Holocene glacier activity in the central British Columbia Coast Mountains

by

Jillian Elizabeth Harvey
B.Sc. University of Victoria, 2004

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MASTER OF SCIENCE

in the Department of Geography

Supervisory Committee

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Abstract

The intent of the research described in this thesis was to reconstruct and document Holocene glacier activity in the central British Columbia Coast Mountains. Despite ongoing efforts to describe glacier fluctuations in the southern and northern Coast Mountains, only limited attention has been directed to revealing the Holocene histories of glaciers in the central Coast Mountain region. The goals of this research were twofold: firstly, to describe mid-Holocene glacier advances at five remote glacier sites in the central Coast Mountains, and secondly, to detail Little Ice Age (LIA) glacier fluctuations at four glacier sites in the central Coast Mountains.

The mid-Holocene behaviour of Canoe, Fyles, Jacobsen, Tchaikazan and Icemaker glaciers was investigated using dendroglaciological techniques and stratigraphic analysis. Subfossil wood evidence suggests these glaciers were expanding into standing forests prior to 6.63, 4.90 and 4.20 ka. Stratigraphically constrained woody detritus at Fyles Glacier records the progradational history of a Gilbert-type delta forming in response to glacial expansion between 7.02-5.47 ka. Glacial expansion occurring between 7.50-4.00 ka has regional correlatives, suggesting coherent broad-scale climate forcing mechanisms influenced glacial mass balance at this time.

Insight into the LIA behaviour of central Coast Mountain glaciers was provided by conducting lichenometric surveys of *Rhizocarpon* spp. across LIA moraines at Pattullo, Fyles, Deer Lake and Jacobsen glaciers. The presence of a second, lesser known, lichen species at some sites necessitated the construction of a *Xanthoria elegans* growth curve. An assessment of lichenometric measurements from the southern and central Coast Mountains provided the opportunity to build a *X. elegans* growth curve constrained by 18 control points. Lichenometric surveys revealed dominant moraine building episodes at 890-1020, 1280-1320, 1490-1530, 1680-1720, 1780 and 1820-1870 AD, highlighting the complex nature of glacier fluctuations during the LIA.

A regional subalpine fir tree-ring chronology (1610-2010 AD) was developed from four stands located in the central Coast Mountains for dendroclimatological investigations. Correlation analyses show that the radial growth of trees corresponded to variations in the mean June/July air temperature and May 1st snowpack. This relationship was used to reconstruct these climate parameters for the duration of the tree-ring record. Intervals of cooler summer air temperatures and above average snowpack were found to broadly correspond with dominant periods of LIA moraine building from 1610-1930 AD.

This reconstruction of mid-Holocene and LIA glacial history offered insights consistent with the emerging record of glacial activity described for the southern Coast Mountain glaciers. It also provides the first evidence for mid-Holocene glacial expansion in the central and northern Coast Mountains. The application of lichenometry in the central Coast Mountains documents the regional LIA behaviour of glaciers and the construction of a *Xanthoria elegans* growth curve for the Coast Mountains provides a framework for future geobotanical dating using this species.

Table of Contents

Chapter 1: Introduction

Chapter 2: Mid-Holocene glacier expansion between 7.50-4.00 ka in the British Columbia Coast Mountains

Chapter 3: Little Ice Age glacier activity in the central British Columbia Coast Mountains

Chapter 4: Conclusion

List of Tables

List of Figures

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The research conducted for this thesis replaced my usual surroundings of penguins, icebergs and polar bears with the glorious backcountry of the Coast Mountains and a lab full the strongest and most inspiring women I could ever hope to meet. I owe a heartfelt thank you to all the ladies of the UVTRL: Jodi Axelson, Bethany Coulthard, Jess Craig, Kira Hoffman, Kate Johnson, Mel Page, Kyla Patterson, Kara Pitman and Colette Starheim. Thank you for sharing your expertise, laughter, friendship and many G&Ts. I will never forget all the amazing experiences and adventures we shared.

Chapter 1: Introduction

1.1 Introduction

Over the past century glaciers in the British Columbia Coast Mountains experienced substantial volumetric losses (Vanlooy and Forster 2008). Reflecting glaciological responses to rapidly changing mass balance states (Bitz and Battisti 1999), from 1985 glaciers in this region have decreased in area by an average of 12.5% (Bolch et al. 2008). The attendant retreat and downwasting is exposing glacial deposits at many locations that provide opportunities for describing Holocene glacial fluctuations.

The broad-scale application of dendroglaciological, lichenometric and dendroclimatological research techniques in this region has provided substantive insights into Holocene glacier activity and Little Ice Age (LIA) climates (Larocque and Smith 2005; Menounos et al. 2009). Emerging is the recognition of an increasingly complex chronology of glacial activity that is largely constrained to six periods of climate change at 9.00–8.00, 6.00–5.00, 4.20–3.80, 3.50–2.50, 1.60–1.30, and 0.60–0.15 ka[1] (Mayewski et al. 2004).

Despite ongoing efforts to describe Holocene glacier activity in the southern and northern Coast Mountains (Menounos et al. 2009), only limited attention has been directed to revealing the Holocene histories of glaciers found within the central Coast Mountain region (Desloges 1987; Desloges and Ryder 1990; Smith and Desloges 2000). This high mountain region extends southward from the latitude of Eustuk Lake to the Silverthone Icefield. Glaciers in the northern portion of the region are typically small and

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[1] In this thesis ka refers to calibrated years before present: 'a' is from the Latin '*annus*' meaning 'year'; 'k' represents 1000; and, 'ka' functions as an acceptable SI symbol used in dendroglaciology and geochronology research (eg. Aubry et al. 2009; Clague et al. 2009; NACSN 2009).

topographically restricted to cirque and valley bottoms; whereas large outlet glaciers flowing from the Monarch and Silverthrone icefields define the southern extent of the central Coast Mountains (Clarke and Holdsworth 2002).

1.2 Research goals and objectives

This thesis presents the results of my field investigations at three glaciers in the central Coast Mountain region in 2010, and includes unreported findings made by members of the University of Victoria Tree Ring Laboratory from 2002-2006 at Tchaikazan, Icemaker, Fyles and Canoe glaciers. The purpose of the research was twofold: a) to document evidence of mid-Holocene glacier activity; and, b) to determine the LIA behaviour of glaciers within the context of proxy records of climate change in the last 400 years.

The specific objectives of the research were to:

1. Reconstruct the mid-Holocene behaviour of Jacobsen and Fyles glaciers; and,
2. Document the LIA glacial activity of Pattullo, Jacobsen, Fyles and Deer Lake glaciers; and,
3. Construct annually-resolved climate reconstructions from tree ring-width records and describe the attendant glacier-response relationships.

Dendroglaciological lichenometric and dendroclimatological research techniques were applied to accomplish these objectives.

1.3 Thesis format

The thesis consists of four chapters. Following this chapter, Chapters 2 and 3 describe the outcomes of fieldwork at remote glaciers in the central Coast Mountains and are presented as manuscripts for forthcoming submission to refereed journals. Chapter 2 presents dendroglaciological and dendrogeomorphological findings describing mid-Holocene glacial advances at Fyles and Jacobsen glaciers. These research findings are presented in the context of prior published and unpublished research dating to this mid-Holocene interval from other Coast Mountain sites. Chapter 3 reconstructs the LIA behaviour of Pattullo, Fyles, Deer Lake and Jacobsen glaciers, drawing attention to apparent relationships between episodes of glacier expansion and regional climate changes recorded by tree-rings. Chapter 4 summarizes the findings of this research, describes limitations associated with the techniques used, and provides recommendations for future research.

Chapter 2: Mid-Holocene glacier expansion between 7.50-4.00 ka in the British Columbia Coast Mountains

2.1 Introduction

Most glaciers in the British Columbia Coast Mountains reached their maximum Holocene extent between ca. 1200 and 1900 AD during the Little Ice Age (LIA) (Menounos et al. 2009). Evidence of glacial activity prior to the LIA is often difficult to locate as it has remained buried by these recent, and usually more extensive, advances. Historical ice front recession and downwasting in the Coast Mountains is now exposing landforms covered until recently by glaciers. As these landforms erode or are incised by meltwater, the emergence of *in situ* and detrital woody debris buried during earlier periods of glacial expansion provides an opportunity to describe Holocene glacial activity.

The emerging chronology of Holocene glaciation in coastal British Columbia appears constrained to six broadly recognized periods of climate change at 9.00–8.00, 6.00–5.00, 4.20–3.80, 3.50–2.50, 1.60–1.30, and 0.60–0.15 ka (Mayewski et al. 2004). There is limited evidence for a short-lived advance at 8.20 ka (Menuonos et al. 2004), equivocal evidence for three intervals of mid-Holocene glacier expansion at ca. 6.00 ka (Ryder and Thompson 1986), 4.90 ka (Menounos et al. 2009) and 4.20 ka (Menounos et al. 2008), widespread glacier expansion at 3.50 ka (Ryder and Thomson 1986), 2.30 ka (Koehler and Smith 2011) and 1.60 ka (Reyes et al. 2006), and the onset of LIA glacier advances in the 11th Century (Allen and Smith 2007; Koch et al. 2007a).

Despite the considerable progress made in understanding the regional character of these Holocene advances (Menounos et al. 2009), very little is known of the character of glacier expansion during the mid-Holocene from 7.50 to 4.00 ka (Ryder and Thomson

1986; Menounos et al. 2008). There is limited detrital and *in situ* evidence suggesting glaciers may have repeatedly advanced into living forests in the southern Coast Mountains at this time (Mathews 1951; Osborn et al. 2007; Koch et al. 2007b; Koehler and Smith, 2011). Clastic units within lacustrine deposits in the same region may record corresponding episodes of mid-Holocene glacier expansion (Souch 1994; Cashman et al. 2002; Filippelli et al. 2006; Osborn et al. 2007; Menounos et al. 2008).

The intent of this chapter is to present new evidence extending and detailing the regional record of mid-Holocene glacial expansion in the Coast Mountains. These findings are summarized in the context of existing mid-Holocene records in the western Canadian Cordillera and are used to discuss the likely climate forcing mechanisms responsible for glacier expansion at this time.

2.2 Evidence for mid-Holocene glacier advance

Alpine glaciers act as sensitive indicators of climatic variability (Denton and Karlen 1973). Over decadal intervals, the terminal and lateral positions of alpine glaciers fluctuate in response to mass balance adjustments to changing temperature and precipitation regimes (Klok and Oerlemans 2003). In the Coast Mountains persistent positive mass balance episodes during the Holocene resulted in periods of glacial expansion that frequently saw glaciers overrunning living trees to bury them within and beneath glacigenic deposits (Koch et al. 2007b; Osborn et al. 2007). Radiocarbon or tree-ring dating of these glacially-killed trees entombed in growth position either within valley-side or valley-floor deposits provides direct evidence for when a glacier advanced over a specific site (Allen and Smith 2007; Koehler and Smith 2011). Studies of lateral moraine stratigraphy, where buried detrital wood mats separate discrete till units, oftentimes yield a continuous chronology of Holocene glacier expansion (Reyes and Clague 2004; Jackson et al. 2008).

Most proxy records describing mid-Holocene climates in the Coast Mountains indicate a trend towards slightly cooler and possibly wetter conditions than present (Mathewes 1985; Hebda 1995). Based upon dendroglaciological evidence at a limited number of sites, previous researchers concluded that these changing environmental conditions prompted glacier expansion at 6.95-5.62 ka (Ryder and Thomson 1986) and 4.40-4.00 ka (Menounos et al. 2009). Supported by radiocarbon-dated remains and clastic-rich sediment facies within lake cores, these advances appear associated with global climatic events documenting extended intervals of cooler summers and wetter winters (Mayewski et al. 2004; Zhang and Hebda 2005). While there is limited evidence for corresponding advances in the Canadian Rocky Mountains at 7.50-5.50 ka (Luckman et al. 1993; Clague et al. 2009), 4.90 ka (Menounos et al. 2009) and 4.20 ka (Gardner and Jones 1985; Luckman 1995; Wood and Smith 2004), ongoing research efforts in the Coast Mountains confirm these were regional events (Menounos et al. 2009).

2.3 Study sites

The Coast Mountains are a northwest trending mountain belt extending from southwestern British Columbia to the St. Elias Mountains of southwestern Yukon and Alaska (Figure 2.1). Recent terminus retreat at five Coast Mountain glaciers located in the Pacific and Boundary Ranges has exposed the glacially-killed remains of trees buried during mid-Holocene glacier advances. In the Frank Mackie Icefield area adjacent to southern Alaska, the remains of trees recovered between till units provide the first confirmation of mid-Holocene glacier expansion in the northern Coast Mountains.

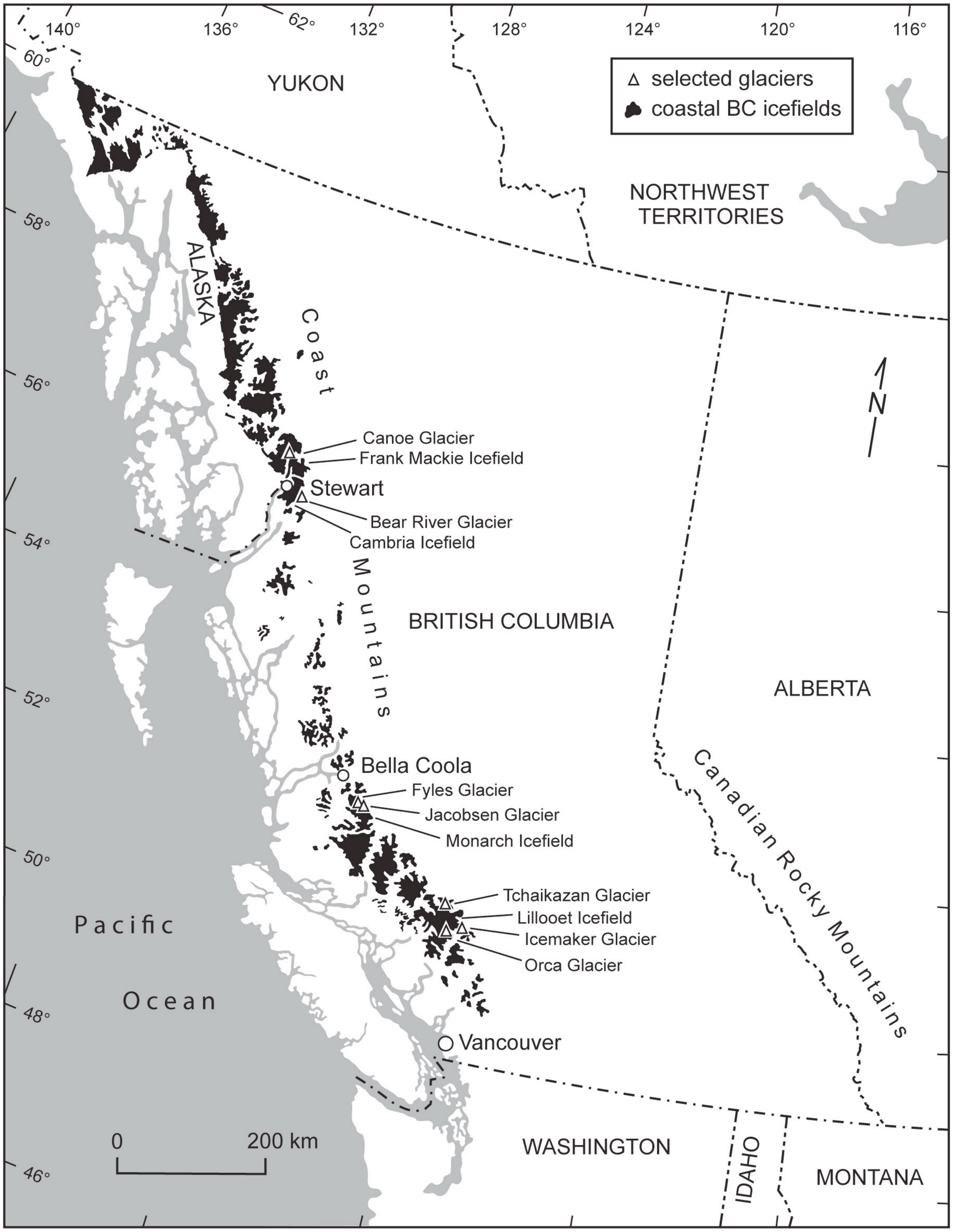

Figure 2.1 Map showing the location of study sites at Canoe, Fyles, Jacobsen, Tchaikazan and Icemaker glaciers.

In the Monarch Icefield area near Bella Coola (Figure 2.1), *in situ* tree stumps and detrital wood provides the first record of mid-Holocene glacier advances in the central Coast Mountains region. In the Lillooet Icefield area (Figure 2.1), detrital wood washed from beneath one glacier and sheared tree stumps in growth position at another site confirm the widespread extent of mid-Holocene advances in the southern Coast Mountains.

The Coast Mountain region is underlain primarily by granitic rocks and geologic structures of the Coast Plutonic Complex marking the transition to the Intermontane Volcanic Belt (Baer 1973). Tertiary and Quaternary glaciations shaped the land surface forming the major valley drainage systems (Tribe 2002). While climates in the vicinity of the Frank Mackie and Monarch icefields are strongly influenced by maritime weather systems originating in the North Pacific Ocean, rainshadow effects and continental climates also have an impact on glaciers in the Lillooet Icefield area. Dominant subalpine tree species in the region include Engelmann spruce (*Picea engelmannii*), mountain hemlock (*Tsuga mertensiana*), subalpine fir (*Abies lasiocarpa*), and whitebark pine (*Pinus albicaulis*). Disturbed and recently deglaciated surfaces host densely growing sitka alder (*Alnus viridis* subsp. *sinuata*) (Meidinger and Pojar 1991; FNA 1997).

2.4 Methods

Reconnaissance investigations were carried out in the summers of 2002 (Fyles and Tchaikazan glaciers), 2005 (Icemaker Glacier), 2006 (Canoe Glacier) and 2010 (Jacobsen Glacier). Each glacier site was systematically surveyed to locate *in situ* and detrital subfossil wood. Disk samples cut with a chainsaw were air-dried and polished using progressively finer grades of sandpaper to define the annual tree-ring boundaries. Tree-ring widths were measured to 0.01 mm on high resolution digital images from perimeter to pith using a WinDENDRO measuring system (v. 2006) (Guay et al. 1992). Multiple radii were measured to reduce the possibility of missing rings and to replicate data (Stokes and Smiley 1968).

