (Koch et al., 2007).

Summary

During terminal Pleistocene deglaciation, glaciers throughout the Pacific Ranges became restricted to high-elevation cirques. There is equivocal evidence for glacier advances at 11.2, 10.8 and 10.2 ka at Bridge, Tiedemann, and Warren glaciers, and limited evidence of early Holocene glacier activity at 9.6 and 8.9 ka in Garibaldi Provincial Park. Sentinel and Sphinx glaciers expanded down valley at 8.5-8.2 ka. A prolonged period of glacier activity occurred from 7.3-5.3 ka although it could be refined to shorter intervals, 7.3-6.0 and 5.4-5.3 ka, based on the probability of the corresponding samples being killed by an advancing glacier. Most glaciers in the Monarch and Lillooet icefields, and in the Mt. Waddington area, appear to have retreated before readvancing between 6.4 and 5.6 ka (c.f., Harvey et al., 2012; Mood and Smith, 2015). Following this interval, glaciers advanced down valley at 4.8-4.6 ka before retreating and expanding down valley from 4.4-4.0 ka into forests (i.e. Franklin Glacier, Chapter Three). Glaciers continued to advance and retreat a number of times over the interval from 3.6 to 2.5 ka (Clague et al., 2009; Menounos et al., 2009; Koehler and Smith, 2011). Following this interval, there is evidence for a subsequent advance at 1.4-1.2 ka, after which some glaciers appear to have retreated and downwasted before advancing into standing forests during the early LIA.

2.7 Discussion

The emerging record of Holocene glacier behaviour in the Coast Mountains shows that in the Pacific Ranges advanced at 8.5-8.2, 7.3-5.3, 4.8-2.5, 1.4-1.2, and 0.8-0.4 ka, whereas in the Boundary Ranges glaciers were expanding from 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.3, 1.7-1.1 and 0.8-0.4 ka (Figure 2.4). As shown by Figure 2.4, there is dendroglaciologic evidence (high, medium, and low probability) of near-continuous glacier activity over the 7,300 years in the Pacific Ranges and 4,100 years in the Boundary Ranges. An advance in the Pacific Ranges from 8.5-8.2 ka contains only equivocal evidence but has been previously reported as a likely glacier response to abrupt climate changes at this time and is briefly described although emphatic evidence does not currently exist. Expansion in the Pacific Ranges from 7.3-5.3 ka contains intervals absent of high probability dendroglaciological samples and could be refined into periods from 7.3-6.0 and 5.4-5.3 ka. Lack of sample depth during this interval prevents the description of definitive intervals of glacier activity and is only briefly referred to in this discussion. Potentially common intervals of glacial expansion in both ranges occurred at 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.5, 1.4-1.2, and 0.8-0.4 ka. Following each advance, many glaciers retreated and downwasted before readvancing. For example, glaciers in Manatee Valley retreated after 3.8 ka, allowing soil and 300-year old trees to develop, before glacier again expanded at 2.4 ka (Koehler and Smith, 2011). Additionally, Franklin Glacier advanced and subsequently retreated at 4.6 and 4.1 ka (Mood and Smith, 2015). As such, the broad

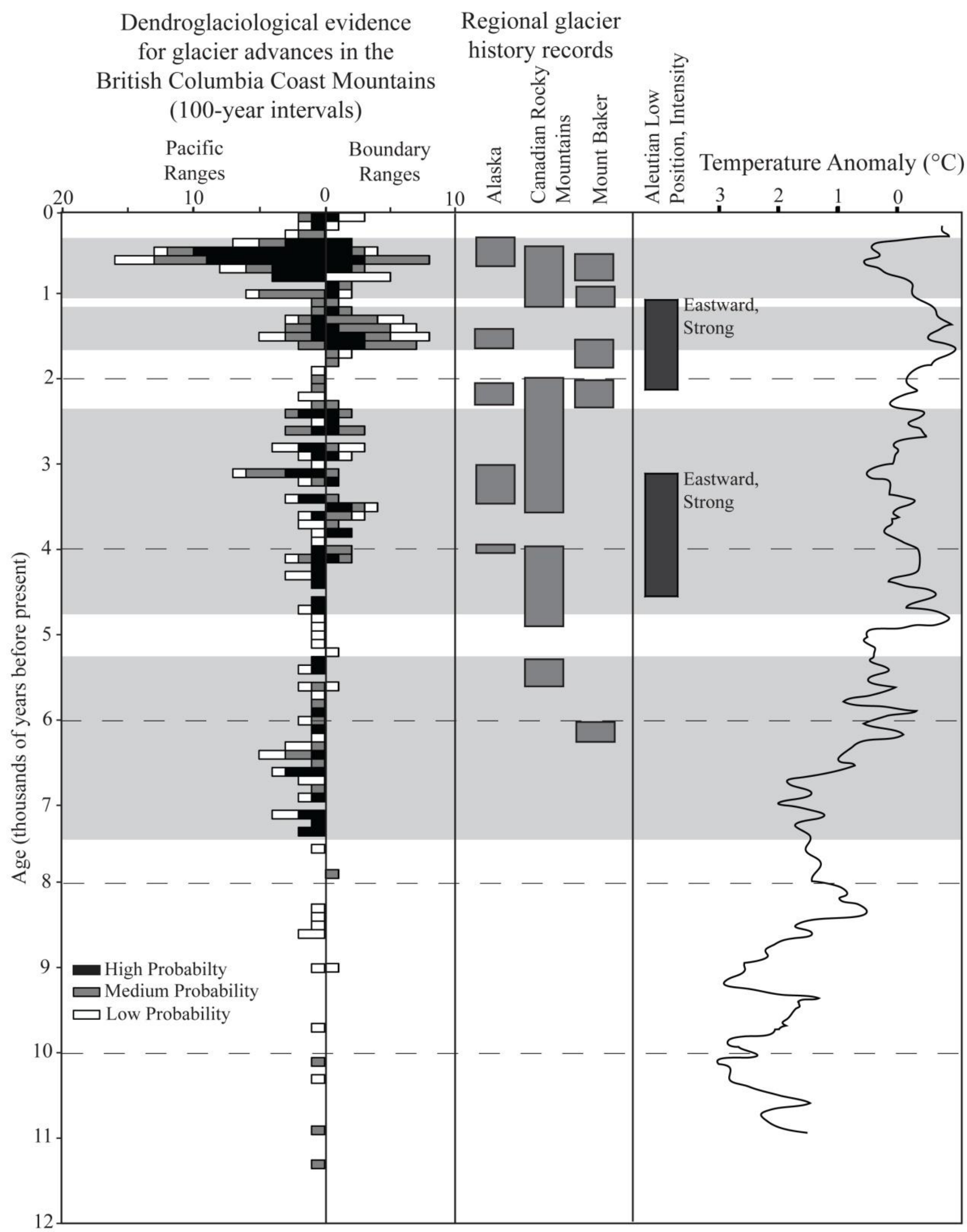

Figure 2.4 – Summary diagram of dendroglaciological evidence in the Coast Mountains. Frequency of radiocarbon dated evidence is grouped into 100-year intervals and divided into three categories determined by the sample description and location (see Table 2.1): high probability (e.g., in situ sheared stumps and wood mats with underlying paleosols),

medium probability (boles pressed into paleosol or in situ within a moraine complex, wood mats with no paleosol, and sheared stumps near growth position), and low probability (detrital boles, snags, branches found retransported by non-glacial activity). Regional glacier history records were obtained for Alaska (Barclay et al., 2009), the Canadian Rocky Mountains (Menounos et al., 2009), and Mount Baker (Osborn et al., 2012). Shown is the relative position of the Aleutian Low pressure centre (Galloway et al., 2013) and the absolute temperature anomalies from Eleanor Lake, British Columbia (based on the 0-2 ka base period in 300-year windows; Gavin et al., 2011). Light grey blocks highlight broad intervals of near-continuous glacier activity in the British Columbia Coast Mountains based on high probability dendroglaciological samples.

scale records of activity may be refined into shorter intervals of expansion according to documented instances of glacier advance-retreats within the same interval.

Each broad interval of glacier activity is described using the median date rounded to the nearest 0.1 ka. Errors associated with median age range from ±20 to ±549 years with an average possible error of ±110 years based on the calibration process (see Table 2.1). Of the 282 radiocarbon dates, only 20 exceed a maximum error potential of ±200 years. These are primarily confined to radiocarbon dated evidence from earlier literature (e.g., Preston et al., 1955; Ryder and Thomson, 1986). This suggests that the true date of a sample within a range may not be in the 100-year interval that it was placed in. It could range from ±1 to ±5 100-year intervals from its median calibrated date. Furthermore, the description of many samples does not indicate the presence of bark or perimeter wood. Glacier response times are also not accounted for in the discussion because size, mass, and mass balance of each glacier cannot be determined. Response times of glaciers in Alaska are approximately 40-years on average (Arendt et al., 2002) indicating further potential error in the Coast Mountains. As a result of the accumulated error, only broad intervals of glacier activity may be described with confidence. Descriptions of events lasting less than 0.2 ka may be misinterpreted but are speculated upon. Common periods of glacier activity between the Pacific and Boundary Ranges are short-lived and I cannot emphatically define concert glacier expansion during these intervals.

Figure 2.4 shows that glaciers in the Pacific and Boundary ranges possibly expanded in concert at about 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.5, 1.4-1.2, and 0.8-0.4 ka based on high probability dates. Prior to these intervals, a possible glacier advance at approximately 8.5-8.2 ka likely occurred in response to a regional climate shift that resulted in decreased summer temperatures but this interval does not contain high probability samples (Figure 2.4; Menounos et al., 2004; Gavin et al., 2011). A discontinuous record of high probability dendroglaciological samples from 7.3-5.3 ka corresponds with termination of early Holocene dry conditions (Galloway et al., 2009, 2010) and the shift toward a near modern wet and cool climate by 6.7 ka in British Columbia (Hebda, 1995; Galloway et al., 2009; Gavin et al., 2011). After 5.3 ka, glacier activity diminished in concert with ameliorating climate conditions in coastal British Columbia (c.f., Spear and Cwynar, 1997; Spooner et al., 1997, 2002; Gavin et al., 2011). Regional temperatures from 5.3 to 4.8 ka appear to decrease activity until an abrupt change that resulted in a return to cool and moist conditions that possibly initiated a number of glacier advances from 4.8 to 2.5 ka. Glaciers were possibly advancing in the Pacific Ranges at 4.8-4.6 ka, after which they likely retreated before expanding at 4.4-4.0 and again at 3.8-3.4 ka.

This activity occurred during a time when an intense, eastward-positioned Aleutian Low pressure centre was established, resulting in generally wet conditions (Figure 2.4; Galloway et al., 2009, 2011, 2013; Babalola et al., 2013; Mood and Smith, 2015). During periods of glacier growth, the Aleutian Low pressure centre migrated closer to Pacific North America, changing storm trajectories and producing high winter precipitation totals in the Coast Mountains (Christoforou and Hameed, 1997). While the Aleutian Low record lies within the 4.8-2.5 ka interval of activity, it also occurs during a period of inactivity from 2.1 to 1.7 ka. These contrasting relationships indicate that its role in glacier activity in the Coast Mountains remains equivocal.

Glaciers in the Boundary Ranges also expanded at 4.1-4.0 and 3.7-3.4 ka likely in response to the same climate changes. Following this glacier advances, in both settings at 2.8-2.5 ka, likely occurred in response to increased precipitation (Galloway et al., 2010). Only medium and low probability evidence for glacier expansion exists from 2.5 to 1.7 ka in the Coast Mountains. Glaciers elsewhere have been documented as advancing at this time in Pacific North America (c.f., Barclay et al., 2009; Menounos et al., 2009; Osborn et al., 2012). This interval of low activity may be a response to slightly-increasing temperatures at this time (Figure 2.4; Gavin et al., 2011).

Glacier expansion at about 1.4-1.2 ka in the Pacific and Boundary Ranges is contemporaneous with climate cooling in southern Alaska and coastal British Columbia (Hu et al., 2001; Reyes et al., 2006). Glaciers in the Boundary Ranges show a more prolonged interval of expansion during this time, from approximately 1.7 to 1.1 ka. A subsequent 200-year interval of warmer-than-average temperatures (Koch and Clague, 2011; Steinman et al., 2014) preceded initiation of early LIA advances from about 0.8 until 0.4 ka (Larocque and Smith, 2005a, 2005b; Allen and Smith, 2007; Osborn et al., 2014). During the 1.4-1.2 and 0.8-0.4 ka periods, documented temperature anomalies may highlight continual expansion of glaciers from 1.5-0.5 ka (Figure 2.4; Gavin et al., 2011). Many of the documented glaciers during this period appear to have continued to advance following expansion from 1.4-1.2 ka as there is no evidence of retreat prior initiation of the LIA. Following this interval most glaciers likely continued to expand, reaching their maximum Holocene extents during the early 18th to late 19th centuries (Munday, 1931; Larocque and Smith, 2005a; Koch et al., 2007). Since the end of the LIA, most Coast Mountain glaciers have downwasted and receded (Schiefer et al., 2007; VanLooy and Forster, 2008; Tennant et al., 2012), and in some instances these glaciers are currently the same size as they were during the mid-Holocene (Harvey et al., 2012; Koch et al., 2014).

The record of Holocene glacier activity in the Coast Mountains is similar to that documented elsewhere in Pacific North America (Figure 2.4; Barclay et al., 2009; Menounos et al., 2009; Osborn et al., 2012; Solomina et al., 2015). Elsewhere in British Columbia, Castle Glacier in the Cariboo Mountains advanced at about 9.0-8.6, 5.6, 5.0-4.5, 2.7-2.5, 1.9-1.7 and 1.5-1.4 ka (Mauer et al., 2012). In Alaska, glaciers were expanding into mature standing forests at approximately 4.5-4.0, 3.9-3.3, 2.2-2.0, 1.6-1.3 ka, and during the LIA (Barclay et al., 2009). On Mount Baker in Washington State, glaciers were also expanding at about 6.0, 2.2, 0.9 and 0.4 ka (Osborn et al., 2012). Contemporaneous glacier activity has also been documented in the southern Canadian Rocky Mountains at approximately 5.6-5.3, 4.5-4.0, 3.5-2.9, 2.6, 1.8-1.3, and during the LIA (Figure 2.4; Menounos et al., 2009).

The record of glacier activity in the Coast Mountains is similar to that documented during the Holocene at many locations globally (Winkler and Matthews, 2010; Solomina et al., 2015). Early Holocene activity at 8.5-8.2 ka occurred at the same time that glaciers expanded in the Northern and Southern hemispheres (Holzhauser and Zumbul, 1999; Nesje et al., 2001; Matthews and Dresser, 2008; Nesje, 2009; Solomina et al., 2015). Cooling and resultant glacial advances at 8.5-8.2 ka have been repeatedly attributed to decreased solar insolation resulting from volcanic and sunspot activity as well as periodic glacial lake drainages into the North Atlantic (Teller et al., 2002; Musheler et al., 2004; Solanki et al., 2004; Gavin et al., 2011; Morrill et al., 2013; Solomina et al., 2015).

Coast Mountains glacier expansion during the 4.8-2.5 ka interval at about 4.8-4.6 ka in the Pacific Ranges is synchronous with advances in Iceland and Sweden, and to a lesser extent in western Norway, perhaps due to lower-than-normal solar irradiance (Nesje et al., 2001; Bakke et al., 2005; Kirkblade and Dugmore, 2006; Nesje, 2009; Solomina et al., 2015). Following this, glacier expansion in the Pacific Ranges from approximately 4.4-4.0 ka and the Boundary Ranges at about 4.1-4.0 ka is coincident with expansion in the Canadian Rocky Mountains and Alaska (Figure 2.4; Barclay et al., 2009; Menounos et al., 2009). This interval corresponds to cooling in the North Atlantic (4.3 ka), a volcanic eruption (4.0 ka), lowered solar activity (4.4-4.3 ka) (Bond et al., 2001; Bay et al., 2006; Solomina et al., 2015); and an eastward shift in the Aleutian Low pressure centre (Galloway et al., 2013).

