Little Ice Age Investigations in Strathcona Provincial Park, Vancouver Island, B.C.

by
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We accept this thesis as conforming to the required standard

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Abstract

Dendroglaciological and lichenometric techniques were used to establish the Little Ice Age glacial history of two glaciers (Septimus and Colonel Foster) in Strathcona Provincial Park, Vancouver Island, British Columbia. This record was placed in the context of the regional Little Ice Age record of the Pacific North American Cordillera, and evaluated in the context of proxy mass balance and climatic records from 1600 to 1994 AD.

The Strathcona Little Ice Age chronology compares well with regional moraine records from coastal British Columbia, Washington, Alaska, and the Canadian Rocky Mountains. Three generally synchronous episodes of glacier activity are noted: late 1600's to early 1700s, and late 1800s, and mid 1930's. Dating results suggest two levels of glacier response: (*i*) asynchronous responses to local factors such as microclimate, topography, and glacier geometry, and (*ii*) synchronous responses to larger-scale climatic forcing.

Dendroclimatological analysis indicates that ice-proximal mountain hemlock are also responding synchronously, albeit inversely with mass balance, to regional climate variability. A proxy glacier mass balance (MB) record was constructed from a mountain hemlock ring-width chronology derived from climate-sensitive trees growing adjacent to each site. A good comparison between the proxy and regional mass balance records for Pacific Northwest glaciers (1966-1994: $R^2 = 50\%$) allowed for the reconstruction of MB anomalies from 1600 – 1994 AD. In addition, a significant relationship was revealed between the mass balance history of glaciers on Vancouver Island and large-scale climatic variability shown to be associated with the Pacific Decadal Oscillation.

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None of this work would have been possible without permission from B.C. Parks to undertake research in one of the most beautiful settings in Canada. Mass balance data was provided by Mike Demuth at the GSC and Roy Krimmel of the USGS. I'd also like to thank Brian Luckman, David Barclay, Greg Wiles and Kurt Nicolussi for comments, suggestions, and for responding to numerous paper requests.

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Dedication

This thesis is dedicated to my mother, and the memory of my father, in appreciation of their assistance and support.

CHAPTER 1. INTRODUCTION

Rapidly changing global climates have fostered renewed interest in understanding the nature of climatic change within the last millennium. In order to increase our understanding of present and future climate change we require a better knowledge of past climatic fluctuations; one that can be best achieved through a multi-proxy approach to paleoenvironmental reconstruction.

It is widely recognised that mountainous regions are highly sensitive to climate changes that alter the fundamental character of the ecological, geomorphological, and glacial processes in this environment (Slaymaker 1990; Haeberli 1998; Luckman 1998). Previous research has shown that climatic change indices can be developed through the study of variations in both the growth characteristics of high-elevation climatically sensitive trees, and the historical mass balance variations of alpine glaciers (LaMarche and Fritts 1971; Matthews 1977; Graumlich and Brubaker 1986; Villalba *et al.* 1990; Luckman *et al.* 1993; Smith *et al.* 1995).

Although long-term meteorological data are scarce for most alpine settings, particularly coastal British Columbia, effective proxy climate records can be reconstructed through dendroclimatological interpretations of annual tree-ring chronologies (Fritts 1976; Schweingruber 1996). Similarly, geomorphic evidence of historical periods of glacial advance and retreat has proven a useful indicator of local and regional climate change, as these glacier responses are a function of changes in glacier mass balance (McClung and Armstrong 1993). Comparison of glacier mass balance changes with the well-dated terminal positions of previous glacial extent can place restraints on the timing and magnitude of glacial response to specific climate change (Burbank 1982).

Climate conditions during the Holocene (*ca.* last 10 000 years) were originally thought to include two general phases, an earlier warm interval (Hypsithermal) followed by an extended cooling period (Little Ice Age). The term 'Little Ice Age' (LIA) was initially used to describe an extended cool period of renewed but moderate glaciation following the climatic optimum of the early Holocene (Matthes 1939). The timing of the LIA was originally thought to have encompassed the last *ca.* 3500 years (Matthes 1939), an interval now termed the Neoglacial (Denton and Karlén 1973; Luckman 2000). However, more recent studies of Holocene environments have revealed a greater degree of climate variability, particularly during the late Neoglacial (Denton and Karlén 1973; Haeberli and Beniston 1998; Ogilvie and Jónsson 2001). The latter stage of the Holocene, therefore, did not constitute a lengthy, uninterrupted cool period, but consisted of a far more complex climate of relatively long warm periods marked by several intervals of glacial expansion (Bradley and Jones 1993; Barlow 2001; Ogilvie and Jónsson 2001). Subsequently, the term 'Little Ice Age' has been primarily used to describe two main phases of Late Neoglacial glacier advances. The 'early' phase occurred between *ca.* 1100 AD to 1300 AD, followed by a relatively inactive glacial period until the later 'classic' LIA advances between *ca.* 1700 AD and the early 1900's AD (Luckman and Villalba 2001). During the latter phase of the LIA, glaciers in many parts of the world expanded and fluctuated around more advanced positions, reaching their maximum extent during the 18th and 19th centuries (Grove 1988, 2001). Consequently, many glacial forefields in western Canada and the Pacific Northwest contain a series of well-developed moraines representing pronounced LIA fluctuations (Luckman 1993; Smith *et al.* 1995).

Research has shown that episodes of glacial activity can be determined from lateral and terminal moraine sequences, where datable material has been preserved, or where tree and lichen colonisation rates can be established (Porter 1981a; Smith *et al.*

1995). However, a single site may not represent a complete LIA record, as extensive glacial advances in the latter stages of the LIA tended to override, rework, and erase evidence of pre-existing LIA glacial deposits. Such a condition presents difficulties for reconstructing glacial histories prior to the 1700's (Smith 1994; 1995; Pelfini 1999).

In conjunction with moraine chronologies, tree-ring series of temperature-sensitive trees often reveal synchronous declines in ring-width during, or prior to moraine building intervals (LaMarche and Fritts 1971; Matthews 1977; Luckman *et al.* 1993; Nicolussi and Patzelt 1996). In maritime locations in Pacific North America, a combination of cool (warm) summers and wet (dry) winters favours glacial expansion (recession) yet limits (enhances) radial growth in mountain hemlock, resulting in below (above) average ring-widths (Graumlich and Brubaker 1986; Smith and Laroque 1998b; Peterson and Peterson *in press*). The congruence between these two natural archives suggests similar but inverse responses to climatic inputs, thereby providing corresponding evidence for the reconstruction of LIA related climatic change (LaMarche and Fritts 1971; Luckman *et al.* 1993; Smith *et al.* 1995; Nicolussi and Patzelt 1996).

Researchers in western Canada, Alaska, and northwestern Washington State have used direct glacial and geomorphological evidence, in conjunction with proxy climate records, to document and reconstruct decade- to century-scale climate fluctuations over the last millennium (Porter and Denton 1967; Burbank 1981; Porter 1981b; Burbank 1982; Ryder and Thomson 1986; Ryder 1987; Desloges and Ryder 1990; Wiles and Calkin 1990; Clague and Mathewes 1996; Smith and Laroque 1996; Wiles *et al.* 1999b; Smith and Desloges 2000). Absolute and relative age dating methods, such as dendroglaciology and lichenometry, have been used in conjunction with dendroclimatology and glacial mass balance studies to provide independent corroboration of the magnitude, timing, and direction of climate fluctuations associated with the LIA

(LaMarche and Fritts 1971; Matthews 1977; Porter 1981a; Burbank 1982; Villalba *et al.* 1990; Bhattacharyya and Yadav 1996; Nicolussi and Patzelt 1996).

The primary objectives of this thesis are to:

(1) Establish the glacial history of two glaciers in Strathcona Provincial Park, Vancouver Island, British Columbia,

(2) Place LIA glacier activity on Vancouver Island in the context of the larger LIA record from the North American Cordillera, and

(3) Evaluate this behaviour in the context of proxy climatic and mass balance records, and determine the feasibility of using mountain hemlock (*Tsuga mertensiana* [Bong.] Carr.) ring-width chronologies as proxy mass balance indicators.

This thesis consists of 8 chapters and 2 appendices. Following the Introduction, Chapter 2 provides an overview of LIA glacier studies in Pacific North America. Chapter 3 describes the general study area of Strathcona Provincial Park and the individual study sites; Chapter 4 describes the field and laboratory methodologies. Chapters 5 and 6 present preliminary dating controls used in dating moraines, and results of the fieldwork undertaken at Colonel Foster and Septimus glaciers, respectively. Chapter 7 provides an overview of glacier-climate relationships and presents the results of reconstructed mass balance anomalies from annual growth rings of mountain hemlock. Finally, Chapter 8 presents a summary and conclusion of the research.

CHAPTER 2. LITTLE ICE AGE IN PACIFIC NORTH AMERICA

2.1 Glaciers and Climate Change

The global climate signal of the last 100 years as recorded by alpine glaciers is manifested as an extensive retreat of terminal positions from maximum LIA extents (Oerlemans 1994b). This marked response is linked to atmospheric conditions through internal glacial filters, such as pronounced memory and climate signal enhancement, thus providing one of the more unmistakable signals of global climate trends in the last century (Haeberli 1998).

Alpine glaciers are dynamic, thermal bodies that depend on interactions between regional climate, local accumulation and ablation, and ice deformation for their existence (Porter 1981a; Paterson 1995). As they are a product of regional climate, and their persistence is a function of specific combinations of climatic variables, glaciers have historically been used as proxy indicators of past climate conditions (Meier 1965; Porter 1981a; Haeberli and Beniston 1998; Oerlemans 1998).

2.2 Glacial Response to Climate Change

Glacier mass balance records are considered to be an undelayed and unfiltered climate signal (Haeberli 1998; Reichert *et al.* 2001). Adequate mass balance records, however, are limited to less than 40 glaciers worldwide (Dyurgerov and Meier 1997). The brevity of these records, along with the fact that mass balance studies are expensive, time consuming and labour intensive, limits their use in studying glacier-climate interactions prior to the mid 1900s. As a result, other glacier parameters such as changes in volume, area, and length, have been used in place of mass balance measurements to determine climate change (Oerlemans 1994b).

Although volumetric measurements provide the most accurate information regarding glacial change, they require an in-depth knowledge of ice thickness and underlying topography, and are thereby limited to the few glaciers that have been intensively monitored (Oerlemans 1994b, 1998). Variations in glacier area are, therefore, often substituted for volume-based measurements. Changes in glacier length are the most common area-based measurements and often provide the longest records (Oerlemans 1994b). Glacier length is one of the easiest glacier parameters to measure, as it can be derived through inexpensive and less labour intensive methods such as photogrammetry and historical records. In addition, this method can be applied to remote glaciers and those that have not previously been monitored, therefore, adding much-needed climate-glacier data to the global glacier database (Haeberli 1998; Oerlemans 1998).

The influence of climate change is particularly obvious at the glacier snout, where the advance and retreat of the ice front and corresponding moraine formation represents significant mass balance variations (Letréguilly and Reynaud 1989; McClung and Armstrong 1993). Although fluctuations in snout position have been linked to interannual climate variability, the most significant changes are the result of longer-term (decade to century) climate events (Oerlemans 1986, 1994b). In contrast to mass balance, the position of the glacier front is considered a filtered and delayed, although enhanced, response to changing climate (Haeberli 1998; Reichert *et al.* 2001). Although glacial chronologies derived from areal fluctuations can provide information on the time and relative extent of glacial advances, they do not provide data that can be directly translated into quantitative records of temperature and precipitation (Porter 1981a; Luckman 2000). Glacier fluctuations integrate the response of several climatic inputs over a period of time and, therefore, are ideally suited for inferring general climate conditions *(i.e.,* decadal to interdecadal trends) such as cool/wet or warm/dry conditions (Porter 1981a; Luckman 2000). As moraines mark the maximum extent of glacial advance, they reflect the shift from an interval of climate conditions related to a positive mass balance (+MB) regime (*i.e.,* cool/wet) to conditions favouring a negative mass balance (–MB) regime (*i.e.,* warm/dry) (Porter 1981a). Therefore, if the age of moraines can be accurately determined, the dates can be used to infer the chronology of climate transitions from cooler, wetter climates associated with possible +MB, to warmer, drier climate conditions associated with a –MB regime.

Unfortunately, reconstructions derived from glacial deposits provide evidence of only select, large magnitude climatic events, resulting in a discontinuous low-resolution record (Luckman and Villalba 2001). This problem is further complicated by additional discontinuities in the glacial record that arise from glacial activity associated with the latter stages of the LIA. These advances were often the most extensive, and in some cases have obliterated or reworked deposits associated with earlier advances (Pelfini 1999). For example, through the analysis of tree-ring chronologies developed from trees growing adjacent to the glacial forefield, both Matthews (1977) and Karlén (1984) were able to reconstruct 'missing' moraines that had been reworked by subsequent advances and were not preserved in the present moraine chronosequence.

Despite such limitations, reconstructions based on glacial deposits remain a widely accepted method of inferring paleoclimatic/paleoenvironmental conditions in alpine environments (Porter 1981a; Smith *et al.* 1995; Wiles *et al.* 1996a; Luckman

2000). Although they do not allow for the reconstruction of a detailed climate history, well-dated glacier chronologies can be used to verify and calibrate reconstructions based on other proxy indicators. Glacial chronosequences, if carefully and accurately dated, can also provide information about the direction and magnitude of climate change (Porter 1981a; Luckman 2000).

Temperate alpine glaciers are highly sensitive indicators of paleoclimatic changes (Yarnal 1984; Brugman 1992; Haeberli and Beniston 1998), and small cirque glaciers in particular have been shown to respond quickly to historical climate change (Porter 1981a; Burbank 1982; Luckman 1986; Smith *et al.* 1995; Haeberli and Beniston 1998; Oerlemans 1998). Such relatively quick response times can be attributed to the following characteristics: (*i*) the low basal shear stress of small cirque glaciers allows for a rapid response to both high and low frequency climate forcing; (*ii*) a warm thermal regime and close proximity to melting conditions requires only a small change in climate parameters to induce a marked change in glacier characteristics; and (*iii*) as a result of their small size, climate induced mass balance perturbations can travel through the glacier relatively quickly resulting in more noticeable changes in volume and area (Oerlemans 1994a; Paterson 1995; Haeberli and Beniston 1998; Oerlemans 1998). Unlike large valley glaciers and icefields that respond to climatic change in a delayed and 'smoothed' manner, smaller cirque glaciers, which respond almost immediately to present climate and mass balance changes, can provide a reasonable high-resolution climate signal (Haeberli and Beniston 1998).

2.3 LIA Glacial Activity in the northwestern Cordillera

Glacial activity during the LIA left a distinct footprint in the mountains of western North America. Moraines associated with the LIA are found down valley from many contemporary glaciers and, if accurately dated, can provide evidence of the timing and extent of the LIA. Lichenometry, radiocarbon dating and the dendrochronological crossdating of glacially sheared stumps and detrital logs have been the primary sources for dating early LIA (*ca.* 12^{th} - 13^{th} century) events, whereas dendrochronological and lichenometric methods are the most applicable for dating LIA deposits of the last 500-600 years.

A composite record of dated glacier advances and moraines from 320 glaciers throughout the coastal mountains of Alaska, British Columbia, and Washington State (Figure 2.1), is compared to data compiled by Luckman (2000) for the interior glaciers of the Canadian Rocky Mountains.

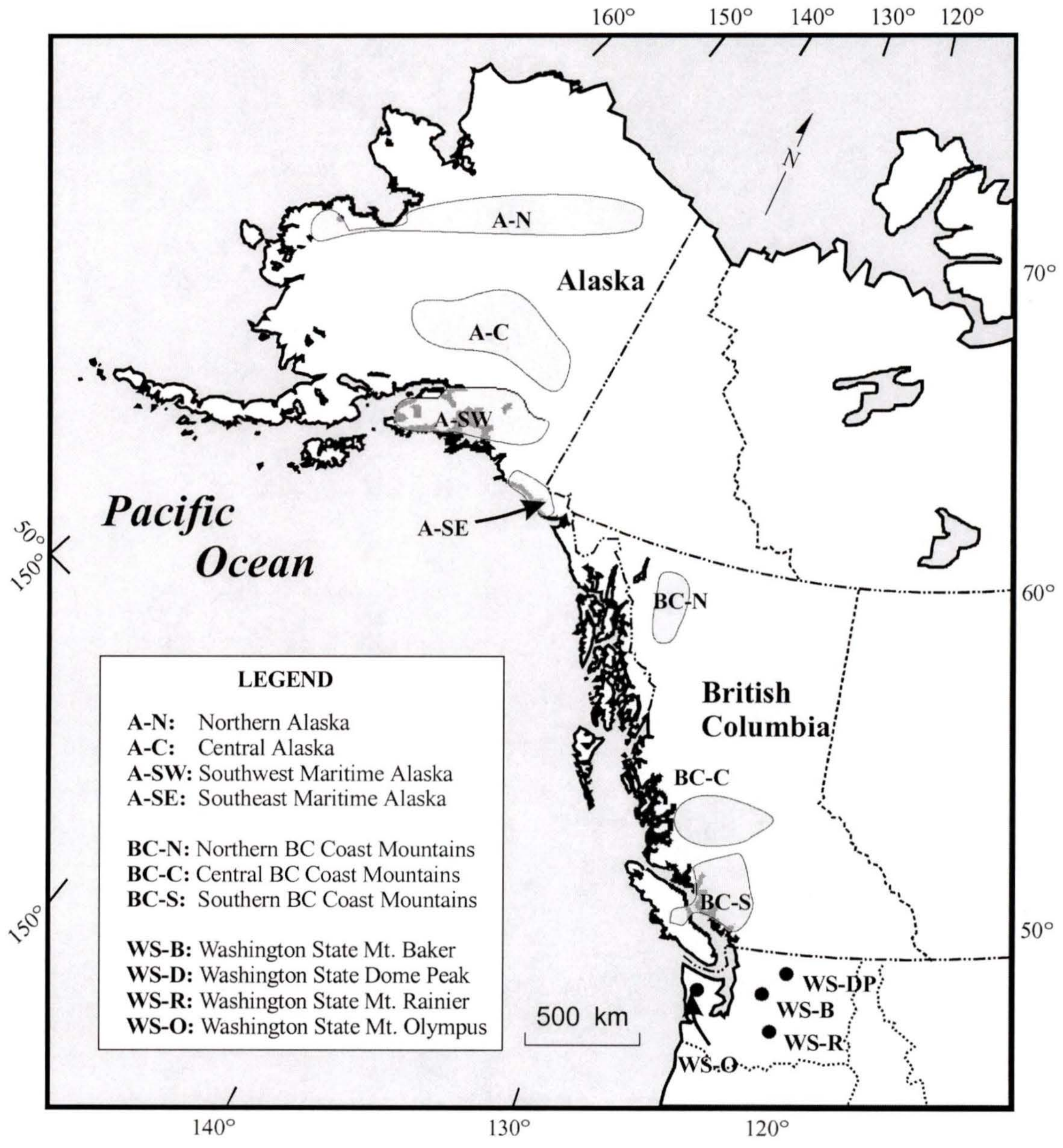


Figure 2.1. Map of the Pacific North America highlighting general areas with LIA moraine data.