Where the provenance suggested samples were likely of similar age, an attempt was made to construct floating chronologies using the visual cross-dating program CDendro v.4.1.1 (Larsson 2003). Following this, the chronologies were quality-checked with COFECHA v. 6.06 (Grissino-Mayer 2001), using 30 year dated segments lagged successively by 15 years with the critical level of correlation set at 0.4 (99% one-tailed confidence interval; Homes 1983).

Perimeter wood from selected samples was submitted to Beta Analytic Inc. for conventional radiocarbon dating. The ^{14}C ages were calibrated using INTCAL 09 (Reimer et al. 2009) and the 2σ ranges reported as cal yr BP (ka). Calendar ranges (ka) were assigned to include the entire area under the probability distribution and rounded to the nearest 0.01 ka. These ranges were used to anchor the ^{14}C samples within the locally-constructed floating chronology.

2.5 Results

2.5.1 Canoe Glacier

Canoe Glacier is an eastward flowing glacier located in the Boundary Ranges 50 km north of Stewart (53º 23’N, 130º 03’W; Figure 2.1). The glacier extends 7 km downvalley from the Frank Mackie Icefield to terminate at Tippy Lake. Like nearby Knipple and Frank Mackie glaciers, Canoe Glacier is assumed to have extended across and dammed the Bowser River during the late-Holocene (Clague and Mathews 1992). Upvalley deposits indicate that Frank Mackie Glacier expanded prior to 3.00 ka, between

2.95-2.54 ka and 1.64-1.39 ka, and during the 10th and 11th centuries to create glacially-dammed Tide Lake (Clague and Mathews 1992).

Three wood mats separated by till were located along the north-facing lateral moraine at Canoe Glacier ca. 1.3 km upvalley from the 2006 icefront. While the uppermost wood mat was inaccessible, perimeter wood from a log (UVTRL06-CAN01; Figure 2.2) lodged ca. 5 m below within a wood-rich laterally continuous organic mat dates to 3360±50 ^{14}C yr BP (3.72-3.47 ka; Table 2.1). A third wood mat located 3.5 m below contained an assemblage of broken boles visible over a 25 m long contact (Figure 2.2). Perimeter wood from a large bole cross-section (UVTRL06-CAN03) dates to 4570±50 ^{14}C yr BP (5.45-5.05 ka; Table 2.1).

This radiocarbon evidence suggests Canoe Glacier was expanding downvalley and overwhelming mature valley-side forests at 5.45-5.05 and 3.72-3.47 ka. While neither advance is recorded in adjacent Bowser Valley glacier histories (Clague and Mathews 1992; Clague and Mathewes 1996), the latter event is recorded by detrital wood in till dating to 4.22-3.85 ka at Bear River Glacier in the nearby Cambria Icefield (Jackson et al. 2008).

2.5.2 Fyles Glacier

Fyles Glacier (52°06’ N, 126°13’ W; Figures 2.1 and 2.3) spills from the Monarch Icefield into the upper Noeick River, where it historically formed a glacial dam that led to the creation of Ape Lake (Gilbert and Desloges 1987). Sometime after 1892 AD Fyles Glacier began retreating at rates ranging from 15 to 50 m/yr (Gilbert and Desloges 1987; VanLooy and Forster 2008), eventually thinning and permitting enhanced subglacial seepage that resulted in catastrophic glacier outburst floods in 1984 and 1986

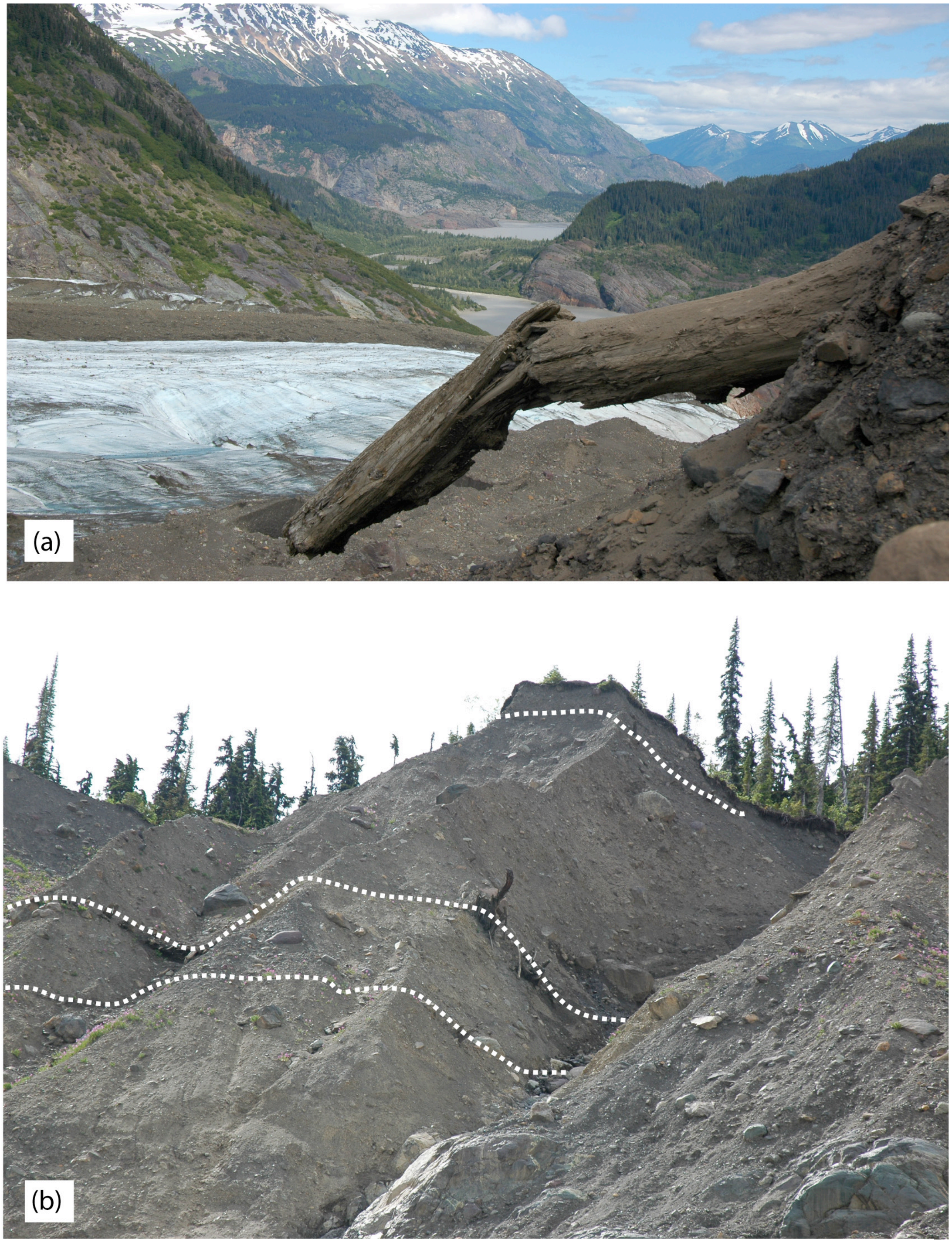

Figure 2.2 Canoe Glacier study site. (a) Sample UVTRL-06CAN01 (b) Location of three wood mats at Canoe Glacier. Woody mats are highlighted with white broken lines. The uppermost mat was inaccessible. Sample UVTRL-06CAN01 was located in the middle mat dating to 3.72-3.47 ka and the lower woody mat dates to 5.45-5.05 ka (UVTRL-06CAN03).

Table 2.1 Summary of mid-Holocene (7.50-4.00 ka) radiocarbon dates obtained at sites in the Coast Mountains, Canadian Rocky Mountains, and Interior Ranges, British Columbia. Locations are arranged from north to south.

Laboratory no. [a]	Sample ID [b]	Reference	Description	^{14}C Age (yr BP)	2 σ Calibration[c] (ka)	Lat.	Long.	Elevation (m asl)
Coast Mountains								
Bear River Glacier						56° 06'N	129° 38'W	
Beta-181857	BR03-801	Jackson et al. 2008	Log in moraine	3680±60	4.22-3.85			435
Beta-185808	BR03-806	Jackson et al. 2008	Log below moraine	3340±60	3.71-3.41			435
Canoe Glacier						56 ° 23'N	130 ° 04'W	
Beta-242989	06CAN03	This study	Bole in moraine	4570±50	5.45-5.05			750
Beta-242988	06CAN01	This study	Log in moraine	3360±50	3.72-3.47			750
Fyles Glacier						52° 06'N	126° 12'W	
Beta-171554	FG02-16A	This study	Branch/root, topsets	4860±60	5.73-5.47			1370
Beta-171550	FG02-06	This study	Branch/root, bottomsets	5640±80	6.63-6.29			
Beta-171551	FG02-07	This study	Branch/root from forests	5980±80	7.02-6.64			
Beta-171555	FG02-19	This study	Large bole, bottomsets	5570±70	6.50-6.22			
Beta-185809	FG02-18	This study	Branch, topsets in organics	5240±70	6.27-5.80			
Jacobsen Glacier						52° 04'N	126° 07'W	
Beta-283598	JAC0301	This study	*In situ* stump	5760±60	6.68-6.41			1421
Beta-283599	JAC0504	This study	Detrital log in till	5630±60	6.55-6.30			1451
Tiedemann Glacier						51° 19'N	124° 59'W	
Beta-220941		Menounos et al. 2009	Log in moraine	5010±40	5.89-5.65			
UCIAMS-40663		Menounos et al. 2009	*In situ* stump	3865±20	4.41-4.19			
UCIAMS-40660		Menounos et al. 2009	*In situ* stump	3820±20	4.29-4.12			
Beta-220940		Menounos et al. 2009	*In situ* stump	3760±60	4.39-3.93			
Beta-220936		Menounos et al. 2009	*In situ* stump	3690±50	4.22-3.89			
Goddard Glacier						51° 06'N	124° 10'W	
GSC-6046		Menounos et al. 2008	Detrital wood in forefield	4120±60	4.83-4.45			
Bridge Glacier						50° 48'N	123° 25'W	
GSC-3219		Blake 1983	Wood washed from glacier	5500±70	6.45-6.12			1935
Tchaikazan Glacier						50° 48'N	123° 25'W	
Beta-171556	TG02-07	This study	Detrital wood	5380±70	6.30-5.99			
Beta-172651	TG02-24	This study	Detrital wood	4660±70	5.59-5.08			
Beta-172650	TG02-10	This study	Detrital wood	4440±70	5.29-4.87			
Beta-171557	TG02-17	This study	Detrital wood	3730±60	4.28-3.90			

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Laboratory no. [a]	Sample ID [b]	Reference	Description	^{14}C Age	2 σ Calibration[c]	Lat.	Long.	Elevation
				(yr BP)	(ka)			(m asl)
Icemaker Glacier						50° 46’N	123° 21’W	
Beta-208440	IM05-01	This study	*In situ* stump	5680±60	6.64-6.32			1730
Orca Glacier						50° 35’N	123° 40’W	1600
Beta-227635	06-MC03	Koehler and Smith 2011	Detrital wood	4270±60	5.03-4.62			
Spearhead Glacier						50° 05’N	122° 50’W	
Beta-157268		Osborn et al. 2007	Detrital wood	3900±80	4.53-4.09			1995
Beta-168423		Osborn et al. 2007	Detrital wood	3900±60	4.52-4.15			1995
Overlord Glacier						50° 01’N	122° 50’W	
Beta-170665		Osborn et al. 2007	Detrital wood	6170±70	7.25-6.90			1640
Beta-170660		Osborn et al. 2007	Detrital wood	5890±70	6.89-6.51			1625
Beta-170667		Osborn et al. 2007	Detrital wood	5980±70	7.00-6.66			1610
Helm Glacier						49° 57’N	122° 59’W	
Beta-186523		Koch et al. 2007b	Detrital stump	4080±40	4.81-4.44			
Sphinx Glacier						49° 55’N	122° 58’W	
Beta-186509		Koch et al. 2007b	Detrital snag	5830±60	6.78-6.49			
Beta-208685		Koch et al. 2007b	Detrital log	4280±70	5.04-4.59			
Beta-186510		Koch et al. 2007b	Detrital snag	3560±70	4.08-3.65			
Sentinel Glacier						49° 54’N	122° 59’W	
Beta-186508		Koch et al. 2007	Detrital snag	6040±60	7.16-6.73			
GSC-1477		Lowden and Blake 1973	Detrital branch	6170±150	7.42-6.73			
GSC-2027		Lowden and Blake 1975	*In situ* stump	5300±70	6.27-5.93			
Warren Glacier						49° 52’N	123° 00’W	
Beta-148789		Koch et al. 2007b	*In situ* stump	6370±70	7.42-7.17			
Beta-148790		Koch et al. 2007b	*In situ* stump	6360±80	7.43-7.03			
Beta-148788		Koch et al. 2007b	*In situ* stump	5780±70	6.74-6.41			
Beta-168424		Koch et al. 2007b	Partially buried log	5700±50	6.64-6.36			
Lava Glacier						49° 49’N	122° 57’W	
Beta-168426		Koch et al. 2007b	*In situ* stump	6170±60	7.25-6.91			
Beta-168427		Koch et al. 2007b	*In situ* stump	6050±50	7.15-6.75			
Beta-186521		Koch et al. 2007b	*In situ* stump	5760±60	6.71-6.41			
Beta-186520		Koch et al. 2007b	Detrital stump	5130±40	5.99-5.75			
Y-140 bis		Preston et al. 1955	*In situ* stump	5850±180	7.16-6.30			
Y-140 bis		Stuiver et al. 1960	*In situ* stump	5260±200	6.43-5.60			

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Laboratory no. [a]	Sample ID [b]	Reference	Description	^{14}C Age	2 σ Calibration[c]	Lat.	Long.	Elevation
				(yr BP)	(ka)			(m asl)
Stave Glacier						49° 46’N	122° 32’W	
Beta-170668		Koch et al. 2007b	Detrital stump	6250±70	7.32-6.97			
Rocky Mountains								
Robson Glacier						53° 08’N	119° 06’W	
Beta-187090		Luckman 2007	Detrital log	4780±60	5.61-5.33			
Beta-65381		Luckman 1995	Root in palesol	3710±70	4.28-3.85			
Beta-65382		Luckman 1995	snag in gravels	3650±60	4.15-3.83			
Dome Glacier						52° 12’N	117° 15’W	
Beta-33007		Luckman et al. 1993	Detrital wood	6380±80	7.46-7.16			
Beta-33008		Luckman et al. 1993	Detrital wood	6120±60	7.17-6.80			
Boundary Glacier						51° 12’N	117° 11’W	
WAT-1182		Gardner and Jones 1985	*In situ* stump	4050±70	4.83-4.41			
Beta-160362		Wood and Smith 2004	*In situ* stump	3880±40	4.42-4.16			
WAT-1183		Gardner and Jones 1985	Peat below till	3880±60	4.51-4.10			
Interior Ranges								
Castle Glacier						53° 03’N	120° 26’W	
GSC-6709		Menounos et al. 2008	Rooted stump	4210±80	4.96-4.46			
UCIAMS-40543		Menounos et al. 2008	Rooted stump	3720±20	4.15-3.99			
GSC-6700		Menounos et al. 2008	Detrital log	3710±80	4.35-3.84			
UCIAMS-40544		Menounos et al. 2008	Detrital wood	3690±20	4.09-3.93			
Haworth Glacier						51° 42’N	117° 54’W	
GSC-6772		Menounos et al. 2008	*In situ*	3870±60	4.44-4.09			
Downie Glacier								
GSC-169		Ryder and Thomson 1986	Detrital log	3760±70	4.41-3.93	51° 18’N	118° 01’W	2104

Notes: [a] Radiocarbon Laboratory: Beta = Beta Analytical Inc. Miami, FL; GSC = Geological Survey of Canada, UCIAMS = University of California, WAT = University of Waterloo, Y = Yale University

[b] Sample ID provided for samples from the University of Victoria Tree-Ring Laboratory

[c] Radiocarbon dates converted to calendar ages BP (2σ) using CALIB 6.0 (Reimer et al. 2009). Ages were rounded to the nearest 0.01 ka.

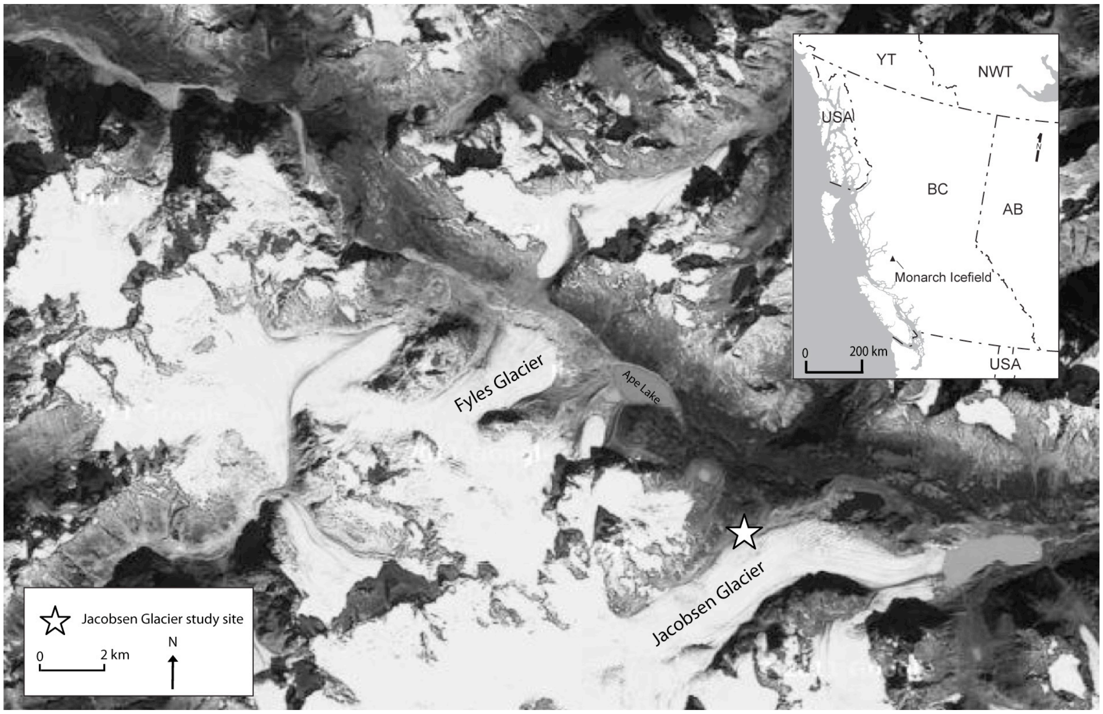


Figure 2.3 Map of the northeast corner of the Monarch Icefield. Inset shows the location of Monarch Icefield in British Columbia. The figure is based on 2011 satellite imagery (Google Earth).