Possible synchronous glacier expansion at about 3.7-3.4 and 3.1 ka in the Coast Mountains coincides with similar intervals of heightened activity in Alaska and the Canadian Rocky Mountains as well as in Iceland, Sweden, Norway, the Swiss Alps and the Southern Alps (Barclay et al., 2009; Menounos et al., 2009; Winkler and Matthews, 2010; Solomina et al., 2015). These expansion events have been previously attributed to decreased solar activity at 3.9-3.8 and 3.5-3.1 ka (Rothisberger, 1986; Holzhauser and Zumbuhl, 1999; Nesje et al., 2001; Renssen et al., 2006; Solomina et al., 2015).

Glacier advances at about 2.8-2.5 and 1.4-1.2 ka in the Coast Mountains occurred following intervals of decreased solar activity at 2.8-2.6 and 1.4-1.2 ka and heightened volcanic activity at 1.7 ka (Bond et al., 2001; Renssen et al., 2006; Sigl et al., 2013; Solomina et al., 2015). Similarly, LIA expansion in the Coast Mountains from approximately 0.8-0.4 ka coincides with glacier advances at many locations in the Northern Hemisphere in response to reduced solar activity (Matthews and Briffa, 2005; Winkler and Matthews, 2010; Solomina et al., 2015).

2.8 Conclusions

Our appreciation of Holocene glacier activity within the Coast Mountain region has expanded significantly since the review provided by Menounos et al. (2009). Research within the Pacific Ranges indicates glaciers were likely expanding at approximately 7.3-5.3, 4.8-2.5, 1.4-1.2, and 0.8-0.4 ka (e.g., Coulthard et al., 2013; Craig and Smith, 2013; Mood and Smith, 2015). Recent research within the Boundary Ranges illustrates glaciers were probably expanding down valley at about 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.3, 1.7-1.1, and 0.8-0.4 ka (e.g., Jackson et al., 2008; Johnson et al., 2011; Harvey et al., 2012; Hoffman et al., 2013; Osborn et al., 2013)..

This updated record of Holocene glaciation within the British Columbia Coast Mountains highlights the spatial synchrony in glacier activity over the past 4,500 years. Potential common intervals of activity in the Coast Mountains at approximately 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.5, 1.4-1.2, and 0.8-0.4 ka correspond to events in Alaska, the Canadian

Rocky Mountains, and on Mt. Baker. This finding and congruent paleoenvironmental studies indicate that glacier expansion during these events was most likely a glaciological response to climate shifts from warm to cool and/or wet conditions.

The global record of glacier change at 4.4-4.2, 3.8-3.4, 3.3-2.8, 2.6, 1.7-1.6, 1.5-1.4, 1.2-1.0, and 0.7-0.5 ka corresponds to that described within the Coast Mountains which focuses attention on the relationship between solar variability and Holocene glacial histories (Solomina et al., 2015). In this paper, I identify the Aleutian Low pressure centre, driven by solar variability, as playing a possible role in regional ice expansion although no single mechanism is singularly driving glacier change. Eastward shifts in the position of the Aleutian Low pressure centre, during periods of decreased solar activity, resulted in increased winter precipitation throughout Pacific North America (Babalola et al., 2013; Galloway et al., 2013). The combination of periodic cooling and high winter precipitation, influenced by the Aleutian Low, likely nourished glacier growth in the study region for prolonged intervals and resulted in distinct expansion events. While the relationship between the Aleutian Low and glacier activity appears strong from 4.4 to 3.0 ka, its relationship is not strong during other intervals of expansion over the last 4800 years. A strong, eastward Aleutian Low is documented to initiate at 2.1 ka but glaciers do not appear to start advancing again until 1.7 ka. Furthermore, glaciers were likely advancing from 3.0 to 2.5 ka without a strong, eastward Aleutian Low.

The expanded Coast Mountain glacier chronology presented in this chapter provides a useful indicator of regional climate change during the Holocene. The record serves to showcase marked environmental changes that are often underestimated by other paleoclimate proxies (Solomina et al., 2015). Continued research efforts will serve to enhance this record of Holocene glacier activity and highlight the magnitude of Holocene environmental change in the Coast Mountains.

2.9 Supplementary Materials

Table 2.1 – Summary of Holocene radiocarbon dated wood samples indicating glacier activity in the British Columbia Coast Mountains. Glacier latitude and longitude are provided.

Icefield Name / Glacier Name/ Sample No.	Radio-carbon Age	Calibrated Range (YBP)[2]	Area Under Curve	Calibrated Age (Median)[3]	Potential Error (years)[1] Min	Max	Mean	Calibrated Age (ka)	Latitude (Decimal Degrees) / Sample Descripti-on [4]	Longitude (Decimal Degrees) / Elevation (m asl) [4]	Source
JUNEAU ICEFIELD											
Llewellyn									59.06	-134.09	
GSC-6634	780 ± 70	640 – 804	0.88	740	64	100	82	0.7	Tree stump	785	Clague et al. (2010)
Beta-244623	1560 ± 40	1367 - 1540	1	1450	83	90	86.5	1.5	Tree stem	760	Clague et al. (2010)
Beta-245591	1730 ± 40	1547 - 1727	1	1640	87	93	90	1.6	Tree stem	718	Clague et al. (2010)
Tulsequah									58.86	-133.87	
AA46375	1228 ± 39	1063 - 1265	1	1160	97	105	101	1.1	Tree stem	190	Clague e al. (2010)
AA46376	1272 ± 52	1168 - 1291	0.76	1230	61	62	61.5	1.2	Tree stem	190	Clague et al. (2010)
ANDRE ICEFIELD AREA											
Scud									57.31	-131.23	
UVTRL06 Scd44	80 ± 50	11 - 151	0.66	116	35	105	70	0.1	Detrital bole		Smith (*unpublished data*)
UVTRL06 Scd44	80 ± 50	0 - 290	1	150	140	150	145	0.1	Detrital bole		Smith (*unpublished data*)
S-2297	455 ± 65	421 - 559	1	490	69	69	69	0.5	In situ sheared stump		Ryder (1987)
S-2298	625 ± 140	425 - 801	1	610	185	191	188	0.6	In situ sheared stump		Ryder (1987)
UVTRL06 Scd11	880 ± 50	702 - 915	1	800	98	115	106.5	0.8	Detrital bole		Smith (*unpublished data*)
UVTRL06 Scd18	900 ± 40	735 - 916	1	830	86	95	90.5	0.8	Detrital bole		Smith (*unpublished data*)

UVTRL06 Scd15	900 ± 50	726 - 926	0.99	820	94	106	100	0.8	Detrital bole		Smith (*unpublished data*)
UVTRL06 Scd52	1570 ± 50	1354 - 1556	1	1460	96	106	101	1.5	Detrital bole		Smith (*unpublished data*)
UVTRL06 Scd43	2760 ± 50	2760 - 2962	1	2860	100	102	101	2.9	In situ sheared stump		Smith (*unpublished data*)
Sphaler									57.09	-131.41	Smith (*unpublished data*)
UVTRL05-SPLP	6990 ± 60	7694 - 7937	1	7820	117	126	121.5	7.8	In situ log		Smith (*unpublished data*)
UVTRL05-SPUP	2980 ± 60	2970 - 3272	0.9	3150	122	180	151	3.2	In situ wood mat		Smith (*unpublished data*)
SG05-01	7980 ± 40	8699 - 8998	0.97	8860	138	161	149.5	8.9	Forest litter		Smith (*unpublished data*)
UVTRL05 Scr02	1530 ± 40	1343 - 1526	1	1430	87	96	91.5	1.4	In situ sheared stump		Smith (*unpublished data*)
Glacier B									57.06	-131.05	
S-2296	595 ± 60	524 - 665	1	590	66	75	70.5	0.6	Ah material from paleosol		Ryder (1987)
More									57.05	-130.94	
UVTRUL05-Mor06	1450 ± 50	1284 - 1416	0.96	1350	66	66	66	1.4	In situ bole		Smith (*unpublished data*)
Meringue									57.03	-131.18	
UVTRL05-MER02	605 ± 15	583 - 648	0.79	620	28	37	32.5	0.6	In situ bole		Smith (*unpublished data*)
UVTRL05 MGNC1	1490 ± 50	1300 - 1424	0.78	1380	44	80	62	1.4	In situ bole		Smith (*unpublished data*)
Forest Kerr									56.9	-130.93	
UVTRL04-FKS21	970 ± 60	739 - 977	1	870	107	131	119	0.9	In situ bole		Smith (*unpublished data*)
UVTRL04-FKS24	1480 ± 60	1295 - 1445	0.8	1380	65	85	75	1.4	In situ log		Smith (*unpublished data*)
UVTRL06 FKa05	1570 ± 50	1354 - 1556	1	1460	96	106	101	1.5	In situ sheared stump		Smith (*unpublished data*)
UVTRL06 FKb01	1600 ± 50	1382 - 1574	0.96	1480	94	98	96	1.5	In situ sheared stump		Smith (*unpublished data*)
UVTRL06 FKb04	1630 ± 40	1412 -	1	1530	81	118	99.5	1.5	In situ sheared		Smith (*unpublishe*

		1611							stump		*d data*)
UVTRL06 FKa04	1670 ± 50	1515 - 1705	0.88	1580	65	125	95	1.6	In situ sheared stump		Smith (*unpublished data*)
UVTRL06 FKa06	1670 ± 50	1515 - 1705	0.88	1580	65	125	95	1.6	In situ sheared stump		Smith (*unpublished data*)
UVTRL04-FKS34	1690 ± 50	1518 - 1718	0.96	1600	82	118	100	1.6	In situ bole		Smith (*unpublished data*)
UVTRL04-FKS43	1720 ± 50	1527 - 1740	0.98	1630	103	110	106.5	1.6	In situ bole		Smith (*unpublished data*)
UVTRL04-FKS60	1750 ± 60	1545 - 1819	1	1660	115	159	137	1.7	In situ bole		Smith (*unpublished data*)
UVTRL04-FKS66	2570 ± 60	2456 - 2788	0.99	2630	158	174	166	2.6	In situ bole		Smith (*unpublished data*)
FRANK MACKIE ICEFIELD											
Charlie									56.46	-129.97	
UVTRL-KG13-26	440 ± 30	459 - 533	0.95	500	33	41	37	0.5	In situ sheared stump	1228	St-Hilaire (2014)
UVTRL-KG13-31	2450 ± 30	2360 - 2544	0.55	2530	14	170	92	2.5	Organic mat layer	1040	St-Hilaire (2014)
UVTRL-KG13-30	2630 ± 30	2728 - 2786	1	2750	22	36	29	2.8	4m long bole in organic mat layer	1040	St-Hilaire (2014)
UVTRL-KG13-41	3190 ± 30	3361 - 3460	1	3410	49	50	49.5	3.4	Spill from ice pressed logs	873	St-Hilaire (2014)
UVTRL-KG13-01	3250 ± 30	3438 - 3561	0.84	3470	32	91	61.5	3.5	Standing, ice pressed into bedrock	1080	St-Hilaire (2014)
UVTRL-KG13-12	3260 ± 30	3442 - 3565	0.92	3490	48	75	61.5	3.5	In situ wood mat	977	St-Hilaire (2014)
Canoe									56.39	-130.11	
UVTRL06 CAN01	3360 ± 50	3459 - 3710	1	3600	110	141	125.5	3.6	Log in moraine	750	Harvey et al. (2012)
UVTRL06 CAN03	4570 ± 50	5047 - 5202	0.47	5220	18	173	77.5	5.2	Bole in moraine	750	Harvey et al. (2012)
Frank Mackie									56.33	-130.12	
UVTRL05 FMb03	230 ± 50	135 - 223	0.35	210	13	75	44	0.2			St-Hilaire (2014)

UVTRL05 FMb06	520 ± 50	497 - 567	0.69	540	27	43	35	0.5		705	St-Hilaire (2014)
UVTRL05 FMa04	850 ± 40	685 - 803	0.84	760	43	75	59	0.8		715	St-Hilaire (2014)
UVTRL05 FMa10	840 ± 40	680 - 800	0.9	750	50	70	60	0.8		715	St-Hilaire (2014)
UVTRL05 FMa03	1770 ± 50	1568 - 1817	1	1690	122	127	124.5	1.7	Log in moraine	705	St-Hilaire (2014)
UVTRL05 FMNL4	2470 ± 40	2420 - 2717	0.93	2570	147	150	148.5	2.6	In situ wood mat	755	St-Hilaire (2014)
UVTRL05 FMNL3	2400 ± 40	2343 – 2514	0.8	2440	74	97	85.5	2.4	In situ wood mat	757	St-Hilaire (2014)
Frank Mackie (Tide Lake)									56.33	-130.12	
GSC-1372	2730 ± 170	2363 - 3244	0.99	2860	384	497	440.5	2.9	Wood from drillhole	650	Lowdon and Blake (1973)
TO-2205	2700 ± 60	2741 - 2929	0.99	2810	69	119	94	2.8	Conifer needles	595	Clague and Mathews (1992)
TO-2204	2650 ± 60	2699 - 2888	0.95	2770	71	118	94.5	2.8	Conifer needles	605	Clague and Mathews (1992)
GSC-5349	1640 ± 60	1399 - 1637	0.91	1540	97	141	119	1.5	Log	616	Clague and Mathews (1992)
TO-2898	1600 ± 40	1394 - 1567	1	1480	86	87	86.5	1.5	Conifer needles	604	Clague and Mathews (1992)
GSC-5386	1520 ± 50	1318 - 1526	1	1410	92	116	104	1.4	Wood fragment	603	Clague and Mathews (1992)
TO-2897	1440 ± 40	1291 - 1398	1	1340	49	58	53.5	1.3	Wood fragment	612	Clague and Mathews (1992)
Salmon									56.17	-130.08	
UVTRL-SG13-01	4900 ± 30	5588 - 5663	0.97	5630	33	42	37.5	5.6	Buried log	883	St-Hilaire (2014)
TODD ICEFIELD											
Horsefly									57.27	-129.63	
UVTRL04-HF03	3300 ± 15	3479 - 3569	1	3520	41	49	45	3.5	Detrital bole		Smith (*unpublished data*)
Surprise									56.19	-129.63	
UVTRL04-SG804	600 ± 50	533 - 662	1	600	62	67	64.5	0.6	In situ log	900	Jackson et al. (2008)
UVTRL04-SG829	1400 ± 60	1229 - 1409	0.95	1320	89	91	90	1.3	In situ branch	870	Jackson et al. (2008)