2.3.1 Alaska

Tree-ring, lichen, and radiocarbon data from 167 glaciers (375 moraines) throughout Alaska provide one of the most comprehensive data sets for reconstructing the chronology of LIA glacier events in the northwestern Cordillera. Alaskan LIA glacier studies are grouped into the following four general regions: Northern Alaska, Central Alaska, Southwest Maritime Alaska, and Southeast Maritime Alaska (Figure 2.1).

The LIA moraine chronology from northern Alaska is based on moraine dates from 100 glaciers, primarily in the Brooks Range. Lichen dates provide evidence for seven episodes of LIA moraine stabilisation within the LIA time frame (Haworth 1988, in Calkin and Wiles 1992; Calkin *et al.* 1998). The moraine histogram in Figure 2.2a suggests three major, or extensive, moraine building events centred around the early 16th, 17th, and 19th centuries, and four minor, or less extensive episodes, centred around the late 12th, mid 13th, late 14th, and early to mid 18th centuries.

In Central Alaska, radiocarbon dates provide evidence of advancing glaciers as early as the 13th century (Denton and Stuiver 1967; Denton and Karlén 1977; Calkin and Wiles 1992). Lichenometric dates from the central Alaska Ranges, Wrangell, and Northeast St. Elias Mountains indicate maximum extent in this area was reached in the mid 16th century (Figure 2.2b). Subsequent moraines were deposited in the mid and late 17th, early 19th centuries, and most extensively in the late 19th and early 20th centuries (Reger and Péwé unpublished, in Viereck 1967; Denton and Karlén 1977; Calkin 1988; Calkin and Wiles 1992). There is a marked disparity in timing of LIA events in the southwestern and southeastern Maritime regions (Figures 2.2c and 2.2d).

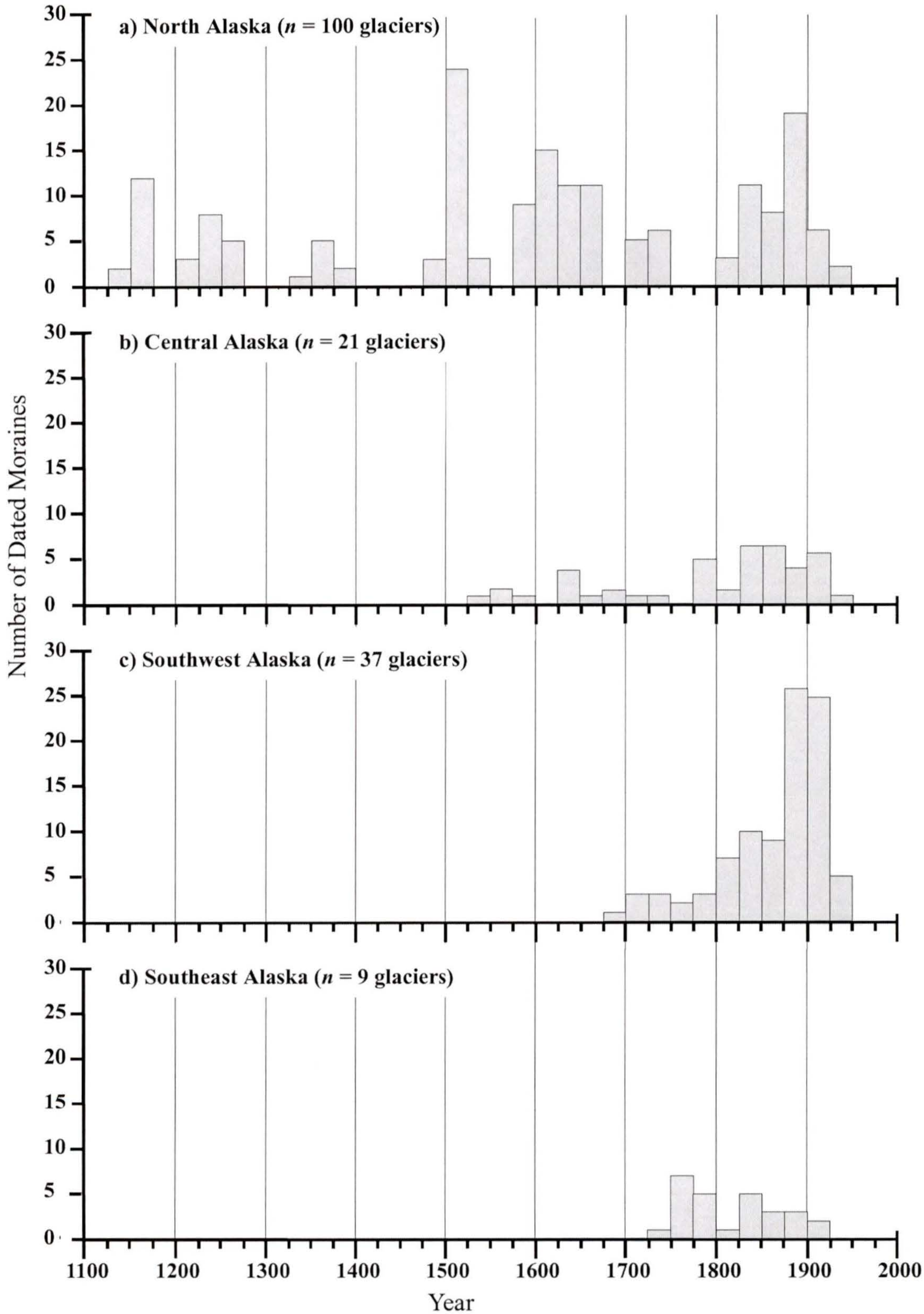


Figure 2.2. Moraine frequency histograms for the four Alaska regions outlined in the text: (a) Northern, (b) Central, (c) southwest Maritime, (d) southeast Maritime.

Although overrun forests in both the Southwest and Southeast Maritime regions indicate a generally synchronous early LIA advance centred on the mid 13th century, the timing of moraine deposition during the mid and late LIA was asynchronous. The main period of LIA moraine formation in the Southwest Maritime region occurred in the mid to late 19th century (Viereck 1967; Wiles and Calkin 1990; Wiles *et al.* 1999a). Within the Southwest Maritime region, however, there was greater temporal variability in the timing of LIA glacier activity. Tree-ring dates reveal a difference in the timing of glacier retreat on the western and eastern sides of the Kenai Peninsula (Wiles and Calkin 1994). Although the onset of the LIA in this region appears to have been synchronous, glaciers began retreating from their LIA maximum on the western side of the mountains by the early 18th century, whereas glaciers on the eastern slopes began retreating almost 200 years later, in the late 19th to early 20th century (Wiles and Calkin 1994) (Figure 2.2c and 2.2d).

Although moraines from the Southwestern Maritime region are predominantly mid to late 19th century in origin, moraines from the Southeastern Maritime region were deposited almost a century earlier (Figure 2.2d). A widely synchronous advance, which is also the LIA maximum extent, was dated to the mid to late 18th century for glaciers descending from the Juneau Icefield. A subsequent readvance culminated with the deposition of mid 19th century moraines (Lawrence 1950b, 1958; Egan 1971, in Calkin and Wiles 1992).

2.3.2 British Columbia Coast Range

LIA glacier research in the British Columbia (B.C.) Coast Mountains is divided into three general regions: North Coast, Central Coast, and South Coast (Figure 2.1).

In the Stikine-Iskut area of the northern B.C. Coast Ranges, dendrochronology and ^{14}C dating provide three general episodes of glacial activity (Ryder 1987). Radiocarbon dates from overridden trees and till samples indicate that glaciers were expanding in this region by the 14th century (Ryder 1987). The maximum LIA extent was dated to the late 17th to early 18th centuries, with subsequent readvances dated to the late 19th century (Figure 2.3a).

In the central B.C. Coast Mountains, radiocarbon dates indicate glacier expansion was underway by the 13th and 15th centuries, with little activity recorded again until early 18th century advances (Desloges 1987; Desloges and Ryder 1990; Smith and Desloges 2000). Lichen and tree data indicate that most terminal moraines from this region were deposited between 1860 and 1900 AD (Desloges and Ryder 1990); however, recent findings by Smith and Desloges (2000) indicate the maximum LIA down-valley extent in this region was reached 200 years earlier, in the early 1700's (Figure 2.3b).

There are four general periods of glacier activity in the southern B.C. Coast Mountains centred on the 13th, 15th, 18th, and late 19th centuries (Mathews 1951; Ryder and Thomson 1986; Ryder unpublished, in Desloges and Ryder 1990). Radiocarbon dates suggest early LIA advances were underway by the 12th century (Ryder and Thomson 1986), and a widely synchronous 18th century advance was recorded at glaciers in Garibaldi Provincial Park (Mathews 1951; Ricker and Tupper 1979). Terminal moraine dates collected from over 30 glaciers in the southern Coast Ranges suggests the maximum LIA extent was reached by the mid 18th century, with a subsequent advance dated to the mid to late 19th century (Munday 1936, in Mathews 1951; Ryder unpublished, in Desloges and Ryder 1990).

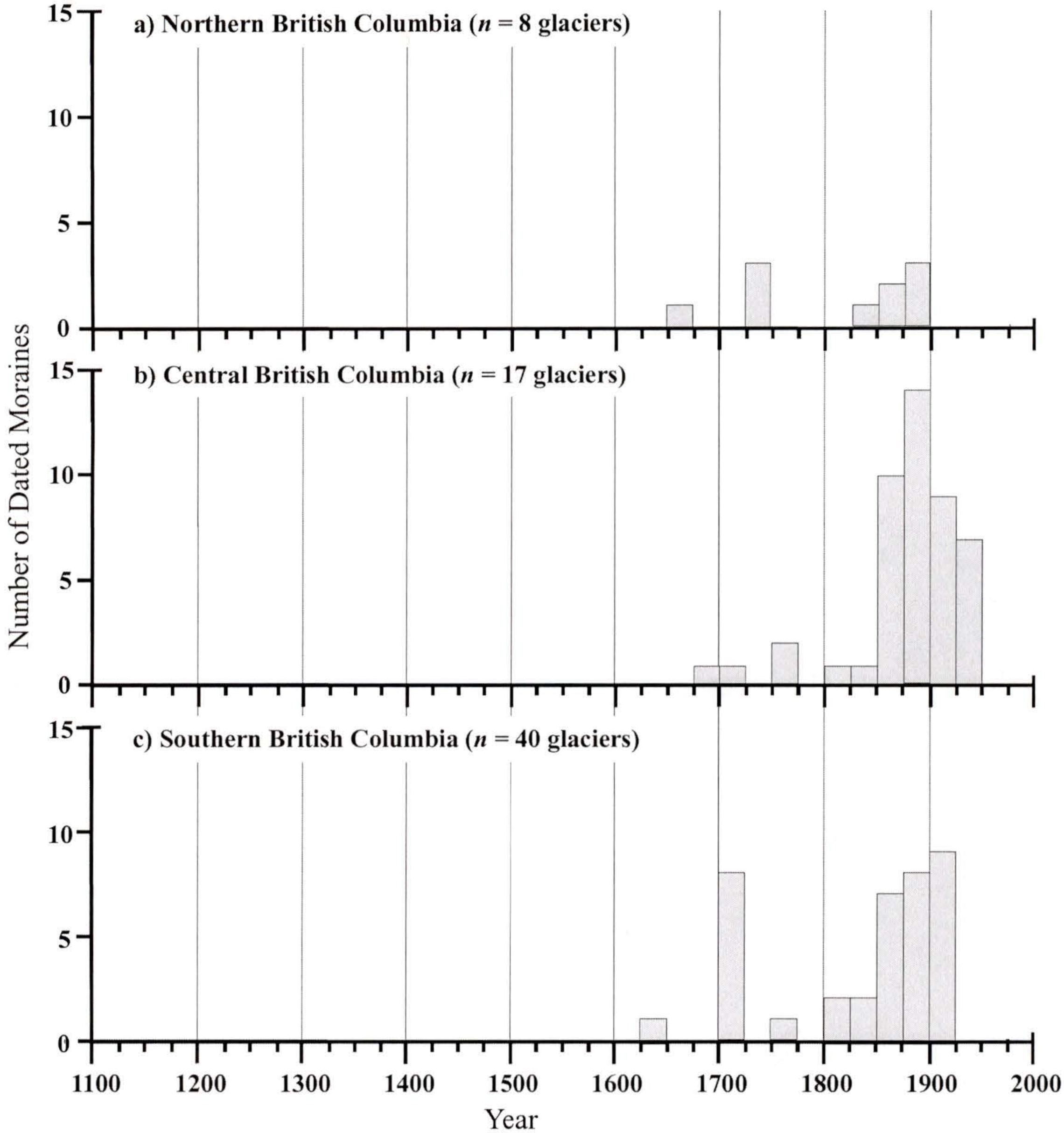


Figure 2.3. Moraine frequency histograms for British Columbia Coast Mountain regions outlined in the text: (a) North Coast, (b) Central Coast, and (c) South Coast.

Until this study, the only dated LIA advance from the Insular Mountains of Vancouver Island was from Smith and Laroque (1996) at Moving Glacier in Strathcona Provincial Park. Glacially sheared stumps and detrital logs suggest the glacier was advancing in the early 1700's AD, and reached its maximum LIA extent sometime after 1818 AD.

Evidence of pre-18th century moraines in the Coast Mountains is rare. Although the 19th century moraines are more extensive and often larger, at sites where both 18th and 19th century moraines were found, the 18th century moraines marked the LIA maximum downvalley extent.

Culmination of the 18th century advance occurred earlier in the southern Coast Mountains (early 1700's AD), whereas 18th century moraines in the central and north Coast Mountains were deposited *ca.* 2-3 decades later. A more extensive 'second-phase' or late LIA episode of moraine deposition, however, was generally synchronous in all three regions of the B.C. Coast Mountains (Figure 2.3).

2.3.3 Washington State

The moraine record from Washington State is comprised primarily of tree-ring data collected from four areas: Mount Baker, Dome Peak and Mount Rainier in the North Cascade Mountains, and Mount Olympus on the Olympic Peninsula (Figure 2.1).

Tree-ring data from forested moraines on Mount Baker suggests five general periods of moraine deposition (Figure 2.4a). The maximum LIA position is dated to the early 16th century, with subsequent readvances occurring in the mid 17th, mid 18th, early 19th, and early 20th centuries (Long 1953, 1955; Easterbrook and Burke 1972; Fuller *et al.* 1983; Fuller 1980, in Heikkinen 1984).

In the Dome Peak area of the northern Cascades, two main intervals of moraine deposition (1600's and late 1800 to early 1900's) are interspersed with a scattering of moraine dates (Figure 2.4b)(Miller 1969). Tree-ring dates indicate that at least one glacier reached its LIA maximum extent in the 13th century, whereas most other glaciers reached their LIA maxima in the 16th century (Figure 2.4b). Late LIA activity, however, was widely synchronous, as all glaciers studied had subsequent advances culminate in the mid to late 19th century, and early 20th century (Miller 1969; Beget 1984). Research in the nearby Entiat Mountains indicates that glaciers there had also reached their LIA maximum in the 16th century, followed by episodes of moraine deposition in the mid 18th and late 19th centuries (Long 1953).

Tree-ring dated moraines from Mount Rainier provide evidence of a varied early LIA moraine chronology with two generally synchronous late LIA events (Figure 2.4c) (Sigafoos and Hendricks 1961; Crandell and Miller 1964; 1969; 1972; 1974).

The two synchronous late LIA advances occur in the early 18th and mid 19th centuries, and are recognised at all but one glacier (Figure 2.4). The maximum LIA downvalley extent, however, was much more variable, as the timing of the maximum extent was different for seven of the twelve glaciers studied. Maximum LIA moraines were dated to the 13th, 14th, 16th, 17th, 18th and 19th centuries (Figure 2.4c).