Figure 2.4 Incised exposure of delta deposits at Fyles glacier. The study site was located approximately 1 kilometre from the 2002 Fyles Glacier ice front. The exposure is approximately 10 metres thick for scale.

(Jones et al. 1985; Desloges et al. 1989; Desloges and Church 1992). Draining of Ape Lake and subsequent subaerial incision exposed sedimentary units in a previously submerged delta complex (Figure 2.4). Containing stratigraphically constrained woody detritus, the deposit offered an opportunity to investigate the progradation history of the delta as it expanded into Ape Lake.

Field investigations in the summer of 2002 within a 200 m horizontal section revealed laterally contiguous stratigraphic units (Figure 2.5). Observations of unit thicknesses and sediment character were used to identify the continuity of different facies. Where organic horizons or oxidized sediments units were identified, their position and lateral extent was determined. Samples of *in situ* and detrital woody debris were collected, dried and submitted for radiocarbon dating.

Unit 1

Unit 1 consists of a minimum 3 m thick rhythmic sequence of horizontally-bedded silt and fine sand lenses that coarsen upwards from river level to Unit 2 (Figure 2.5). Scattered 2 to 5 cm diameter dropstones appear in the uppermost 25 cm of the unit. Ripples and climbing ripples distinguish the contact with Unit 2.

Two wood samples were retrieved from Unit 1 at the mid-point of the exposure. A sample (FG02-19) located 1 m below the contact with Unit Two contained 141 annual rings and yielded a perimeter age of 5570±70 ^{14}C yr BP (6.50-6.22 ka; Table 2.1). A branch fragment with eight rings (FG02-06) found 0.5 m below the contact dates to 5640±80 ^{14}C yr BP (6.63-6.29 ka; Table 2.1).

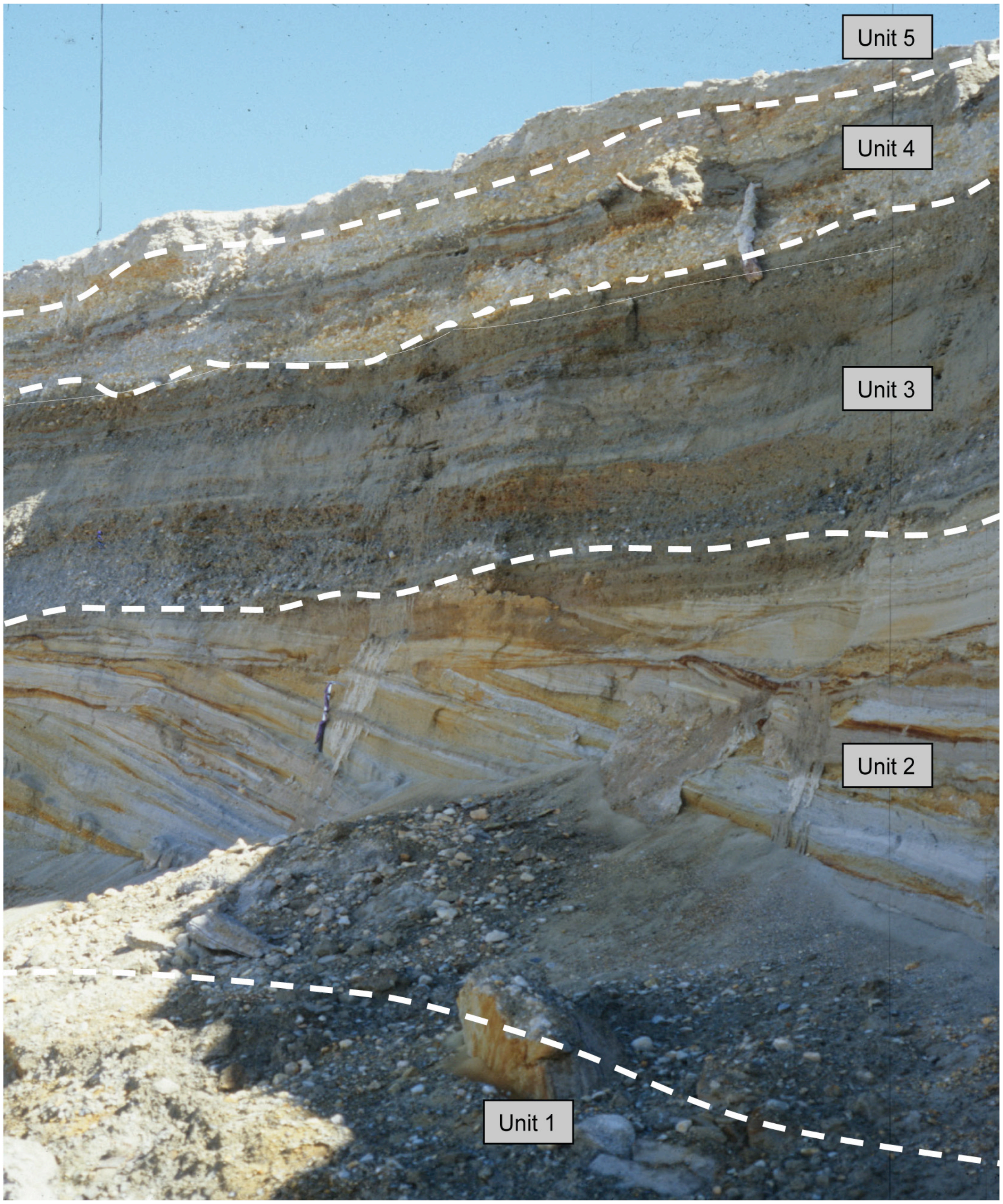


Figure 2.5 Delta complex at Fyles Glacier shows five distinct stratigraphic units. Unit 3 is 2 metres thick for scale.

Unit 2

Unit 2 contains a 3 m thick sequence of steeply-dipping (25°) interbedded coarse sand and gravel layers (Figure 2.5). Both fining- and coarsening-upwards sequences were noted within the unit. A branch fragment (FG02-07) was collected 1 m above the contact within Unit 1 close to the westernmost extent of the exposure. Containing 13 annual rings, the branch dates to 5980±80 ^{14}C yr BP (7.02-6.64 ka; Table 2.1).

Unit 3

Overlying Unit 2 is a 2 m thick horizontally-bedded unit whose lower extent is distinguished by interbedded coarse sand and gravel lenses (Figure 2.5). The upper extent of Unit 3 contains sloping beds (<10°) of fine sand (max. 40 cm thick) interrupted by a thick (>10 cm) planar organic horizon containing the partially-decomposed remains of small tree branches. A branch fragment (FG02-18) from the organic horizon dates to 5240±70 ^{14}C yr BP (6.27-5.80 ka; Table 2.1).

Unit 4

Unit 4 decreases in thickness (1 to 2 m) from west to east, suggesting that the proximal extent of the unit is truncated at the surface. At its thickest point the unit contains two horizontally-bedded gravel facies with occasional coarse sand lenses separated by laterally contiguous inclusions of fine sand and organics. A log (FG02-07) found protruding from the uppermost gravel layer contained 62 annual rings and dates to 4860±60 ^{14}C yr BP (5.73-5.47 ka; Table 2.1; Figure 2.5).

Unit 5

Unit 5 is capped by a 1 m thick deposit of matrix-supported sediment ranging in size from fine sands to large boulders (Figure 2.5). Annual recessional moraines on the upper surface of Unit 5 suggest it represents an ice-contact deposit.

Interpretation

The lowermost units exposed at Fyles Glacier following catastrophic release of floodwater from Ape Lake in 1984, contain topset, foreset and bottomset beds typical of a Gilbert-type delta (Bates 1953). The delta architecture indicates progradation towards the present-day position of Ape Lake. The delta deposits are overlain by braided stream deposits and till.

The radiocarbon dates assigned to detrital wood fragments sampled within the delta deposit indicate Fyles Glacier expanded to blockade upper Noeick River prior to 7.02-6.64 ka (Table 2.1). Progradation into Ape Lake was underway at this time burying a branch (FG02-07) within the foreset beds close to the western extent of the delta complex. The detrital character of the branch and the other woody debris recovered from the section suggest snow avalanches, or possibly debris flows, were depositing wood debris at the glacier terminus and within Ape Lake.

Delta progradation persisted for an extended period, as wood samples collected from the contemporaneous bottomset unit (ca. 70 m upvalley) date to 6.63-6.29 and 6.50-6.22 ka (Table 2.1). Dropstones in Unit 1 indicate that either Ape or Fyles glacier was calving into Ape Lake at this time. The progradation of foreset over bottomset beds is demarcated by climbing ripples, indicating that deposition occurred at a mid to distal delta position (Clemmensen and Houmark-Nielsen 1981). A branch fragment contained within the overlying topset bed (Unit 3) confirms delta formation continued at least until 6.27-5.80 ka (Table 2.1).

The delta is overlain by 1 to 2 m of braided stream deposits (Unit 4) that date to 5.73-5.47 ka (Table 2.1). Indicative of a subaerial depositional environment, the unit may signal that Ape Lake had drained by this time or that Fyles Glacier had advanced closer to the site. The surface of Unit 4 is truncated and overlain by a more recently deposited and modified till.

The delta complex at Fyles Glacier describes a mid-Holocene advance that dammed the Noeick River allowing Ape Lake to fill from 7.02 to 5.80 ka. While subsequent Holocene-age fluctuations of Fyles Glacier may have allowed Ape Lake to drain and fill several times, there is no evidence of these events within the delta complex.

2.5.3 Jacobsen Glacier

Jacobsen Glacier is a large northeastward-flowing outlet glacier of the Monarch Icefield (Figures 2.1 and 2.3). Prominent lateral moraines of late-Holocene and LIA origin flank the glacier (Desloges and Ryder 1990). Since 1915 AD the terminus of Jacobsen Glacier has retreated over 6 km (Desloges 1987), at rates averaging 30-40 m/yr (VanLooy and Forster 2008). Accompanying downwasting of the glacier surface exceeded 150 m.

In July 2010 field surveys were completed on a portion of the Jacobsen Glacier north lateral moraine marking the divide between the Talchako and Noeick river valleys, approximately 7 km east of Ape Lake (Figures 2.3 and 2.6). An *in situ* stump rooted in a paleosol, ice pressed logs, and multiple detrital wood samples were located at four sites close to ice level within the proximal moraine face.

Figure 2.6 Jacobsen Glacier study site.
(a) Dendroglaciological study sites at Jacobsen Glacier.
(b) At Site A (1451 metres asl), the location of four detrital bole fragments including radiocarbon dated sample JAC0504 (6.55-6.30 ka) is shown. (c) Sample JAC0301 (6.68-6.41 ka), one of two ice-pressed logs at Site B and radiocarbon dated (1420 metres asl).

Four detrital bole fragments oriented downvalley were excavated from till at Site A (1451 m asl; Figure 2.6). Cross-sectional disks were cut and perimeter wood from one sample (JAC0504) containing 63 rings had a radiocarbon age of 5630±60 ^{14}C yr BP (6.55-6.30 ka; Table 2.1).

The remains of two additional trees were located 200 m downslope from Site A adjacent to the glacier margin at Site B (1420 m asl; Figure 2.6). Pressed into the proximal face of a projecting piece of bedrock and encased by till, perimeter wood containing 52 rings from JAC0301 dates to 5760±60 ^{14}C yr BP (6.68-6.41 ka; Table 2.1).

One hundred metres downvalley at Site C the remains of a second rooted stump surrounded by detrital branches and small bole fragments were found in till at 1458 m asl (Figure 2.7). At Site D 130 m downvalley, a small glacially-sheared stump rooted in a steeply-sloping paleosol was located beneath a till veneer close to where a small stream was eroding through the proximal moraine face (1420 m asl, Figure 2.7). Additional woody detritus was found in till several metres downslope.

In total, 23 samples of detrital wood and stump remnants were collected at the four sites (Table 2.2). Nine of the samples crossdate to form a 177 year-long floating tree-ring chronology (r= 0.403; Table 2.2).

Interpretation

Expansion of Jacobsen Glacier from 6.68-6.30 ka resulted in the burial of a mature subalpine forest and paleosol located beneath a deposit of till. The crossdated remains of sheared rooted stumps and detrital boles within 0–20 m vertically of the 2010 ice surface suggest Jacobsen Glacier is morphologically similar in size and shape today as it was during this mid-Holocene event.

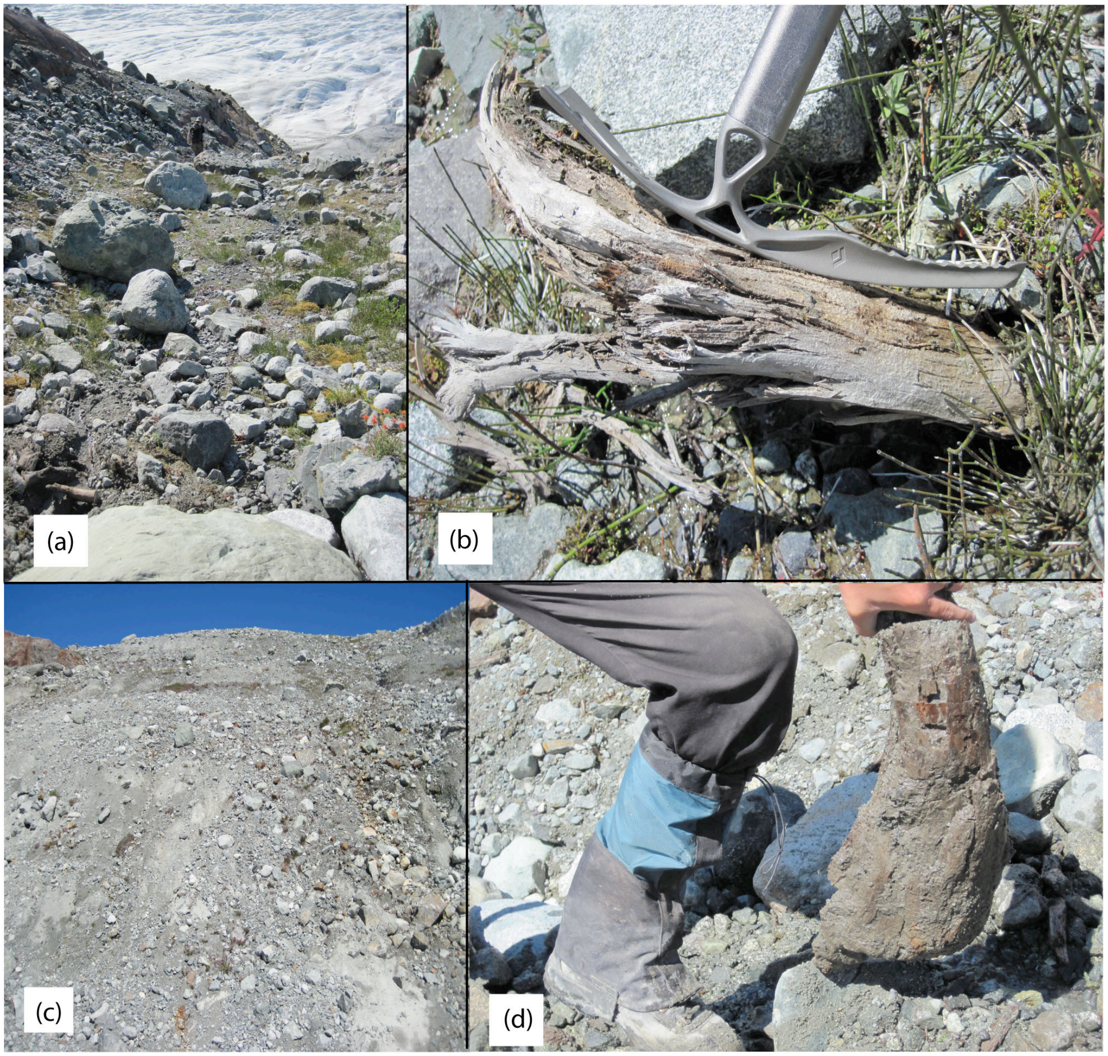


Figure 2.7 Jacobsen Glacier study site. (a) Site C (1458 metres asl) (b) glacially sheared stump located (JAC0101) at Site C (c) Site D (1420 metres asl) (d) Sample JAC0602 located at Site D.

Table 2.2 Tree-ring statistics for subfossil floating tree-ring chronology.

Sample ID	**^{14}C yr BP**	**Number of Rings**	**Correlation with master**
JAC0101A		81	0.622
JAC0102A		92	0.553
JAC0102B		96	0.630
JAC0302A	5760±60	114	0.332
JAC0303A		101	0.323
JAC0501A		152	0.329
JAC0503B		98	0.322
JAC0504B	5630±60	115	0.327
JAC0601A		76	0.377
JAC0602A		95	0.318
		Length	Interseries Correlation
Master Chronology		177 years	0.403

2.5.4 Tchaikazan Glacier

Tchaikazan Glacier is a large 7 km long valley glacier located at the northern limit of the Lillooet Icefield (51 º 02’N, 123º 48’W; Figure 2.8). The glacier drains into Taseko Lake via Tchaikazan River and has retreated 2 km upvalley at rates ranging from 0.9 to 31.0 m/yr (1951-2002; Ricker 1976; Lixvar 1983) from nested LIA terminal moraines dating to 1700, 1823 and 1928 AD (Smith, *unpublished notes*).

In July 2002, during moraine surveys in the upper headwaters of the Tchaikazan Valley, quantities of large detrital bole fragments were observed washing from beneath the snout of Tchaikazan Glacier at 1900 m asl. Appearing freshly-broken, this woody detritus rafted downstream to be deposited on the surface of a broad outwash plain.

After drying, cross-sections of samples stranded on the outwash surface were examined and identified as the remnants of mature whitebark pine trees. Perimeter wood from four samples dates to: TG02-7 = 5380±70 ^{14}C yr BP (6.30-5.99 ka); TG02-24 = 4660±70 ^{14}C yr BP (5.59-5.08 ka); TG02-10 = 4440±70 ^{14}C yr BP (5.29-4.87 ka); and, TG02-17 = 3730±60 ^{14}C yr BP (4.28-3.90 ka) (Table 2.1).