UVTRL04-SG828	1690 ± 60	1473 - 1731	0.93	1600	127	131	129	1.6	In situ log	850	Jackson et al. (2008)
SG03-801	2960 ± 70	2941 - 3275	0.91	3120	155	179	167	3.1	In situ log	800	Jackson et al. (2008)
Todd									56.18	-129.77	
UVTRL04-TG802	120 ± 60	2 - 153	0.58	130	23	128	75.5	0.1	In situ sheared stump	1046	Jackson et al. (2008)
UVTRL-TG02-36	270 ± 70	257 - 501	0.75	330	73	171	122	0.3			Jackson et al. (2008)
TG04-879	360 ± 60	302 - 509	1	410	99	108	103.5	0.4	In situ sheared stump	1090	Jackson et al. (2008)
TG04-872	410 ± 60	419 - 532	0.62	450	31	82	56.5	0.4	Glacier pushed detrital pile of stumps	990	Jackson et al. (2008)
UVTRL04-TG801	440 ± 60	420 - 551	0.74	480	60	71	65.5	0.5	In situ bole	975	Jackson et al. (2008)
UVTRL04-TG871	650 ± 70	527 - 695	1	610	83	85	84	0.6	In situ bole	950	Jackson et al. (2008)
UVTRL04-TG838	630 ± 60	535 - 675	1	600	65	75	70	0.6	In situ bole	925	Jackson et al. (2008)
UVTRL04-TG805	690 ± 60	550 - 710	0.97	650	60	100	80	0.6	In situ bole	960	Jackson et al. (2008)
UVTRL04_TG874	1680 ± 60	1472 - 1716	0.9	1590	118	126	122	1.6	In situ sheared stump	1010	Jackson et al. (2008)
TG03-815	1540 ± 60	1317 - 1546	1	1440	106	123	114.5	1.4	Log on surface of bedrock outcrop	1100	Jackson et al. (2008)
Sage									56.18	-129.77	
TG03-806	2300 ± 60	2148 - 2473	0.99	2310	162	163	162.5	2.3	Detrital bole	930	Jackson et al. (2008)
CAMBRIA ICEFIELD											
Bear River									56.09	-129.7	
B10B-04	1040 ± 50	910 - 1060	1	990	70	80	75	1	Detrital, wood mat		Osborn et al. (2013)
B8-04	3310 ± 70	3380 - 3700	1	3540	160	160	160	3.5	In situ, wood mat		Osborn et al. (2013)
B5-03	3330 ± 60	3435 - 3695	1	3560	125	135	130	3.6	Detrital, wood mat		Osborn et al. (2013)
UVTRLBR 03-806	3340 ± 60	3445 - 3719	0.99	3580	135	139	137	3.6	Derital bole	435	Jackson et al. (2008)
B3B-04	3380 ± 60	3780 - 3820	1	3800	20	20	20	3.8	Detrital on grey-brown		Osborn et al. (2013)

									contact		
B9-04	3410 ± 60	3480 - 3860	1	3670	190	190	190	3.7	Derital in grey till		Osborn et al. (2013)
B7-04	3540 ± 60	3670 - 3980	1	3820	150	160	155	3.8	In situ, wood mat		Osborn et al. (2013)
BR03-801	3680 ± 60	3846 - 4156	0.98	4020	136	174	155	4	Log from lower woody mat	435	Jackson et al. (2008)
B1-03	3680 ± 60	3850 - 4160	1	4010	150	160	155	4	Detrital, wood mat		Osborn et al. (2013)
B4B-04	3710 ± 70	3860 - 4250	1	4060	190	200	195	4.1	In situ, wood mat		Osborn et al. (2013)
B3-03	3790 ± 70	3820 - 4380	1	4100	280	280	280	4.1	Detrital, wood mat		Osborn et al. (2013)
Bromley									55.94	-129.73	
UVTRL011 -KHS2S2	830 ± 30	688 - 789	1	740	49	52	50.5	0.7	Wood in moraine		Hoffman and Smith (2013)
UVTRL011 -KHS1S6	1480 ± 30	1306 - 1411	1	1360	51	54	52.5	1.4	Wood in moraine	930	Hoffman and Smith (2013)
UVTRL011 -KHS2S6	1850 ± 30	1715 - 1865	1	1780	65	85	75	1.8	Wood in moraine	960	Hoffman and Smith (2013)
UVTRL011 -KHS3S1	2410 ± 30	2350 – 2497	0.85	2430	67	80	73.5	2.4	Wood in moraine	890	Hoffman and Smith (2013)
UVTRL011 -KHS1S11	2470 ± 30	2425 - 2717	0.96	2580	137	155	146	2.6	Wood in moraine	895	Hoffman and Smith (2013)
South Flat									55.84	-129.47	
UVTRL08-SF18A	610 ± 60	530 - 669	1	600	69	70	69.5	0.6	Lake sediments		Johnson and Smith (2012)
UVTRL08-SF16A	830 ± 40	676 - 798	0.95	740	58	64	61	0.7	Lake Sediments		Johnson and Smith (2012)
UVTRL08-SF12	960 ± 40	786 - 938	0.99	860	74	78	76	0.9	Lake sediments		Johnson and Smith (2012)
UVTRL08-SF07A	1090 ± 50	924 - 1090	0.93	1000	76	90	83	1			Johnson and Smith (2012)
UVTRL08-SF01A	1280 ± 50	1171 - 1294	0.82	1220	49	74	61.5	1.2	Lake sediments		Johnson and Smith (2012)
White									55.8	-129.52	
UVTRL09-WG09-02B	1360 ± 50	1220 - 1354	0.87	1290	64	70	67	1.3	Wood detritus		Johnson and Smith (2012)

UVTRL09-WG10	1360 ± 40	1233 - 1343	0.91	1290	53	57	55	1.3	Wood in moraine		Johnson and Smith (2012)
UVTRL08-WG04A	1370 ± 40	1238 - 1350	0.95	1290	52	60	56	1.3	Wood in moraine		Johnson and Smith (2012)
UVTRL09-WG03-01A	1460 ± 40	1292 - 1413	1	1350	58	63	60.5	1.3	Wood in moraine		Johnson and Smith (2012)
UVTRL09-WG07-14A	1530 ± 40	1343 - 1526	1	1430	87	96	91.5	1.4	Wood detritus		Johnson and Smith (2012)
UVTRL09-WG05-03A	1550 ± 40	1356 - 1532	1	1460	72	104	88	1.5	Wood in moraine		Johnson and Smith (2012)
MONARCH ICEFIELD											
Purgatory									52.16	-126.25	
GSC-4191	480 ± 50	451 - 561	0.89	506	55	55	55	0.5	Wood roots		Desloges and Ryder (1990)
S-2978	630 ± 65	531 - 678	1	600	69	78	73.5	0.6	Forest litter with paleosol		Desloges and Ryder (1990)
S-2976	1110 ± 70	912 - 1185	0.97	1050	135	138	136.5	1	Wood		Desloges and Ryder (1990)
S-2977	785 ± 70	645 - 804	0.87	720	75	84	79.5	0.7	Forest litter with paleosol		Desloges and Ryder (1990)
GSC-4028	460 ± 50	429 - 556	0.9	490	61	66	63.5	0.5	Wood		Desloges and Ryder (1990)
Fyles									52.1	-126.2	
UVTRL-FG02-16A	4860 ± 60	5469 - 5727	1	5600	127	131	129	5.6	Branch/root, topsets	1370	Harvey et al. (2012)
UVTRL-FG02-9	5110 ± 70	5707 – 5992	0.97	5840	133	152	142.5	5.8	Branch/root, topsets		Harvey et al. (2012)
UVTRL-FG02-18	5240 ± 70	5892 - 6210	0.99	6020	128	190	159	6	Branch, topsets in organics		Harvey et al. (2012)
UVTRL-FG02-19	5570 ± 70	6270 - 6497	0.97	6360	90	137	113.5	6.4	Large bole, bottomsets		Harvey et al. (2012)
UVTRL-FG02-8	5560 ± 80	6193 - 6505	0.99	6360	145	167	156	6.4			Harvey et al. (2012)
UVTRL-FG02-6	5640 ± 80	6292 - 6570	0.94	6430	138	140	139	6.4	Branch/root, bottomsets		Harvey et al. (2012)
UVTRL-	5980	6636	0.98	6820	184	198	191	6.8	Branch/ro		Harvey et

FG02-7	± 80	- 7018							ot from forests		al. (2012)
Jacobsen									52.05	-126.07	
UVTRL10_ JACA02	1060 ± 40	923 - 1057	1	970	47	87	67	1			Harvey et al. (2012)
UVTRLJA C0504	5630 ± 60	6296 - 6544	1	6410	114	134	124	6.4	Detrital log in till	1451	Harvey et al. (2012)
UVTRLJA C0301	5760 ± 60	6409 - 6677	0.99	6560	117	151	134	6.6	In situ stump at 2010 ice margin	1421	Harvey et al. (2012)
GSC-4155	2470 ± 50	2376 - 2718	0.98	2560	158	184	171	2.6	Wood	1370	Desloges and Ryder (1990)
HA-ILTZUK ICEFIELD											
Klinaklini									51.45	-125.85	
S-1566	400 ± 45	420 - 519	0.64	450	30	69	49.5	0.5	Tree root from paleosol	530	Ryder and Thomson (1986)
S-1567	900 ± 40	735 - 916	1	830	86	95	90.5	0.8	Tree stump in growth position	400	Ryder and Thomson (1986)
MT. WADDING -TON AREA											
Whitesadd le Glacier									51.59	-124.85	
01B809A	540 ± 60	502 - 654	1	560	58	94	76	0.6	In situ stump		Larocque and Smith (2005)
01B814	760 ± 50	651 - 784	0.98	670	19	114	66.5	0.7	In situ stump		Larocque and Smith (2005)
Scimitar									51.44	-125.25	
UVTRL011 -S11/04/03	230 ± 30	265 - 317	0.42	200	65	117	26	0.2	Wood fragment	1287	Craig and Smith (2013)
UVTRL011 -S11/03/03	1610 ± 30	1413 - 1558	1	1490	68	77	72.5	1.5	Wood fragment	1187	Craig and Smith (2013)
UVTRL011 -S11/04/02	2650 ± 30	2740 - 2798	0.94	2760	20	38	29	2.8	Detrital bole	1236	Craig and Smith (2013)
UVTRL011 -S11/03/02	2740 ± 30	2765 - 2885	0.97	2830	55	65	60	2.8	Detrital bole	1150	Craig and Smith (2013)
UVTRL011 -S11/03/04	2910 ± 30	2961 - 3084	0.74	3050	34	89	61.5	3	Detrital bole	1159	Craig and Smith (2013)
Tiedemann									51.34	-125.1	

Beta-2020941	5010 ± 40	5653 - 5797	0.63	5750	47	97	72	5.7	Log		Menounos et al. (2009)
UCIAMS-40663	3865 ± 20	4234 - 4409	0.97	4300	66	109	87.5	4.3	In situ sheared stump		Menounos et al. (2009)
UCIAMS-40660	3820 ± 20	4149 - 4288	0.99	4200	51	88	69.5	4.2	In situ sheared stump		Menounos et al. (2009)
Beta-220940	3760 ± 60	3962 - 4298	0.96	4130	168	168	168	4.1	In situ sheared stump		Menounos et al. (2009)
Beta-220936	3690 ± 50	3887 - 4155	0.99	4030	125	143	134	4	In situ sheared stump		Menounos et al. (2009)
UCIAMS-40662	2820 ± 20	2861 - 2970	1	2920	50	59	54.5	2.9	In situ stump		Menounos et al. (2009)
Beta-220939	2710 ± 40	2751 - 2878	1	2810	59	68	63.5	2.8	In situ stump		Menounos et al. (2009)
Beta-220937	2670 ± 50	2735 - 2872	1	2790	55	82	68.5	2.8	In situ stump		Menounos et al. (2009)
Beta-220938	2520 ± 50	2432 - 2750	0.99	2590	158	160	159	2.6	In situ stump		Menounos et al. (2009)
UCIAMS-40664	365 ± 20	425 - 497	0.59	440	15	57	36	0.4	Wood mat		Menounos et al. (2009)
GSC-948	2250 ± 130	1948 - 2543	0.92	2250	293	302	297.5	2.3	Peat	825	Ryder and Thomson (1986)
GSC-938	2940 ± 130	2787 - 3383	1	3100	283	313	298	3.1	Peat	825	Ryder and Thomson (1986)
S-1470	3345 ± 115	3351 - 3887	1	3590	239	297	268	3.6	Wood fragment	1360	Ryder and Thomson (1986)
GSC-939	9510 ± 150	10411 - 11214	1	10830	419	384	401.5	10.8	Basal peat	825	Ryder and Thomson (1986)
S-1471	2355 ± 60	2303 - 2541	0.78	2410	107	131	119	2.4	Log with bark, near growth position	990	Ryder and Thomson (1986)
S-1473	1330 ± 65	1171 - 1351	0.9	1250	79	101	90	1.3	Log (transported)	1360	Ryder and Thomson (1986)
GSC-977	1270 ± 140	920 - 1417	0.98	1180	237	260	248.5	1.2	Peat	825	Ryder and Thomson (1986)
S-1474	300 ± 60	271 - 501	0.93	380	109	121	115	0.4	Log (transported)	1360	Ryder and Thomson (1986)
S-1472	65 ± 100	-160	0.59	130	30	130	80	0.1	Log with bark, near	1015	Ryder and Thomson

									growth position		(1986)
Beta-170671	8650 ± 60	9521 - 9788	0.99	9620	99	168	133.5	9.6			Koch et al. (2007b)
UCIAMS-40661	3940 ± 20	4343 - 4439	0.77	4410	67	29	48	4.4	In situ sheared stump		Menounos et al. (2009)
Confederation									51.32	-125.46	
UVTRL-B04C	4850 ± 50	5566 - 5663	0.63	5590	24	73	48.5	5.6	Detrital bole		Coulthard et al. (2013)
UVTRL-C01B	3470 ± 70	3568 - 3911	1	3740	171	172	171.5	3.7	Bole		Coulthard et al. (2013)
UVTRL08-C04I	3220 ± 60	3338 - 3585	0.98	3450	112	135	123.5	3.4	Detrital Bole		Coulthard et al. (2013)
UVTRL08-C06B	3370 ± 50	3475 - 3722	0.98	3610	112	135	123.5	3.6	In situ sheared stump		Coulthard et al. (2013)
UVTRL08-C07E	630 ± 40	549 - 665	1	600	51	65	58	0.6			Coulthard et al. (2013)
Franklin									51.29	-125.35	
S-1568	835 ± 45	676 - 801	0.89	750	51	74	62.5	0.7	Tree root from paleosol	1170	Ryder and Thomson (1986)
UVTRL12-FG03	1600 ± 30	1412 - 1551	1	1480	68	71	69.5	1.5	In situ sheared stump	1470	Mood and Smith (2015)
UVTRL-FG13-05	2410 ± 30	2350 - 2497	0.85	2430	67	80	73.5	2.4	In situ bole	1410	Mood and Smith (2015)
UVTRL-FG13-25	2930 ± 30	2975 - 3168	1	3080	88	105	96.5	3.1	In situ bole	1290	Mood and Smith (2015)
UVTRL12-FG01	3730 ± 30	3981 - 4154	1	4080	74	99	86.5	4.1	In situ bole	1480	Mood and Smith (2015)
UVTRL-FG13-17	4570 ± 30	5267 - 5323	0.49	5290	23	33	28	5.3	In situ wood mat	1450	Mood and Smith (2015)
UVTRL-FG13-20	4610 ± 30	5375 - 5458	0.6	5400	25	58	41.5	5.4	In situ wood mat	1460	Mood and Smith (2015)
UVTRL-FG13-01	5470 ± 30	6259 - 6310	0.63	6280	21	30	25.5	6.3	In situ bole	1220	Mood and Smith (2015)
UVTRL-FG13-23	10920 ± 50	12700 - 12912	1	12770	70	142	106	12.8	In situ wood mat	1570	Mood and Smith (2015)
UVTRL-FG13-24	10880 ± 50	12691 - 12832	1	12750	59	82	70.5	12.8	In situ wood mat	1380	Mood and Smith (2015)