The moraine chronology from Mt. Olympus is limited to only four glaciers, but shows two generally synchronous advances dated to the early 19th and early 20th centuries (Figure 2.4d). A 13th century moraine marks the maximum downvalley extent at one glacier, whereas the others appear to have reached their LIA maxima during the 18th century (Heusser 1957; Spicer 1989).

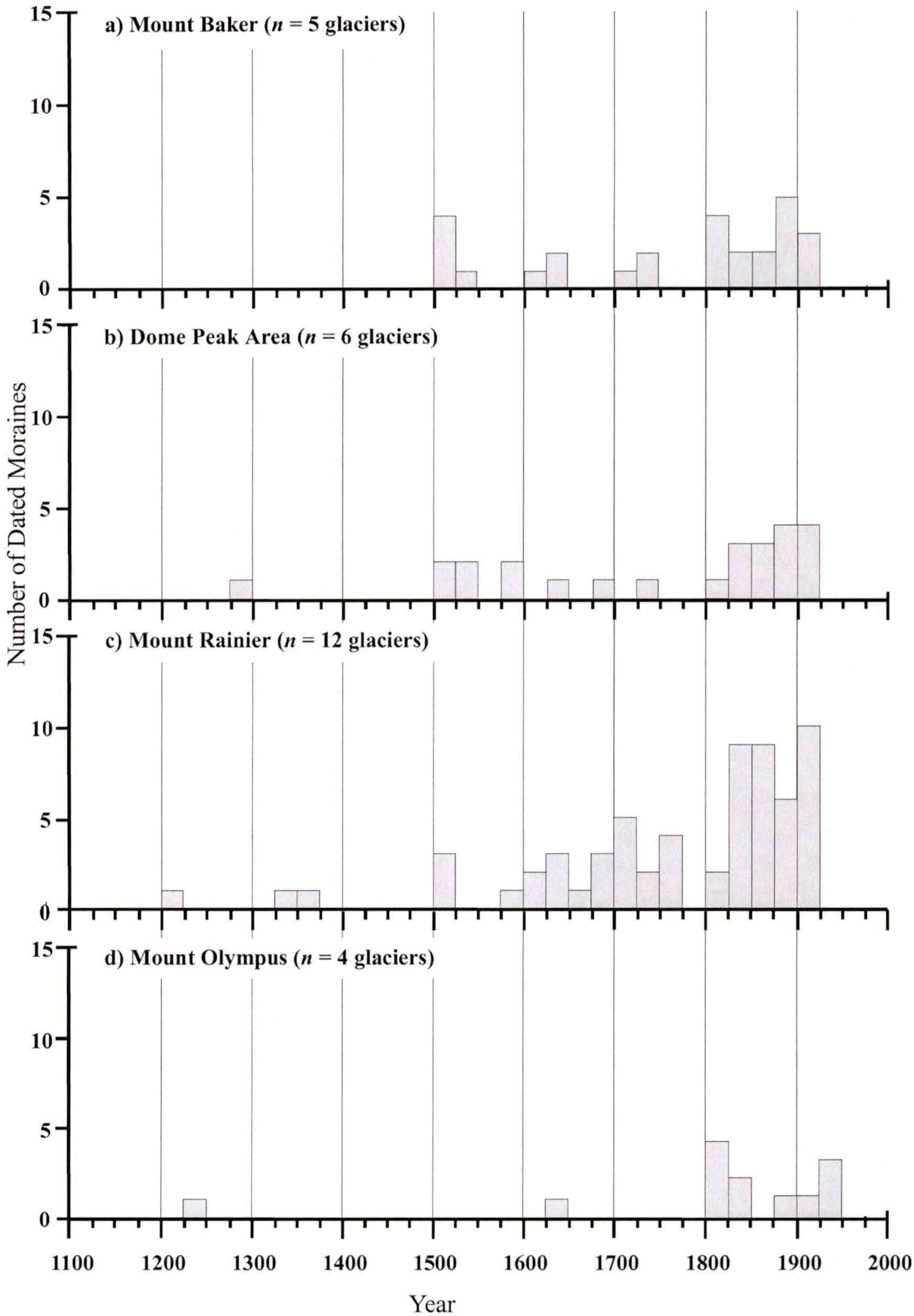


Figure 2.4. Moraine frequency histograms for Washington State regions outlined in the text: (a) Mt. Baker, (b) Dome Peak, (c) Mt. Rainier, (d) Mt. Olympus.

2.3.4 Summary of Maritime LIA Record

Moraine records from 260 glaciers in the Pacific Northwest provide sufficient data to reconstruct a regional history of maritime glacier activity from Alaska to Washington State for the last 800 years. The Alaskan record itself provides the longest and most complete LIA record within this region, providing moraine dates back to the 12th century AD (Figure 2.5a).

Moraine histograms in Figure 2.5 show the maritime moraine records from Alaska, British Columbia, and Washington State are generally synchronous at longer timescales (50-100 yrs), but lack synchroneity at shorter time scales (10-25 yrs). All three regions show a widely synchronous pattern, with peak moraine formation dated to the late 1800's to early 1900's. Alaska and Washington State also show a second 19th century peak between 1825 and 1850 (Figure 2.5). A secondary peak in the early 18th century is also recorded in all three regions. Moraine data prior to the 1700's is rare in British Columbia, yet is evident in the regions to the north and south. Although the early LIA moraine record from Washington State is limited, there is a degree of synchroneity between these moraines and the early LIA moraines from the much deeper Alaskan record.

A compilation of the Maritime moraine record (Figure 2.6b) suggests seven general episodes of glacial activity in the Northwestern Cordillera during the last millenium. Broadly synchronous early LIA events are dated to the 12th, 13th and 14th centuries; with mid LIA events in the early 16th and mid 17th centuries followed by two extensive "classic" late LIA advances, which culminated in early 18th and late 19th centuries (Figure 2.6b).

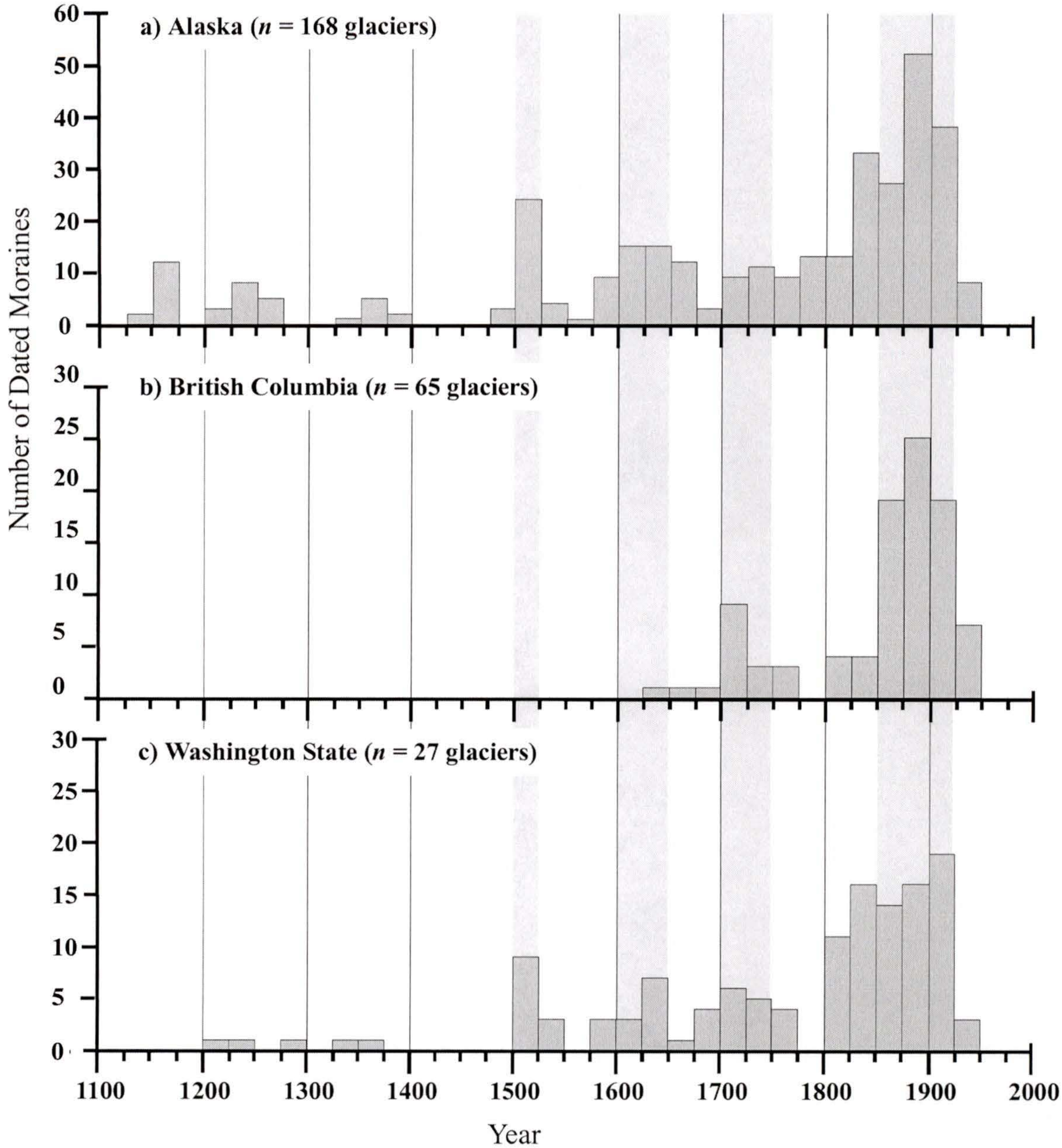


Figure 2.5. Moraine frequency histograms for maritime glaciers of the Pacific Northwest. (a) Alaska record, (b) coastal British Columbia record, and (c) Washington State record. Vertical grey bars illustrate episodes of general synchroneity between regions. Note the scale change for Alaska histograms.

In comparing the Maritime moraine record with the record from the Canadian Rocky Mountains, four broadly synchronous episodes of glacier activity emerge (Figure 2.6). The congruence between the maritime and interior glacier records indicates that, although there may be a lack of synchroneity at shorter time scales, LIA glacier activity in western North America was synchronous at longer time scales.

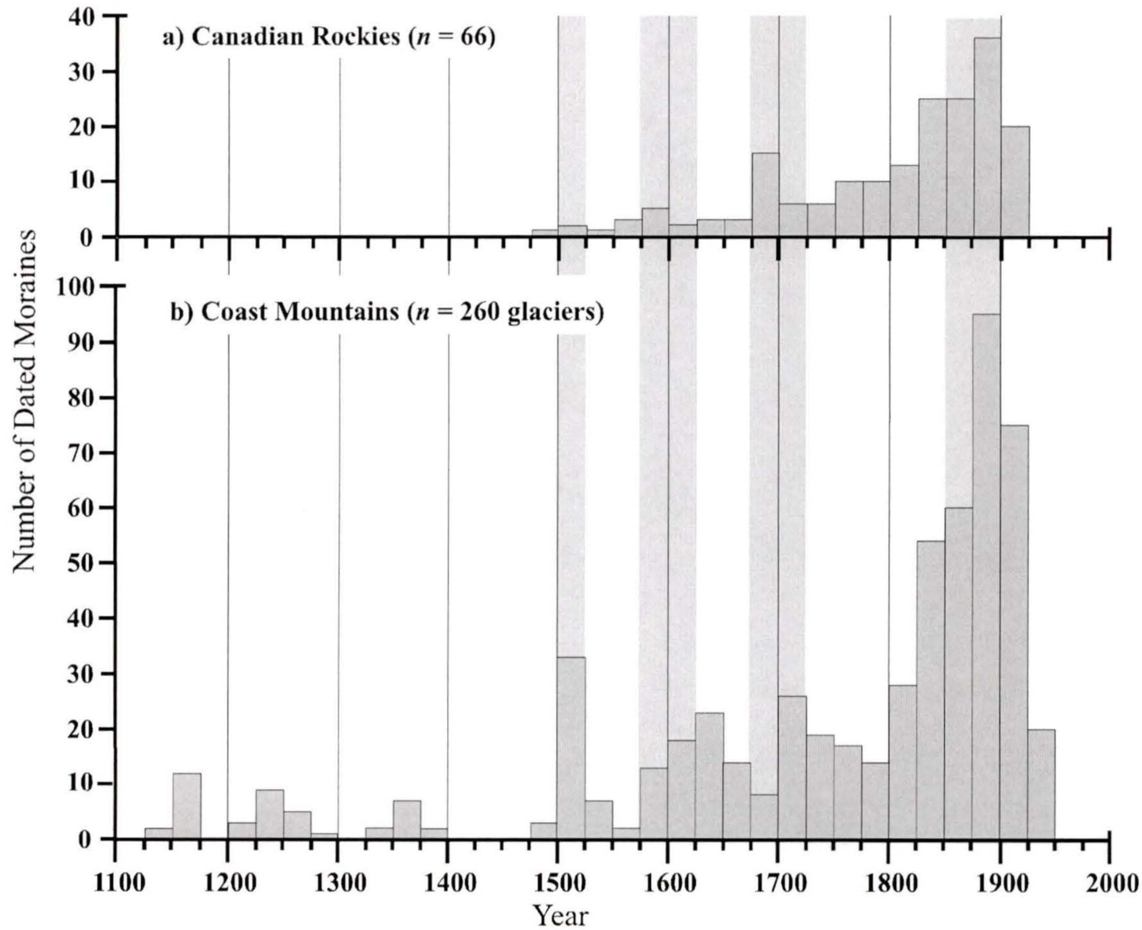


Figure 2.6. Moraine frequency histograms for maritime glaciers of the Pacific Northwest and the interior glaciers of the central and southern Canadian Rockies: (a) Canadian Rockies, and (b) Coastal Mountains. Vertical grey bars illustrate episodes of general synchroneity between regions.

CHAPTER 3. RESEARCH SETTING

3.1 Introduction and location

Field research was conducted in Strathcona Provincial Park (hereafter Strathcona Park) on central Vancouver Island (49° 40' N, 125° 40' W). The park extends over a predominantly mountainous area of approximately 2090 km^2. Entrance to the park is 48 km west of the Campbell River townsite along Highway 28 (Figure 3.1). Strathcona Park straddles the Vancouver Island Ranges and contains the tallest mountains on Vancouver Island. This includes peaks such as the Golden Hinde (2228 m), Elkhorn Mountain (2195 m), Mount Colonel Foster (2134 m), and Mount Septimus (1962 m) (Hnytka 1990).

Glaciers are not uncommon at higher elevations in the park, with more than 200 alpine glaciers have been documented among the peaks of Vancouver Island (Ommaney 1972). Although Strathcona Park contains the largest glaciers on Vancouver Island, most are relatively small and have experienced significant retreat within the last century.

3.1.1 Physiography

The mountains within Strathcona Park belong to a much larger, discontinuous mountain belt that reaches from the St. Elias Mountains in the extreme northwest of the province, southwards to the North Cascades near the Canadian – U.S. border. Included in this belt are the Insular Mountains of the Queen Charlotte Islands and Vancouver Island, and the western-most reaches of the Coast Mountains of British Columbia (Holland 1964).