Interpretation

The glacially-broken detrital wood flushed from unknown locations beneath Tchaikazan Glacier appears associated with three distinct mid-Holocene advances. The oldest sample (TG02-7) is presumed related to a mid-Holocene advance of Tchaikazan Glacier in ca. 6.00 ka, with the three other radiocarbon-dated samples likely providing evidence for glacier advances identified regionally at 4.90 ka (Menounos et al. 2009) and at 4.20 ka (Menounos et al. 2008).

Figure 2.8 View of north facing Tchaikazan Glacier.

Figure 2.9 Icemaker Glacier. The black circle indicates the location of the rooted glacially sheared stump IM05-01 dated to 6.64-6.32 ka.

2.5.5 Icemaker Glacier

Icemaker Glacier is small cirque glacier located on the north face of Icemaker Mountain (50º 46’N, 123º 21’W; Figure 2.9). Meltwater from the snout of the glacier at 1770 m asl flows into the headwaters of McPurlon Creek and Bridge River east of the Lillooet Icefield. Within the last century Icemaker Glacier has retreated 1.5 km from nested terminal moraines constructed in ca. 1650, 1770 and 1810 AD (Smith, *unpublished notes*).

During a reconnaissance survey in July 2005, two glacially-sheared stumps in growth position were located 200 m from the ice margin at 1793 m asl (Figure 2.9). Partially-buried by till and rooted in a well-developed paleosol, the stumps remained beneath the glacier until sometime after 1947 AD (BC477:39). A perimeter sample from IM05-01 dates to 5680±60 ^{14}C yr BP (6.64-6.32 ka) and indicates that Icemaker Glacier was expanding into mature forests following an ice-free interval of sufficient length for soils to develop.

2.6 Discussion

The discovery and dating of mid-Holocene age dendroglaciological samples from several new Coast Mountain locations serve to emphasize that glaciers expanded and retreated at least three times between 7.50-4.00 ka. Evidence from Fyles, Jacobsen and Icemaker glaciers suggests regional expansion began prior to 6.63-6.29, 6.68-6.41 and 6.64-6.32 ka, respectively. Continued delta progradation into proglacial Ape Lake at 6.27-5.80 ka, along with the discovery of detrital wood killed by an advance of Tchaikazan Glacier at 6.30-5.99 ka, suggests glaciers in the region remained in advanced positions until sometime before 5.47 ka. Braided stream deposits truncating delta topset beds at Ape Lake date to 5.73-5.47 ka and may suggest Fyles Glacier was retreating at this time.

A subsequent interval of mid-Holocene glacier expansion between 5.59-4.87 ka is discerned from detrital evidence discovered at Tchaikazan Glacier and from bole fragments recovered from the lateral moraine at Canoe Glacier. Only at Tchaikazan Glacier was evidence located corroborating glacier expansion also occurred between 4.28-3.90 ka.

2.6.1 Regional comparisons

A growing body of evidence from the southern Canadian Cordillera describes a complex history of mid-Holocene glaciation (Figure 2.10). An early mid-Holocene episode of glacier expansion (the “Garibaldi Phase” of Ryder and Thompson [1986]) is restricted to the interval from 6.95-5.62 ka. Previous researchers describe concurrent glacier advances in Garibaldi Provincial Park at 7.30-5.80 ka (Mathews 1951; Koch et al. 2007b; Osborn et al. 2007), as well as in the Canadian Rocky Mountains at 7.46-6.80 ka (Luckman 2007) and 5.61-5.32 ka (Menounos et al. 2009). The *in situ* and detrital dendroglaciological evidence collected during surveys at Fyles, Icemaker, Jacobsen and Tchaikazan glaciers dating to 6.63-5.90 ka constrains the timing of this advance and lends considerable support for interpretations of it as regional in extent (Koch et al. 2007b; Menounos et al. 2009).

The discovery of a second interval of mid-Holocene glacier expansion 5.59-5.05 ka at Tchaikazan and Canoe glaciers has only a few regional correlatives (Menounos et al. 2009). Combined with the discovery of detrital wood in till in Garibaldi Provincial Park (5.04–4.59 ka; Osborn et al. 2007), a glacially-killed *in situ* stump in the Canadian

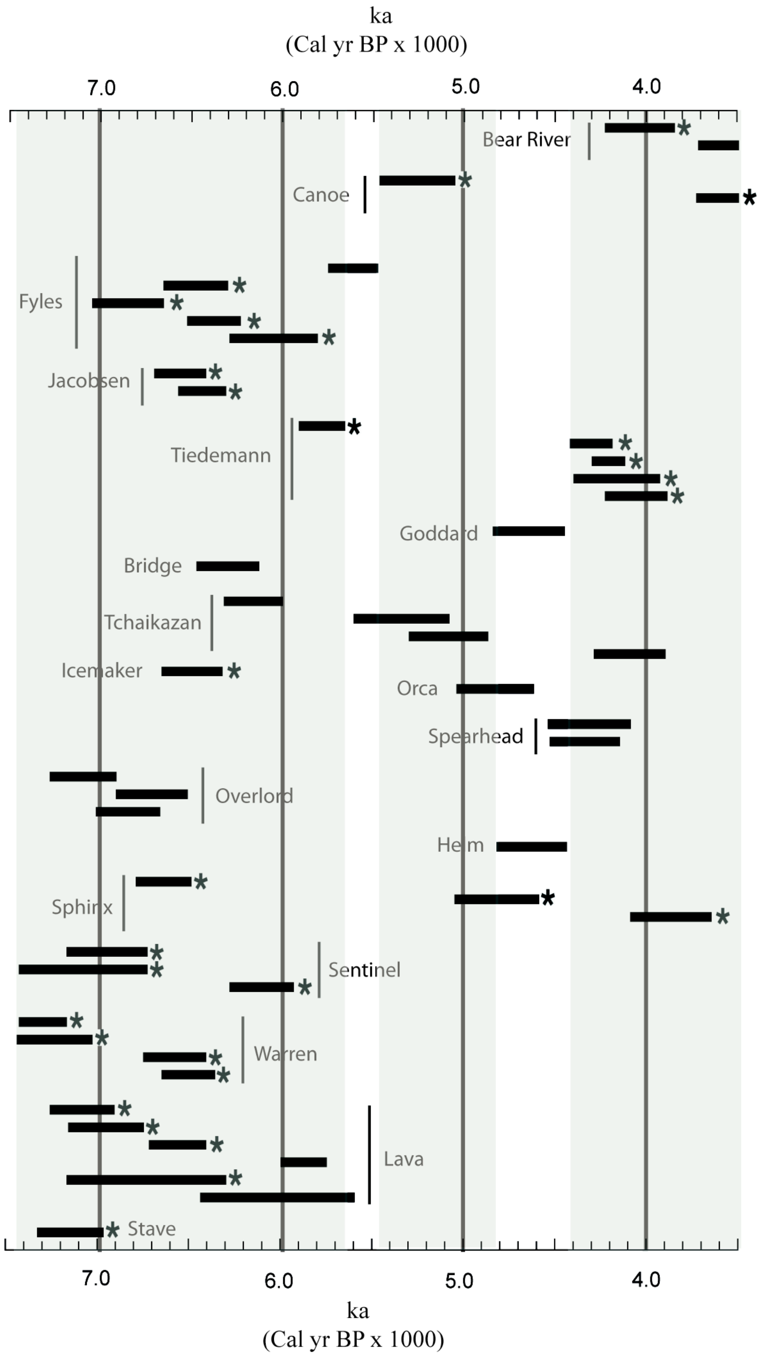


Figure 2.10 Summary of calibrated radiocarbon ages pertaining to glacier advances during the 7.50-4.00 ka mid-Holocene interval in the Coast Mountains. 2σ calibrated ranges are presented. The stars indicate samples that directly date ice margin expansion. The vertical grey bars indicate broad episodes of glacier advance. Glacier sites are arranged north to south.

Rockies dating to 5.61-5.33 ka (Luckman 2007), and detrital and *in situ* wood in the Cariboo Mountains at 5.58-5.46 (Menounos et al. 2009), these dendroglaciological findings substantiate lake sediment records of a regional glacier advance prior to ca. 4.90 ka (Menounos et al. 2008).

A third mid-Holocene glacier advance at 4.4-4.0 ka has previously been documented at several sites in the southern Coast Mountains (Koch et al. 2007b; Osborn et al. 2007; Menounos et al. 2008; Koehler and Smith 2011) and in the Canadian Rocky Mountains at 4.82-3.83 ka (Gardner and Jones 1985; Luckman 1995; Wood and Smith 2004). Detrital wood washing from the snout of Tchaikazan Glacier dating to 4.28-3.90 ka may also record this event and expands the regional extent of this advance.

This emerging record of mid-Holocene glacier expansion suggests three positive mass balance episodes at 6.68-5.99, 5.59-5.05 and 4.4-4.0 ka. Cooler summer temperatures and increased winter precipitation at these times likely lead to these events (Hebda 1995). Huesser et al. (1985) argue that this change in climate conditions was inaugurated as the subtropical Pacific anticyclones that dominated the early Holocene to produce warm and dry climates in the Pacific Northeast, were replaced by cooler and wetter conditions as the influence of the Aleutian low pressure centre gradually intensified. The discovery of glacially-killed trees in Garibaldi Provincial Park suggests this climatic shift was inaugurated by 7.4 ka in the southern Coast Mountains when glaciers began expanding into established forests (Mathews 1951; Koch et al. 2007b; Osborn et al. 2007). Evidence of continued glacier expansion at Icemaker, Fyles and Jacobsen glaciers dating to 6.63-6.30 ka confirms the persistence of this shift and the regional nature of positive glacier mass balance states at the beginning of the mid-Holocene in the southern and central Coast Mountains.

2.7 Conclusion

Mid-Holocene glacier expansion was documented at five glaciers in the Coast Mountains. Radiocarbon dates assigned to *in situ* and detrital subfossil wood located in glacial forefields, delta deposits and lateral moraines constrains the timing of three glacier advances between 7.50 and 4.00 ka. The findings support and improve upon existing chronologies describing mid-Holocene glacier activity in the British Columbia Coast Mountains, firmly establishing the regional nature of advances that began prior to 6.63, 4.90 and 4.20 ka. Documentation of these discrete glacier advances indicates that mid-Holocene climates were likely more varied than reported (Hebda 1995) and were likely precipitated by climate changes related to intensification of the Aleutian low pressure center in the region (Huesser et al. 1985). Future investigations of Holocene glaciation in the central and northern Coast Mountains and in the Canadian Rocky Mountains will benefit from this more complete characterization of mid-Holocene glacier fluctuations.

Chapter 3: Little Ice Age glacier activity in the central British Columbia Coast Mountains

3.1 Introduction

Glaciers in Pacific North America (PNA) fluctuated in size and extent during the Holocene in response to changing climate regimes (Denton and Karlen 1973). Intervals of glacial expansion during the Holocene are broadly synchronous across the Cordillera of PNA (Luckman and Villalba 2000), with most glaciers reaching their maximum Holocene extent during the Little Ice Age (LIA) (Luckman 2000; Calkin et al. 2001; Menounos et al. 2009).

The LIA in PNA is characterized by a prolonged interval of glacial expansion beginning as early as the 11th century (Ryder and Thompson 1986; Luckman 2000; Allen and Smith 2007; Koch et al. 2007a; Clague et al. 2009), with distinct early-LIA moraine building episodes recorded in the 12th and 13th centuries (Ryder and Thompson 1986; Larocque and Smith 2003; Allen and Smith 2007; Koch et al. 2007). Following this most glaciers in PNA receded and downwasted before late-LIA expansion during the 17th century led to regional moraine building episodes in the 18th and 19th centuries (Larocque and Smith 2003; Menounos et al. 2009).

Over the last century glaciers in the British Columbia Coast Mountains have undergone considerable volumetric losses (Larsen et al. 2007; Bolch et al. 2008; Vanlooy and Forster 2008), exposing previously buried glacial deposits and allowing for opportunities to establish a chronology of LIA glacial fluctuations (Menounos et al. 2009). Following recent research, LIA glacier activity in the southern Coast Mountains is now relatively well documented (Larocque and Smith 2003; Koch et al. 2007a; Menounos et al. 2009; Koehler and Smith 2011). Evidence dating to this interval from the northern and central Coast Mountains remains limited, pointing to this part of PNA as a region where focused LIA-related research is warranted (Menounos et al. 2009).

This chapter presents the results of field investigations at glaciers in the central Coast Mountain region. Fieldwork completed in 2002 and 2010 focused on determining the LIA history of four glaciers by employing lichenometric and dendroglaciologic research techniques. The findings of the research are discussed in the context of existing LIA evidence from PNA and are used to understand late-LIA glacier-climate interactions.

3.2 Study Area

The Coast Mountains are a northwest trending mountain belt extending from southwestern British Columbia to the St. Elias Mountains of southwestern Yukon and Alaska. The central Coast Mountain study area extends south from the latitude of Eustuk Lake (Figure 3.1; 53° 14’N, 126° 34’W) to the Silverthone Icefield (Figure 3.1; 51° 29’N, 125°56’W). The region is underlain primarily by granitic rocks and geologic structures of the Coast Plutonic Complex marking the transition into the Intermontane Volcanic Belt (Baer 1973). Quaternary continental and alpine glaciations shaped the land surface in the region forming the major coastal fjords and valley drainage systems of the Bella Coola, Atnarko and Talchako rivers (Baer 1973; Holland 1976). The Cordilleran Ice Sheet covered the region during the late Wisconsinan, reaching its greatest extent by 18.25 to 16.68 ka (Clague et al. 1988). Following this, the ice sheet downwasted, stagnated and, by 11.50 ka the extent of alpine glaciers in the region was comparable to present (Clague and James 2002).

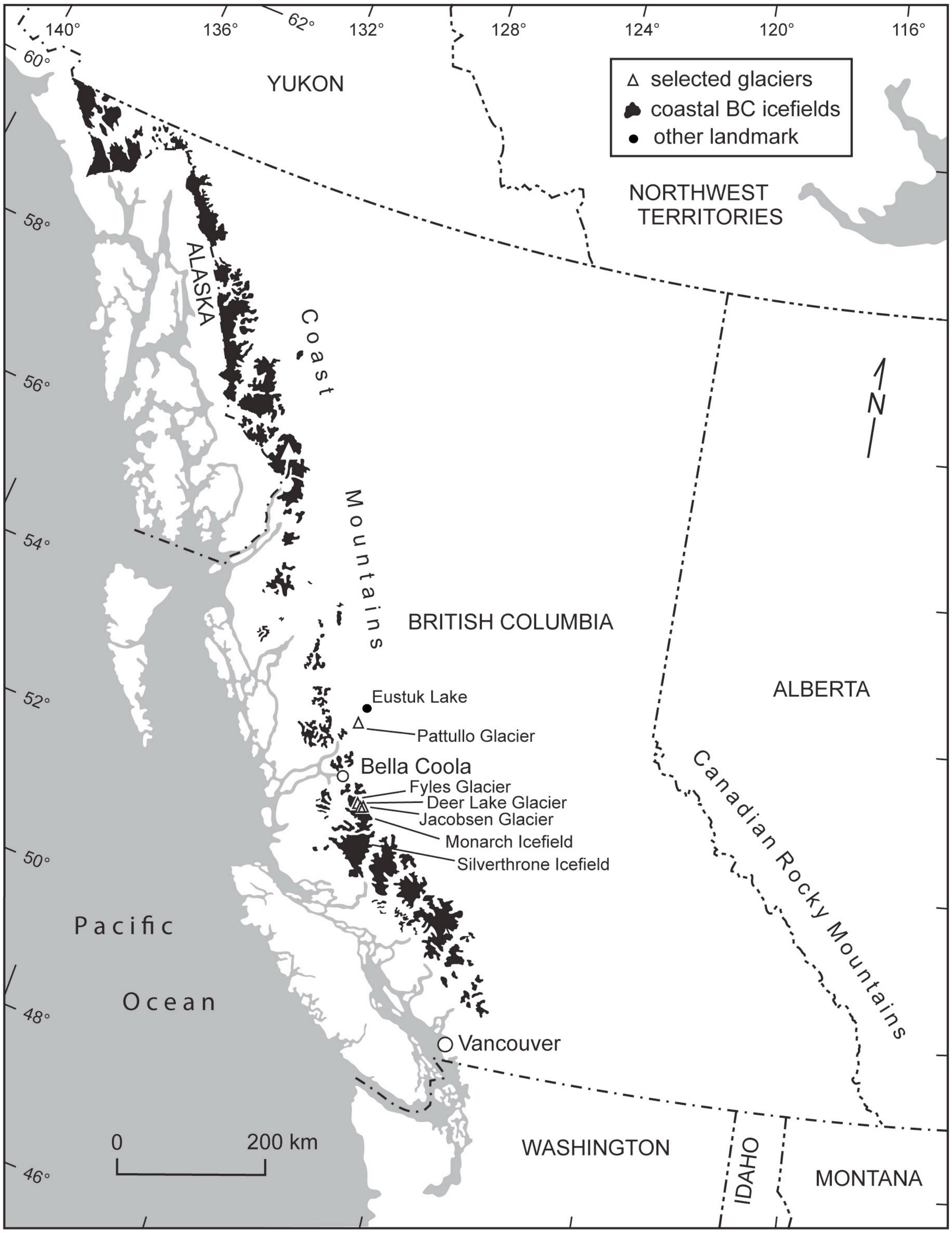


Figure 3.1 Map of the Coast Mountains study region. The central Coast Mountains includes the area between Eustuk Lake and Silverthrone Icefield.

The central Coast Mountain region lies within the Pacific coast maritime climate system, characterized by orthographic lifting of weather systems originating in the northern Pacific Ocean (Wood 2001). The northwest trending Pacific and Kitimat ranges influence regional precipitation regimes causing a gradient of decreasing precipitation from west to east. The dominant subalpine tree species in the region reflect this gradient, with mountain hemlock (*Tsuga mertensiana*) forests common on windward Pacific slopes, and subalpine fir (*Abies lasiocarpa*) and whitebark pine (*Pinus albicaulis*) dominating the eastern slopes adjacent to the interior plateau (Meidinger and Pojar 1991). Densely growing sitka alder (*Alnus viridis* subsp. *sinuata*) characteristically populate disturbed sites near stream channels and avalanche chutes.