UVTRL-FG13-09	4120 ± 30	4528 - 4714	0.72	4650	64	122	93	4.7	In situ sheared stump	1450	Mood and Smith (2015)
UVTRL08-C05G	520 ± 50	497 - 567	0.69	540	27	43	35	0.5	In situ sheared stump		Coulthard et al. (2013)
UVTRL08-C01A	530 ± 50	501 - 568	0.62	550	18	49	33.5	0.5	Root in paleosol		Coulthard et al. (2013)
Jambeau									51.23	-125.28	
UVTRL09-JG01	230 ± 40	140 - 220	0.38	210	10	70	40	0.2	In situ stump		Coulthard et al. (2013)
UVTRL09-JG05	2780 ± 50	2769 - 2996	1	2880	111	116	113.5	2.9	Detrital bole		Coulthard et al. (2013)
UVTRL09-JG02	3030 ± 60	3059 - 3378	0.99	3230	148	171	159.5	3.2	Detrital bole		Coulthard et al. (2013)
HOMATHKO ICEFIELD											
Queen Bess									51.2	-124.54	
UVTRL12-QB05	820 ± 30	686 - 784	1	730	44	54	49	0.7	Colluvial debris from wood mat		Mood and Smith (2015)
UVTRL12-QB08	790 ± 30	672 - 748	0.98	710	38	38	38	0.7	In situ sheared stump		Mood and Smith (2015)
UVTRL12-QB04	860 ± 30	694 - 800	0.87	760	40	66	53	0.8	Colluvial debris from woodmat		Mood and Smith (2015)
UVTRL12-QB11	1470 ± 30	1306 - 1404	1	1350	44	54	49	1.4	Detrital bole in soil		Mood and Smith (2015)
LILLOOET ICEFIELD AREA											
Gilbert									50.88	123.17	
S-1572	2040 ± 40	1919 - 2116	0.96	2000	81	116	98.5	2	Peat (litter)	1450	Ryder and Thomson (1986)
S-1459	2220 ± 75	2038 - 2353	0.98	2220	133	182	157.5	2.2	Log (transported)	1430	Ryder and Thomson (1986)
S-1461	2175 ± 75	1998 - 2339	1	2180	159	182	170.5	2.2	Log (transported)	1450	Ryder and Thomson (1986)
S-1462	3415 ± 70	3544 - 3842	0.93	3670	126	172	149	3.7	Log (transported)	1460	Ryder and Thomson (1986)
S-1460	4360 ± 75	4825 - 5087	0.82	4960	127	135	131	5	Branch (transported)	1435	Ryder and Thomson (1986)

Bridge									50.83	-123.54	
BG03-801	430 ± 60	420 - 543	0.7	470	50	73	61.5	0.5	In situ branch on paleosol		Allen and Smith (2007)
UVTRLBG 03-09	620 ± 60	533 - 672	1	600	67	72	69.5	0.6	In situ sheared stump in paleosol		Allen and Smith (2007)
UVTRL-BG02-5	690 ± 50	618 - 705	0.62	650	32	55	43.5	0.6	In situ sheared stump		Allen and Smith (2007)
UVTRL-BG02-9	650 ± 50	547 - 675	1	610	63	65	64	0.6	In situ sheared stump		Allen and Smith (2007)
UVTRL-BG02-01	1190 ± 60	977 - 1194	0.84	1120	74	143	108.5	1.1	Detrital bole		Allen and Smith (2007)
UVTRL-BG02-10	1500 ± 50	1305 - 1445	0.75	1390	55	85	70	1.4	In situ sheared stump		Allen and Smith (2007)
UVTRL03-BG57	2920 ± 60	2917 - 3229	0.96	3070	153	159	156	3.1	In situ wood mat	1690	Allen and Smith (2007)
UVTRLBG 03-91	1360 ± 60	1176 - 1386	1	1280	104	106	105	1.3	In situ detrital bole		Allen and Smith (2007)
BG03-802	1930 ± 70	1705 - 2045	1	1880	165	175	170	1.9	In situ detrital bole pressed into paleosol		Allen and Smith (2007)
BG03-890	1040 ± 50	899 - 1062	0.9	960	61	102	81.5	1	Buried detrital bole		Allen and Smith (2007)
GSC-3219	5500 ± 70	6179 - 6445	0.98	6300	121	145	133	6.3	Wood washed out from glacier	1935	Blake (1983)
S-1464	6590 ± 135	7246 - 7699	0.99	7480	219	234	226.5	7.5	Charcoal from paleosol	1480	Ryder and Thomson (1986)
GSC-3173	9070 ± 130	9858 - 10569	0.97	10230	339	372	355.5	10.2	Wood washed out from glacier	1935	Blake (1983)
S-1570	9810 ± 160	10710 - 11808	1	11240	530	568	549	11.2	Basal peat	1390	Ryder and Thomson (1986)
S-1468	685 ± 60	547 - 708	0.98	640	68	93	80.5	0.6	Detrital bole	1490	Ryder and Thomson (1986)
S-1463	680 ± 50	551 - 691	1	640	51	89	70	0.6	Log close to growth position	1750	Ryder and Thomson (1986)
S-1467	655 ±	540 -	1	610	70	76	73	0.6	Log	1465	Ryder and

	60	686							(transported)		Thomson (1986)
S-1465	530 ± 65	482 - 660	0.99	550	68	110	89	0.6	Branch (transported)	1465	Ryder and Thomson (1986)
S-1571	540 ± 45	507 - 567	0.58	550	17	43	30	0.6	Root in paleosol	1750	Ryder and Thomson (1986)
S-1569	115 ± 40	8 - 152	0.61	120	32	112	72	0.1	Peat	1525	Ryder and Thomson (1986)
Tchalkazan									50.8	-123.42	
UVTRL-TG02-17	3730 ± 60	3897 - 4248	0.99	4080	168	183	175.5	4.1	Detrital wood		Harvey et al. (2012)
UVTRL-T602-10	4440 ± 70	4870 - 5148	0.66	5070	78	200	139	5.1	Detrital wood		Harvey et al. (2012)
UVTRL-T602-24	4660 ± 70	5279 - 5586	0.95	5400	121	186	153.5	5.4	Detrital wood		Harvey et al. (2012)
UVTRL-TG02-7	5380 ± 70	5993 - 6296	1	6160	136	167	151.5	6.2	Detrital wood		Harvey et al. (2012)
Icemaker									50.77	-123.35	
IM05-01	5650 ± 60	6302 - 6566	0.98	6430	128	136	132	6.4	In situ stump	1730	Harvey et al. (2012)
Lillooet									50.76	-123.52	
Wk-12311	3030 ± 40	3140 - 3356	0.95	3230	90	126	108	3.2	Wood fragment from paleosol		Reyes and Clague (2004)
GSC-6746	2960 ± 60	2953 - 3260	0.94	3120	140	167	153.5	3.1	Log on paleosol		Reyes and Clague (2004)
GSC-6756	2490 ± 60	2378 - 2738	1	2570	168	192	180	2.6	Log in till		Reyes and Clague (2004)
Wk-12307	2440 ± 40	2356 - 2545	0.61	2500	45	144	94.5	2.5	Wood fragment from paleosol		Reyes and Clague (2004)
Wk-12313	2090 ± 50	1927 - 2158	0.95	2060	98	133	115.5	2.1	Charcoal from paleosol		Reyes and Clague (2004)
Wk-12306	1720 ± 40	1544 - 1715	1	1630	86	85	85.5	1.6	Charcoal from paleosol		Reyes and Clague (2004)
GSC-6767	1700 ± 80	1474 - 1813	0.92	1610	136	203	169.5	1.6	Log on paleosol		Reyes and Clague (2004)
TO-9754	1600 ± 70	1343 - 1627	0.97	1490	137	147	142	1.5	Branch from top of		Reyes and Clague (2004)

									paleosol		
Wk-12309	1550 ± 50	1344 - 1546	1	1450	96	106	101	1.5	Wood fragment from paleosol		Reyes and Clague (2004)
Wk-12310	1530 ± 40	1343 - 1526	1	1430	87	96	91.5	1.4	Twig from paleosol		Reyes and Clague (2004)
GSC-6760	1390 ± 50	1240 - 1387	0.96	1310	70	77	73.5	1.3	Log on paleosol		Reyes and Clague (2004)
Wk-12308	1090 ± 50	924 - 1090	0.93	1000	76	90	83	1	Wood fragment from paleosol		Reyes and Clague (2004)
GSC-6606	1090 ± 50	924 - 1090	0.93	1000	76	90	83	1	Branch on peat		Reyes and Clague (2004)
Manatee									50.64	-123.59	
UVTRL_MC06-07	3440 ± 60	3564 - 3857	1	3700	136	157	146.5	3.7	Detrital wood		Koehler and Smith (2011)
Orca									50.64	-123.59	
UVTRL06-MC03	4250 ± 60	4782 - 4965	0.53	4800	18	165	91.5	4.8	Detrital wood		Koehler and Smith (2011)
UVTRL_MC06-27	3500 ± 60	3631 - 3925	0.98	3770	139	155	147	3.8	Detrital wood		Koehler and Smith (2011)
Beluga									50.64	-123.59	
UVTRL06-MC23	2350 ± 70	2299 - 2544	0.71	2410	111	134	122.5	2.4	In situ sheared stump		Koehler and Smith (2011)
UVTRL06-MC04	260 ± 70	254 - 497	0.69	310	56	187	121.5	0.3	Standing snag		Koehler and Smith (2011)
Goddard									51.1	-124.17	
GSC-3219	5500 ± 70	6123 - 6145	1	6300	155	177	11	6.3			Blake (1983)
GSC-6046	4120 ± 60	4515 - 4829	0.96	4660	145	169	157	4.7	Detrital wood from forefield		Menounos et al. (2008)
GARIBALDI PROVINCIAL PARK											
Mystery									50.18	-122.75	
Beta-170664	710 ± 50	627 - 731	0.76	660	33	71	52	0.7	Stick on surface		Koch et al. (2007b)
Spearhead									50.08	-122.83	
Beta-157268	3900 ± 80	4090 -	1	4320	207	230	218.5	4.3	18-cm-long wood	1995	Osborn et al. (2007)

		4527							fragment		
Beta-168423	3900 ± 60	4152 - 4445	0.95	4330	115	178	146.5	4.3	14-cm-long wood fragment	1995	Osborn et al. (2007)
Decker									50.07	-122.83	
Beta-157265	3200 ± 70	3316 - 3579	0.91	3430	114	149	131.5	3.4	In situ sheared stump	2040	Osborn et al. (2007)
Beta-157262	2960 ± 50	2963 - 3250	0.98	3120	130	157	143.5	3.1	Detrital log	2030	Osborn et al. (2007)
Beta-157263	2960 ± 40	2985 - 3234	1	3120	114	135	124.5	3.1	Rooted snag	2010	Osborn et al. (2007)
Beta-157264	2920 ± 50	2925 - 3212	0.99	3060	135	152	143.5	3.1	In situ sheared stump	2010	Osborn et al. (2007)
Overlord									50.02	-122.83	
Beta-170665	6170 ± 70	6896 - 7249	1	7070	174	179	176.5	7.1	27-cm-long wood fragment	1640	Osborn et al. (2007)
Beta-170660	5890 ± 70	6530 - 6889	0.99	6710	179	180	179.5	6.7	Detrital bole	1625	Osborn et al. (2007)
Beta-170667	5980 ± 70	7250 - 6900	0.99	7120	130	220	175	7.1	8-cm-diameter branch	1610	Osborn et al. (2007)
Helm									49.95	-122.98	
Beta-186523	4080 ± 40	4496 - 4655	0.63	4580	75	84	79.5	4.6	In situ sheared stump		Koch et al. (2007a)
Beta-168430	8900 ± 40	9896 - 10190	1	10030	134	160	147	10	Partially buried log		Koch et al. (2007a)
Beta-208681	810 ± 60	662 - 804	0.88	730	68	74	71	0.7	Branch on surface		Koch et al. (2007b)
Beta-208282	690 ± 60	550 - 710	0.97	650	60	100	80	0.6	In situ sheared stump		Koch et al. (2007b)
Beta-168428	500 ± 40	495 - 559	0.9	530	35	29	32	0.5	Log on surface		Koch et al. (2007b)
Beta-186522	490 ± 60	428 - 568	0.77	520	48	92	70	0.5	In situ sheared stump		Koch et al. (2007b)
Beta-186526	430 ± 50	423 - 539	0.76	480	57	59	58	0.5	In situ sheared stump		Koch et al. (2007b)
Beta-208684	380 ± 40	420 - 508	0.57	470	38	50	44	0.4	In situ sheared stump		Koch et al. (2007b)
Beta-208683	330 ± 60	289 - 505	0.98	390	101	115	108	0.4	In situ sheared stump		Koch et al. (2007b)
Beta-	300 ±	271 -	0.93	380	109	121	115	0.4	Log on		Koch et al.

186524	60	501							surface		(2007b)
Beta-186525	270 ± 60	265 - 492	0.78	330	65	162	113.5	0.3	Branch between tills		Koch et al. (2007b)
Sphinx									49.92	-122.97	
Beta-185509	5830 ± 60	6490 - 6758	0.97	6640	118	150	134	6.6	Snag		Koch et al. (2007a)
Beta-208685	4280 ± 70	4783 - 5042	0.72	4850	67	192	129.5	4.9	Log		Koch et al. (2007a)
Beta-186510	3560 ± 70	3685 - 4000	0.93	3850	150	165	157.5	3.9	Snag		Koch et al. (2007a)
GSC-6770	7720 ± 80	8373 - 8649	0.99	8510	137	139	138	8.5	Branch		Menounos et al. (2004)
Y-347	460 ± 40	453 - 549	0.95	510	57	39	48	0.5	In situ sheared stump		Barendsen et al. (1957)
GSC-1993	7640 ± 80	8318 - 8596	1	8450	132	146	139	8.4	Wood (transported)	1650	Lowdon and Blake (1975)
Beta-186511	1570 ± 40	1378 - 1547	1	1460	82	87	84.5	1.5	Snag		Koch et al. (2007a)
Beta-186512	580 ± 60	518 - 660	1	590	70	72	71	0.6	In situ sheared stump		Koch et al. (2007b)
Y-347	460 ± 70	420 - 561	0.72	500	61	80	70.5	0.5	In situ sheared stump		Koch et al. (2007b)
Beta-186513	370 ± 40	420 - 504	0.53	430	10	74	42	0.4	Log in outermost moraine		Koch et al. (2007b)
Sentinel									49.9	-122.98	
Beta-186508	6040 ± 60	6732 - 7028	0.95	6890	138	158	148	6.9	Snag		Koch et al. (2007a)
GSC-1477	6170 ± 150	6726 - 7342	0.96	7050	292	324	308	7.1	Branch		Lowdon and Blake (1973)
GSC-2027	5300 ± 70	5927 - 6215	0.95	6090	125	163	144	6.1	In situ sheared stump		Lowdon and Blake (1975)
Beta-148787	7720 ± 70	8390 - 8610	0.99	8500	110	110	110	8.5	Branch		Menounos et al. (2004)
Beta-157267	7470 ± 80	8156 - 8418	0.97	8280	124	138	131	8.3	Branch		Menounos et al. (2004)
Beta-148786	7380 ± 80	8025 - 8359	1	8200	159	175	167	8.2	Branch		Menounos et al. (2004)
Warren									49.87	-123	