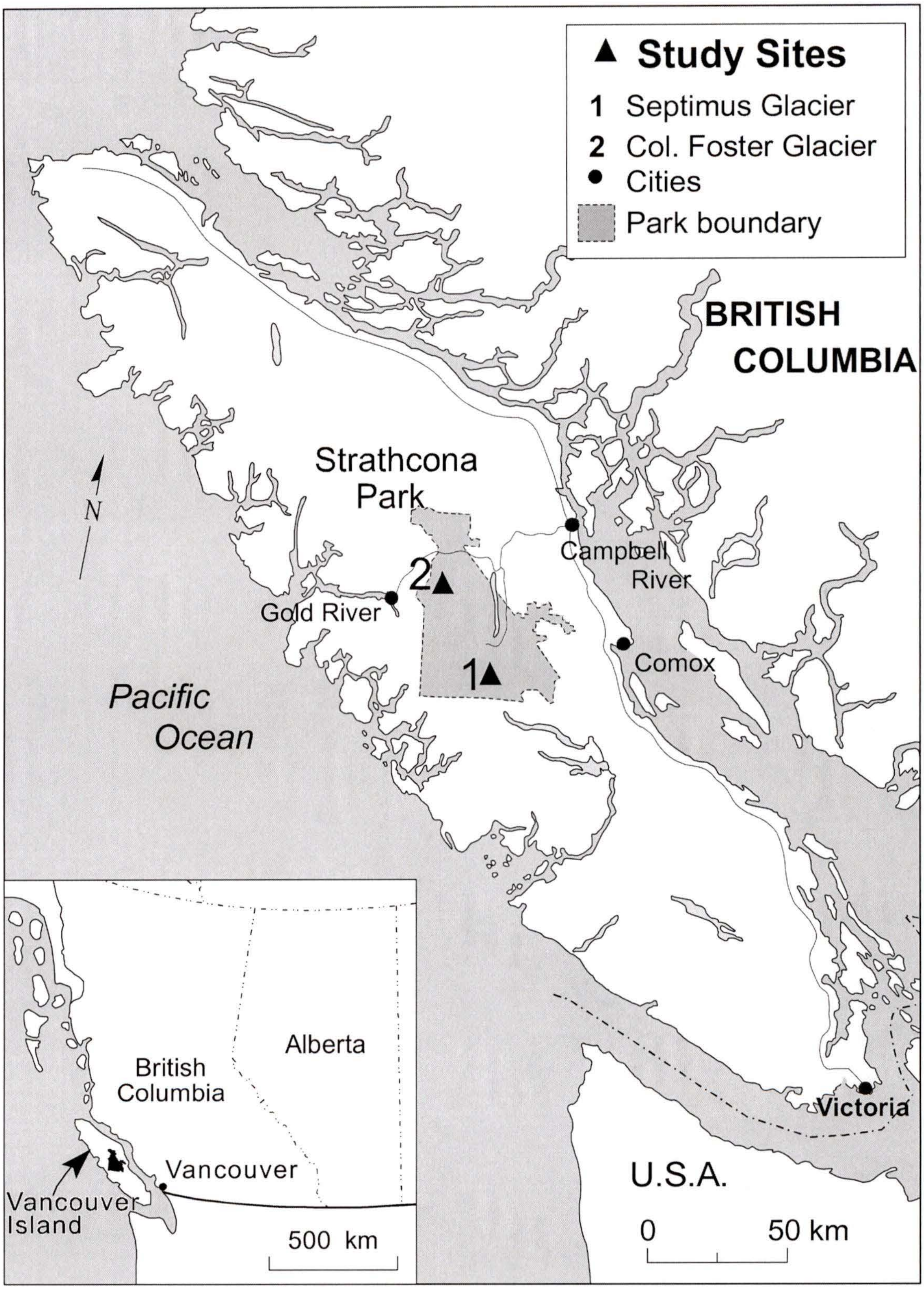
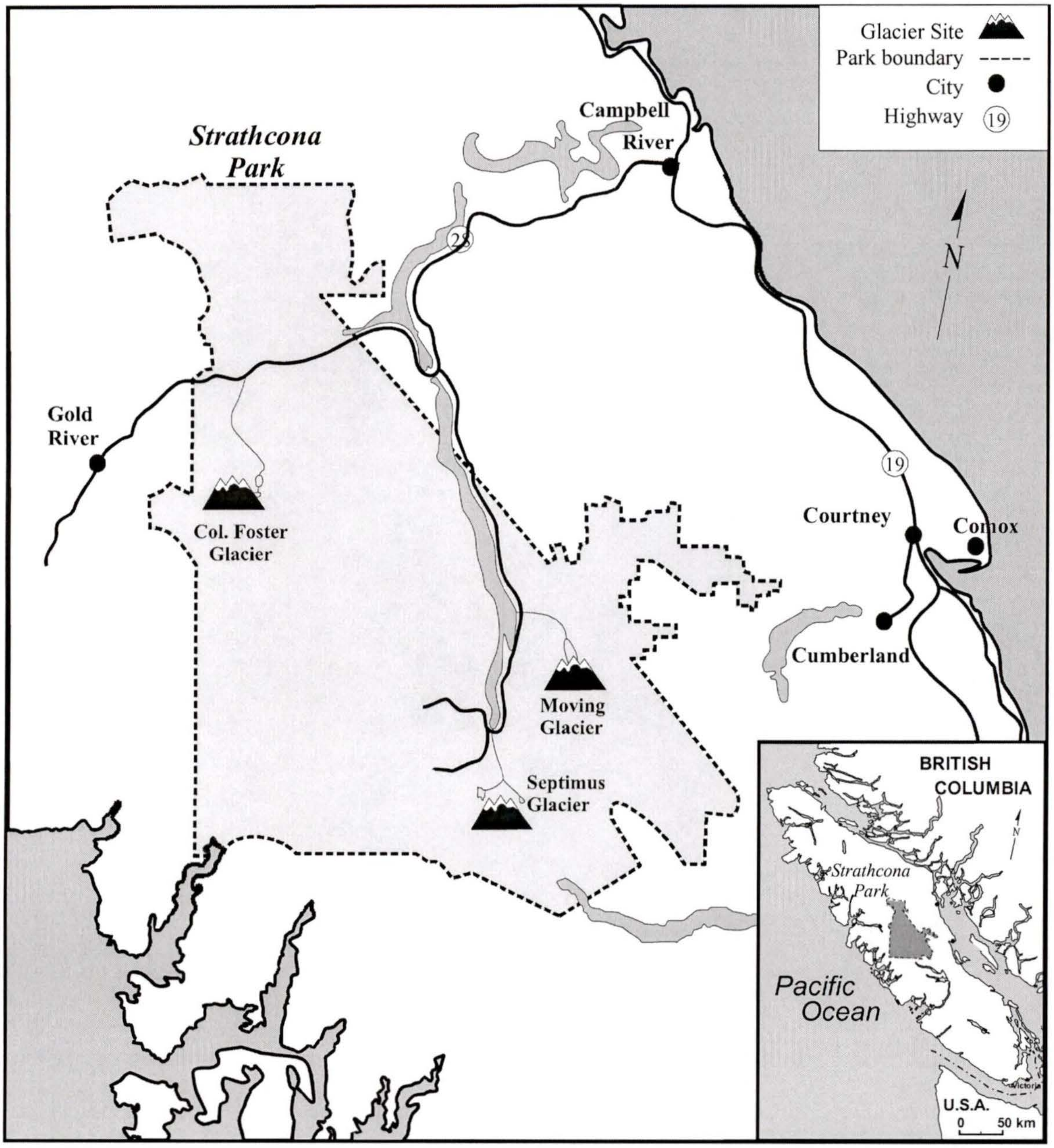


Figure 3.1. Location map of study area and study sites on Vancouver Island, British Columbia

The predominant mountain-building activity in Strathcona Park was a result of extensive volcanic activity during the early Devonian, the Middle Triassic, and the Lower Jurassic Periods, followed by considerable erosion and dissection during the Cretaceous (Surdam 1968; Grant and Logan 1995). Pre-Pleistocene uplift and dissection produced the rugged topography of the central and northern parts of the Island that exists today (Surdam 1968).

3.1.2 Glacial History

Strathcona Park experienced at least three major glacial events during the Pleistocene. Glaciation in the park was topographically controlled and followed the classical model of alpine glaciation (see Sugden and John 1982; Hicock 1986). Cirque glaciers spilled over their sills and into valleys to form tributary lobes that advanced downvalley to coalesce and form main trunk glaciers such as the Buttle lobe (Hicock 1986). Eventually the accumulating valley ice built up and overtopped the inter-valley ridges. During the maximum extent of the Fraser glaciation (*ca.* 15 000 BP), Coast Mountain ice coalesced and overrode the entire area, covering all but the highest peaks (Hicock 1986). The elevation of the ice surface during this time was at least 2000 m asl, as evidenced by both the glacial striae preserved on mountainsides and drumlinoid features found on glacially rounded ridges (Hicock 1986; Grant and Logan 1995). Deglaciation occurred primarily through differential downwasting in response to an ameliorating climate that commenced prior to *ca.* 13 000 BP. By 9500 BP, most of the land was free from ice (Howes, 1981).



igure 3.2. Map of Strathcona Park showing locations of Colonel Foster and Septimus Glacier from this study, and Moving Glacier from Smith and Laroque (1996).

Little is known about the Holocene glacial history of the park, with limited data for Moving Glacier provided by Smith and Laroque (1996). Their study shows that Moving Glacier has retreated *ca.* 950 m since its LIA maximum extent, and has lost over

95 percent of its surface area. Aerial photographic analysis indicates that other glaciers in the park have also retreated or downwasted substantially over the last 65 years.

3.1.3 Vegetation

The vegetation of Strathcona Park is predominantly evergreen coniferous forest. Distinct vegetation zones within the park reflect the biogeoclimatic influences of topography and a climate dominated by the Pacific Ocean (Kojima and Krajina 1975). Most of the vegetation at lower elevations in the park (< 900 m) falls within the Coastal Western Hemlock (CWH) Zone. This zone is characterized by dense, continuous forest cover dominated by western hemlock (*Tsuga heterophylla* [Raf.] Sarg.). Douglas-fir (*Pseudotsuga menziesii* [Mirb.] Franco) and western red-cedar (*Thuja plicata*) are found throughout the zone, whereas yellow-cedar (*Chamaecyparis nootkatensis* [D. Donn.] Spach.) and amabilis fir (*Abies amabilis* [Dougl.] Forbes) are found in the wetter CWH subzones (Hnytka 1990).

The vegetation above 900 m asl is almost entirely within the Mountain Hemlock (MH) Zone, dominated by the climax species mountain hemlock (Kojima and Krajina 1975). The MH Zone in Strathcona Park occurs between 900 m and 1500 m asl on the windward side of the mountains, and between 1100 m and 1800 m on the leeward side (Brett and Klinka 1997; Egan 1997).

At its lower limits, the MH Zone is characterized by continuous forest cover of the Mountain Hemlock Maritime Forested subzone; mixed stands of old growth subalpine forest composed largely of mountain hemlock, yellow-cedar and amabilis fir. The Maritime Forested MH subzone is further subdivided into an East and West Vancouver Island Variant based on local climatic variables. The West Vancouver Island

Variant is slightly wetter and milder, whereas the East Vancouver Island Variant has a drier, slightly subcontinental climate (Hnytka 1990). The MH Forested subzones are characterized by abundant tree regeneration and dense shrub growth under the canopy. White-flowered rhododendron (*Rhododendron albiflorum*), false azalea (*Menziesia ferruginea*) and *Vaccinium sp.* are the most common shrubs associated with these subzones (Egan 1997).

At its upper limits, the MH Zone is characterized by the discontinuous forest cover, tree islands and supalpine meadows of the Mountain Hemlock Parkland subzone. Undergrowth in the MH Parkland subzone is typically sparse due to poor soil development, increased slope angles and exposed bedrock outcrops, and is often limited to dwarf evergreen shrubs, heathers, clubmosses, and lichen (Egan 1997).

3.1.4 Climate

The climate of the MH Zone in Strathcona Park ranges from hyper-maritime on the west side of the mountains, to a drier sub-maritime climate on the eastern side. The MH Zone is characterized by short, cool summers and long, cool and wet winters (Egan 1997). The growing season is often short, with only 1.7 months of the year having a mean temperature greater than 10.0°C. Mean annual temperature for the MH Zone is 3.0°C, with the coldest month averaging –5.1°C and the warmest month 11.1°C (Klinka *et al.* 1991).

The majority of precipitation falls between the beginning of October and the end of May, with an average monthly total in the wettest months of 414 mm, and 62 mm for the driest summer months. Annual precipitation can reach up to 5000 mm a year, with an annual average of 2620 mm (Klinka *et al.* 1991; Egan 1997). With up to 70 percent of this precipitation falling as snow in the colder winter months, total snow accumulations can reach as high as 900 cm (Egan 1997; British Columbia Ministry of Environment Lands and Parks 2000). Late-lying snow cover is common into the months of July and August. Soils within the MH Zone remain unfrozen throughout the year, as air temperatures do not drop low enough to overcome the insulating effect of deep snowpacks (Klinka *et al.* 1991).

3.2 Strathcona Park Study Sites

3.2.1 Introduction

Although there are well constrained glacial records available from other areas in the North American Cordillera (*i.e.*, Canadian Rocky Mountains), comparatively little research of this nature has been undertaken in the southern Coast Mountains of British Columbia (*i.e.*, Mathews 1951; Ryder and Thomson 1986), and the only LIA glacier research on Vancouver Island is that of Smith and Laroque (1996) at Moving Glacier. The lack of baseline glacier knowledge for southwestern Canada, and particularly Vancouver Island, indicates that a focused research program is necessary in order to determine the full extent of mass balance changes in response to Holocene climate change in the Pacific Northwest.

Analysis of aerial photographs and reconnaissance trips to various glacier forefields in Strathcona Park between 1995 and 1998 led to the discovery of nested terminal moraine complexes at two sites: Septimus Glacier and Colonel Foster Glacier (*unofficial names*). The preservation of well-defined moraines in conjunction with the presence of tree and lichen growth on the moraine surfaces provided an opportunity to construct a LIA moraine chronosequence for each site.

3.2.2 Colonel Foster Glacier

Colonel Foster Glacier is located in a northeast-facing cirque at the base of a steep headwall below the summit of Mount Colonel Foster in northern Strathcona Park (49° 46' N, 125° 51' W; Figure 3.2). Access to the site is by the Elk River Trail, which begins at Highway 28 and follows the Elk River Valley for *ca.* 12 km.

Bedrock in the Mount Colonel Foster study area is predominantly volcanic, consisting of pillow basalts, breccias, and sheet flows of the Karmutsen Formation (Vancouver Group) (Massey, 1994). Vegetation at the Mt. Colonel Foster site is classified in the Western Island Variant of the Maritime Forested MH subzone (Hnytka 1990).

Colonel Foster Glacier is presently 0.35 km^2 in area and calves into Iceberg Lake (*unofficial name*) at 980 m asl. A nested moraine complex consisting of eight terminal and recessional moraines is located on the north side of the lake, 200 m from the present glacier terminus (Figures 3.3 and 3.4). The moraines are breached by an outlet stream that divides the complex into eastern and western components. The eastern moraines have been heavily influenced by snow avalanche activity, and consequently have few trees and lichen to assist with relative age dating. In contrast, the western moraines have a more complete and healthy vegetation cover suitable for sampling and dating, and were the focus of study at this site.

The western moraine complex is approximately 100 m wide and converges at an avalanche path at *ca.*1200 m asl against the bedrock face of Mount Colonel Foster. From this location, the moraines follow an arcuate path for *ca.* 560 m along the forest trimline to the outlet stream that drains Iceberg Lake at 980 m asl (Figures 3.3, 3.4, and 3.5).

Figure 3.3. 1978 Aerial photograph of the Colonel Foster Glacier site (NAPL 30BC 78 076 236). The moraine complex is highlighted by a black rectangle.

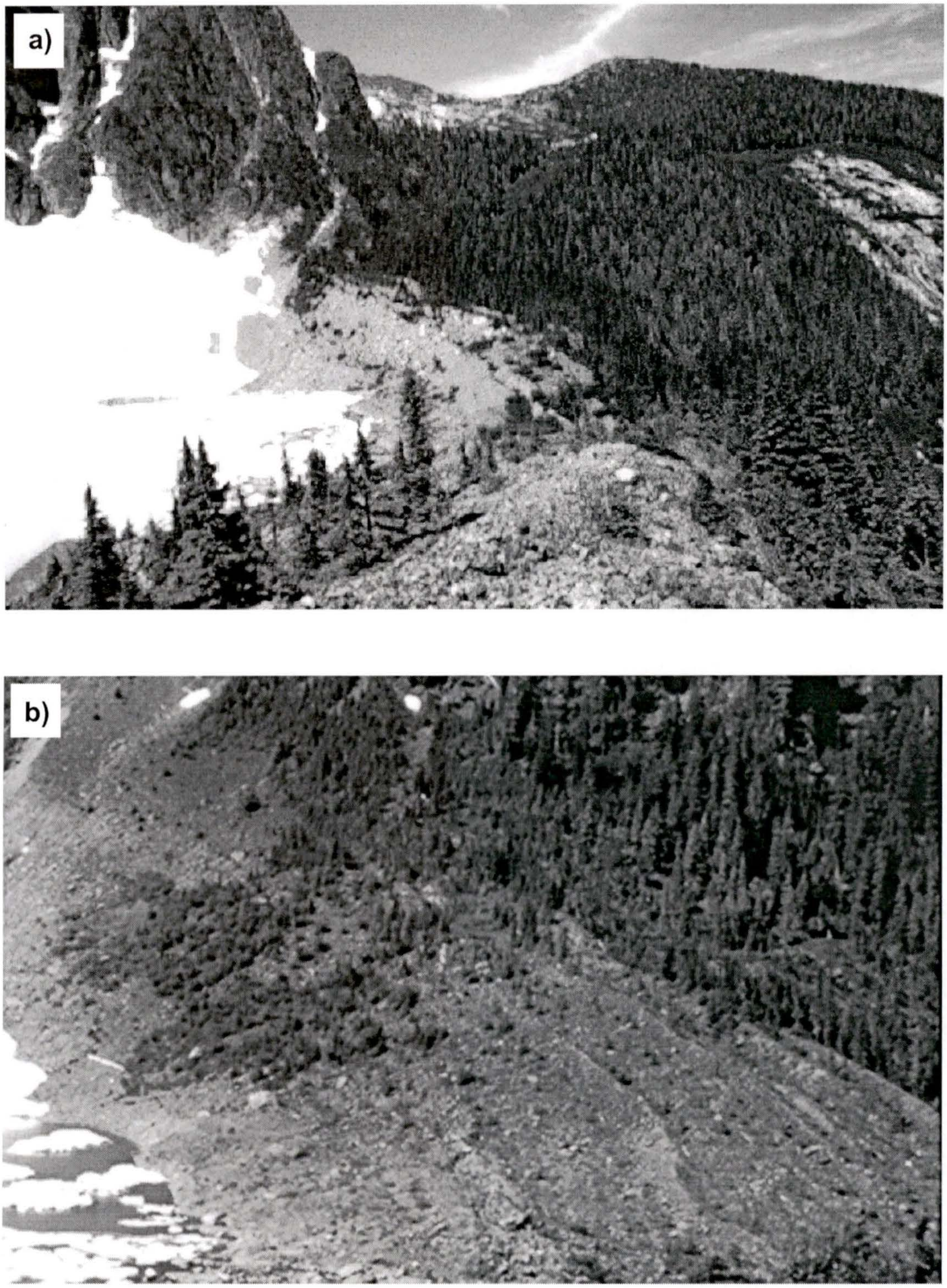


Figure 3.4. Photographs of the western moraine complex at Colonel Foster Glacier: (a) Standing on lower eastern moraines looking across at western complex; (b) Looking down across from high on the eastern moraine complex.