Glaciers north of the Bella Coola River valley are typically small and topographically restricted to cirque and valley bottoms; whereas large outlet glaciers flowing from the Monarch and Silverthrone icefields define the Coast Mountains to the south (Clarke and Holdsworth 2002). Fieldwork was undertaken at Pattullo Glacier in the vicinity of Eustak Lake and at several glaciers along the northern periphery of the Monarch Icefield (Figure 3.1).

3.2.1 Pattullo Glacier

Pattullo Glacier is located along the eastern flank of the Kitimat Ranges adjacent to the western flank of the Nechako Plateau (54º 02'N, 126º 37'W; Figure 3.2). The glacier flows eastward from a small icefield below Tsaydaychuz Peak on the east side of the Pattullo Range in North Tweedsmuir Provincial Park, approximately 70 km northeast of Bella Coola (Figure 3.2). Bedrock in the area consists primarily of mafic volcanics from the lower to middle Jurassic Period (Mahoney et al. 2006). Since 1946 the

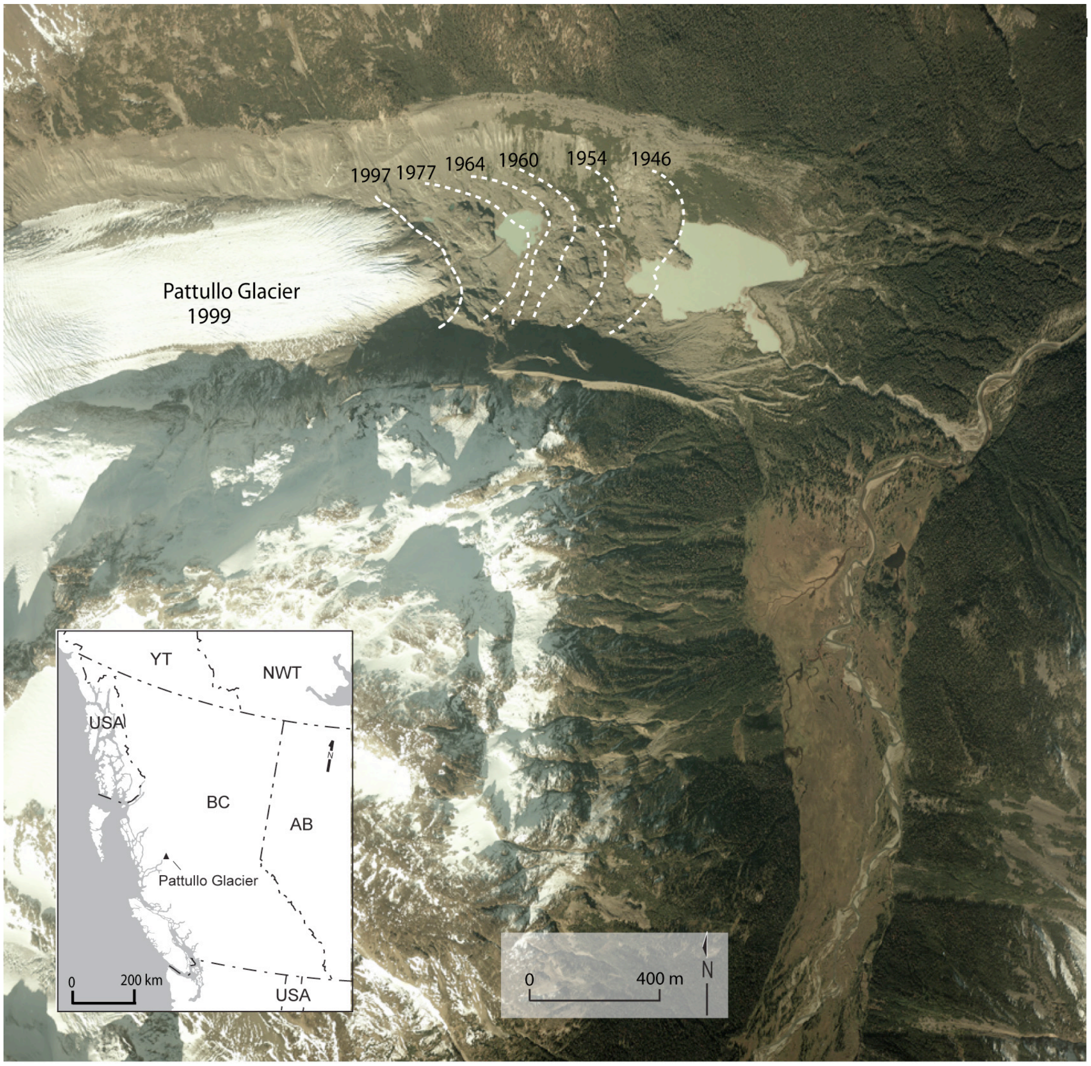


Figure 3.2 Pattullo Glacier, located in the central Coast Mountains. The 1997, 1977, 1964, 1960, 1954 and 1946 ice-front positions are indicated by broken lines. The figure is based on a 1999 aerial photograph (BCC99031:82).

terminus of Pattullo Glacier has retreated ca. 1 km at an average rate of ca. 20 m/year (Table 3.1).

3.2.2 Monarch Icefield

The Monarch Icefield is situated along the eastern flanks of the Pacific Ranges and covers an area of approximately 350 km^2 (Desloges and Ryder 1990). The icefield rests atop lower Cretaceous Monarch andesitic flows and volcaniclastic sedimentary rocks (Mahoney et al. 2006; Mahoney et al. 2009). Jacobsen, Fyles and Deer Lake glaciers flow northeast from the northeast corner of the Monarch Icefield (52º 03'N, 126º 14' W; Figure 3.3).

Jacobsen Glacier is a 7 km long valley glacier that terminates in Ape Creek valley, a drainage divide between the Talchako and Noeick river valleys. Jacobsen Glacier attained its maximum LIA extent as late as 1915 (Desloges 1987), but since that time has retreated ca. 4.5 km upvalley at approximately 50 m/yr (Google Earth, 2011).

Little is known of the LIA behaviour of Jacobsen Glacier (Desloges, 1987). The northern lateral moraine was breached prior to the most recent advance of the glacier, with distinct lobes of ice having encroached into the surrounding subalpine fir forest (Figure 3.4). At the margin of one lobe Desloges and Ryder (1990) located a paleosol below till that dates to 400±60 ^{14}C yr BP (1427-1636 AD).

Deer Lake and Fyles glaciers spill from the Monarch Icefield into the upper Noeick River (Figure 3.3). Fyles Glacier historically formed a glacial dam that led to the creation of Ape Lake (Gilbert and Desloges 1987). Sometime after 1892 AD Fyles Glacier began retreating 15 to 50 m/yr (Gilbert and Desloges 1987; VanLooy and Forster

Table 3.1 Calculated rates of glacier retreat for Pattullo Glacier

Year	**Source**	**Ice Front Position**	**Calendar Dates** (years AD)	**Recession Rate** (m/yr)
1946	Aerial photograph A10605:59	Lat: 53º02’28” N Long: 126º33’58” W Elevation: 1270 m asl		
1954	Aerial photograph BCC1822:31,32	Lat: 53º02’28” N Long: 126º34’09” W Elevation: 1265 m asl	1946-1954	24
1960	Historical photograph Austin Post[1]	Lat: 53º02’30” N Long: 126º34’16” W Elevation: 1280 m asl	1954-1960	25
1964	Historical photograph Austin Post[1]	Lat: 53º02’29” N Long: 126º34’24” W Elevation: 1285 m asl	1960-1964	35
1977	Aerial photograph BCB77044:36	Lat: 53º02’28” N Long: 126º34’27” W Elevation: 1285 m asl	1964-1977	5
1997	Aerial photograph BCC97096:92	Lat: 53º02’29” N Long: 126º34’44” W Elevation: 1370 m asl	1977-1997	16
2011	Cnes/Spot Image Google Earth	Lat: 53º02’29” N Long: 126º34’52” W Elevation: 1410 m asl	1997-2010	10
			Average Rate:	19

[1] http://www.gi.alaska.edu/cgi-bin/gdftp/imageFolio.cgi?img=0&search=Pattullo&cat=all&bool=phrase

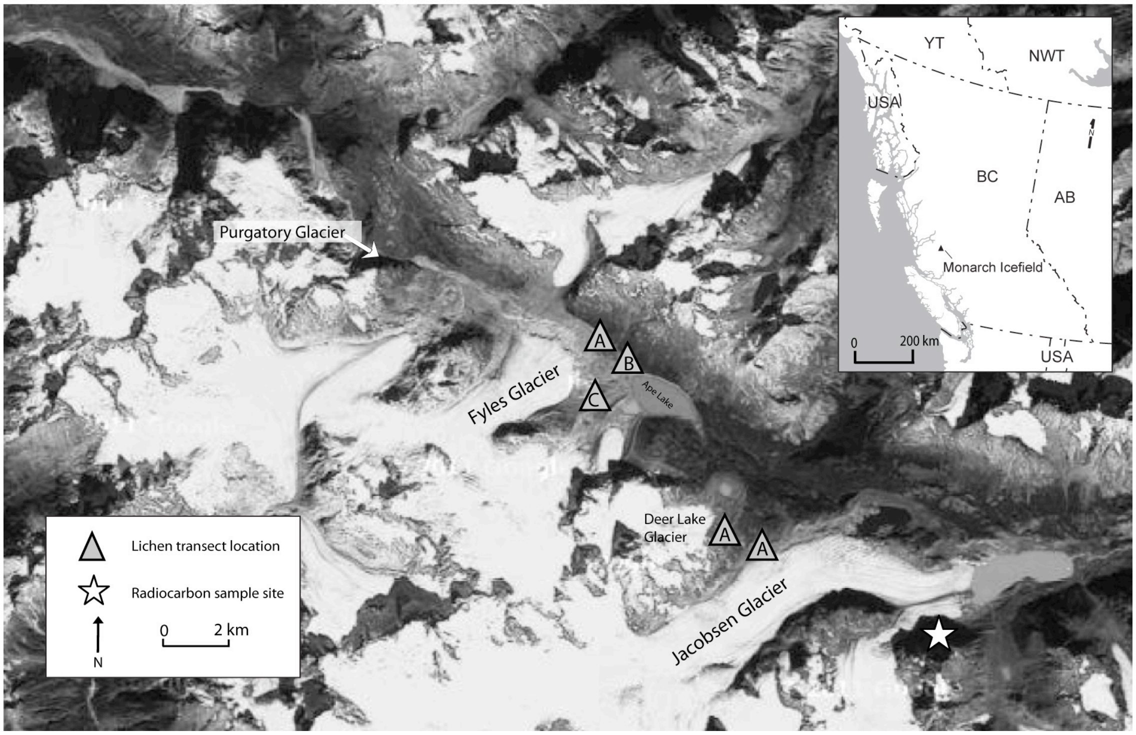


Figure 3.3 Map of the northeastern corner of Monarch Icefield and the location of lichen transects. The figure is based on 2011 satellite imagery (Google Earth).

Figure 3.4 Early-Little Ice Age lobes encroaching on surrounding subalpine forest behind the north lateral moraine at Jacobsen Glacier.

2008), eventually thinning and permitting enhanced subglacial seepage that resulted in catastrophic glacier outburst floods in 1984 and 1986 (Jones et al. 1985; Desloges et al. 1989; Desloges and Church 1992).

Deer Lake Glacier is a small hanging glacier located between Jacobsen and Fyles glaciers whose source is high on the northeastern slopes of Mt. Jacobsen. Prominent terminal and lateral moraines mark its past LIA extents into the surrounding forest (Figure 3.5). Between 1978 and 2011 the terminus of Deer Lake Glacier receded approximately 9 m/yr (BC78124:227; Google Earth 2011).

3.3 Methods

Lichenometric and dendroglaciological dating methods were used to describe the LIA behaviour of the four glaciers studied. Local tree-ring chronologies were collected and combined with archived chronologies to develop proxy records describing late-LIA climates in the central Coast Mountain region.

3.3.1 Lichenometry

Lichenometry is a relative dating technique that uses the radial growth characteristic of lichens to determine the minimum surface age of a deposit (Innes 1985). The largest lichen thallus on a surface is generally considered to provide the best estimate of surface age assuming a linear relationship between the thallus size and the age of the surface (Beschel 1973; Innes 1985). If the interval of time between the exposure of a surface and the colonization by lichens, and the age-growth relationship of the species is known, then a minimum date can be assigned based on the diameter of the lichen (Innes 1985). Where the intent is to develop a chronology of local LIA glacier activity, locally relevant growth curves are constructed from dated control points and are used to assign

Figure 3.5 View of northeast facing Deer Lake Glacier.

relative age dates to prominent moraine crests (*ie.* Larocque and Smith 2004).

Lichenometric data were collected along transects established across nested moraine crests at Fyles Glacier in 2002 and at Pattullo, Jacobsen and Deer Lake glaciers in 2010. Crests were labeled so that the most distal moraine was numbered ‘M1’ with the inner moraines numbered sequentially higher. Where a transect crossed a moraine crest the diameter (a- and b-axis) of the 20 largest thalli present were measured with digital calipers to the nearest ± 0.1 mm. Frequency histograms of the thalli sampled at each location were produced to test for normality. For samples where $n \geq 30$ a Shapiro-Wilk normality test was employed to detect deviations from normality and identify outliers (Larocque and Smith 2004). The single largest lichen thallus was used to assign minimum dates to moraines.

Rhizocarpon spp.

Rhizocarpon spp. is greenish- to yellow-coloured lichen widely used to date glacial deposits in PNA (Wiles et al. 2002; O’Neal and Schoenenberger 2003; Larocque and Smith 2004). *Rhizocarpon* spp. generally prefer silicious substrates, and its slow radial growth makes it attractive for use in lichenometric investigations (Innes 1985). In this instance, the *Rhizocarpon* spp. growth curve constructed by Smith and Desloges (2000) from control points located in the central Coast Mountain region, and more recently constrained by Koehler and Smith (2011), was considered the most appropriate. To confirm this, aerial photographs were examined to identify historical ice front positions. Two locations were visited in 2010 and the diameters of the largest *Rhizocarpon* spp. present were plotted on the growth-curve presented by Koehler and Smith (2011) for comparison.

Lichenometric dates were assigned to surfaces using equations defining the statistical lichen growth-age relationship and 95% confidence intervals for age-specific error estimates (Appendix A, Koehler and Smith 2011). Dates are reported in calendar years AD and provide minimum stabilization ages of moraine surfaces (Innes 1985).

Xanthoria elegans

Xanthoria elegans is relatively fast-growing yellowish-orange foliose to crustose lichen with an affinity for calcareous rocks (Murtagh et al. 2002). *Xanthoria elegans* has circumpolar distribution and is found in many treeless alpine environments in the western North American Cordillera (McCarthy and Smith 1995). Recognizing that the circularity of individual *X. elegans* thalli increases with age (Hill 1984), the species has been used to date landforms and deposits at various locations (Carrera and Andrews 1973; Chen 1989; Serebryanny and Solomina 1989). The only published North American *X. elegans* growth curves were established for sites in the Canadian Rocky Mountains (Osborn and Taylor 1975; McCarthy and Smith 1995).

During reconnaissance surveys in the central Coast Mountains it was recognized that circular *X. elegans* thalli were frequent colonizers of LIA moraines. To establish the usefulness of *X. elegans* for dating surfaces in the Coast Mountains, a regional growth curve was established with reference to dated control points. Minimum surface stabilization dates were assigned by interpolating the relative age of the largest diameter individuals and are reported in calendar years AD.

3.3.2 Tree-ring investigations

Tree-ring chronologies were established from samples of living and dead trees. Chronologies were collected to facilitate the crossdating of glacially-killed trees and to develop proxy climate models. Two increment cores were extracted at ≥90º with a standard 5.2 mm borer from stands of mature living trees (Schweingruber et al. 1988). Additional increment core samples were opportunitively collected from nearby dead standing snags and coarse woody debris in order to extend the earliest portion of the living chronology.

The core samples were allowed to air dry, polished with progressively finer sandpaper and scanned to produce high resolution digital images. A WinDendro™ (v.2006) tree-ring system (Guay et al. 1992) was used to measure the annual ring-widths to the nearest 0.01 mm. CDendro (v.4.1.1) was used to visually crossdate the cores using narrow index years and/or characteristic patterns (Larsson 2003). The software program COFECHA (v. 6.06) (Holmes 1999; Grissino-Mayer 2001) was used to verify the cross-dating using 50-year dated segments with 25-year lags, significant at the 99% critical level of correlation of 0.320 (Fritts 1976). The chronologies were standardized to produce ring-width indices using a double detrending approach in the program ARSTAN (Cook and Holmes 1999). To remove any age-growth trends a negative exponential curve was applied to the data and, where that was not appropriate, a negatively sloped line was used (Fritts 1976). A second detrending was applied to remove the effect of non-climatic factors on radial growth including stand dynamics and defoliation. This was achieved by applying a cubic smoothing spline with a common level of 67% frequency-response cutoff preserving low frequency variance (Cook et al. 1990). The express population signal (EPS) threshold year was calculated for 25 year moving periods to evaluate signal strength through the chronology and identify periods of insufficient sample depth (Wigley et al. 1984).

Dendroglaciology offers an opportunity to calendar date glacial events by cross dating glacially-killed trees to living tree ring chronologies (Luckman 1988, 1995; Smith and Lewis 2007). The field sites were systematically surveyed for subfossil wood and cross-sectional samples cut with a chainsaw.

The dendroglaciological samples were allowed to air dry, sanded and the annual rings measured along individual radia using WinDendro^{TM}. An attempt was first made to crossdate individual samples to the tree-ring chronologies, and where this failed perimeter wood samples were submitted for radiocarbon dating to Beta Analytic (Miami, Florida). The assigned radiocarbon ages (^{14}C yr BP) were calibrated using INTCAL 09 (Reimer et al. 2009) and the 2σ probability range reported in calibrated years AD.