Beta-168425	8050 ± 60	8705 - 9124	0.99	8920	204	215	209.5	8.9	Stick		Koch et al. (2007a)
Beta-148789	6370 ± 70	7171 - 7424	1	7310	114	139	126.5	7.3	In situ sheared stump		Koch et al. (2007a)
Beta-148790	6360 ± 80	7155 - 7434	0.97	7290	135	144	139.5	7.3	In situ sheared stump		Koch et al. (2007a)
Beta-148788	5780 ± 70	6434 - 6736	0.98	6580	146	156	151	6.6	In situ sheared stump		Koch et al. (2007a)
Beta-168424	5700 ± 50	6397 - 6638	0.99	6490	93	148	120.5	6.5	Log		Koch et al. (2007a)
Beta-148791	920 ± 70	697 - 939	0.99	840	99	143	121	0.8	In situ rootstock		Koch et al. (2007b)
Lava									49.82	-123	
Y-140 bls	5850 ± 180	6301 - 7030	0.95	6680		350	364.5	6.7			Preston et al. (1955)
Y-140 bls	5260 ± 200	5598 - 6414	0.99	6030		384	408	6			Stuiver et al. (1960)
Beta-168426	6170 ± 60	6927 - 7246	0.99	7070	143	176	159.5	7.1	In situ sheared stump		Koch et al. (2007a)
Beta-168427	6050 ± 50	6749 - 7018	0.97	6900	118	151	134.5	6.9	In situ sheared stump		Koch et al. (2007a)
Beta-186521	5760 ± 60	6409 - 6677	0.99	6560	117	151	134	6.6	In situ sheared stump		Koch et al. (2007a)
Beta-186520	5130 ± 40	5844 - 5944	0.541	5880	36	64	50	5.9	Stump on surface		Koch et al. (2007a)
Beta-186517	3190 ± 40	3341 - 3483	0.97	3410	69	73	71	3.4	Log in wood mat		Koch et al. (2007a)
Beta-186518	860 ± 70	682 - 917	1	780	98	137	117.5	0.8	Log in wood mat		Koch et al. (2007b)
Beta-157266	640 ± 70	527 - 688	1	610	78	83	80.5	0.6	Log in lateral moraine		Koch et al. (2007b)
Beta-186519	640 ± 50	545 - 672	1	600	55	72	63.5	0.6	Log in lateral moraine		Koch et al. (2007b)
Beta-186516	530 ± 40	506 - 563	0.69	540	23	34	28.5	0.5	In situ sheared stump		Koch et al. (2007b)
Stave									49.75	-122.53	
Beta-170668	6250 ± 70	6969 - 7315	1	7170	145	201	173	7.2	Detrital stump		Koch et al. (2007a)
Beta-171096	1080 ± 60	909 - 1176	0.99	1000	91	176	133.5	1	Small log		Koch et al. (2007a)

Beta-171095	830 ± 30	688 - 789	1	740	49	52	50.5	0.7	Log in lateral moraine		Koch et al. (2007b)
Beta-171094	310 ± 50	289 - 494	0.98	380	91	114	102.5	0.4	Branch		Koch et al. (2007b)
Beta-170669	250 ± 50	263 - 343	0.34	300	37	43	40	0.3	Log in lateral moraine		Koch et al. (2007b)
Garibaldi									44.82	-122.95	Koch et al. (2007b)
Beta-186515	670 ± 70	537 - 709	0.98	630	79	93	86	0.6	In situ sheared stump		Koch et al. (2007b)
Beta-208680	490 ± 70	426 - 572	0.7	520	52	94	73	0.5	In situ sheared stump		Koch et al. (2007b)
Beta-186514	290 ± 60	267 - 500	0.89	370	103	130	116.5	0.4	Log in till		Koch et al. (2007b)

[1] – Based on the difference of the median calibrated date and ±2σ calibrated range.
[2] – Only the highest probability ±2σ calibrated range is provided.
[3] – Rounded to the nearest 10.
[4] - When provided by the corresponding published article.

Chapter Three: Latest Pleistocene and Holocene behaviour of Franklin Glacier, Mt. Waddington area, British Columbia Coast Mountains, Canada

Statement of Co-Authorship

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3.1 Introduction

The latest Pleistocene and Holocene behaviour of glaciers in the British Columbia Coast Mountains reflects mass balance responses to changing regional and global climates (Menounos et al., 2009). Most glaciers in the region attained their maximum late Holocene extent at the end of the Little Ice Age (Larocque and Smith, 2003; Harvey and Smith, 2013), after which they experienced significant volumetric losses (Koch et al., 2007; VanLooy and Forster, 2008; Tennant et al., 2012; Koch et al., 2014). The resultant downwasting and terminus retreat has revealed the remains of glacially overridden and buried forests below till and outwash deposits, enabling description of the spatial and temporal character of latest Pleistocene and Holocene glacier activity (e.g., Koch et al., 2007; Harvey et al., 2012; Hoffman and Smith, 2013).

Previous glacial history investigations in the Mt. Waddington area of the central Coast Mountains (Figure 3.1a) show that glaciers repeatedly advanced into mature standing forests throughout the Holocene (Ryder and Thomson, 1986; Larocque and Smith, 2003; Coulthard et al., 2013; Craig and Smith, 2013; Menounos et al., 2013). Based on present-day mass balance-climate relationships (Larocque and Smith, 2005a;

Wood et al., 2011), these Holocene advances likely occurred in response to persistent intervals of cool summers and wet winters similar to those that resulted in glacial advances during the Little Ice Age (Larocque and Smith, 2005b; Pitman and Smith, 2012). Ameliorating climate following each advance would have caused glaciers to downwaste and retreat, allowing forests to repeatedly colonize the deglaciated valley bottoms and lateral moraine slopes.

Despite a growing appreciation of the Holocene behaviour of glaciers in the Mt. Waddington area, complementary studies undertaken elsewhere in the Coast Mountains indicates that our understanding of this activity remains incomplete (e.g., Menounos et al., 2009). In this paper, I characterize the latest Pleistocene and Holocene behaviour of Franklin Glacier as revealed by radiocarbon and tree-ring analyses of subfossil wood remains collected in the summers of 2012 and 2013. On the basis of this evidence, I describe the behaviour of glaciers in the Mt. Waddington area from 13,000 cal yr BP to present.

3.2 Study area

Geographic setting

Franklin Glacier is an 18-km long valley glacier that originates in a broad icefield at 2150 m asl below the west face of Mt. Waddington (4019 m asl) on the windward side of the Pacific Ranges (51°16’ N, 125°23’ W) (Figures 3.1a, b). Encompassing an area of approximately 175 km^2 (Bolch et al., 2010), Franklin Glacier originates from the confluence of smaller tributary glaciers before flowing down valley to its present terminus at 650 m asl, approximately 15 km from Knight Inlet (Figure 3.1a). The

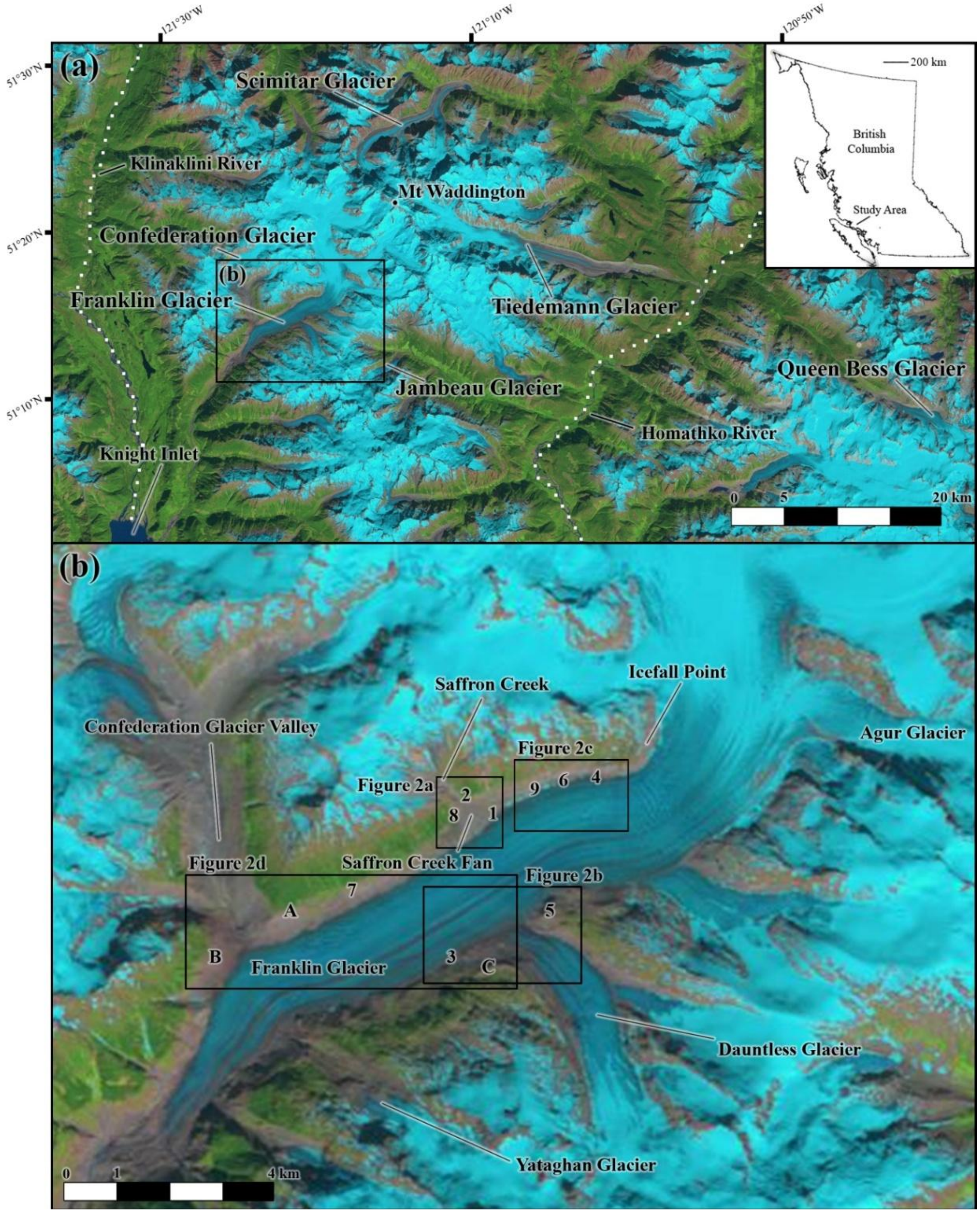

Figure 3.1 – (a) Mt. Waddington area highlighting location of previously studied glaciers in the region and geographic features and (b) Franklin Glacier study area. Numbers indicate sampling sites from this study and letters indicate samples sites from Rutherford et al. (1984), Ryder and Thomson (1986), and Coulthard et al. (2013).

regional climate of the Mt. Waddington area is moderated by close proximity to the Pacific Ocean, with a mean annual temperature of 3.6°C and mean annual precipitation of 1870 mm/yr from 1981 to 2010 (ClimateBC v. 5.00 in Wang et al., 2012). Above 1000 m asl vegetated mountainsides are forested by mature subalpine fir (*Abies lasiocarpa*) and mountain hemlock (*Tsuga mertensiana*) to the local tree line at 1500-1600 m asl.

Research background

In 1927, a party of mountaineers were the first to traverse the length of Franklin Glacier while attempting to summit Mt. Waddington (Munday, 1926-1927). Over the next two decades a succession of expeditions followed the same route, leaving behind a rich descriptive and photographic record providing insight into the response of Franklin Glacier to climate changes in the first half of the 20th Century (Munday, 1926-1927, 1930, 1931a, 1931b, 1933, 1934-1935, 1939).

By 1927 Franklin Glacier had retreated almost 2 km up from terminal moraines deposited during a 'recent' advance (Munday, 1926-1927). Munday (1939: 2-3) reports that Franklin Glacier deposited these "moraines against a mature forest containing trees that must be several centuries old", noting that "the recent advance exceeded any previous ones that have occurred for several centuries". Munday (1931b: 133) also noted that the "[…] ice-front remained nearly stationary [at this point] for some time" with "a number of short retreats, followed by progressively rapid shrinkage. Between June, 1927, and July, 1931, this amounted to approximately 900 feet [275 m] […] along the whole half-mile [800 m] front."

Since 1931, the glacier terminus has retreated over 4 km up valley (Munday, 1934-1935; VanLooy and Forster, 2008) in response to persistent negative mass balance conditions generated by increased summer temperatures (Larocque and Smith, 2005a, 2005b). This period of general retreat has been interrupted by short stillstands and brief intervals of expansion during periods of below average temperatures (Munday, 1939; VanLooy and Forster, 2008).

Confederation Glacier flows southwest down a bedrock-confined valley and, until ca. 1960, was confluent with Franklin Glacier (Coulthard et al., 2013). Photographs from 1927 show the surface of Confederation and Franklin glaciers positioned within 50 m of the Little Ice Age lateral moraine crest at 1255 m asl (Munday 1926-1927). Investigations by Ryder and Thomson (1986) and Coulthard et al. (2013) close to the former Confederation-Franklin glacier confluence describe the early Little Ice Age expansion of Franklin Glacier from 0.8 to 0.6 ka (Sites A and B, Figure 3.1a). At Site A, located at 1220 m asl, Ryder and Thomson (1986) reported an age of 835 ± 45 ^{14}C yr BP (0.8 ka; Table 3.1) on a mountain hemlock root in till (S-1568). Coulthard et al. (2013) obtained an age of 600 ± 50 ^{14}C yr BP (0.6 ka; Table 3.1) on the perimeter rings of a nearby mountain hemlock stump (A01) rooted in a paleosol immediately below a till unit. At Site B, located down valley at 1180 m asl in Franklin Glacier valley, Coulthard et al. (2013) describe a laterally contiguous wood mat (> 30 m wide) exposed in the collapsed proximal face of the lateral moraine. Perimeter rings from a large log that fell from the wood mat to the valley floor (G05) yielded an age of 550 ± 50 ^{14}C yr BP (0.5 ka; Table 3.1). While investigations in Confederation Glacier valley indicate that Confederation

Glacier advanced into standing forests at 5.6, 3.8, 3.5, and 3.4 ka (Coulthard et al. 2013), there are no reports of contemporaneous Holocene advances at Franklin Glacier.

3.3 Methods

The proximal slopes of lateral moraines flanking Franklin Glacier up valley of the Confederation-Franklin glacier confluence were surveyed for dendroglaciological evidence with a handheld GPS in 2012 and 2013. Exposed paleosols were located and representative samples of 30 buried subfossil wood remains were sampled with a chainsaw. When possible, I collected bark and perimeter wood to allow for kill-date determination by tree-ring cross-dating or radiocarbon dating (Coulthard and Smith, 2013). Wood samples were wrapped for transport to the University of Victoria Tree-Ring Laboratory where they were allowed to air-dry before being sanded to a fine polish for tree-ring measurement. A high-resolution scanner (1200 dpi) was used to create a digital image of each sample and the annual radial growth rings were measured using WinDendro software (v. 2012b; precision = 0.001 mm; Guay et al., 1992). Bark and anatomical characteristics were used for species identification (Hoadley, 1990).

Floating tree-ring chronologies were developed by internally cross-dating and compiling ring-width series from individual logs to produce site-specific series using standardized methodologies and COFECHA chronology quality control software (Fritts, 1976; Grissino-Mayer, 2001). Attempts were made to cross-date undated ring-width series to local living and floating tree-ring chronologies developed by Coulthard et al. (2013). Where cross-dating failed, representative perimeter wood samples were radiocarbon dated using conventional methods by Beta Analytic Inc. to assign a relative kill date. The ^{14}C ages from this and previous studies were calibrated using INTCAL13 (Calib v. 7.02; Stuiver and Reimer, 1993; Reimer et al., 2013) with ±2σ error limits, with the calibrated ^{14}C age designated as the minimum kill date (Coulthard and Smith, 2013).