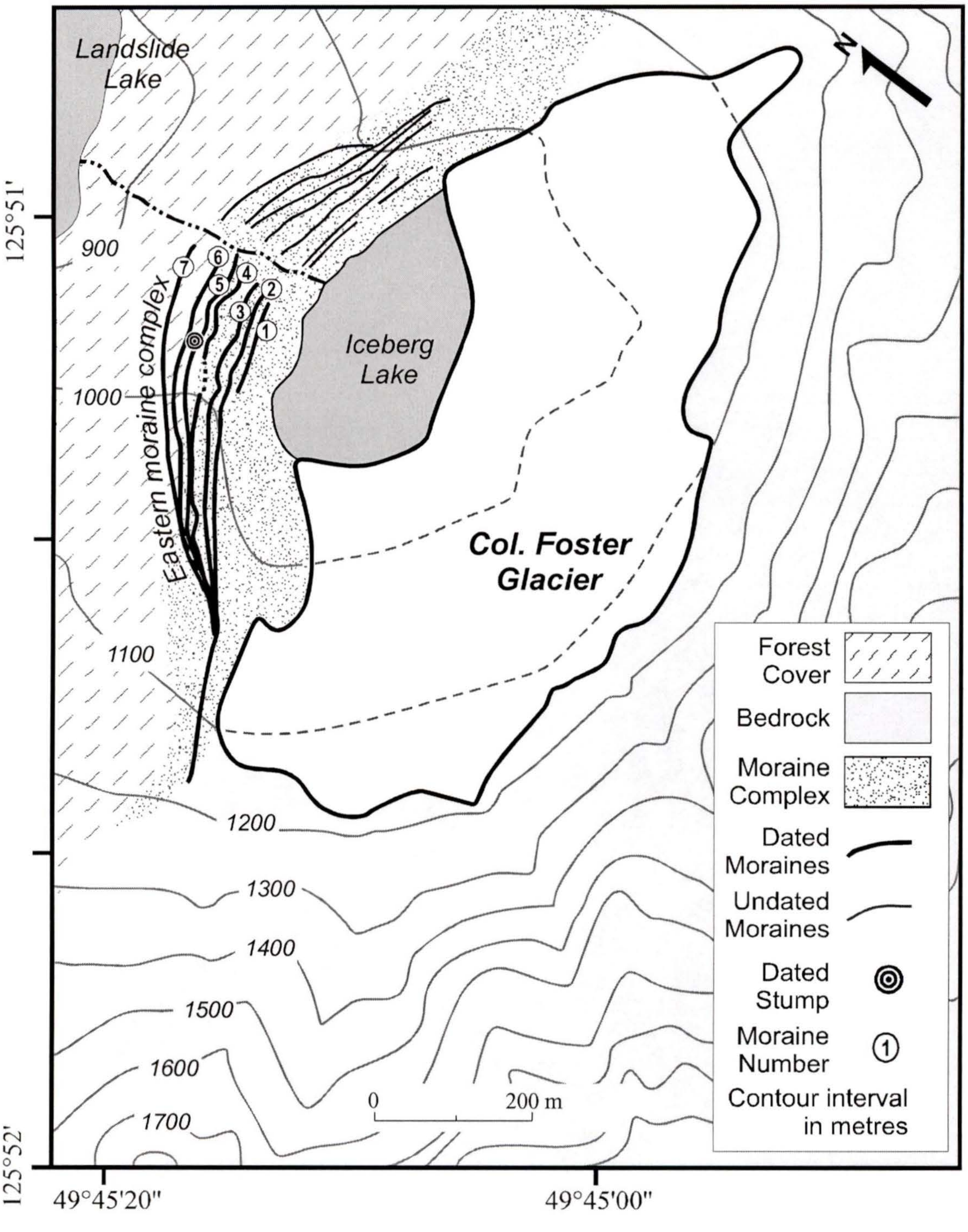


Figure 3.5. Site map of the Colonel Foster Glacier study area highlighting the location of the western moraine complex. Dated western moraines are shown as heavy black lines, and the undated eastern moraines are shown as thin lines.

The western moraine complex can be separated into two distinct moraine suites based on morphology and stratigraphic position: an inner suite of five relatively young and sparsely treed moraines (Moraines 1- 5), and an outer, older suite of three mainly forested moraines (Moraines 6 - 8) (Figure 3.6).

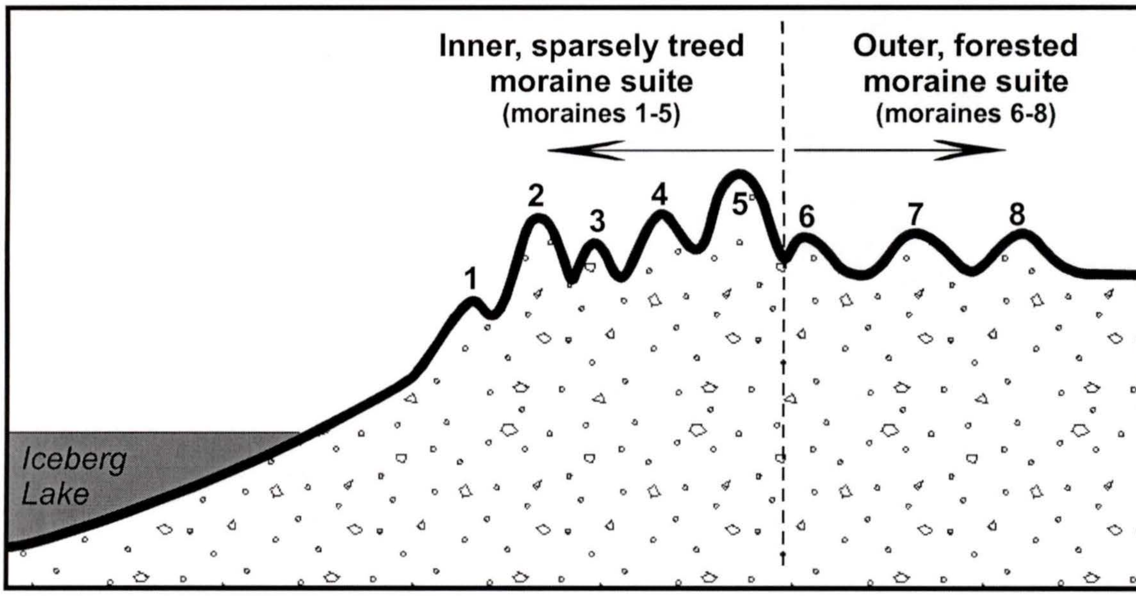


Figure 3.6. Schematic cross-section of the inner and outer eastern moraine suites at Colonel Foster Glacier. Image is not to scale.

Tree growth on the inner ridges (Moraines 1- 5) is characterized by immature mountain hemlock, a limited number of supalpine fir (*Abies lasiocarpa*), and fewer yellow-cedar. The steepest eastern and western limits of the moraine complex are dominated by alder (*Alnus sp.*) and willow (*Salix sp.*) thickets. With the exception of Moraines 1 and 2, lichen thalli (*Rhizocarpon sp.*) are found on larger boulders throughout the moraine complex.

Individual ridges of the inner suite (Moraines 1 –5) are well preserved with distinct crests, and have a relatively 'fresh' appearance compared to the moraines of the outer suite (Moraines 6 – 8). Among the five distinct ridges of this suite are numerous small recessional and/or annual push moraines. The general composition of the inner moraine suite consists of gravel, angular stones and boulders, and several large blocks that range up to 5 m in diameter. A general change in moraine composition occurs *ca.* 140 m up (west) from the stream outlet. From this point up into the forested western flank of Mount Colonel Foster a change from relatively finer materials to coarser and larger clasts and boulders occurs as a result of the proximity to the parent material.

The innermost ridge (Moraine 1) is a small ridge, *ca.* 0.5 – 1 m high, 1 –1.5 m wide, composed of smaller, relatively uniform-sized angular debris (0.02-0.1 m diameter) with only a few larger boulders. The proximal slope drops sharply to the lakeshore, whereas the distal slope appears to 'lap' onto the proximal face of the larger Moraine 2. Most of the young trees on this moraine are found on the proximal slope, and lichen presence is sporadic, found only on the few larger clasts and boulders.

The second ridge (Moraine 2) is distinct from Moraine 1 in its morphology and composition. It is composed of large, angular debris (rocks and boulders 0.5 to 2 m diameter) at its westernmost extent, *ca.* 350 m from the outlet stream. It is broader and taller than Moraine 1, and has a sharper crest that rises 2 m above the first moraine. Moraine 2 is difficult to distinguish at its western limit nearest the avalanche path, but is more prominent throughout the lower reaches, becoming a double crest in the lower 150 m near the stream, where its composition becomes markedly finer. Moraine 2 pushes up close to, and overrides parts of the proximal slope of Moraine 3 in a number of locations.

There is a large amount of woody debris mixed in with Moraine 2, particularly at its western limits. However, the degree of weathering and deterioration of the wood was too extensive to preserve any distinguishable ring patterns. As with Moraine 1, most of the trees found on Moraine 2 were growing on the proximal face, and lichen cover was also limited to a few larger boulders.

The third and fourth ridges (Moraine 3 and Moraine 4) are similar in morphology and height to Moraine 1, but Moraine 3 has a generally larger rock and boulder content (0.2 –1.0 m diameter), with several large boulders (1-3 m) along the crest line. Moraine 3 splits into a double crest approximately halfway down towards the outlet stream. A 3-4 m diameter lichen-covered boulder embedded in the moraine marks the location where this bifurcation occurs. Tree growth is found on all aspects of Moraine 3 (crest, distal and proximal faces), but is limited to the proximal face of Moraine 4. Lichen cover on Moraine 3 is more prominent than on Moraine 2, particularly the distinct large lichen-covered block *ca.* 200 m up (west) from the outlet stream. Moraine 4 is discontinuous at its easterly extent, and in places appears to override the proximal face of Moraine 5. The region between Moraine 2 and 5 is a slight depression, approximately 1-2 m lower than the surrounding moraines, with a greater moss cover than the other areas.

The prominent outer ridge (Moraine 5) of the inner suite is similar in morphology to Moraine 2 in that it is broader and taller than the other moraines. However, the morainic debris is an order of magnitude larger, consisting of mostly large clasts (1–2 m diameter) with a few larger boulders interspersed along the crest. Moraine 5 becomes progressively more forested towards its westerly extent, beyond which it is influenced by avalanche activity. The distal slope of Moraine 5 spills down *ca.* 3 - 4 m, where it directly overlies Moraine 6 of the outer and older moraine suite (Figure 3.6).

In contrast to the well-defined moraines of the inner suite (Moraine 1 – Moraine 5), the outer set of ridges (Moraine 6 – Moraine 8) have a heavily weathered, 'aged' appearance and a well-rounded morphology lacking distinct crests. The extent of this suite (Moraine 6 – Moraine 8) is not as great as the inner suite (Moraine 1 – Moraine 5), as the moraines converge and are overridden by the inner suite approximately 450 m up (west) from the outlet stream. Moraines 6, 7 and 8 also have limited soil development (grey Podzol), and are also more heavily forested with mature mountain hemlock, interspersed with yellow-cedar, and subalpine fir.

3.2.3 Septimus Glacier

Septimus Glacier is located 40 km southeast of Mount Colonel Foster on the northern flank of Mount Septimus near the southern boundary of Strathcona Park (49° 29' N, 125° 32' W). Access to the site is by the 12 km Bedwell Lake/Cream Lake trail at the south end of Buttle Lake (Figure 3.2).

Bedrock at the Mount Septimus site is primarily Sicker Group volcanics, consisting of andesitic and basaltic flows, as well as volcanic breccia, tuff, and volcanic sandstone. Minor intrusions of quartz diorite, gabbro, and gneiss of the Westcoast Complex are found throughout the study area (Surdam 1968). As a result of its higher average elevation (*ca.* 300 m higher than the Colonel Foster moraines), the Mount Septimus site falls within the Maritime Parkland MH subzone, with the higher reaches in the Coastal Alpine Tundra Zone (Hnytka 1990).

Situated in a northwest-facing cirque at 1350 m, Septimus Glacier has a present area of *ca.* 0.08 km^2. A complex of four nested moraines is located directly above a prograding delta on the southeast shore of Cream Lake at 1261 m asl (Figures 3.7, 3.8, and 3.9). An outlet stream breaches the moraines, dividing the complex into eastern and western components. The eastern moraines are moderately well-defined, but appear to have been influenced by either snow avalanche activity, or snowloading. Consequently they lack adequate tree cover for dating. In contrast, the western moraines are much smaller and discontinuous, yet have two small stands of trees growing on protected bedrock outcrops. Although there is sufficient lichen growth for dating purposes, it is more abundant on the eastern moraine complex where the larger substrate provides more suitable surfaces for lichen growth.

Along the northeast side of the cirque there is a set of four distinct lateral moraines (Moraine 1-Moraine 4) that extend *ca.* 650 – 700 m from the terminal moraine complex at 1315 m asl up to 1500 m asl, where they converge and become a single ridge that terminates against the bedrock wall of Mount Septimus (Figures 3.6, 3.7 and 3.8).

The innermost ridge, Moraine 1, is *ca.* 0.5 m tall by 75 m in length, situated near the bottom of the proximal face of Moraine 2, and composed of smaller angular gravels and stones (0.05 – 0.20 m diameter) in a fine matrix. There are no trees present, and lichen growth is limited to a few large clasts near the southerly most 25 m of the moraine.

Moraine 2 descends from the main complex at 1500 m asl, and extends down (north) towards the terminal complex for 250 m. Two sections of the crest appear to have been disturbed or partially reworked, giving the appearance of a dual crest line.

Figure 3.7. 1980 Aerial photograph of the Septimus Glacier site (NAPL 30BC 80 095 044). The moraine complex is highlighted by a black rectangle.

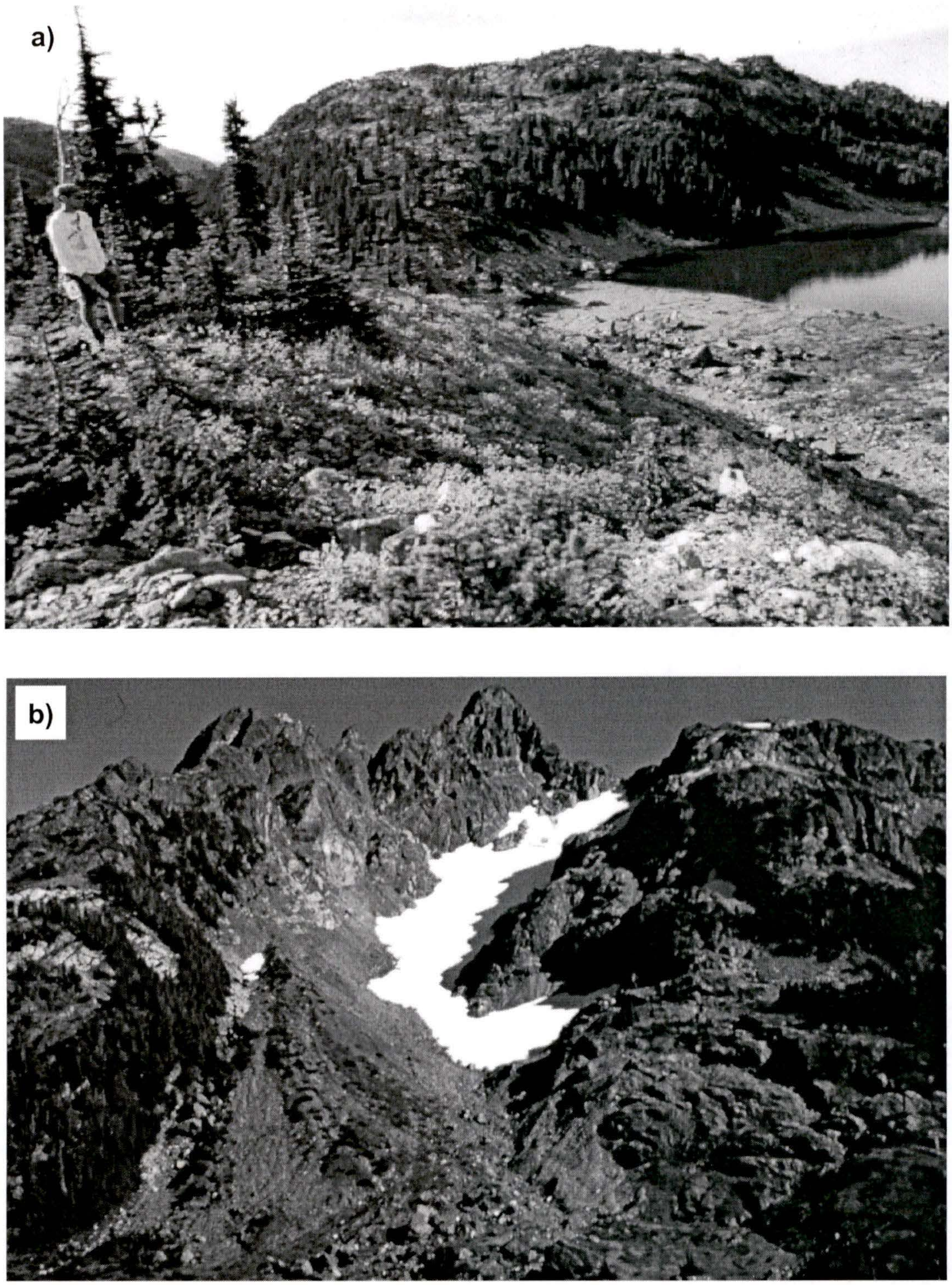


Figure 3.8. Photographs of the eastern moraine complex at Septimus Glacier: (a) Looking north down the eastern moraines from Tree Site A to Tree Site B; (b) Looking southeast to the eastern moraine complex.