3.3.3 Climate Reconstructions

Historic climate indicators, including summer temperature and snow pack, were examined for their contribution to ring-width variance using correlation analyses. Climate data from the long-term meteorological station (51º 40'N, 124º 24'W) and snow survey station (51º 38'N, 124º 19'W) at Tatlayoko Lake (Table 3.2) was obtained from Environment Canada, Adjusted Homogenized Canadian Climate Database (1930-2004; AHCCD 2011) and British Columbia River Forecast Centre (1964-1998; BC RFC 2011) and compared to standardized tree-ring indices using Pearson's correlation coefficients. Significant relationships ($p \leq 0.05$) confirmed the presence of climate signals in tree growth measurements and these variables were then used to construct proxy records of climate. Simple linear regression was used to construct a model between climate (predictor variable) and tree-ring width. The relationship was independently verified using the leave-one-out method, where linear regressions were applied to the relationship as many times as the station record was years long, with a different value excluded each time. The output of the leave-one-out process was compared to the actual record to test the predictive capacity of the model (Fritts 1976; Gordon 1982). Models with sufficient R^2 values and significant correlations between instrumental records and predicted records were used for proxy climate reconstructions.

3.4 Results

3.4.1 Lichenometric results

Growth-curve calibration

The *Rhizocarpon* spp. growth-curve presented by Koehler and Smith (2011) was derived from 22 control points. To ensure the curve measured radial growth trends observed within this study area, the largest lichen present at the ice front positions of Pattullo Glacier in 1946 (A10605:58) and 1954 (BCB1822:31) were measured in 2010. The two locations provide reliable control points (56 years: 27.3 mm; 64 years: 28.1 mm) consistent with those previously documented for thalli of this size (Larocque and Smith 2004; Koehler and Smith 2011).

Figure 3.6 presents the *X. elegans* curve developed for application in this study. Three control points were identified at Pattullo Glacier where both *Rhizocarpon* spp. and *X. elegans* cohabitate the same moraine crest. The substrate surface age was established using the regional *Rhizocarpon* spp. growth curve (Koehler and Smith 2011) and this age was assigned to the largest *X. elegans* thallus found (Table 3.3). An additional 13 control

Table 3.2 Meteorological and snow survey stations used in climate reconstructions.

Station	Data	Station ID	Record Length	Latitude/Longitude	Elevation (m asl)
Tatlayoko Lake	Meteorological	1088010	1930-2004	51°40’N, 124°24’W	870
Tatlayoko Lake	Snow survey	3A13	1964-1998	51°38’N, 124°19’W	1710

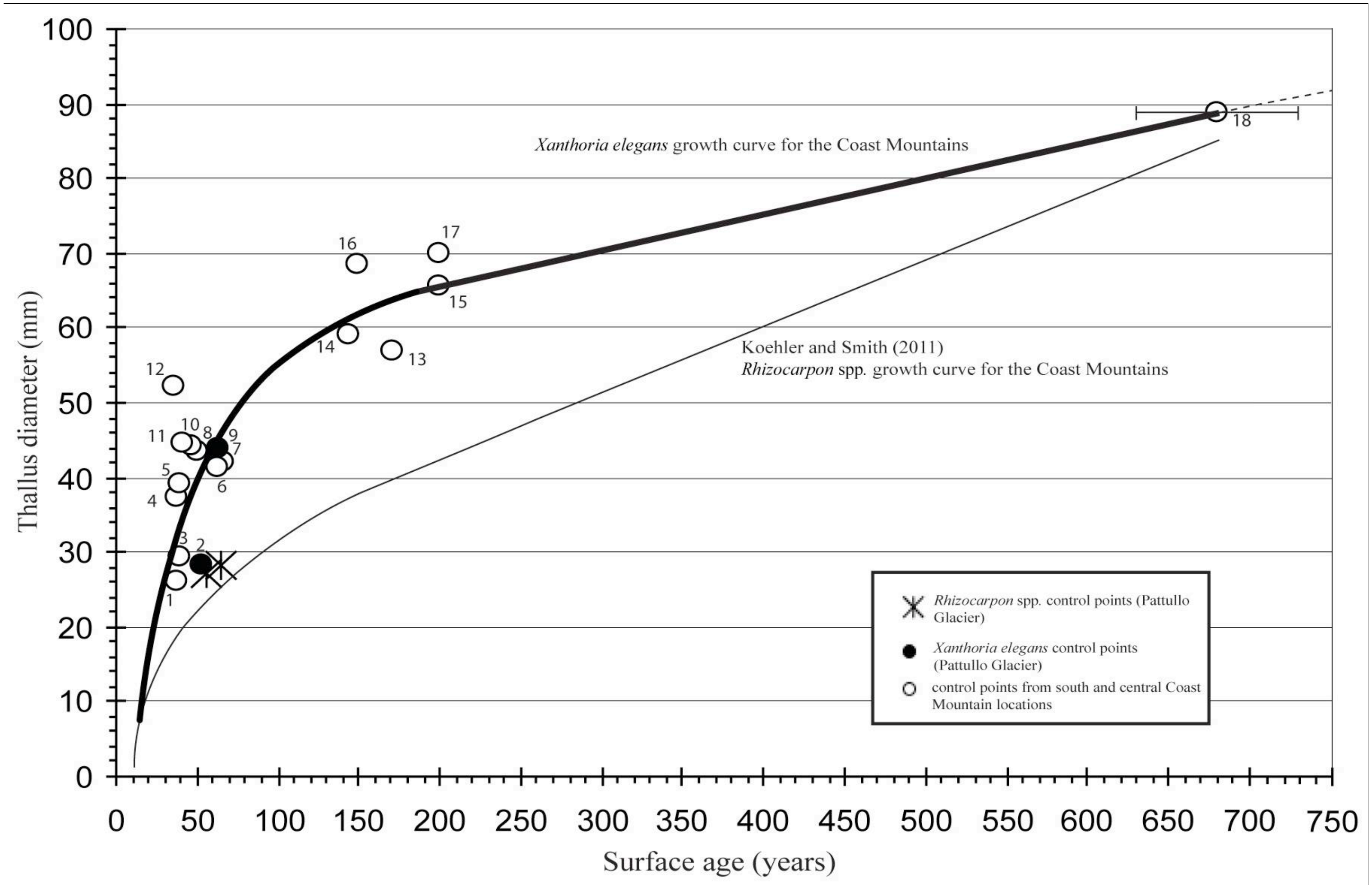


Figure 3.6 *Xanthoria elegans* growth curve with control points developed for the Coast Mountains. Shown for comparison is the *Rhizocarpon* spp. growth curve updated by Koehler and Smith (2011). Error bars for Point #18 incorporate ± 50 years of radiocarbon dating error (Larocque and Smith 2004).

points were compiled from sites in the Mt. Waddington area where both *Rhizocarpon* spp. (Larocque and Smith 2004) and *X. elegans* (Smith *unpublished notes*) were found colonizing moraine crests in 2001 and 2002. The surface age of each of those sites was established using the *Rhizocarpon* spp. growth curve presented by Koehler and Smith (2011), after which the largest *X. elegans* found at each site was assigned a relative age (Table 3.3).

The *X. elegans* growth curve presented in Figure 3.6 represents the age-growth relationship of this species in the Coast Mountains. The curve suggests that *X. elegans* experiences an interval of rapid radial growth until individual thalli reach an age of approximately 100 years. Following this, radial growth occurs at a slower rate (0.07 mm/yr), with weathering and habitat characteristics allowing relatively few individuals to attain ages exceeding 300 years (eg. Chen 1989; McCarthy and Smith 1995).

Pattullo Glacier

At Pattullo Glacier four lichen surveys were completed across the northern and southern lateral moraines (Figure 3.7). Transect A-a’ extends across five distinct lateral moraine crests at 1350 m asl. M1 was composed of large rounded boulders partially-colonized by mature subalpine fir trees. The largest *Rhizocarpon* spp. thallus (87.5 mm) found indicates M1 stabilized prior to 1321 AD (Table 3.4). M2 and M3 were located in close proximity and were similarly composed of large rounded boulders. The largest *Rhizocarpon* spp. on the moraines (65.7 and 68.5 mm) indicate they were deposited in ca. 1562 and 1531 AD, respectively (Table 3.4). The most proximal and largest moraines (M4 and M5) were formed in the late-18th and early-19th centuries (Table 3.4).

Table 3.3 Control points used to develop the *Xanthoria elegans* growth curve. Unless otherwise noted, control points were applied using *Rhizocarpon* spp. dated surfaces.

No.	**Location**	**Surface Age** (yr)	**Maximum *Rhizocarpon* spp. thallus** (mm)	**Maximum *X. elegans* thallus** (mm)	**Elevation** (m asl)
1	Byamee Glacier	37	19.4	26.0	1855
2[1]	Pattullo Glacier	56		28.4	
3	Pagoda Glacier	39	19.9	29.4	1620
4	Liberty Glacier	37	19.2	37.2	1800
5	Escape Glacier	40	18.9	39.0	1800
6	Pattullo Glacier	63	26.6	41.2	
7	Pattullo Glacier	67	27.5	42.0	
8	Baby Pagoda Glacier	51	23.7	43.4	
9[2]	Pattullo Glacier	64		44.0	
10	Razor Creek Glacier	47	22.6	44.3	1755
11	Razor Creek Glacier	41	20.8	44.7	1785
12	Liberty Glacier	36	18.7	52.2	1800
13	Siva Glacier	171	40.6	57.0	
14	Pattullo Glacier	144	38.2	58.9	
15	Whitesaddle Glacier	199	43.1	65.4	
16	Oval Glacier	150	38.7	68.3	1310
17	Liberty Glacier	199	43.1	69.8	1800
18	Whitesaddle Glacier	680	90.6	88.7	

[1] Control point derived using 1954 aerial photograph BCB1822:31

[2] Control point derived using 1946 aerial photograph A10605:58

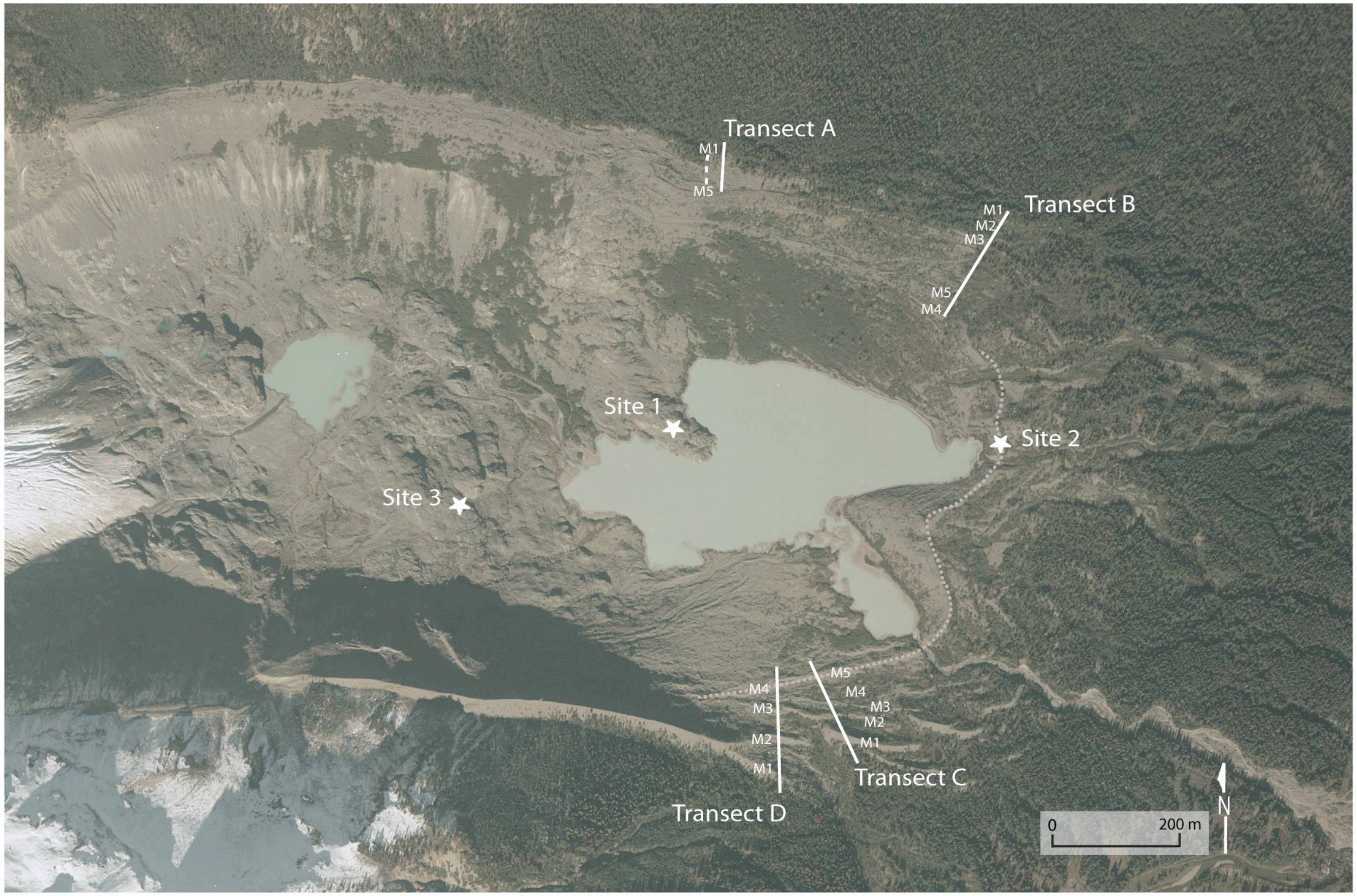


Figure 3.7 Location of lichen transects at Pattullo Glacier. Sites 1 and 2 indicated locations where the *Xanthoria elegans* growth curve was used to apply minimum surface ages. To establish which *Rhizocarpon* spp. growth curve was applicable to the study region, aerial photographs were examined to identify historical ice front positions at Sites 1 and 3.

Table 3.4 Lichenometric dates derived using *Rhizocarpon* spp. for lateral moraine stabilization at Pattullo, Fyles, Deer Lake and Jacobsen glaciers.

Location	**Elevation**	**Maximum Thallus**	**Average of 5 largest Thalli**	**Surface Age**[1]	**Moraine Date**	**95 % CI**[1]
	(m asl)	(mm)	(mm)	(yr)	(year AD)	(year AD)
Pattullo Glacier						
Transect A-a'	1350					
M1		87.5	82.8	689	1321	1165-1436
M2		65.7	64.2	448	1562	1461-1633
M3		68.5	63.3	479	1531	1424-1608
M4		45.0	43.2	220	1790	1744-1820
M5		43.2	38.6	200	1810	1768-1836
Transect B-b'	1275					
M1		91.0	59.3	727	1283	1117-1405
M2		66.7	59.8	459	1551	1448-1624
M3		61.0	53.4	396	1614	1526-1676
M4		39.3	35.8	157	1853	1821-1871
M5		33.2	30.7	102	1908	1910-1906
Transect C-c'	1275					
M1		52.0	48.5	297	1713	1648-1757
M2		23.1	18.0	49	1961	1963-1959
M3		28.5	26.9	73	1937	1939-1935
M4		19.5	17.4	38	1972	1974-1970
M5		39.6	34.0	160	1850	1852-1848
Transect D-d'	1290					
M1		57.8	46.1	361	1649	1569-1704
M2		55.5	44.1	335	1675	1601-1725
M3		40.1	31.0	165	1845	1847-1843
M4		34.9	33.3	115	1895	1897-1893
Fyles Glacier						
Transect A-a'	1480					
M1		81.1	72.7	618	1384	1244-1486
M2		77.4	58.4	577	1425	1294-1520
M3		57.2	51.5	354	1648	1570-1702
M4		48.4	45.1	257	1745	1689-1781
M5		41.8	39.6	184	1818	1779-1841
M6		37.7	35.8	139	1863	1835-1878
Transect B-b'	1400					
M1		63.4	59.9	423	1579	1485-1646
M2		54.2	52.3	321	1681	1610-1729
Transect C-c'	1560					
M1		51.2	48.3	288	1714	1651-1756
M2		38.9	35.5	152	1850	1819-1867
Deer Lake Glacier						
Transect A-a'	1505					
M1		72.2	68.4	520	1490	1373-1575
M2		67.2	57.0	465	1545	1441-1620
M3		53.9	49.5	318	1692	1623-1740
M4		32.4	28.3	96	1914	1916-1912
Jacobsen Glacier						
Transect A-a'	1520					
M1		69.8	63.4	493	1517	1406-1596
M2		69.0	58.3	484	1526	1417-1603

[1] Calculated using single maximum thallus and equations from Koehler and Smith (2011), Appendix A.

Transect B-b’ crossed five moraine crests and was located near the eastern extent of the northern forefield at 1275 m asl. M1 is partially-colonized by subalpine fir trees (<200 years old). The largest *Rhizocarpon* spp. thallus located on M1 (91.0 mm) describes a period of moraine construction prior to the late 13th century (Table 3.4). *Rhizocarpon* spp. on M2 (66.7 mm) and M3 (61.0 mm) describe nearly contemporaneous moraine stabilization events at 1551 and 1614 AD, respectively (Table 3.4). Lichen thalli on M4 and M5 record episodes of ice expansion dating to ca. 1853 and 1908 AD (Table 3.4).

Transect C-c’ is located 0.75 km south of Transect B-b’ (Figure 3.7) where five moraine crests are present at 1275 m asl. M1 contains large boulders and is mantled by mature subalpine fir trees and colonized *Rhizocarpon* spp. indicates the outermost moraine at C-c’ was deposited in ca. 1713 AD. M2, M3 and M4 were truncated by stream incision and fluvial disturbance. The few thalli found (n= <8) were small and describe stabilization dates of ca. 1961, 1937 and 1972 AD respectively. Numerous lichen on M5 (largest thallus = 39.6 mm) indicate it was deposited in ca. 1850 AD (Table 3.4).

Transect D-d’ crosses four lateral moraine crests on the south side of the Pattullo Glacier forefield at 1290 m asl (Figure 3.7) The largest *Rhizocarpon* spp. on M1 (57.8 mm) indicates the moraine stabilized prior to ca. 1649 AD (Table 3.4). M2 (55.5 mm) was deposited on the proximal slope of M1 in ca. 1675 AD (Figure 3.7). Lichen thalli on M3 and M4 describe moraine building episodes dating to the mid- and late-19th centuries (Table 3.4).

The *X. elegans* curve was used to assign minimum surface ages at two locations (Figure 3.7). At Site 1 a thalli on a bedrock promontory near the large proglacial lake measured 51.9 mm, providing a surface stabilization date of ca. 1930 AD. Site 2 was associated with a moraine crest also intersected by transects C-c' (M5) and D-d' (M4). *Rhizocarpon* spp. thalli suggest moraine stabilization occurred from 1850 to 1895 AD. *Xanthoria elegans* (58.9 mm) at Site 2 indicate the moraine surface dates to ca. 1885 AD.