3.4 Results

Dendroglaciological samples were collected from sites along both flanks of Franklin Glacier, from Icefall Point along the south-facing lateral moraine across the Saffron Creek fan to the confluence with Confederation Glacier valley (sites 1, 2, 4, 6, 7, 8, 9), and from Agur Glacier to Yataghan Glacier along the north-facing lateral moraine (sites 3 and 5) (Figure 3.1a; Table 3.1).

Site 1

Site 1 is located ca. 20 m above the 2013 ice surface within sediments exposed by stream erosion on the Saffron Creek fan along the northern flank of Franklin Glacier at 1340 m asl (Figure 3.2a; Table 3.1). The remains of two logs (FG13-23, FG13-24) were found pressed into the organic surface of a buried paleosol located in a gully sidewall ca. 100 m below the Little Ice Age maximum ice level position and adjacent to the proximal face of bedrock outcrop. The paleosol dips toward Franklin Glacier and includes a black Oh or Ah horizon (< 5 cm) with visible fir needles, a thin (< 1 cm) underlying grey-coloured incipient eluviated horizon, and a red-brown B horizon (10-15 cm) developed on the surface of a sandy diamict unit. The logs were found within 2 m of each other below a remnant cap of diamict (>1.5-2 m thick) mantled by recently-deposited alluvial gravels. The diamict units are inferred to be tills but detailed sedimentological data were not obtained to support this interpretation. Both boles were oriented down valley towards

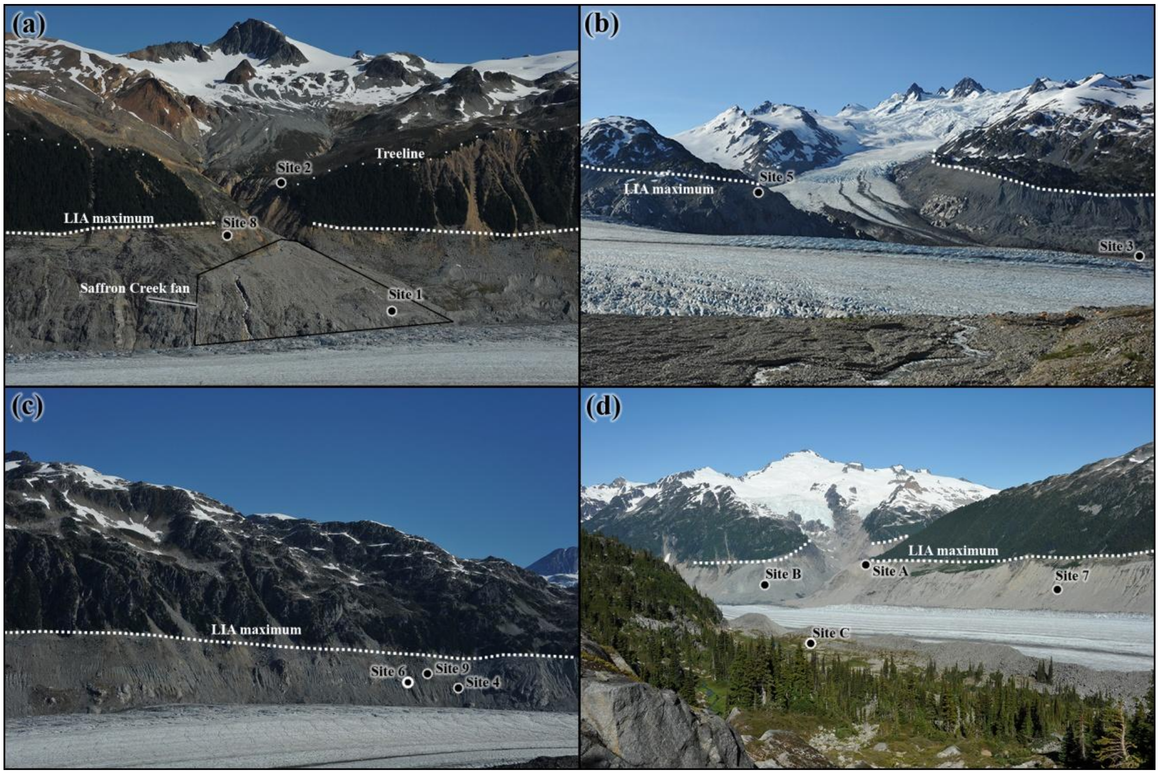

Figure 3.2 – Site photographs showing location of study sites above the 2013 ice surface. Shown in: (a) northern flank of Franklin Glacier in vicinity of the Saffron Creek fan. Position of contemporary treeline and Little Ice Age trimline highlighted; (b) the Dauntless-Franklin glacier confluence showing Little Ice Age trimline; (c) northern flank of Franklin Glacier down valley from Icefall Point; and, (d) foreground shows location of drained moraine-dammed lake adjacent to Dauntless-Franklin glacier confluence. Background illustrates northern flank of Franklin Glacier in vicinity of former confluence with Confederation Glacier. Highlighted is position of Little Ice Age trimline.

Table 3.1 – Summary of Holocene radiocarbon dated wood and peats from the Mt Waddington area. LIA moraine and 2013 ice surface elevations only provided for Franklin Glacier.

Glacier name/Lab no. [1]	Sample no.	Description [2]	14C Age[3]	Calibrated age (cal yr BP) [4]	Source
Franklin Glacier					
Beta-362290	FG13-23	detrital bole pressed into paleosol	10920 ± 50	12910 - 12700	This study
Beta-364401	FG13-24	detrital bole pressed into paleosol	10880 ± 50	12830 - 12690	This study
Beta-367878	FG13-11	detrital root encased in a paleosol	8910 ± 40	10190 - 9910	This study
Beta-362284	FG13-01	detrital bole in till	5530 ± 30	6360 - 6280	This study
Beta-362288	FG13-20	wood mat at diamict contact	4680 ± 30	5470 - 5320	This study
Beta-3622287	FG13-17	wood mat at diamict contact	4600 ± 30	5330 - 5280	This study
Beta-362286	FG13-09	in situ sheared stump	4150 ± 30	4770 - 4580	This study
Beta-327741	FG12-01	wood mat pressed into paleosol	3790 ± 30	4260 - 4080	This study
Beta-362290	FG13-25	masticated bole	2960 ± 30	3210 - 3020	This study
Beta-362285	FG13-05	wood mat at diamict contact	2460 ± 30	2620 - 2380	This study
Beta-327742	FG12-03	in situ sheared stump	1630 ± 30	1570 - 1480	This study
S-1568	NA	root in paleosol	835 ± 45	800 - 680	Ryder and Thomson (1986)
Beta-250500	E07	upper gullies	670 ± 40	600 - 560	Coulthard et al. (2013)
Beta-250496	A01	in situ sheared stump	600 ± 50	660 - 530	Coulthard et al. (2013)
Beta-250498	G05	detrital bole	550 ± 50	570 - 510	Coulthard et al. (2013)
S-1475	NA	wood from gully exposure	150 ± 70	290 - 0	Rutherford et al. (1984)
Confederation Glacier					
Beta-257238	UVTRL-B04C	detrital bole	4850 ± 50	5660 - 5570	Coulthard et al. (2013)
Beta-259835	UVTRL-C01B	bole	3470 ± 70	3910 - 3570	Coulthard et al. (2013)
Beta-250499	UVTRL08-C06B	stump in growth position	3370 ± 50	3720 - 3480	Coulthard et al. (2013)
Beta-250497	UVTRL08-C04I	detrital bole	3220 ± 60	3580 - 3340	Coulthard et al. (2013)

Jambeau Glacier					
Beta-262874	UVTRL09-JG02	detrital bole	3030 ± 60	3380 - 3060	Coulthard et al. (2013)
Beta-262875	UVTRL09-JG05	detrital bole	2780 ± 50	3000 - 2770	Coulthard et al. (2013)
Beta-262873	UVTRL09-JG01	rooted stump	230 ± 40	220 - 140	Coulthard et al. (2013)
Queen Bess Glacier					
Beta-327746	UVTRL12-QB11	detrital bole on soil	1470 ± 30	1400 - 1310	Smith (*unpublished data*)
Beta-327743	UVTRL12-QB04	colluvial debris from wood mat	860 ± 30	800 - 690	Smith (*unpublished data*)
Beta-327744	UVTRL12-QB05	colluvial debris from wood mat	820 ± 30	780 - 690	Smith (*unpublished data*)
Beta-327745	UVTRL12-QB08	in situ stump	790 ± 30	750 - 670	Smith (*unpublished data*)
Scimitar Glacier					
Beta-310688	UVTRL011-S11/03/04	detrital bole	2910 ± 30	3080 - 2960	Craig and Smith (2013)
Beta-310686	UVTRL011-S11/03/02	detrital bole	2740 ± 30	2890 - 2760	Craig and Smith (2013)
Beta-310689	UVTRL011-S11/04/02	detrital bole	2650 ± 30	2800 - 2740	Craig and Smith (2013)
Beta-310687	UVTRL011-S11/03/03	wood fragment	1610 ± 30	1560 - 1410	Craig and Smith (2013)
Beta-310690	UVTRL011-S11/04/03	wood fragment	230 ± 30	3200 - 140	Craig and Smith 2013)
Tiedemann Glacier					
GSC-939	NA	basal peat	9510 ± 150	11210 - 10410	Fulton (1971)
TO-10757	NA	wood	8760 ± 100	10190 - 9530	Arsenault et al. (2007)
Beta-2020941	NA	log	5010 ± 40	5800 - 5650	Menounos et al. (2009)
UCIAMS-40661	NA	in situ stump	3940 ± 20	4440 - 4340	Menounos et al. (2009)
UCIAMS-40663	NA	in situ stump	3865 ± 20	4410 - 4230	Menounos et al. (2009)
UCIAMS-40660	NA	in situ stump	3820 ± 20	4288 - 4150	Menounos et al. (2009)
Beta-220940	NA	in situ stump	3760 ± 60	4300 - 3960	Menounos et al. (2009)
Beta-220936	NA	in situ stump	3690 ± 50	4160 - 3890	Menounos et al. (2009)
S-1470	NA	wood mat	3345 ± 115	3890 - 3350	Ryder and Thomson (1986)
CAMS-100585	NA	needles	3290 ± 60	3690 - 3380	Arsenault et al. (2007)

GSC-938	NA	peat	2940 ± 130	3380 - 2790	Ryder and Thomson (1986)
UCIAMS-40662	NA	in situ stump	2820 ± 20	2970 - 2860	Menounos et al. (2009)
Beta-220939	NA	in situ stump	2710 ± 40	2880 - 2750	Menounos et al. (2009)
Beta-220937	NA	in situ stump	2670 ± 50	2870 - 2740	Menounos et al. (2009)
TO-10756	NA	needles	2530 ± 50	27520 - 2360	Arsenault et al. (2007)
Beta-220938	NA	in situ stump	2520 ± 50	2750 - 2430	Menounos et al. (2009)
S-1471	NA	log with bark	2355 ± 60	2410 - 2300	Ryder and Thomson (1986)
TO-10755		needles	2290 ± 50	2360 - 2150	Arsenault et al. (2007)
GSC-948	NA	peat	2250 ± 130	2540 - 1950	Ryder and Thomson (1986)
CAMS-100584	NA	needles	1880 ± 40	1920 - 1710	Arsenault et al. (2007)
CAMS-101001	NA	needles	1850 ± 35	1870 - 1710	Arsenault et al. (2007)
S-1473	NA	masticated log	1330 ± 65	1350 - 1170	Ryder and Thomson (1986)
GSC-977	NA	basal peat	1270 ± 140	1420 - 920	Ryder and Thomson (1986)
UCIAMS-40664	NA	wood mat	365 ± 20	500 - 420	Menounos et al. (2009)
CAMS-100583	NA	needles	335 ± 40	500 - 300	Arsenault et al. (2007)
S-1474	NA	masticated log	300 ± 60	500 - 270	Ryder and Thomson (1986)
S-1472	NA	log with bark	65 ± 100	160 - 0	Ryder and Thomson (1986)

[1] – Prefixes: Beta = Beta Analytics Inc; S = Saskatchewan Research Council; GSC = Geological Survey of Canada; UCIAMS/CAMS = University of California; TO = Isotrace Laboratory, University of Toronto.
[2] – The term in situ is used to describe samples entrained in glacial sediments and not spilled or washed to a secondary location. Detrital refers to the possibility of the sample being transported prior to being buried by sediments. Sample descriptions from other studies are taken directly from their respective text. Inferences on the nature or state of each sample in other studies cannot be made.
[3] – All dates were obtained using conventional radiocarbon dating methods.
[4] – ±2σ calibration using INTCAL2013 (Stuiver and Reimer, 1993; Reimer et al., 2013).

the glacier terminus. Bark remained on the underside of the trunks and anatomical characteristics indicate both were the remains of young subalpine fir trees (ca. 90 and 52 years old; Table 3.1). There was insufficient evidence to determine whether the trunks were the remains of trees transported by snow avalanches or, alternatively, are the remains of young subalpine forest that once colonized the buried paleosol. Thirty perimeter rings from FG13-23 yielded a ^{14}C age of 10920 ± 50 (12.8 ka); 28 perimeter rings from FG13-24 yielded ^{14}C age of 10880 ± 50 (12.8 ka; Table 3.1). Attempts to cross-date the samples failed due to the limited number of annual rings.

Site 2

Site 2 is located within sediments exposed in the scarp-face of a rotational slump at 1560 m asl, 100 m above the Little Ice Age maximum ice level position (Figures 3.2a, 3.3a). The slump exposed a >5 m thick vertical section of diamict, inferred to be till, overlain by 15-30 cm thick layer of weakly-bedded silts and fine sands with occasional subangular stones. A modern soil varying in thickness from 0.75-1.0 m overtops the sequence. Detrital subalpine fir remains of unknown origin are buried within the bedded colluvial unit (<15 cm above the diamict section) and a sample with ca. 40 annual rings was collected (FG13-11). Twenty-seven perimeter rings from FG13-11 yielded a ^{14}C age of 8910 ± 40 (10.2 ka; Table 3.1).

Site 3

Site 3 is 900 m down valley from the confluence of Dauntless and Franklin glaciers at 1220 m asl (Figures 3.2b, 3.3b). Ice-marginal stream erosion exposed >5 m of bedded sands and subangular gravels overlain by >2 m of matrix-supported, diamict with

Figure 3.3 – Dendroglaciological sample sites at Franklin Glacier, Shown in: (a) Site 2 illustrating modern soil and position of FG13-10 exposed above modern treeline by recent rotational slump adjacent to Saffron Creek; (b) large trunk exposed by ice-marginal erosion >5 m below the summer 2013 ice surface at Site 3; (c) location and gully characteristics at site 4 south of Icefall Point showing where sample FG13-17 was collected; (d) rooted and glacially sheared stump (FG13-09) found in growth position at Site 5 on southern lateral moraine of Franklin Glacier; (e) vertical gully at Site 6 along the proximal face of the northern lateral moraine where FG12-01 was found resting on a weakly-expressed paleosol; (f) Site 8 illustrating glacially-pressed logs in gully. Logs are oriented down valley (right to left) and lie below a mantle of diamict; (g) in situ glacially-sheared stump at Site 9 down valley from Icefall Point; (h) Bedded-lacustrine sands and silts deposited distal to Little Ice Age moraine at Site C.

sub-rounded clasts, inferred to be till. I sampled a large (ca. 1 m diameter) detrital subalpine fir trunk (FG13-01; Table 3.1) protruding into the abandoned channel at the unit contact. While the origin of the trunk is unknown, its orientation and position immediately below diamict are suggestive of burial by advancing ice. Perimeter wood yielded an age of 5530 ± 30 ^{14}C yr BP (6.3 ka; Table 3.1)

Site 4

Site 4 is located at 1450 m asl downstream of Icefall Point, 40 m below the Little Ice Age lateral moraine crest and 60 m above the 2013 ice surface (Figures 3.2c, 3.3c). A large gully bisects a laterally continuous wood mat separating two matrix-supported diamict units thought to be tills; 11 trunk samples were collected (0.25 to 1 m in diameter, up to 4 m in length) from the wood mat (FG13-12 to FG13-22). With the exception of one mountain hemlock trunk (271 rings, FG13-17), the samples came from mature subalpine fir trunks (>70 - 185 rings). Thirty perimeter rings from FG13-20 and FG13-17 returned ages of 4680 ± 30 ^{14}C yr BP (5.4 ka) and 4600 ± 30 ^{14}C yr BP (5.3 ka; Table 3.1), respectively. Samples from the wood mat cross-date to form a floating chronology spanning 308 years (Figure 3.4a).