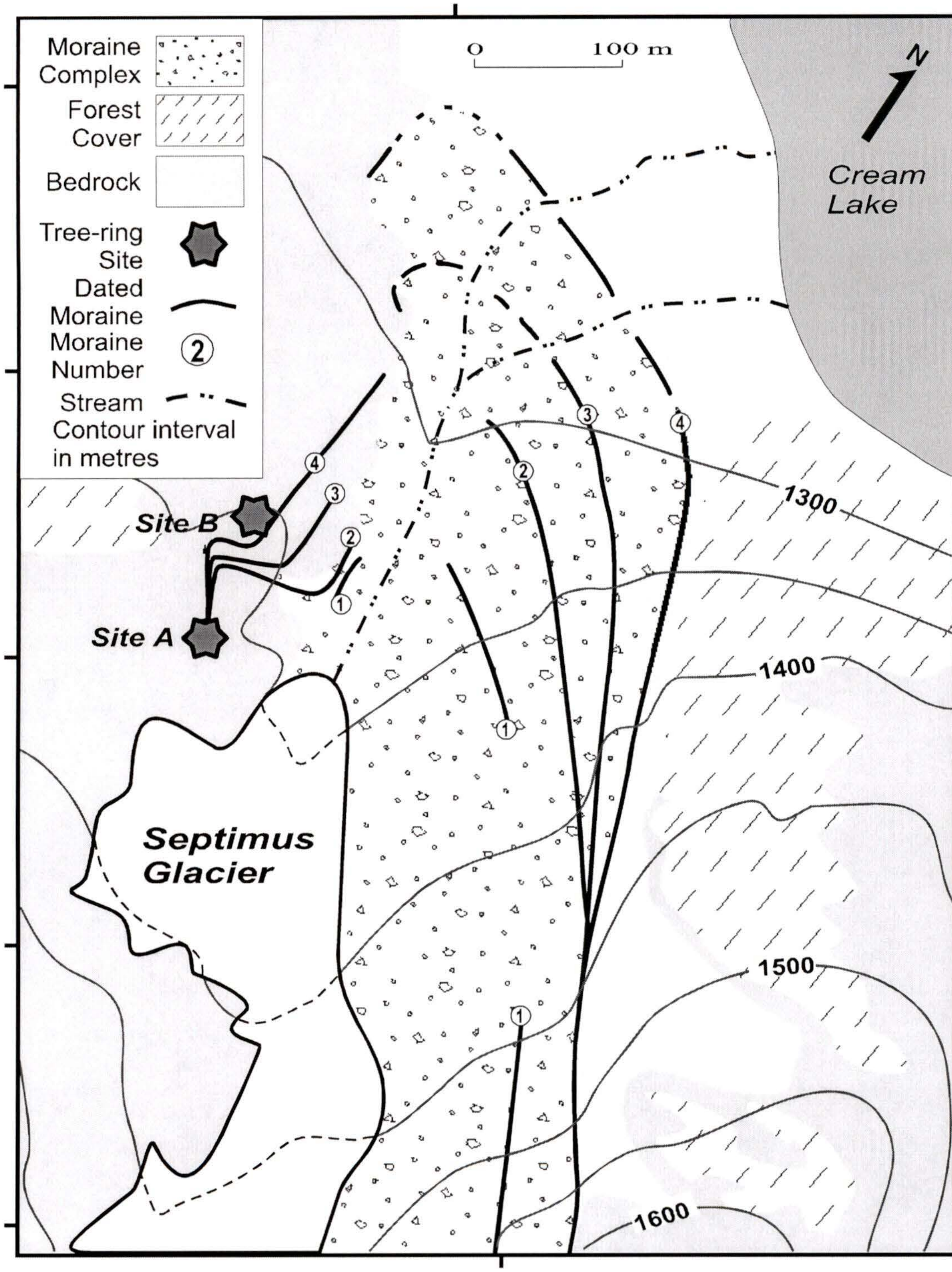


Figure 3.9. Site map of the Septimus Glacier study area showing the location of the moraines and tree-ring dating sites.

Moraine 2 is larger than Moraine 1, with a taller, more rounded appearance, and is composed of larger, coarse debris (0.2 - 0.9 m diameter), which is predominantly clast-supported. Tree growth is limited to snow damaged krummholtz, and lichen growth is present on the larger debris. The proximal face has a dense cover of moss and heath.

Moraine 3 is located along the 'spine' of the eastern complex. It is similar in composition to Moraine 2, with the exception of several large embedded boulders (1 – 3 m diameter). This moraine extends for the entire length of the complex, but is washed out at its lower northerly reaches. Vegetation cover is similar to that of Moraine 2, although the lichen presence is greater as a result of the larger clasts and boulders.

The outermost ridge (Moraine 4) branches away from the main complex to the east, and is located 10 and 20 m outside of Moraine 3. The composition of Moraine 4 is similar to that of Moraine 3, with the addition of several very large blocks and boulders up to 5 m in diameter. Tree growth here is also limited to snow-damaged krummholtz.

Unlike the eastern moraines, moraines along the western perimeter of the forefield are discontinuous, and broken by a large, steep bedrock outcrop that begins 250 m upglacier from the terminal moraine. Fragments of Moraine 3 and Moraine 4 are preserved on, and between, the bedrock outcrops at 1350 m and 1425 m asl, whereas Moraines 1 and 2 are not well preserved at all.

Lichen (*Rhizocarpon sp.*) is present on all moraines, although sparse on Moraines 1 and 2 of the eastern suite. Tree growth on the eastern moraines is limited to young, shrubby mountain hemlock, subalpine fir and yellow-cedar krummholtz interspersed with alder (*Alnus sp.*) seedlings and saplings. All trees on the eastern moraines show signs of significant snow damage. However, the western moraine complex contains two isolated stands of trees above bedrock outcrops, 250 m and 350 m upglacier from the terminal moraines, respectively. Stands of mature, old growth (500 to 700 years) mountain hemlock are found on the north and west sides of Cream Lake, and along the ridge to the east of the western moraines (Figure 3.7).

CHAPTER 4. DATING LIA GLACIER DEPOSITS

4.1 Introduction

In mid and high-latitude mountainous regions, the most obvious evidence of previous glaciation is the sediment and landforms remaining after glacial retreat. The relatively 'fresh' appearance of LIA moraines found in the forefield of many contemporary glaciers is often the most prominent landform in such environments (Luckman 2000). The ultimate value of paleoenvironmental or paleoclimatic time series derived from such deposits rests on the degree to which the deposits can be accurately dated (Porter 1981a). Accurate dating of the surface record provides chronological control for glacial and climatic events, information about the timing and relative extent of glacial advances, date of the maximum advance, and recessional history of the most recent events.

Glacial deposits can be used to reconstruct glacier histories, assess the synchroneity of glacier advances, and to compare the glacial history with both a tree-ring and climate record. This requires the determination of both the temporal and spatial extent of former ice front positions, which can be determined through detailed mapping of landforms, deposits, and vegetation that have been formed or influenced, directly or indirectly, by previous glacial activity (Porter 1981a). Such features are best determined from topographic maps, aerial and historical photographs, and on site field investigations (Luckman 1988). An approach combining aerial photography with field observations often provides the best morphologic and topographic evidence without the "cartographic interpretation" inherent in many topographic maps (Robinson 1998). Historical and aerial photographs can also be used to identify relatively recent (< 70 years) terminal ice positions not recorded by moraines or glacially induced trimlines.

4.2 Landform Mapping

In this study, glacial landforms were initially identified from aerial photographs and verified through field investigations. Although aerial photographic coverage began in the early 1930's, adequate photographs were limited due to late-lying snow cover that obscured most glacial deposits. From an initial inventory of 30 aerial photographs, five photos were selected for producing base maps for field mapping (Table 4.1).

In addition to the aerial photographs, Terrain Resource Information Management (TRIM) maps were used to construct individual site maps (Table 4.1). Aerial photographs and TRIM maps were enlarged to scales of 1:10,000 to 1:5000, digitized, and information such as moraine ridges and sampling areas were plotted.

Table 4.1. Aerial photographs and maps used in this study

1. Aerial Photographs			
Flight Line	***Photo Number***	***Scale***	***Year***
A4011	24	1:19,500	1931
30BC 78 076	236	1:20,000	1978
30BC 80 072	045[1], 111	1:20,000	1980
30BC 80 095	044[1]	1:20,000	1980
2. Terrain Resource Information Management (TRIM) Maps			
Map Number	***Study Site***	***Scale***	***Year***
92F.043[1]	Colonel Foster Glacier	1:20,000	1993
92F.071[1]	Septimus Glacier	1:20,000	1992
3. NTS Topographic Maps			
Map Number	***Title***	***Scale***	***Year***
92F/5	Bedwell River	1:50,000	1989
92F/12	Buttle Lake	1:50,000	1989
92F/13	Upper Campbell Lake	1:50,000	1974

[1]Denotes map or photo used to construct base maps for study sites

4.3 Landform Dating

Several techniques are available for dating glacial deposits, but may not be applicable at any one particular site due to a lack of suitable dating material. Methods of dating control that have proven widely applicable to alpine glacial deposits of the Late Neoglacial are primarily radiocarbon dating, historical data, and geobotanical methods such as dendrochronology and lichenometry (Fritts 1976; Shroder 1980; Schweingruber 1988; Smith *et al.* 1995; Luckman and Villalba 2001).

Despite the advantages of radiocarbon dating, unavoidable problems associated with long-term fluctuations in atmospheric carbon limit the applicability of this method for providing precise dates of deposits in the last 1000 years. Due to substantial variation in atmospheric ^{14}C, there is a non-linear relationship between radiocarbon years and calendar years. Consequently, it is generally not possible to obtain a unique and unambiguous calendar age from a ^{14}C date for late LIA deposits (Stuiver 1978; Stuiver and Becker 1993). Figure 4.1 illustrates how a radiocarbon date of 220 ± 50 yr. B.P. could correspond to a calendar date in the range 150 to 420 years B.P. (Porter 1981a).

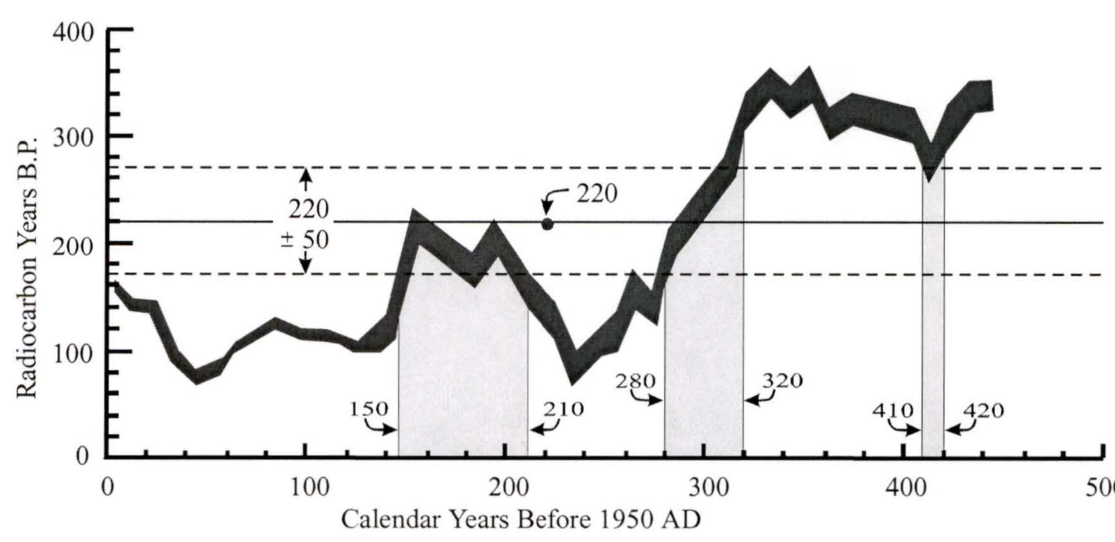


Figure 4.1. ^{14}C dating problems for LIA deposits (from Porter 1981a).

When available, documentary records offer some of the most reliable information for glacier variations during the latter stages of the LIA. In the European Alps, documentary accounts, paintings, tax records, sketches and lithographs have provided evidence of ice front positions as far back as the seventeenth century (Grove 1988). However, direct observations before the 1900's are rare for Canadian alpine glaciers. As with most regions that have short, or poorly-recorded documentary histories of ice front position, the age determination of LIA moraines in the Vancouver Island mountains must rely on evidence preserved in natural archives such as trees and glacial deposits.

Where documentary sources are limited, biological techniques such as dendrochronology and lichenometry provide the most suitable means of dating LIA deposits. Dendrochronological techniques are preferable below treeline, as dating materials (*i.e.*, trees and stumps) are usually present. The growth cycles of most alpine trees encompass the timing of the LIA, and thus, the annual ring-widths provide a continuous high-resolution record of environmental events. Lichenometry is the only technique applicable to sites above treeline on Vancouver Island; however, treeline can reach the summit of all but the tallest peaks. Since most glaciers on Vancouver Island are located in the treeline-glacier zone, dendrochronological dating methods appear to provide the most suitable means for dating glacial activity on Vancouver Island.

4.4 Dendrochronological Dating

In forested mountain regions, it is possible to date glacial advances directly or indirectly through dendrochronological techniques when a glacier has advanced below the treeline (*e.g.*, Lawrence 1950a; Heusser 1956; Bray and Struik 1963; Sigafoos and Hendricks 1969; Luckman 1988; Smith and Laroque 1996). Conventionally, there are four types of dendrochronological evidence used to date glacial fluctuations and their associated deposits: (*i*) the age of the oldest tree growing on a landform within the glacial forefield; (*ii*) dating of abrupt changes in growth rates or ring symmetry of trees that were tilted or in close proximity to the glacier; (*iii*) the date of ice-damage corrasion scars on living trees produced by prior glacial contact; and, (*iv*) the age of trees that were killed by an advancing glacier (Table 4.2) (Lawrence 1950a; Bray and Struik 1963; Alestalo 1971; Shroder 1980; Schweingruber 1988; Wiles *et al.* 1996a).

Table 4.2. Dendroglaciological dating methods

Type of Evidence	Dating Precision	Information Provided	Potential Limitations
Trees growing on landforms in the glacier forefield.	-Within 5 years or greater.	- Age of the oldest tree provides minimum estimate for surface age.	- Ecesis interval can be difficult to estimate. - Assumption that the oldest tree has been sampled.
Abrupt change in ring symmetry as a result of tilting.	- Exact calendar date of tilting event.	- Date of eccentric or abnormal growth indicates onset of event.	- Event may be severe enough to kill the tree and would require crossdating.
Ice-contact scars.	- Exact calendar date of damage or ice-contact event.	- Damage date indicates glacier position at a specific time	- Scarred, living trees are often difficult to find - Dead trees may require crossdating.
Trees killed by glacier.	- Exact calendar date of tree death.	- *In-situ*: kill date indicates glacier position at a specific time. - Detrital wood provides limiting date for glacial event.	- Requires crossdating to determine tree age or kill date. - Dating precision for an event depends on preservation of wood and loss of outer rings.

adapted from Robinson (1998)

The first three dating methods utilize living trees to provide dating control, whereas the last method involves the combination of both living and dead trees. When using live trees as a dating control, a count of the annual growth-rings can be used to determine the minimum age of the deposit, the year the tree was tilted, or the date of the damage to the stem (Lawrence 1950a). Trees growing on ice-marginal deposits within the glacial forefield not only provide minimum dates for the substrate on which they are growing, thus providing limiting dates for landform formation, but they also date the transition to a climate trend that favoured glacial retreat (Porter 1981a). In exceptional cases, trees that were directly affected by an advancing glacier (*i.e.*, tilting or ice-contact scarring) and continued to live, provide the most precise dating of glacial events, and, therefore, can be used to date the time of maximum glacier extent. Although it is possible to date the maximum extent, it is much more difficult to obtain the initial date of an advance, and thereby determine the time of climatic change that led to a positive mass balance, or define the duration of the advance (Porter 1981a; Luckman 2000).

In using dendrochronology to determine the dates of glacial deposits, moraine ages are commonly estimated by counting the growth-rings of the oldest tree growing on the surface, with an adjustment for the colonisation (ecesis) period (Porter 1981a; Luckman 2000). In response to ameliorating climate conditions, the glacier front recedes and 'fresh' substrate is continuously exposed to a progressive colonisation by tree seedlings (Sigafoos and Hendricks 1969). This process implies that trees will colonise the moraines in the order that the moraines were exposed and stabilized, resulting in increasingly younger tree ages towards the glacier snout. Consequently, the germination date of the oldest tree growing on a substrate provides a minimum estimate of the surface age, and therefore the minimum date for emplacement (Heusser 1956; Sigafoos and Hendricks 1969; McCarthy and Luckman 1993).