Fyles Glacier

Lichenometric surveys were completed along three transects at Fyles Glacier (Figure 3.3). Transect A-a' intercepted the crest of six nested moraines approximately 1.3 km east of the contemporary margin of the glacier at 1480 m asl (Figure 3.3). The largest *Rhizocarpon* spp. located on the forested surface of M1 (81.1 mm) indicates the outermost moraine was deposited prior to ca. 1384 AD (Table 3.4). The partially-vegetated surface of M2 (77.4 mm) was assigned a moraine stabilization date of ca. 1425, and the largest thalli on M3 and M4 suggest moraine deposition during the mid-17th and mid-18th centuries respectively (Table 3.4). *Rhizocarpon* spp. on M5 (41.8 mm) and M6 (37.7 mm) date the moraine surfaces to ca. 1818 and 1863 AD (Table 3.4).

Transect B-b' was located about 500 m south of Transect A-a' and crosses two moraine crests at 1400 m asl (Figure 3.3). The largest *Rhizocarpon* spp. on M1 (63.4 mm) indicate the moraine was deposited prior to 1579 AD (Table 3.4). Lichen on the surface of M2 indicates it was deposited in ca. 1681 AD (Table 3.4).

Transect C-c' crosses the crests of two large treeless lateral moraines flanking the southern margin of Fyles Glacier at 1560 m asl (Figure 3.3). *Rhizocarpon* spp. colonizing the surface of two crests date to the early-18th (M1, 51.2 mm) and mid-19th (M2, 38.9 mm) centuries (Table 3.4).

Deer Lake Glacier

Four prominent, treeless, nested lateral moraines demarcate the southern extent of recent glacier activity at Deer Lake Glacier (Figure 3.8). Transect A-a’ crosses the moraine crests between 1505-1451 m asl. *Rhizocarpon* spp. thalli indicate M1 (72.2 mm) stabilized prior to 1490 AD and M2 (67.2 mm) was deposited in ca. 1545 AD (Table 3.4). Lichens on M3 and M4 indicate they were deposited following glacier advances prior to 1692 and 1914 AD respectively (Table 3.4).

Jacobsen Glacier

At Jacobsen Glacier transect A-a’ intercepts two small partially-vegetated terminal moraine crests at 1520 m asl (Figure 3.8). Resulting from the lateral spillover and extension of trunk valley ice into the surrounding forest, *Rhizocarpon* spp. on the crest of M1 (69.8 mm) and M2 (69.0 mm) indicate this advance occurred before ca. 1517 and 1526 AD.

The *X. elegans* curve was used to assign minimum surface ages at one site at Jacobsen Glacier. A thalli growing on bedrock approximately 250 m north of the contemporary ice margin measured 26.7 mm, indicating a surface stabilization date of ca. 1980 AD.

3.4.2 Dendroglaciology findings

Numerous detrital wood fragments were located below gullies eroded into the proximal face of Jacobsen Glacier’s southern lateral moraine at ca. 1260 m asl (Figure 3.9). Perimeter wood from a small branch spilled from the moraine has an age of 1070±40 ^{14}C yr BP (Table 3.5; 892-1023AD).

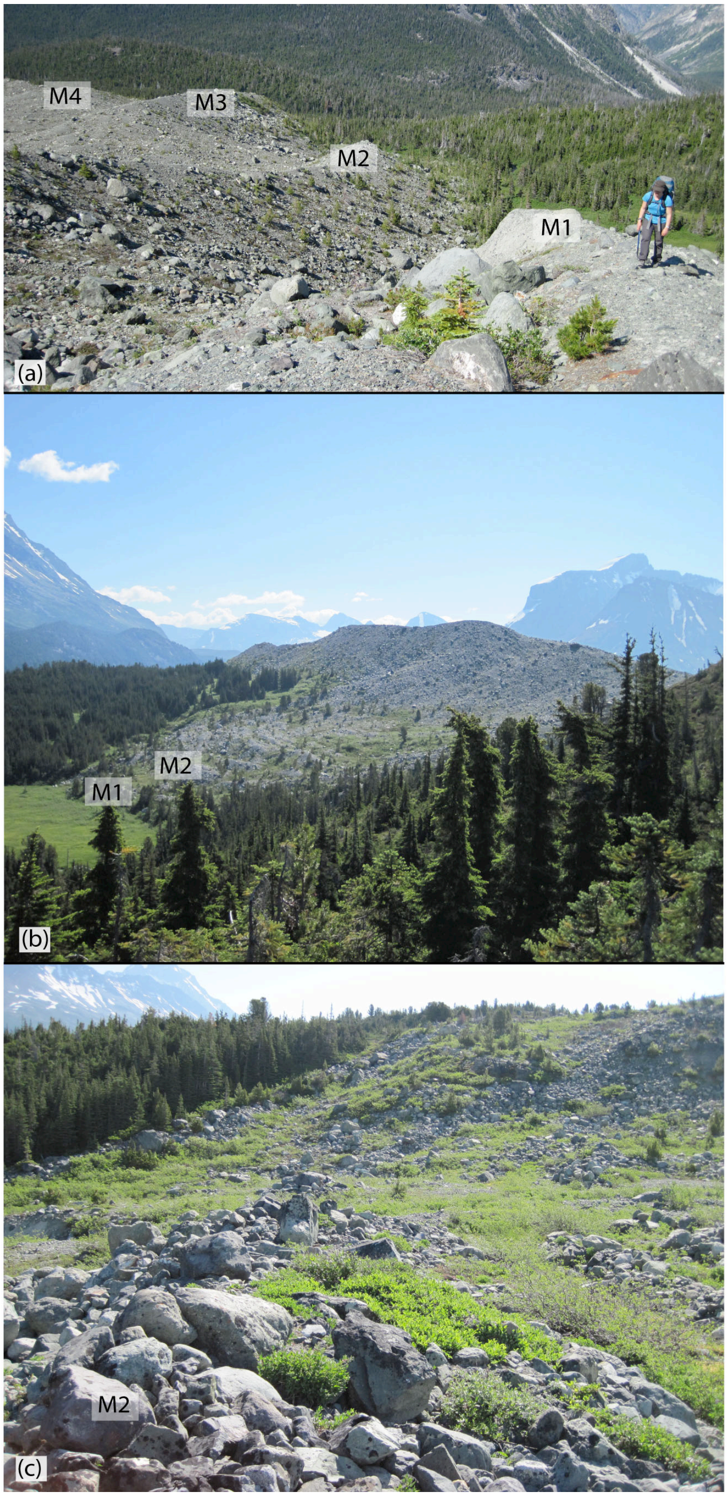


Figure 3.8 (a) The moraines included in the Deer Lake Glacier lichen transect A-a’. (b) The location of lichen transect A-a’ at Jacobsen Glacier. (c) The crest of moraine M2 in lichen transect A-a’ at Jacobsen Glacier.

Figure 3.9 Star indicates the location of the detrital branch dated to 1070±40 ^{14}C yr BP (892-1023AD) spilled from the southern lateral moraine at Jacobsen Glacier.

Table 3.5 Summary of Monarch Icefield area Little Ice Age radiocarbon dates.

Laboratory No.	Description	^{14}C Age (yr BP)	2 σ Calibration ^{a}	Lat./Long	Elevation (m asl)
Jacobsen Glacier					
S-2979[1]	Ah horizon of paleosol	400±60	1427-1636 AD	52º 04'N, 126º 08'W	1495
Beta-286104[2]	Detrital fragment	1070±40	892-1023 AD	52º 02'N, 126º 04'W	1351
Purgatory Glacier					
S-2977[1]	Forest litter from paleosol	785±70	1146-1306 AD	52º 10'N, 126º 21'W	595
S-2978[1]	Forest litter from paleosol	630±65	1272-1420 AD	52º 10'N, 126º 21'W	595

[a] Radiocarbon dates converted to calendar ages BP (2σ) using CALIB 6.0 (Reimer et al. 2009)

[1] Desloges and Ryder (1990), S = Saskatchewan

[2] This study, Beta = Beta Analytical Inc. Miami, FL.

3.4.3 Dendroclimatological findings

Increment cores were collected from mature subalpine fir trees within four stands in the central Coast Mountains between 1997 and 2010 (Table 3.6). Tree-ring sampling was generally conducted in mixed stands of mountain hemlock, subalpine fir and whitebark pine. Site-specific subalpine fir chronologies were constructed and combined to create a master chronology (n = 122 trees) spanning 478 years (1532-2010 AD).

Correlation analyses between the master chronology and instrument record from Tatlayoko Lake revealed strong significant ($p \leq 0.05$) relationships with two climate variables. For the instrumental, record May 1st snow water equivalents (SWE) correlated most strongly with the residual tree ring chronology ($r = -0.511$) and a significant positive relationship was observed between standardized tree ring chronology and mean June/July air temperature ($r = 0.493$).

A positive relationship between subalpine fir radial growth and June/July air temperature has been previously documented (Villalba et al. 1994; Parish et al. 1999; Luckman 2000; Spelchtna et al. 2000; Peterson et al. 2002). Warmer spring and summer temperatures accelerate melting of residual winter snowpacks and lengthen the growing season (Peterson and Peterson 1994). A negative relationship between radial growth and May 1st snowpack is explained by a reduced growing season and a negative impact on seedling establishment (Peterson et al. 2002).

Proxy climate records were developed to 1610 AD by creating statistical models describing linear regression relationships between May 1st SWE ($R^2 = 0.26$), mean June/July air temperature ($R^2 = 0.24$), and the master chronology (Table 3.6). Explaining 24 and 26% of the measured variation in annual radial growth, the models derived from

Table 3.6 Tree-ring chronology sampling locations and statistics.

Chronology	Lat./ Long	Elevation	Sample year	No. of series /trees	Interseries correlation	Total length	Mean Sensitivity
		m asl				(yr)	
Pattullo G.	53º02'N, 126º33'W	1282	2010	94/48	0.539	451	0.192
Tzeetsaytsul G.	52º35'N, 126º22'W	1260	1997	25/13	0.525	269	0.190
Jacobsen G.	52º04'N, 126º08'W	1477	2010	66/33	0.567	477	0.193
Hammer Lake	52º12'N, 126º19'W	1291	2010	50/28	0.553	283	0.197
Master[1]				235/122	0.508	478	0.193

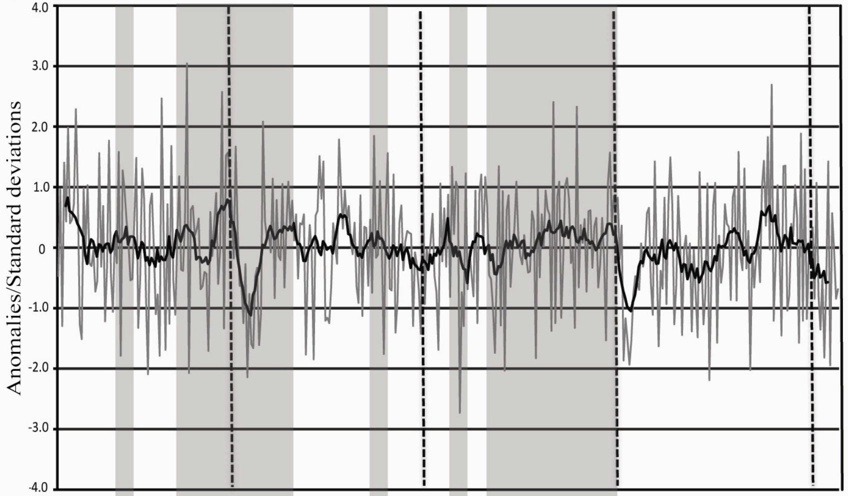
[1] Master Chronology includes chronologies from: Pattullo Glacier, Tzeetsaytsul Glacier, Jacobsen Glacier, and Hammer Lake.

these relationships provide high resolution and complimentary records of late-LIA climates over the past four centuries (Figure 3.10). Sustained intervals with lower than average June/July temperature occurred from ca. 1620-1630, 1680-1710, 1720-1790, 1860-1905, 1960-1970, 1975-2005 AD (Figure 3.10). Intervals with above average seasonal snowpacks occurred between ca. 1620-1630, 1670-1680, 1690-1710, 1720-1740, 1760-1770, 1810-1820, 1860-1880, 1960-1965, and 1970-2000 AD (Figure 3.10).

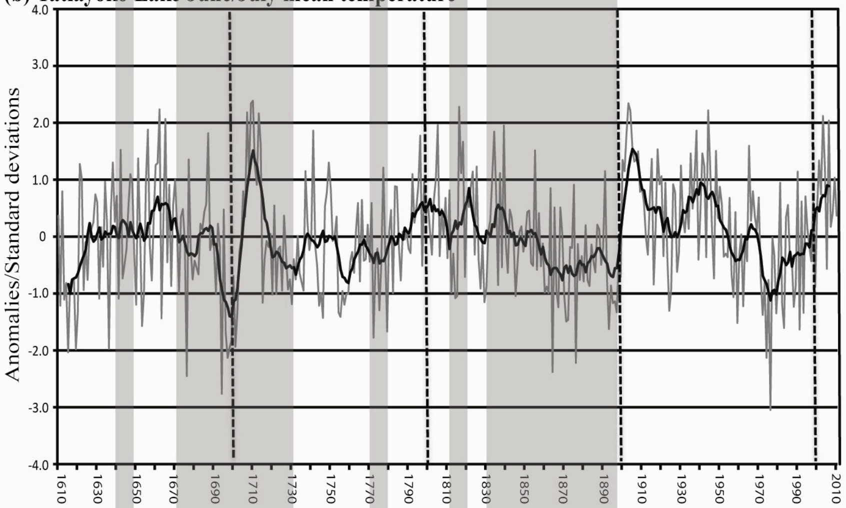
3.5 Synthesis and Interpretation

The emerging record of LIA glacier activity in PNA suggests glacier expansion occurred during intervals characterized by decreased ablation season temperatures and increased winter precipitation (Luckman 2000; Koch et al. 2007a). Moraine building episodes identified in this research corroborate the findings of complimentary lichenometric and dendrochronologic studies in the region where the LIA is commonly reported to have been initiated between the 11th and 13th centuries (Calkin et al. 2001; Menounos et al. 2009). There is, however, equivocal evidence for an earlier interval of glacial expansion following the well-documented first millennium advance centered on 400-700 AD (Reyes et al. 2006). Allen and Smith (2007) and others (Menounos et al. 2009) report that some Coast Mountain glaciers were expanding into mature forests at ca. 1000 AD. Evidence supporting this advance is recorded in the study area by the detrital branch dating to 892-1023 AD spilled from the southern lateral moraine at Jacobsen Glacier (Table 3.5). Its suspected burial position, approximately 30 m below the moraine crest, suggests Jacobsen Glacier was close in size to its maximum LIA dimensions at this time.

(a) Tatlayoko Lake SWE May 1st



(b) Tatlayoko Lake June/July mean temperature



(c) Minimum moraine dates for the central Coast Mountains and sunspot minima

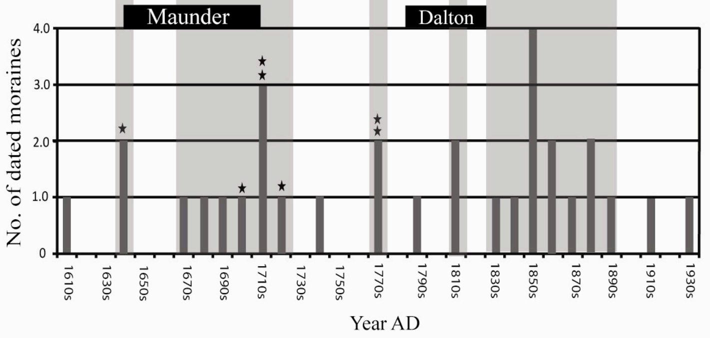


Figure 3.10 Reconstructions of anomalies of May 1st SWE (a) and June/July air temperature (b). Black lines represent a 10-year running mean of the data. (c) Frequency distribution of minimum dates of moraines in the central Coast Mountains between 1610 and 1930 AD. Includes dates assigned in this research as well as dates assigned in the central Coast Mountains by Desloges (1987) and Smith and Desloges (2000). Two intervals of sunspot minima (Raspopov et al. 2008) are indicated with black horizontal bars. Stars indicate outermost moraines. The vertical grey bars highlight moraine building episodes that are broadly synchronous with lower than average summer temperatures and higher than average snowpack conditions.

Early-LIA glacier activity in the study area is recorded by moraines dating to 1283 and 1321 AD at Pattullo Glacier, and 1384 AD at Fyles Glacier. Similar 12th and 13th centuries early-LIA expansion dates are documented in both the northern (Ryder and Thomson 1986; Jackson et al. 2008) and southern Coast Mountains (Larocque and Smith 2003; Koch et al. 2007a); as well as between 1240 and 1300 AD at glaciers in the Northern Cascade Mountains (Crandell and Miller 1964) and in the mountains of coastal Alaska (Calkin et al. 2001).

A moraine building period during the late 15th and the 16th centuries was identified at all four glaciers studied. These findings corroborate the report of Desloges and Ryder (1990) who dated a paleosol (Sample J1) entombed by Jacobsen Glacier in 400±60 ^{14}C yr BP (Table 3.5; 1427-1636 AD) beneath moraines now lichenometrically dated to 1517 (M1) and 1526 (M2) at the same location (Figure 3.8; Transect A-a'). At Pattullo Glacier a contemporary moraine was deposited in ca. 1490 AD. Corresponding evidence of glacier expansion at this time from the southern Coast Mountains (1506-1524 AD; Larocque and Smith 2003), the Northern Cascades (Heikkinen 1984; Sigaoos and Hendricks 1972) and in Alaska (Denton and Karlen, 1977; Wiles and Calkin 1994; Barclay et al. 2001) indicate this was a regional event.