Site 5

Site 5 is located between two large bedrock outcrops up valley from the Dauntless-Franklin Glacier confluence at 1410 m asl (Figures 3.2b, 3.3d). Sixteen perimeter rings recovered from a rooted and glacially sheared stump (FG13-09) found in growth position on diamict provided an age of 4150 ± 30 ^{14}C yr BP (4.6 ka; Table 3.1).

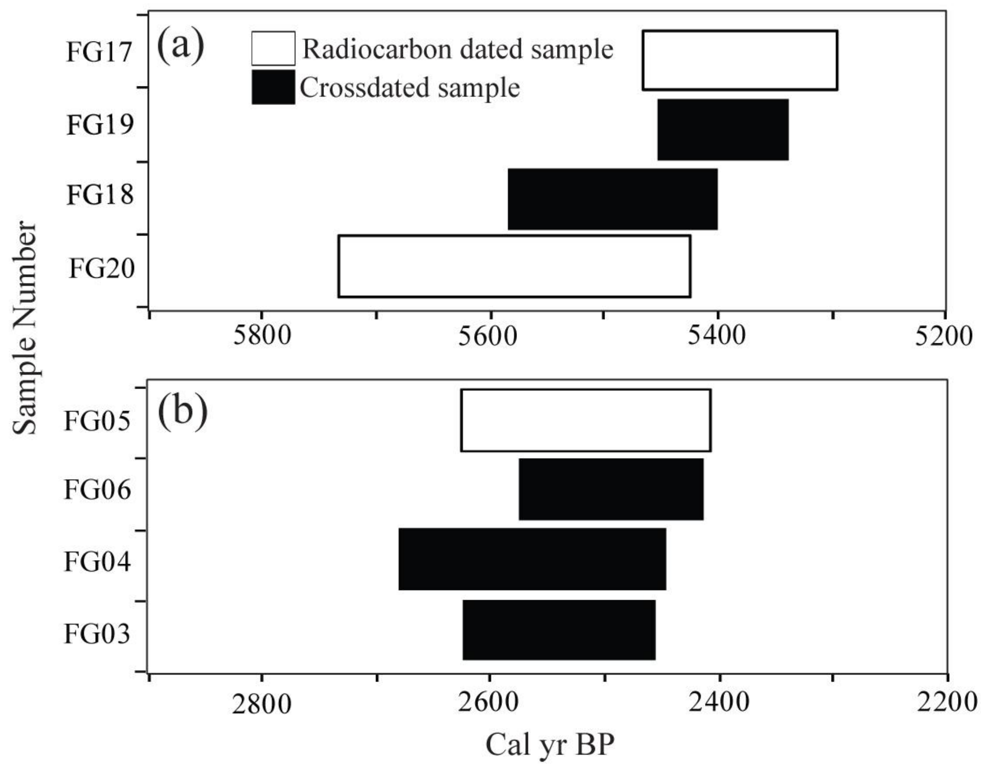

Figure 3.4 – Floating tree-ring chronologies from: (a) Site 4 (series intercorrelation = 0.425; mean sensitivity = 0.209); and (b) Site 7 (series intercorrelation = 0.512; mean sensitivity = 0.226).

Site 6

Site 6 is located downstream from Icefall Point at 1480 m asl (Figures 3.2c, 3.3e), ca. 50 m down valley from site 4. An approximately 70 m long vertical gully bisects the lateral moraine and exposes the remains of numerous small-diameter (<30 cm) trunks on a weakly-expressed paleosol at the contact between two matrix-supported diamict units of different colour, interpreted to be two different till units. Perimeter rings from FG12-01 returned an age of 3790 ± 30 ^{14}C yr BP (4.1 ka; Table 3.1).

Site 7

Site 7 is located 1000 m down valley from the Saffron Creek fan at ca. 1300 m asl (Figure 3.2d). A gully cuts through a lateral moraine adjacent to bedrock and has exposed numerous down valley oriented, small-diameter (<15 cm) subalpine fir trunks with bark, pressed into the surface of sloping red-coloured paleosol below a matrix-supported diamict unit. Perimeter rings from FG13-25 yielded an age of 2960 ± 30 ^{14}C yr BP (3.1 ka; Figure 3.2b; Table 3.1).

Site 8

Site 8 is located at 1430 m asl in a steep gully above the Saffron Creek fan and ca. 10 m below the Little Ice Age moraine crest (Figures 3.2a, 3.3f). Five subalpine fir logs (0.25-0.5 m in diameter; >161 - 179 rings) were found partially damming a gully (FG13-03 to 13FG-07). The logs are oriented down valley, lie below a mantle of diamict inferred to be till, and are glacially-pressed into an underlying diamict unit. Twenty-four perimeter rings from FG13-05 yielded an age of 2460 ± 30 ^{14}C yr BP (2.4 ka; Table 3.1). A floating chronology developed from the five logs spans 264 years (Figure 3.4b).

Site 9

Site 9 is a bedrock outcrop at 1520 m asl located downstream of Icefall Point between sites 4 and 6 (Figures 3.2c, 3.3g). Several small sheared subalpine fir stumps were found rooted on rock at this site and a perimeter sample from FG12-03 yielded an age of 1630 ± 30 ^{14}C yr BP (1.5 ka; Table 3.1).

3.5 Synthesis and regional correlation

The latest Pleistocene history of Franklin Glacier prior to the Younger Dryas is presumed to correspond to that of other trunk valley glaciers in this region (Margold et al., 2013). Based upon interpretation that the pedogenic surface at site 1 developed during an ice-free interval before 12.8 ka, prior to this time Franklin Glacier would have retreated from Knight Inlet to at least the current ice level at site 1 (< 1340 m asl). While it is possible that the trunks found pressed into paleosol at site 1 represent the remains of eroded deposits transported downslope and buried beneath colluvial diamict, their condition and down valley orientation, rather than downslope, is consistent with them having been buried by an expanding glacier in 12.8 ka during a well-documented period of regional cooling in coastal British Columbia (Mathewes et al., 1993; Lacourse et al., 2012) that corresponds to the cool, humid phase of the Younger Dryas (Bakke et al., 2009; Broecker et al., 2010; Murton et al., 2010). This period of changing environmental conditions corresponds to a time when glaciers were expanding worldwide (Golledge, 2010; MacLeod et al., 2011; Lohne et al., 2012; Osborn et al., 2012), and is benchmarked by changes in δ^{18}O levels and temperatures in the Greenland GISP2 ice cores and sea surface temperatures in tropical and high-latitude regions (Alley, 2000; Lea et al., 2003; Figure 3.5).

If Franklin Glacier was expanding down valley in 12.8 ka as posited, it was doing so at a time when trunk valley glaciers located elsewhere in the Coast Mountains had already reached their maximum Younger Dryas positions (Clague, 1985; Friele and Clague, 2002a, 2002b). This differential behaviour is assumed to reflect the overriding influence of regional topography on ice dynamics as the decaying Cordilleran Ice Sheet transitioned to ice lobes in major valleys (Menounos et al., 2009; Margold et al., 2013). In the case of Franklin Glacier, the high-elevation Mt. Waddington massif may have prevented resurgent trunk valley glaciers in the surrounding Homathko and Klinaklini river valleys from contributing to the flow of Franklin Glacier during the Younger Dryas. The local source area and size of the tributary glaciers contributing to Franklin Glacier suggests it likely experienced exaggerated post-glacial retreat before responding to the Younger Dryas cold event and readvancing down valley (Friele and Clague, 2002a). If the samples were killed and transported by mass wasting, although unlikely from my interpretations, Franklin Glacier's ice level conditions may have been still been substantially lower that other documented locations in the Coast Mountains with trees being established on proximal moraine slopes. It remains to be determined whether Franklin Glacier continued to thicken after 12.8 ka to deposit the diamict observed at Site 2. By 10.2 ka, however, the terrain above 1560 m asl was ice-free and colonized by subalpine fir trees at or above the present-day treeline. This colonization occurred at a time when local treelines elsewhere in the Coast Mountains were at least 60 m higher

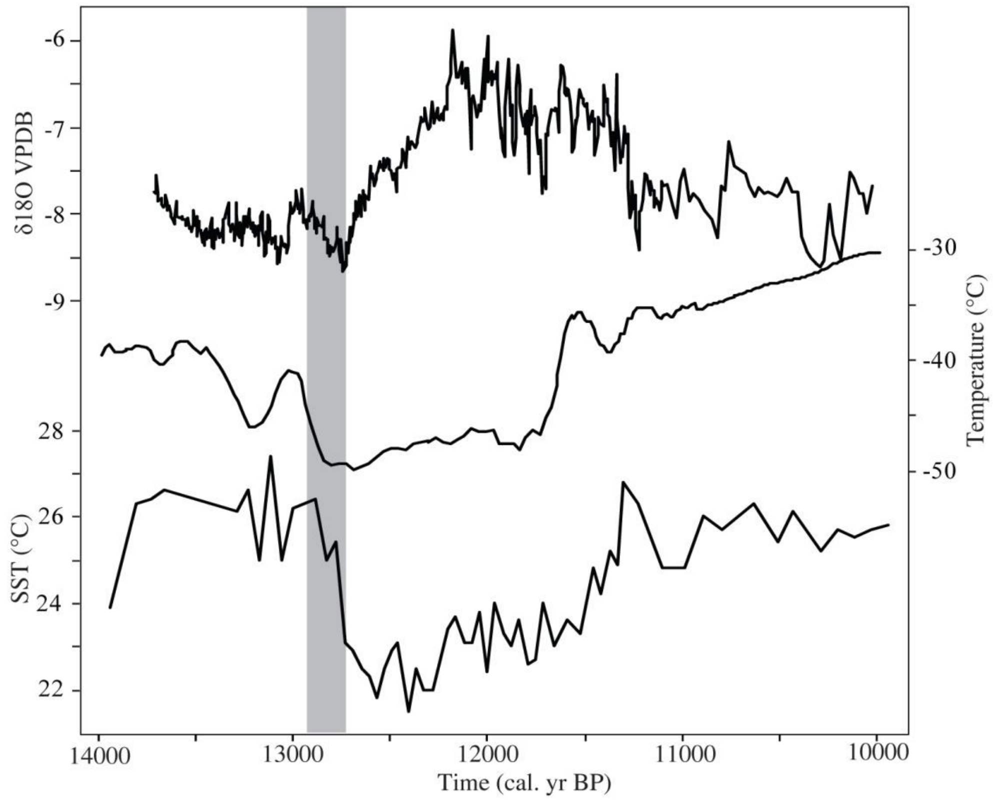

Figure 3.5 - Reconstructed paleoclimate indicators from ca. 14.0 to 10.0 ka. The grey bar presents the calibrated radiocarbon age (±2σ) of Younger Dryas-age samples (FG13-23, FG13-24) from Franklin Glacier. The $\delta^{18}O$ and temperature record is derived from Greenland GISP2 ice cores (Alley, 2000) and the sea surface temperature record based on paleo-forminifera from Cariaco Basin, northern Venezuelan Shelf (Lea et al., 2003).

than present as a result of higher-than-present temperatures during the early Holocene (Clague and Mathewes, 1989).

Following the Younger Dryas, during the early Holocene warm period, most Coast Mountain glaciers decayed rapidly, disappearing entirely (Clague and James, 2002) or retreating to terminal positions close to those reached during the Little Ice Age (Menounos et al., 2009). For example, by 10.6 ka the Squamish and Mamquam tributary glacier valleys in the southern Coast Mountain region were completely ice-free (Friele and Clague, 2002a).

The first of several documented middle Holocene advances at Franklin Glacier occurred at 6.3 ka when the glacier buried the remains of large tree trunks below diamict at site 3. This interval is regionally characterized by lower-than-present July temperatures (Gavin et al., 2011) and long, cool springs (Galloway et al., 2011) (Figure 3.6), conditions that led to glaciers expanding into standing forests elsewhere in the Coast Mountains (Koch et al., 2007; Harvey et al., 2012).

By 5.3 ka, Franklin Glacier was expanding into a forest of mature trees at 1450 m asl (site 4), as were nearby Confederation and Tiedemann glaciers (Menounos et al., 2009; Coulthard et al., 2013) (Table 3.1). This period of glacial expansion is recorded at several other Coast Mountain glaciers (Harvey et al., 2012) and was presumably a response to the establishment of cool, moist conditions (Galloway et al., 2011; Gavin et al., 2011) which also saw decreased forest fire activity in coastal and interior British Columbia (Hallet et al. 2003; Figure 3.6).