4.4.1 Dendrochronological Field Methods Used in This Study

Tree-ring data were collected through four field-sampling programmes: (*i*) sampling of living trees on the surface of glacial forefields to provide minimum dates of moraine stabilization; (*ii*) collecting cores and discs from trees growing on a surface of known age to determine the ecesis interval for mountain hemlock; (*iii*) sampling trees that were either killed or disturbed (*i.e.*, tilted) by advancing ice; and (*iv*) collecting cores from mature mountain hemlock forests adjacent to each site for constructing master tree-ring chronologies. Master ring-width chronologies are used for crossdating, positioning glacial activity in the context of a qualitative proxy climatic record (*e.g.*, general climate regime), and determining the feasibility of using mountain hemlock ring-widths as a proxy mass balance indicator.

4.4.2 Live Tree Dating

The objective of sampling in tree-ring dating is to recover a complete radial path, including the pith, from the oldest tree growing on the moraine surface. However, coring itself may not reveal the establishment year for a tree (*i.e.*, germination date or surface stabilisation date) because the increment corer may have missed the pith or the core was extracted too high up the tree (Sigafoos and Hendricks 1972). Tree ages obtained without addressing these potential sampling errors can result in an underestimation of surface dates; any reconstructions based on these uncertainties will, therefore, be of limited value. It is essential then, that potential dating errors are addressed and appropriate corrections applied so that episodes of moraine emplacement can be placed more accurately in the context of past climate and mass balance conditions.

Possible sampling uncertainties that need to be addressed include: (*i*) whether or not the oldest tree on the moraine has been sampled; (*ii*) the number of years (tree-rings) lost when cores are removed higher up the stem; (*iii*) the number of years to the pith if the pith was not reached during sampling; and, (*iv*) the ecesis interval - the time between surface exposure and colonisation of the first tree (Table 4.3) (Lawrence 1950a; Sigafoos and Hendricks 1969; Shroder 1980; Heikkinen 1984; McCarthy *et al.* 1991; 1993; Wiles *et al.* 1996a).

Table 4.3. Potential tree-ring dating errors and associated correction factors

Potential Dating Error	**Error Term**	**Method of Calculation**
OT = Oldest Tree	= 0 if all trees are sampled	
HE = Height-Age Error	= 0 if sampled at root crown = *n*, depends on height sampled above the root crown	Estimate time to grow to sampling height: - use mean apical growth rate of young trees - sample at given intervals up the stem
PE = Pith Error	= 0 if pith sampled = *n*, depends on years from pith	Estimate distance to pith in years based on ring curvature from similar aged trees
EI = Ecesis Interval	= 1…*n*, depends on seed source, microclimate, and substrate	Calculate ecesis interval by subtracting seedling age from a known substrate age

An estimation for the minimum date of a glacial deposit (*i.e.*, moraine) can be summarised by:

Landform Date = *Germination Date* + EI

Where *Germination Date* = Age of the Oldest Tree = **ER** + **HE** + **PE**
And,
ER = year of Earliest Ring
HE = Height Error
PE = Pith Error

adapted from McCarthy (1985)

Another factor that needs consideration, but is more difficult to determine, is whether the trees on the moraine are first generation trees (Sigafoos and Hendricks 1969).

If the trees are not first generation, the age of the surface will be greatly underestimated and reconstructions based on these moraines will be of little use.

On heavily forested moraines, the oldest tree is often hard to identify, as the relationship between tree age and girth, height, or branch spacing is uncertain and varies greatly between species. The harsh climate of a glacial forefield can severely limit tree growth, and it is not uncommon to find a 200-year-old tree with the girth and height of a 20-year-old sapling found in more favourable conditions (Franklin *et al.* 1971; Robinson 1998). Although there is no absolute method for ensuring the oldest tree has been identified, sampling all of the trees on the landform, or at least an adequately large sub-sample of the population, will increase the probability of identifying the oldest tree.

All living trees growing in both glacier forefields were sampled in order to determine the oldest tree, and therefore the minimum age of moraine establishment. Core samples were removed from each tree using an 18" Hagloff increment borer. Trees used for dating were cored as close to the base of the tree as possible, with the increment borer angled down towards the root crown. Sampling height was recorded for each tree so that an age-height correction could be calculated for those trees that were cored above the root crown.

The location of each tree sampled was mapped and recorded as either growing on the proximal face (PF), the crest (C), or the distal face (DF). Elevation was determined from a hand-held altimeter and later verified on 1:50,000 NTS topographic maps and 1:20,000 TRIM maps. Coring height, pith accuracy, the number of cores taken, and tree species were also recorded. All cores were sealed in plastic straws, labelled, and returned to the University of Victoria Tree-Ring Laboratory for preparation and analysis.

4.4.3 Sampling Corrections

4.4.3.1 Pith correction

Dating uncertainty can be introduced when non-destructive sampling (increment coring) is used to determine tree ages; particularly with older trees, as the possibility of missing the pith increases with the size of the tree (Sigafoos and Hendricks 1972). Asymmetric growth, or the presence of branches or other obstacles, may also result in an off-pith sample. This can be resolved by either removing successive cores until the pith is recorded, or by applying a correction factor to approximate the pith based on ring-width and ring curvature data (Norton *et al.* 1987).

For each tree sampled, the extracted cores were examined to determine if the pith had been reached. If the pith was not sampled (*i.e.,* off-pith), and the tree appeared to be one of the older trees on the moraine, additional cores were extracted from the tree until the pith was reached or could be determined with a degree of certainty. If the pith was sampled, or if the tree was significantly younger than other trees sampled on the same moraine, no further cores were taken from the tree. For core samples that were close to, but did not pass through the centre of the tree, a years-to-pith (YTP) correction factor was estimated based on the curvature and early ring-width pattern of the off-pith core sample and from similarly aged trees sampled nearby. An appropriate YTP correction factor was assigned to the off-pith core, which was then subtracted from the date of the earliest ring to determine the estimated pith date (Table 4.4 and Figure 4.2).

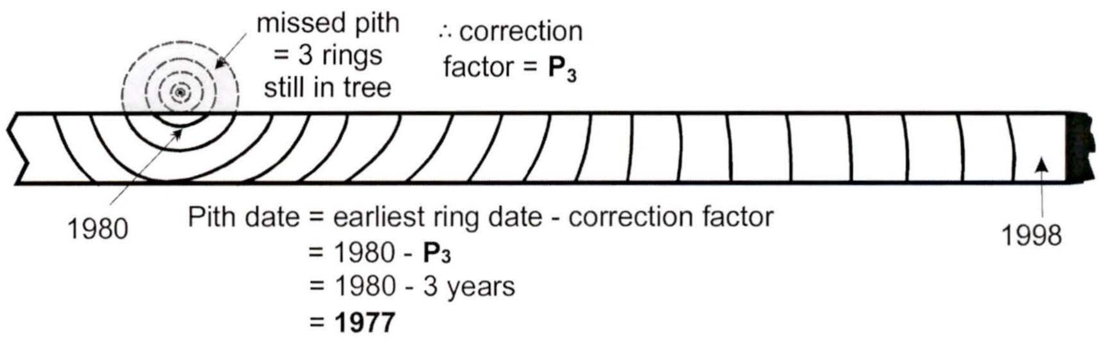


Figure 4.2. Schematic diagram illustrating the correction method used for core samples that were close to, but did not include, the pith.

Table 4.4. Years-to-Pith correction factors

Proximity to Pith	Classification	Correction Factor (yrs)
0 yrs (contains pith)	P	0
n yrs (missed pith)	P_n	1, 2, 3…***n***

Example: If the proximity to the pith is estimated at 3 years then $\mathbf{P}_n = \mathbf{P}_3$
A correction factor of three years is then added to the ring-count of the tree.

4.4.3.2 Sampling Height-Age Correction

Another consideration to be addressed when using dendrochronology to date tree-covered deposits is the sampling height-age relationship: the time required for the tree to grow to the level at which it was sampled. Tree ages are most accurately determined by counting the annual growth-rings from multiple radial paths on a cross section taken from the root crown level as close to the soil surface as possible (McCarthy *et al.* 1991). However, destructive sampling is not always desired or permitted, and branches or other obstacles may force sampling higher on the stem. In such cases, the tree can be cored above the obstacle and a correction added to the ring count to compensate for the time it took the tree to grow to the sampling height (McCarthy *et al.* 1991).

In order to determine a coring height-age correction factor for rings lost due to sampling above the root crown, a general rate of apical growth from trees growing in similar conditions is required. A coring height-age correction factor was established from a sample of ten young mountain hemlocks from the Colonel Foster Glacier forefield. The height of ten saplings and seedlings (*ca.* 1 m tall) was measured and the trees were cored at the root crown to determine their age and to estimate a general age-height relationship.

Where appropriate, the sampling height-correction factor was applied to the age of trees corresponding to the height at which the core was extracted. As most cores were taken in the lower 8 cm of the stem with the increment corer angled towards the root crown, the coring height correction factor was limited to only a small number of trees.

4.4.4 Ecesis Interval

Once tree ages are determined, the ecesis, or interval between deglaciation (stabilisation of the moraine surface) and the germination of the first seedling (*i.e.*, oldest tree on the surface), has to be determined. This time lag is the main source of error in dating a glacial deposit using tree ages (Heusser 1956; Sigafoos and Hendricks 1969; McCarthy and Luckman 1993). The ecesis interval for a given tree species is a function of the local environmental conditions, including the existence of a sufficient seed source, characteristics of the seedbed, and the climate conditions during the ecesis interval (Sigafoos and Hendricks 1969). A number of techniques have been used to determine ecesis intervals, however, the method used by Sigafoos and Hendricks (1969) remains the most widely accepted method.

Ecesis intervals have been determined for a number of tree species in the Pacific Northwest and Canadian Rocky Mountains and vary considerable from one site to the next, ranging from 1 to 100 yrs (Sigafoos and Hendricks 1961, 1969, 1972; Desloges and Ryder 1990; McCarthy *et al.* 1991; Smith *et al.* 1995; Wiles *et al.* 1999b; Luckman 2000). With such a large discrepancy in ecesis estimates, it is necessary to determine a local ecesis interval that reflects local growing conditions. The key to determining a representative ecesis interval is the presence of a precisely dated surface in close proximity to the study area. Such a site is located 900 m north of the Colonel Foster Glacier study site (Figures 4.3, 4.4, and 4.5) .

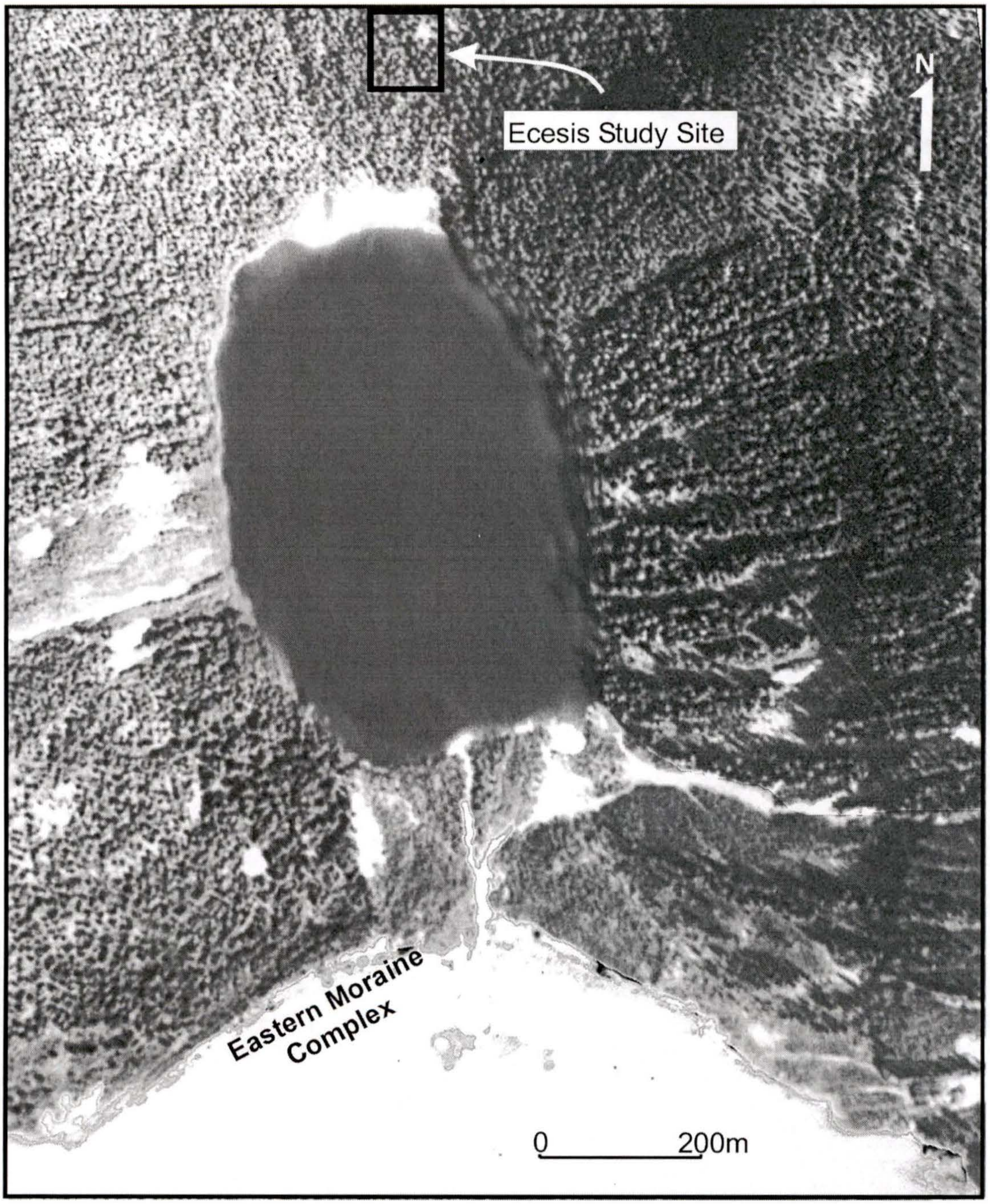


Figure 4.3. 1931 Aerial photograph showing the location of the ecesis study site before the 1946 earthquake and its relative position to the moraine study site (NAPL A4011-24). The black rectangle outlines the ecesis study area.

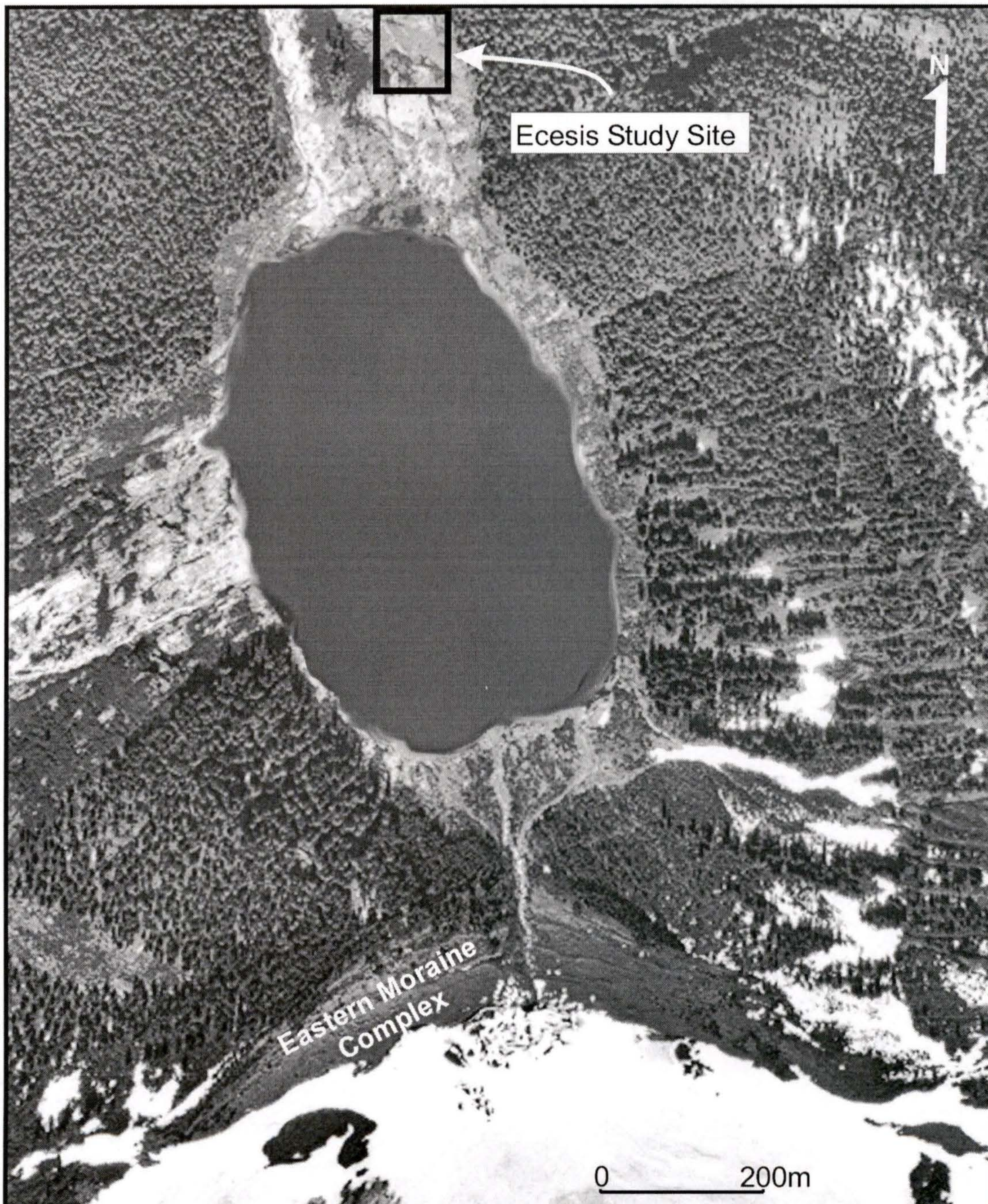


Figure 4.4. 1978 Aerial photograph showing the location of the ecesis study site after the 1946 earthquake and its relative position to the moraine study site (NAPL 30BC78 076 235). The black rectangle outlines the ecesis study area.