Evidence for the late-LIA expansion and retreat of glaciers in the study area is recorded by moraines constructed after the late-17th century at several sites (Table 3.4). Prominent terminal and lateral moraines were constructed at Fyles (1681 AD), Deer Lake (1692 AD), Tzeetsaytsul (1700 AD; Smith and Desloges 1990), Pattullo (1713 AD), and Borealis (1722 AD; Desloges 1987) glaciers (Table 3.4 and Figure 3.11). Following this

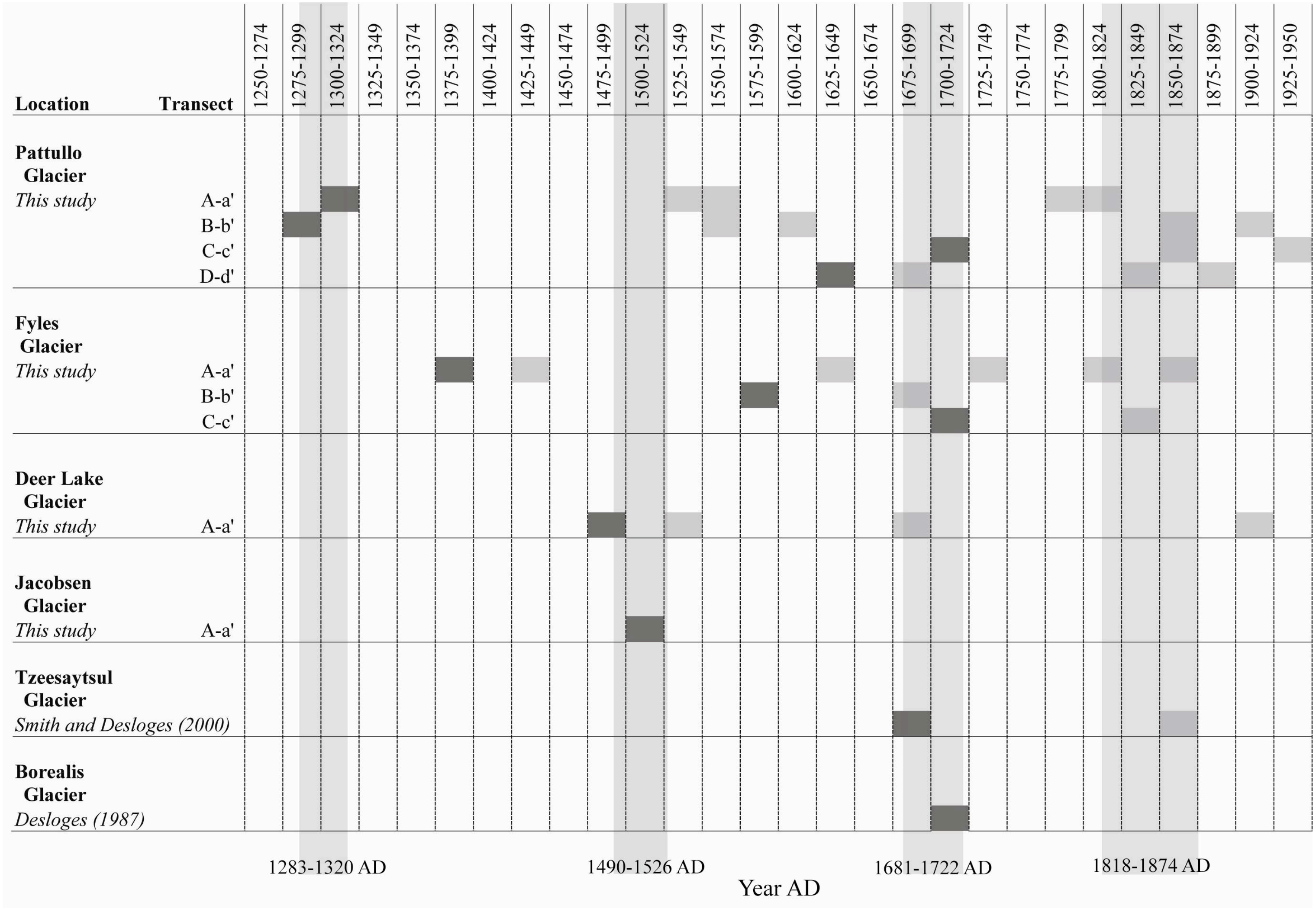


Figure 3.11 Summary of lichenometrically dated moraines from the central Coast Mountains. Dates are presented in years AD. The dark grey boxes represent outermost moraine dates and the light grey boxes indicate inner moraine dates. The shaded grey vertical bars indicate dominant periods of moraine building.

advance glaciers throughout the study area receded before readvancing to moraines constructed in the mid-19th Century at Fyles (1818-1863 AD), Pattullo (1845-1853 AD), Tzeetsaytsul (1853 AD; Smith and Desloges 1990) and Deer Lake (1874 AD; Desloges 1987) glaciers (Tables 3.4 and 3.7). At Pattullo (1895-1908 AD) and Deer Lake (1914 AD) glaciers, the subsequent historical retreat of glaciers in the study area was interrupted by a minor readvance or stillstand (Figure 3.11). Corresponding periods of late-LIA moraine building recorded at many PNA glaciers (Larocque and Smith 2003; Menounos et al. 2009) suggest that climate forcing mechanisms originating in the North Pacific Ocean likely contributed to these regionally synchronous glacier fluctuations (Larocque and Smith 2003).

Larocque and Smith (2005) present a late-LIA glacier mass balance record from the nearby Mt. Waddington area describing intervals of positive mass balance in the 1700s, 1750s, 1820s-1830s and 1970s that correspond to intervals of moraine building in the southern Coast Mountains (Larocque and Smith, 2003). Similar conditions likely prevailed in the central Coast Mountain region, where the deep seasonal snow packs and cool summer temperatures reconstructed from tree-ring records (Figure 3.10) would have resulted in positive mass balance conditions leading to moraine building episodes (Figure 3.10).

Wiles et al. (2004), Menounos et al. (2009) and Gavin et al. (2011) draw attention to the temporal synchronicity between diminished sunspot activity, cooler summer temperatures, positive glacial mass balance and moraine building episodes. Periods of low sunspot activity have been identified using tree-ring proxies in ca. 1020 AD (Oort), 1300 AD (Wolf), centered at ca. 1450 AD (Spörer), from the late-1600s to the early 1700s AD (Maunder) and during the early decades of the 1800s AD (Dalton) (Raspopov et al. 2008). These intervals correspond closely to the intervals of glacier expansion recorded in the central Coast Mountain area in 890-1020, 1280-1320, 1490-1530, 1680-1720 and 1820-1870 AD (Figure 3.10). This linkage suggests that in this setting both glaciers and trees are sensitive indicators of climate changes driven by large-scale solar forcings.

3.6 Conclusion

The findings of this research provide a detailed characterization of LIA glacier activity in the central Coast Mountain region. Moraine building episodes at Pattullo, Fyles, Deer Lake and Jacobsen glaciers in the central Coast Mountains were identified using lichenometric and dendroglaciological research techniques. Tree-rings were used as a proxy to reconstruct climate parameters in order to gain insights into mid- to late-LIA climates. The presentation and application of a preliminary regional based growth curve for *X. elegans* provides an additional means for dating LIA deposits in this region.

Table 3.7 summarizes the outer moraines and inner moraines of all transects included in this study and from previous research conducted in the central Coast Mountains. This research indicates that most glaciers in the central Coast Mountains reached their maximum downvalley extent prior to 1775 AD. The dominant moraine building episodes identified between 1280-1320, 1490-1530, 1680-1720 and 1820-1870 AD broadly correspond to intervals characterized by lower summer temperature and above average winter precipitation.

The LIA glacier histories described in this study corroborate the findings of research in other regions of PNA. Recognition of broad regional synchronicity of glaciers along the coast of PNA suggests that alpine glaciers are responding to one or more large scale climate forcings. The major periods of glacial advance in the past millennium in the central Coast Mountains correspond with times of low sunspot numbers, suggesting that glacial activity may in part be forced by fluctuations in solar activity.

Chapter 4: Conclusion

4.1 Summary and Conclusions

The intent of the research described in this thesis was to reconstruct and document Holocene glacier activity and to describe late-LIA climates in the central Coast Mountains. Previous research undertaken in this region describes Holocene glacial advances during the middle- to late-Neoglacial (3.50-1.00 ka; Desloges and Ryder 1990) and during the LIA (Desloges, 1987; Desloges and Ryder 1990; Smith and Desloges 2000). While these insights provide a perspective on the glacial history of the region, spatially-extensive fieldwork recently completed in the southern Coast Mountains has led to the documentation of a complex history of ice advance and retreat within the Holocene (Koch et al. 2007a; Koch et al. 2007b; Menounos et al. 2009). At least six distinct Holocene glacier advances have been identified, as well as the recognition that multiple ice front oscillations likely characterized each distinct advance (Koch et al. 2007b; Menounos et al. 2009).

The goal of my research was to explore recently deglaciated forefields in the central Coast Mountains region to locate the remains of forests buried during intervals of Holocene glacier expansion. The primary objectives of the research were to increase understanding about glacier activity during the mid-Holocene and to establish whether the emerging chronology of LIA glacier activity in PNA was represented in the central Coast Mountains. A third objective of the research was to examine the apparent synchronicity between solar forcings, region-specific climate variations and glacial chronologies throughout the Coast Mountains.

Dendroglaciological research methodologies were applied during field work completed in July 2010 at Pattullo and Jacobsen glaciers. Following the discovery of mid-Holocene aged deposits at Jacobsen Glacier, the scope of my investigations expanded to include the analysis of previously unreported dendroglaciological samples from Fyles (2002), Tchaikazan (2002), Canoe (2006) and Icemaker (2006) glaciers. An outcome of the research is enhanced understanding and resolution of mid-Holocene glacial activity between 7.50-4.00 ka.

In situ remains from Canoe Glacier substantiate previous reports of distinct mid-Holocene episodes of glacial expansion at 5.45-5.05 and 3.72-3.47 ka. Stratigraphically constrained woody debris at Fyles Glacier describes the progradational history of a Gilbert-type delta deposited during an expansion of the glacier between 7.02-5.47 ka. Ice-pressed trees, *in situ* stumps and detrital wood at Jacobsen Glacier indicate the glacier was advancing downvalley between 6.68-6.30 ka and was morphologically similar in size and shape as it is today. Detrital wood washed from beneath Tchaikazan Glacier describe multiple pulses of glacial expansion 6.30 and 3.90 ka. Glacially-sheared stumps in growth positon at Icemaker Glacier indicate the glacier was expanding downvalley at 6.64-6.32 ka.

Collectively, the dendroglaciological evidence reported in this thesis confirms the timing of these mid-Holocene advances, lending support for interpretations of their regional extent and character (Koch et al., 2007b; Menounos et al., 2009). This regional perspective creates a better viewpoint from which to consider glacier response to large-scale climate change at this important transition in Holocene climate.

Lichenometric research methodologies were employed during fieldwork in July 2010 to describe the LIA behaviour of Pattullo, Deer Lake and Jacobsen glaciers. The inclusion of previously unreported lichen sampling at Fyles Glacier (2002) and in the Mt Waddington area (2000, 2001) provided additional insights.

The application of lichenometry necessitated the construction of a growth curve for *X. elegans* and an evaluation of whether the existent *Rhizocarpon* spp. growth curve was relevant in this setting. An assessment of lichenometric measurements from the southern and central Coast Mountains provided the opportunity to construct a *X. elegans* growth curve constrained by 18 control points. This regional curve provides a working framework for additional control points and revision. A notable contribution was the inclusion of a 680 year-old lichen that indicates that individual *X. elegans* can attain very old ages and may potentially be a more useful species for lichenomentic dating than previously reported (Osborn and Taylor, 1975; McCarthy and Smith, 1995).

Moraine building episodes were identified at ca. in 890-1020, 1280-1320, 1490-1530, 1680-1720, 1780 and 1820-1870 AD and highlight the complex nature of glacier fluctuations during the LIA. Proxy climate variables, derived from tree-rings, indicate that during the late-LIA the constructon of moraines in the central Coast Mountain area broadly followed periods when cooler and snowier conditions prevailed.

LIA moraine construction in the study area occurred during intervals that are broadly comparable to those identified elsewhere in the Coast Mountains (Larocque and Smith 2003; Menounos et al. 2009). The regional synchronicity in Coast Mountain moraine construction in the LIA appears to describe a widespread glaciological response intimately linked to periods of lower sunspot activity (Wiles et al. 2004; Menounos et al. 2009).

4.2 Research limitations

a) The provenance of subfossil wood

The provenance and stratigraphic position of dated materials requires careful consideration when interpreting radiocarbon ages (Rothlisberger et al, 1980; Osborne 1986; Ryder and Thomson, 1986). In the context of my interpretations:

- While the subfossil wood recovered from the delta complex at Fyles Glacier provided insights into a singular interval of glacier expansion, it was not possible to identify the source of the detrital wood. Nonetheless, the character of the wood samples (detrital twigs and branch fragments) and the time-transgressive depositional history implicit in their stratigraphic placement indicates that snow avalanches were likely responsible for supplying wood to the delta front.
- *In situ* glacially sheared stumps and ice-pressed detrital wood provide strong evidence for an advancing ice margin at the time of tree-death. Fossil trees preserved as boles or branches in or at the surface of moraines without bark or small attached branches are considered to have undergone transport from *in situ* trees at the base of the glacier or within till at the glacier margins of the glacier. In such instances samples must be carefully interpreted taking into consideration the possibility of snow avalanching or debris flows as a mechanism for deposition.

b) Lichenometry

The application of lichenometry to date surfaces has proven applicable in the Coast Mountain region (Smith and Desloges 2000; Larocque and Smith 2003; Allen and

Smith 2007; Koehler and Smith 2011). In the context of my application, however, several issues required consideration:

- The lichenometric curve for *Rhizocarpon* spp. presented by Koehler and Smith (2011) contains control points from both the southern and central Coast Mountains. While the curve has demonstrated regional applicability, it may not account for local radial growth variations resulting from micro-scale topography, aspect and climate.
- Lichen data collected in 2000-2002 and 2010 provided the basis for this research. While all of the lichen measurements were collected using standard practices, errors associated with multiple field personnel collecting data may have been incorporated.
- My interpretations assumed that the largest lichen located on a deposit was the oldest individual present. It remains a possibility that during sampling the largest thalli on a deposit was not present at the transect intersect points and/or that the implicit age-size relationship was incorrect. To account for these possibilities, the age assigned to the deposits were presented as minimum surface stabilization dates.
- Trees and shrubs colonize some lichenometrically-dated moraine surfaces (*ie.* outermost moraines at Pattullo Glacier). While maturation of this vegetation will eventually overwhelm the surface of the moraines resulting in lichen-kill (Innes 1985), at early successional stages canopy shading may result in comparatively greater radial growth than on non-vegetated surfaces (Larocque and Smith, 2003).

4.3 Future research

This research provides new insights into understanding the spatial and temporal resolution of mid-Holocene and LIA glacial activity in the Coast Mountains. Additional research is recommended to:

- evaluate whether the mid-Holocene climate changes recorded by glaciological changes in this region are time-transgressive. The findings of this research suggest a mid-Holocene transition to cooler and wetter conditions began first in the southern British Columbia Coast Mountains and the Cascade Mountains of Washington State before propogating northward.
- evaluate the reliability and applicability of the *X. elegans* radial growth curve. Firmly dated control points should be established (*ie.* from aerial and historical photographs, historical monuments and tombstones) to provide absolute surface ages and ecesis estimates for growth curve calibration.

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Appendix A: *Rhizocarpon* spp. growth curve, and associated equations, for the Coast Mountains presented in Koehler, L. and Smith, D. 2011. Late Holocene glacial activity in Manatee Valley, southern Coast Mountains, British Columbia, Canada. Canadian Journal of Earth Sciences, 48: 603-618.

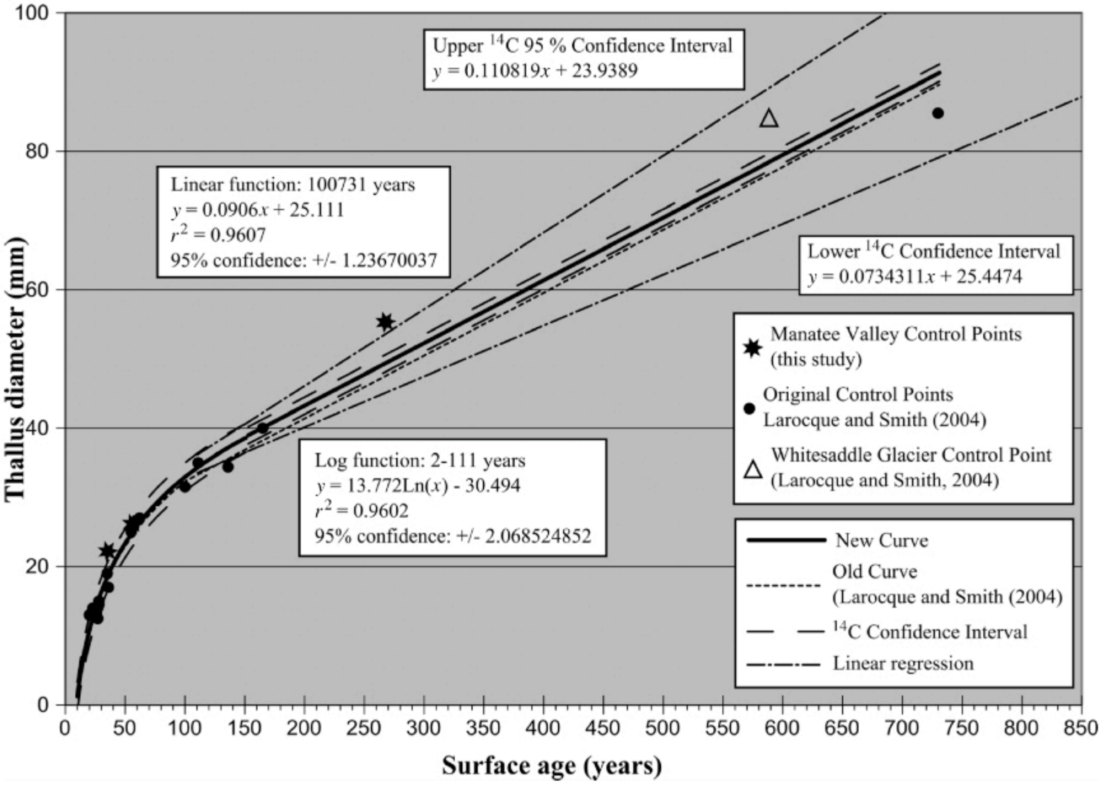

Figure A1 Revised *Rhizocarpon* spp. growth curve for the Coast Mountains presented in Koehler and Smith (2011: 614).