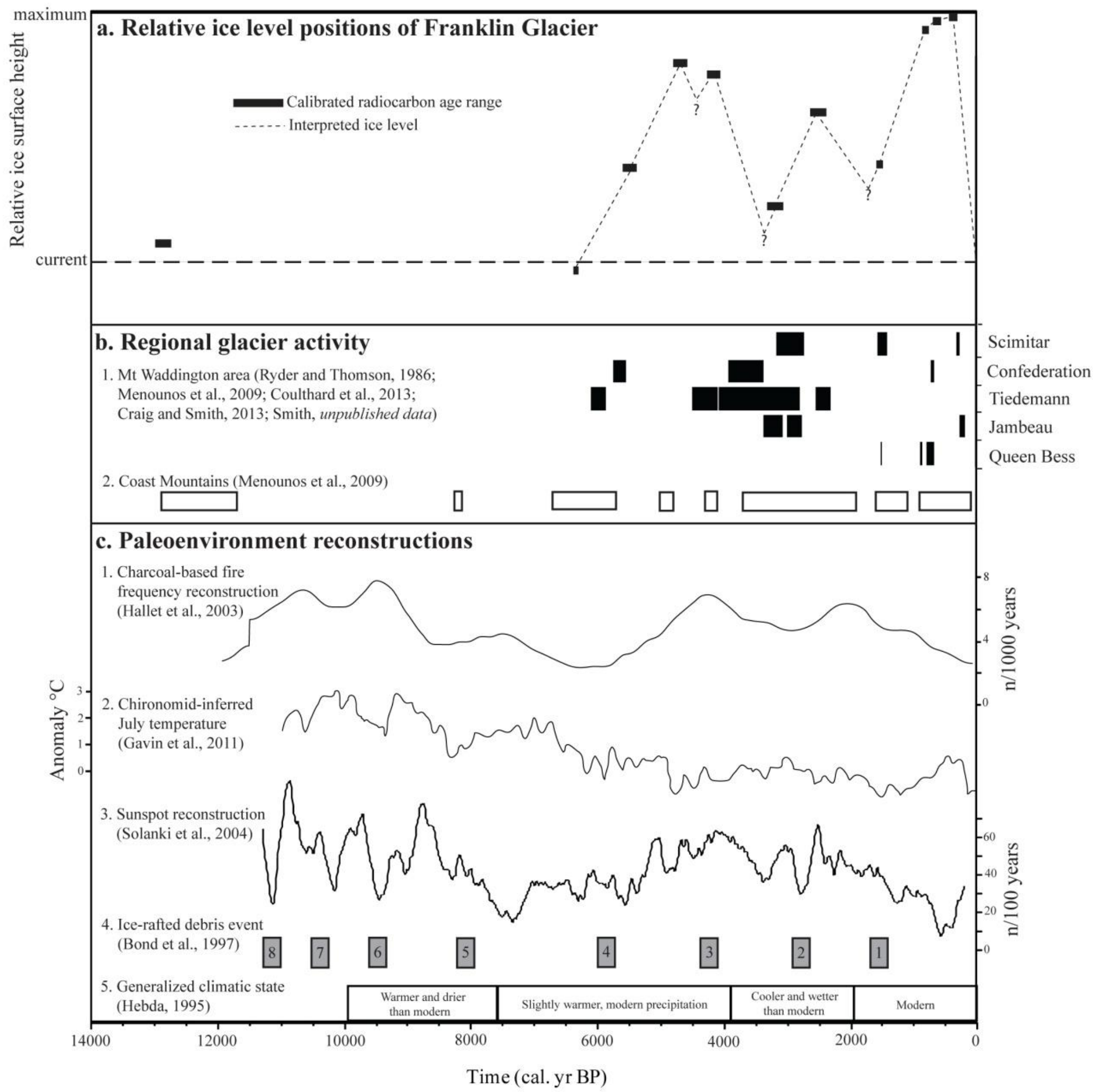

Figure 3.6 - Radiocarbon-dated evidence at Franklin Glacier compared to regional and global proxy reconstructions and other evidence of glacier activity. Shown in: (a) relative ice surface height of Franklin Glacier over the past 14,000 years. Information derived from radiocarbon dated wood samples and positions on the lateral moraine, where current the 2013 ice surface position and maximum is the greatest extent during the Little Ice Age (see Table 1 for details); (b) [1.] dated glacier advances at other sites in the Mt. Waddington area (see Table 1 for details). Black bars represent ±2σ error limits of calibrated ^{14}C dates. [2.] Regionally recognized phases of positive mass balance (Menounos et al., 2009); and (c) [1.] fire frequency record from the southern Coast Mountains (Hallet et al., 2003), [2.] July temperature anomaly record from interior British Columba (Gavin et al., 2011), [3.] global sunspot reconstruction (Solanki et al., 2004), [4.] ice-rafted debris events (Bond et al., 1997), and [5.] generalized climatic states in southwest British Columbia (Hebda, 1995).

Following the advance at 5.3 ka, Franklin Glacier expanded close to its maximum LIA position by 4.6 ka (1410 m asl) during a period distinguished by cooler summer temperatures and increased winter precipitation (Hebda, 1995; Figure 3.6). These conditions were widespread (Galloway et al., 2011) and initiated glacier advances throughout the Coast Mountains (Menounos et al., 2009; Koehler and Smith, 2011; Harvey et al., 2012).

After the 4.6 ka, advance Franklin Glacier thinned before expanding up the valley walls in 4.1 ka during an interval of wetter, cooler climates (Hebda, 1995), characterized by low summer temperatures (Gavin et al., 2011), increased precipitation (Galloway et al., 2010), and decreased sea surface temperature (Bond et al., 1997) (Figure 3.6). The burial of mature trees alongside Franklin Glacier at site 6 (1470 m asl) occurred synchronously with a sustained period of regional glacial expansion (Osborn et al., 2007, 2013; Menounos et al., 2008; Figure 3.6).

Following expansion at 4.1 ka, Franklin Glacier appears to have downwasted significantly until 3.1 ka when it began expanding again in concert with other glaciers in the Mt. Waddington area (Clague et al., 2009; Coulthard et al., 2013) and throughout western Canada (Menounos et al., 2009) during a prolonged interval of cooler temperatures (Gavin et al., 2011) and decreased forest fire frequency (Hallet et al., 2003; Figure 3.6). Following the 3.1 ka advance, Franklin Glacier retreated before readvancing into a mature standing forest in 2.4 ka at site 8 (1430 m asl). There are similar records of glacial expansion at this time at nearby Scimitar and Tiedemann glaciers (Craig and Smith, 2013; Arsenault et al., 2007; Figure 3.6, Table 3.1), as well as at other glaciers in

Pacific North America (Koehler and Smith, 2011; Barclay et al., 2013; Osborn et al., 2013). The regional character of these advances suggests they were possibly a positive mass balance response to a large-scale climate perturbation (Galloway et al., 2011).

Following the 2.4 ka advance, Franklin Glacier downwasted and retreated, with no evidence for subsequent expansion until 1.5 ka (site 9) during a broadly-recognized period of glacier activity in western North America (Reyes et al., 2006; Jackson et al., 2008). In the Mt. Waddington area, contemporaneous advances were underway at Queen Bess and Scimitar glaciers (Table 3.1 and Figure 3.6) during a period of increased precipitation (Steinman et al., 2014). The global signature of ice expansion during this period suggests that climate deterioration related to millennial-scale cycles (Bond et al., 1997) may have played a significant role in contemporaneous glacier activity.

After the 1.5 ka advance, Franklin Glacier downwasted and retreated before expanding into mature valley side forests (>400 years old) during the early Little Ice Age (Ryder and Thomson, 1986; Coulthard et al., 2013). Advances at 0.8 and 0.6ka saw Franklin Glacier thicken to a position close to those achieved during the following late Little Ice Age (see Coulthard et al., 2013). Contemporaneous early Little Ice Age expansion is recorded at many glaciers in the Mt. Waddington area (Larocque and Smith, 2003; Table 3.1), reflecting the cool wet environment in the Coast Mountains at this time (Larocque and Smith, 2005b; Arsenault et al., 2007; Pitman and Smith, 2012; Babalola et al. 2013; Steinman et al., 2014).

Observations and photographs by Munday (1931a; 1934-1935) indicate that Franklin Glacier receded before expanding down valley to reach its maximum Little Ice

Age extent in the mid-19th to early 20th century. Based upon findings at other glaciers in the Mt Waddington area (Larocque and Smith, 2003; Craig and Smith, 2013; Figure 3.6), the late Little Ice Age advance of Franklin Glacier likely began prior to the mid-1700s in response to an extended interval of colder temperatures (Larocque and Smith, 2005a, 2005b; Pitman and Smith, 2012). Erosion of bedded lacustrine sands following draining of a moraine-damned lake at Site c (1357 m asl; Figure 3.3h) exposed a wood fragment indicating that establishment of the late Little Ice Age moraine was complete by 150 ± 70 ^{14}C yr BP (0.1 ka) (S-1475, Table 3.1).

The dendroglaciological record emerging from the Mt. Waddington area indicates that glaciers repeatedly expanded during the middle to late Holocene into forests on the proximal slopes of lateral moraines flanking Franklin Glacier valley. This behaviour is consistent with the record of regional temperature and precipitation fluctuations and corresponding cooling peaks in the North Atlantic, higlighted by ice-rafted debris events (Galloway et al., 2010; 2011; Gavin et al., 2011; Bond et al., 2001),(Figure 3.6). Similar causal relationships have been identified as presumed drivers for glacier expansion elsewhere in Pacific North America (Wiles et al., 2004) and serve to emphasize the role that synergistic climate cycling plays in establishing long-term trends in glacier behaviour.

3.6 Conclusions

Dendroglaciological evidence from Franklin Glacier provides insight about the latest Pleistocene and Holocene behaviour of glaciers in the Mt. Waddington area. The equivocal evidence for ice expansion at Franklin Glacier at 12.8 ka during the Younger

Dryas cold event is in contrast to findings elsewhere in the Coast Mountains and northern British Columbia. Our observations, if correct, may signify that present understanding of Younger Dryas glacier activity in the North American Cordillera remains incomplete.

Following warm conditions during the early Holocene, Franklin Glacier repeatedly advanced into valley side forests. Mid-Holocene advances at 6.3, 5.3, and 4.6 ka show progressive ice thickening followed by minor retreats prior to expansion at 4.1 ka (Figure 3.6). Subsequent advances occurred at 3.1 and 2.4 ka. Franklin Glacier retreated after the 2.4 ka advance, with no record of expansion again until 1.5 ka. The glacier attained its Holocene maximum vertical extent during the Little Ice Age and remained close to this position until the 1920s, after which it subsequently underwent significant downwasting and continued frontal retreat. The record of activity at Franklin Glacier is greatly enhanced by the interpreted ice level position although it does not correspond well with the record elsewhere in the Coast Mountains.

The dendroglaciological record at Franklin Glacier is among the most comprehensive recovered from the Coast Mountains and showcases the complexity of glacial activity in this region over the past ca. 6,000 years. The discovery of multiple episodes of mid- to late Holocene glacier expansion strengthens our understanding of the impact of long-term climate changes on the Holocene behaviour of glaciers in the Coast Mountains. Notable are the corresponding records of relative volumetric changes experienced by Franklin Glacier over the Holocene, which have implications for furthering our understanding of the impact of ongoing climate changes on glaciers throughout this region.

Chapter Four: Summary

The Holocene behaviour of glaciers in the British Columbia Coast Mountains is similar to the global record of activity (c.f., Solomina et al., 2015). Glaciers in the Pacific Ranges were likely advancing down valley at about 8.5-8.2, 7.3-5.3, 4.8-2.5, 1.4-1.2, and 0.8-0.4 ka. In the Boundary Ranges, ice expansion possibly occurred at approximately 4.1-4.0, 3.7-3.4, 3.1, 2.8-2.3, 1.7-1.1 ka, and also during the LIA. Common periods of likely ice expansion at about 4.1-4.0, 3.7-3.4, 2.8-2.5, 1.4-1.2, and 0.8-0.4 ka correspond to events in Alaska, the Canadian Rocky Mountains, and on Mt. Baker. This finding and congruent paleoenvironmental studies indicate that glacier expansion during these events was most likely a glaciological response to climate shifts from warm to cool and/or wet conditions

Dendroglaciological evidence from Franklin Glacier provides insights into the latest Pleistocene and Holocene behaviour of glaciers in the Mt. Waddington area. Ice expansion at 12.8 ka during the Younger Dryas contrasts with evidence of activity elsewhere in the Coast Mountains. Following this period, Franklin Glacier repeatedly advanced down valley at about 6.3, 5.3, and 4.6 ka, at which time it attained a size close to that it achieved in the late LIA. The glacier downwasted after 4.6 ka, after which it readvanced in 4.2, 3.1, and 2.4 ka. No record of ice expansion exists following the 2.4 ka advance until 1.6 ka when Franklin Glacier thickened and advanced into young subalpine fir trees. Activity before the first millennium AD is contemporaneous with regional behaviour in the Pacific Ranges over the last 6,300 years. During the LIA, three advances between 0.7 and 0.5 ka are documented, followed by a mid-19th to early 20th century advance that saw Franklin Glacier attain its maximum Holocene extent. The dendroglaciological record at Franklin Glacier is among the most comprehensive recovered from the British Columbia Coast Mountains, it supplements our current understanding of activity during the Holocene and showcases the complexity of mid- to late Holocene glacier expansion in the region.

Intervals of ice expansion at Franklin Glacier are associated with periods of climatic change and/or deterioration in British Columbia. Documented expansion at 12.8 ka is associated with a global climate reversal and worldwide glacier expansion. Glacier activity during the mid-Holocene at 6.3, 5.3, and 4.6 ka corresponds to regional climate deterioration. This period has been characterized as a sustained period of positive mass balance in the regional glacier chronology. Ice expansion at 3.1 and 2.4 ka are associated with regional glacier activity in the Mt. Waddington area in response to a prolonged interval of cooler temperatures. Following expansion at 2.4 ka, glacier activity at 1.5 ka is recognized within regional and global glacial histories as an interval of ice expansion. During this period, glaciers throughout Pacific North America and the rest of the northern Hemisphere were expanding in response to heightened precipitation and global climate deterioration. Little Ice Age activity at Franklin Glacier is contemporaneous with expansion elsewhere and reflects the cool, wet environment in the Coast Mountains at this time.

Uncertainties and assumptions

Dendroglaciological sample size, location, description, and state influence the uncertainties of their origins and probability of being killed by a glacier. The imprecise nature of dendroglaciological samples indicates possible uncertainties of the dated evidence. Lagged responses to climate by glaciers may also suggest that dendroglaciological evidence post-dates the environmental conditions influencing ice expansion during the Holocene. Furthermore, the presence of bark and/or perimeter wood is paramount when determining the true kill date of wood samples. The lack of either may result in misinterpretation of radiocarbon dates. In Chapter 2, I group dated evidence into 100-year intervals although the potential error may lie outside the designated period by 20 to 549 years (110-year average). In this thesis, I use high probability samples to describe glacier activity over broad intervals. Glacier activity from 4.8-2.5 ka in the Pacific Ranges is based on a near-continuous record of in situ sheared stumps and wood mats with underlying paleosols. However, many of the 100-year intervals only contain one high probability sample. From 7.3-5.3 ka, I describe an additional interval of near continuous glacier activity in the Coast Mountains but there are limited high probability samples. This interval may possibly be refined into smaller intervals (7.3-6.0 and 5.4-5.3 ka) based on the available evidence but this is not possible with the large error margins of dated evidence. From 8.5-8.2 ka glaciers may have been advancing down valley as suggested by Menounos et al. (2004) but evidence for this is equivocal. The cumulative error of dendroglaciological samples and the 100-year blocks used in this thesis highlight that glacier activity should be inferred as broader intervals where possible rather than short-lived events. For example, it is possible that glaciers were advancing at 3.1 ka in the Boundary Ranges but the cumulative error from sample design, glacier response times, and calibration methods indicates it could be ±0.3 ka from the corresponding 100-year interval.

4.4 Future Directions

Continued Research

Three field surveys have now been completed at Franklin Glacier. Older wood deposits, as highlighted by the 6.3 ka sample at Franklin Glacier, have been discovered close to or at the present ice level position. Revisiting glaciers that were previously studied may prove beneficial and serve to increase the sample size of mid- and early-Holocene dendroglaciologic evidence.

Research in the Boundary Ranges since 2008 has been extensive following Menounos et al. (2009)'s proposal that northern regions of the Canadian Cordillera should be primary focus of investigations. Four icefields in the region have records of activity extending back to 4.1 ka and multiple glaciers have been examined within each massif. Contradicting this, many promising icefields in the Pacific Ranges (e.g., Homathko and Ha-Itzuk icefields) have only been briefly surveyed at one glacier. Efforts should focus on developing regional-, or massif-, scale glacier chronologies to determine contemporaneous activity in response to climatic change. The Mt. Waddington area record emphasizes the potential usefulness of massif-scale records where six glaciers exhibit similar responses to environmental conditions throughout the Holocene.

Development of a High-Resolution Paleoclimate Record

Currently available paleoclimate records in coastal British Columbia provide low-resolution, relative descriptions of environmental conditions during the Holocene. Glacier activity in the Coast Mountains continues to be linked to these records but definitive relationships cannot be evaluated. This research, as well as Menounos et al. (2009), suggests that developing continuous Holocene-length tree-ring records would prove beneficial as: (1) dendroglaciological evidence could be cross-dated to the chronology to determine exact calender kill dates of glacially overridden subfossil wood, and; (2) annually-resolved proxy climate reconstructions extending over this duration would provide the opportunity for describing the causal factors that led to spatially and temporally synchronous glacial events in this setting.

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