On June 23, 1946, a large magnitude earthquake (M 7.2) triggered a massive rock avalanche from the north face of Mount Colonel Foster, which deposited a substantial amount of material into Landslide Lake (Evans 1989). The resultant displacement wave rose to a height of 51 m, spilled over the lip of the lake destroying forest vegetation down the Elk River valley as far as 3 km from the lake. Forest regeneration has since occurred as a mixed forest of mountain hemlock, yellow-cedar, and amabilis fir (Figure 4.5).

Figure 4.5. Photograph looking southeast towards the bedrock knoll of the ecesis study site. Photo shows the recent forest regeneration adjacent to a stand of trees that survived the 1946 displacement wave.

A local mountain hemlock ecesis interval was determined from 13 mountain hemlock found growing on the bedrock surface of the landslide scar, 100 m below Landslide Lake. Eleven trees were sampled as close to the root crown as possible, and a cross-section removed from each tree. The remaining two trees were sampled with an increment borer angled down towards the root crown. After preparation in the laboratory, four radial paths were drawn on each cross-section and annual rings were counted along all four paths to determine the age of each tree. Ring counts for each of the increment cores were also recorded. Subtracting the age of the oldest tree from the known age of the landslide surface provided the local mountain hemlock ecesis interval.

4.4.5 Dendrogeomorphological Dating

Optimal sites for dating glacier events are found where the glacier has advanced into forested environments, thereby increasing the possibility of finding tilted, scarred or dead trees. In favourable conditions, it is possible to date the culmination of the advance by determining the age of trees that were scarred or tilted by the glacier but not necessarily killed (Lawrence 1950a; Heusser 1956; Luckman 1988). This can also be obtained by dating abrupt changes in growth patterns as a result of glacial proximity or through the removal of competing trees (Bray and Struik 1963; Porter 1981a; Heikkinen 1984; Villalba *et al.* 1990; Nicolussi and Patzelt 1996). During an advance phase, the glacier, or the ice proximal debris, may disrupt the surface upon which the tree is growing and cause the tree to tilt. This can result in an abrupt change in ring symmetry, or cause direct damage to the stem in the form of a corrasion scar (Shroder 1980).

When a tree is tilted by a glacier advance, either directly or indirectly, and continues to grow, a record of the event will be preserved by a change in the annual growth pattern that coincides with the time of tilting (Alestalo 1971; Shroder 1980). An abrupt change from symmetrical to eccentric growth occurs as the tree attempts to upright itself through the production of reaction wood and increased radial growth (Figure 4.6)

(Alestalo 1971; Fritts 1976). The change in the radial growth pattern can be accurately dated to the year the tree was first disturbed and, therefore, provide evidence for the timing of a glacial advance (Lawrence 1950a).

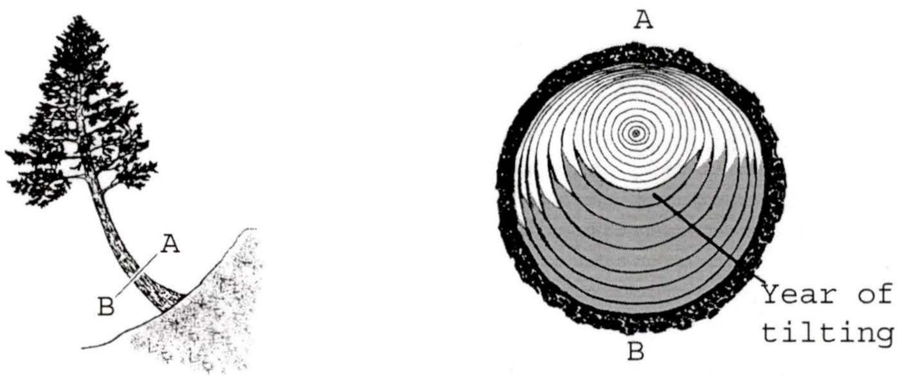


Figure 4.6. Asymmetrical growth pattern of conifers in response to tilting (Fritts, 1976).

To minimise destructive sampling in the park, trees that were tilted and not killed were sampled with an increment corer instead of being felled to obtain a cross section. A minimum of two increment cores, *ca.* 180° apart, were taken from each stem to determine the first year of eccentricity or reaction wood formation in the ring-width pattern. Each tree was cored until the pith was reached or until the distance to the pith could be confidently calculated. All cores sampled from the moraine trees were prepared as outlined in Section 4.5.1. Each core was analysed for distinct changes in annual ring-width patterns, and the date of onset and duration of notable eccentric growth patterns were recorded. Before dating eccentric ring-width patterns, an ecesis interval was added to the ring count, and where required, corrections were made for sampling height and distance from the pith.

4.4.6 Dendroglaciological Dating

When a glacier advances into a forest, trees are often killed, overridden, and/or buried by the advancing ice front. Trees sheared off above the root crown often leave behind a fully rooted *in-situ* stump that can provide information about the position and timing of the ice advance responsible for the tree's death. If all growth-rings are intact, the last year of growth can be used to determine the date and position of the glacier when the tree was killed (Schweingruber 1988). However, the use of subfossil trees to date glacial landforms is slightly more complex. This method requires crossdating the samples of unknown age with a local or regional master (dated) ring-width chronology developed from living trees in order to determine a kill date (Fritts 1976; Schweingruber 1988).

For trees that were pushed over and killed by advancing ice, a complete cross section was removed from the remaining *in situ* stump using a 53 cm bow saw. An attempt was made to sample the portion of the stump that appeared to be the most complete and well preserved. Samples were bound with tape to reduce damage during transport, and to prevent expansion and splitting once they were removed from the stumps and set to dry.

In the laboratory, the longest possible radial path from the pith to the outermost ring that passed through the fewest irregularities (*e.g.,* cracks, reaction wood) was selected for counting and measurement. Annual growth-rings were counted to determine the age of the stumps, and if needed, corrections were made for sampling height and ecesis interval. Undated tree-ring series (floating chronologies) were constructed from each stump and crossdated with a master chronology to determine the pith dates and, therefore, the kill dates.

4.4.7 Site Chronology Data Collection

In order to provide pith and kill dates to the floating chronologies, a local ring-width chronology was constructed from live mountain hemlock in the forest adjacent to each site. Two increment cores, approximately 180° apart, were extracted at breast height from each tree. By removing cores from opposite sides of each tree, ring-width data are replicated and the possibility of missing rings or 'false' rings is reduced (Fritts 1976; Schweingruber 1988; Cook and Kairiukstis 1990). Samples were transported in plastic straws to the University of Victoria Tree-Ring Laboratory where they were prepared, counted, and measured.

4.5 Dendrochronological Laboratory Methods

4.5.1 Sample Preparation and Analysis

Each core sample was prepared according to standard dendrochronological procedures (Stokes and Smiley 1996). After air-drying, each core was glued into a grooved board, labelled, and prepared for analysis by sanding with progressively finer grades of sand paper (100 to 800 grit). Cores were then hand-polished to enhance the definition and contrast of the annual tree-ring boundaries. Solid discs taken from living trees were prepared by removing rough saw cuts with a band-saw and belt-sander, and like core samples, were finished to a clean, polished surface using fine-grit sandpaper. Discs obtained from subfossil logs were soaked in melted paraffin wax to provide additional support.

All samples were counted a minimum of three times using digital and manual measuring systems. Samples were first converted to high-resolution digital images (800 to 2000 DPI) with an Agfa Duoscan flatbed scanner. Annual rings were counted and measured to the nearest ±0.01mm using the WinDENDRO (version 6.4a) digital tree-ring image processing system (Guay *et al.* 1992). Each sample was counted at least twice on a Velmex-type measuring stage using a Wild M3B stereo-microscope. Counts of annual rings were repeated until the number of rings counted could be replicated a minimum of three times. Any significant anomalies in the annual rings, such as scars or distinctly wide or narrow rings, were recorded including the year(s) in which they occurred.

4.5.2 Chronology development

The primary objective in developing tree-ring chronologies is to develop stationary time-series of dimensionless ring-width indices that are directly comparable and appropriate for statistical analysis (Fritts 1976). Chronology development generally follows a three-step procedure: (*i*) *crossdating*: used to check the quality control of the ring-width measurement data for possible dating or measurement problems; (*ii*) *standardisation*: individual ring-width time-series are transformed into stationary time-series with no autocorrelation, stable variance, and a common mean; and (*iii*) *master chronology development*: the transformed time-series are combined into a single representative site chronology.

4.5.3 Crossdating

Crossdating provides the means to assign an absolute date to each ring in a tree-ring series of unknown age by matching the undated ring-width patterns to those of a known dated series (Fritts 1976). Crossdating is possible due to interannual variations in ring-widths produced by trees growing in an area that are influenced by the same limiting growth factors (Fritts 1976).

Each time-series of measured ring-widths were crossdated to a series of narrow marker rings through a preparatory visual examination (Figure 4.7). The crossdated time-series were then quality checked using the International Tree-Ring Data Bank (ITRDB) software program COFECHA to create a master ring-width chronology for each site (Holmes 1983, 1999). This program highlights segments of a ring-width series that correlate poorly with corresponding segments in the master chronology due to natural irregularities (missing or false rings) and human error. Erroneous segments can then be corrected (*i.e.*, remeasured) or deleted from the data set until a statistically significant master chronology is produced (Holmes 1983, 1999).

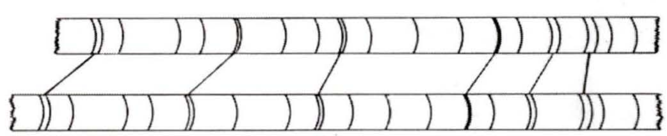


Figure 4.7. Example of crossdating based on distinct marker years

In addition to verifying the measurement data and crossdating among individual ring-width series, COFECHA provides three descriptor values that describe the data set making up the site chronology: series intercorrelation, mean sensitivity, and autocorrelation. Series intercorrelation is a measure of the common signal (degree of homogeneity) contained in a group of trees from a given site. Possible values range from +1.0 for a perfect positive correlation, to a value of -1.0, which indicates a perfect negative correlation. For the 50-year time intervals used in this study, correlation values greater than 0.3281 are significant at the 99 percent confidence level (Holmes 1983, 1999).

Mean sensitivity is a measure of the relative difference in ring-width from one year to the next (Fritts 1976). Values range from 0.0 when there is no difference between adjacent ring-widths (complacency), to 2.0 where a zero value occurs next to a nonzero value (extremely sensitive).

Autocorrelation is a measure of the effect growth in a given year has on growth in the following year. A value of 0.0 indicates no autocorrelation in the time-series, whereas a value of 1.0 indicates that growth in one year completely dictates growth in the following year (Holmes 1983). Ideally, a chronology with high intercorrelation and sensitivity values, in conjunction with a low autocorrelation value, has high potential for dendroclimatological studies (Colenutt and Luckman 1991).

4.5.4 Standardisation and Development of a Master Chronology

Standardisation refers to the estimation and elimination of non-climatic age-related growth trends in the ring-width time-series. In general, ring-widths are widest near the base and centre of the stem and decrease with increasing age and height of the tree. Such changes produce a downward trend in ring-width, resulting in a non-stationary time-series with a declining mean and variance due to intrinsic factors such as tree age and geometry (Fritts 1976). Standardisation allows individual ring-width series obtained from trees of differing ages and growth rates to be averaged into a single dimensionless ring-width index, or site chronology. This procedure reduces the biological-age factor in ring-width variation so that it reflects the extrinsic (environmental) constraints on growth.

The annual radial growth of a tree can be explained with the help of a simplified linear aggregate model (Cook 1987, 1990).

$$R_t = A_t + C_t + \delta D1_t + \delta D2_t + E_t \qquad (4.1)$$

where R_t is the observed ring-width at time t; A_t is the age-size related trend in ring width; C_t is the climate related signal; $D1_t$ and $D2_t$ are local endogenous and stand-wide exogenous disturbance signals respectively; and E_t is the unexplained variability not related to the other signals. The δ associated with $D1_t$ and $D2_t$ is an indicator variable, indicating the presence ($\delta = 1$) or absence ($\delta = 0$) of either the disturbance signal.

The age-related growth trend (A_t) is usually removed with a two-step procedure that involves fitting a combination of two 'best-fit' detrending curves to the ring-width data. The first curve is usually a negative exponential curve that fits the early juvenile portion of the series well, but often under or overestimates the middle or end of the series associated with maturity and old age (Cook and Kairiukstis 1990). The second curve, often a cubic smoothing spline, is then fit to each series to remove $D1_t$ and $D2_t$ related variability not removed by the first curve. The second curve often fits the latter portion of the series well, but is usually not sufficiently flexible to fit growth trends in the early years of the series. Measured ring-widths are then converted to ring-width indices (I) by dividing the observed ring-width (RW) by the expected growth (EG) from the detrending curves as follows:

$$I = \frac{RW}{EG} \tag{4.1}$$

The ITRDB program ARSTAN (Cook and Holmes 1986, 1988) was used to detrend and standardize each ring-width time series into a stationary dimensionless index. Each series of ring-widths was evaluated and a combination of two user-defined detrending curves were applied to maximize the signal to noise ratio. All of the cross-dated, detrended indices were averaged into single master site chronologies using a bi-weight robust mean (Cook and Holmes 1986, 1988). Robust means discount the influence of outliers and consequently result in stronger estimations of the chronology. The convention in dendrochronology is that all observed ring-widths greater than 6 standard deviations are excluded from master chronology. Estimation of the mean is an iterative process that runs until the estimated mean does not change by more than 10^{-3} (Cook *et al.* 1990)

4.6 Lichenometry

4.6.1 Introduction

The use of lichen to determine the age of a substrate (lichenometry) has become widely adopted since its inception in the 1950's (Beschel 1973). Lichenometry has proven to be a particularly useful technique for dating Neoglacial deposits in arctic and alpine environments where suitable materials required for alternative dating techniques (*i.e.*, dendrochronology) are not often present (Beschel 1973; Innes 1985, 1988; McCarroll 1994).

Lichen can be classified into three general categories based on their morphology; crustose (thin, flat, disc-like crust), foliose (flat, with leaf-like lobes), and fruticose (upright, branched, and bushy) (Galun 1988). Only the crustose lichens are used extensively in lichenometry, as their area commonly increases radially with age. This size-age relationship forms the basis of lichenometry - the use of lichen size as an indicator of substrate age.

4.6.2 Moraine Dating

Lichenometric dating is based on the assumption that the largest lichen growing on a substrate is the oldest, and therefore represents the first individuals to colonise the surface in those locations with optimum growth conditions (McCarroll 1994). If the size-age relationship (*i.e.*, growth rate) of a particular lichen species is known, the minimum age of a surface can be inferred from the size of the largest lichens present on that surface. All other thalli are considered either late colonizers or growing in sub-optimal conditions (Beschel 1973).

Identification and classification of lichen used in dating are problems that have received considerable attention. The most commonly used species in lichenometric studies are those within the *Rhizocarpon* subgenus *Rhizocarpon*, a group of greenish-yellow crustose lichens. This lichen is favoured for dating purposes given its slow and relatively circular growth, extreme longevity, global distribution, and abundance in arctic and alpine environments (Carrara and Andrews 1973; Aplin and Hill 1979; Porter 1981b; Benedict 1988; McCarthy and Smith 1995; Calkin *et al.* 1998).

Lichen can often be difficult to identify down to the species level in the field. One of the main difficulties in comparing growth rates, and ultimately landform dates, is the lack of differentiation between species within the subgenus *Rhizocarpon*. This would not be a problem if all species in the subgenus *Rhizocarpon* grew at the same rate. However, some species have faster growth rates, wheras others colonise a substrate more rapidly or compete more aggressively than others in a mature lichen community (Calkin and Ellis 1980; Benedict 1988). Nevertheless, although field classification can be difficult, it is usually possible to identify *Rhizocarpon* down to the section level in the field, and down to the species or subspecies level in a laboratory setting (Table 4.5) (Innes, 1985).

Methodologies and basic assumptions used in lichenometry can be divided into two approaches: (*i*) the 'statistical' approach, and (*ii*) the traditional, or conventional approach (McCarthy 1999). The statistical approach is based on the fundamental

Table 4.5. Recommended *Rhizocarpon* classification levels

Classification Level	Description
1. Rhizocarpon subgenus	-identification is limited, referring to the entire yellow-green *Rhizocarpon* subgenus
2. Section level	-identification has been made to the section level, *e.g.*, *Rhizocarpon* section *Alpicola* or *Rhizocarpon* section *Rhizocarpon* (usually possible in the field)
3. Species level	-identification has been made to the specific label, *e.g.*, *Rhizocarpon geographicum* (L.) DC (usually involves laboratory determination)

from Innes (1985)

assumptions that conditions affecting growth are constant and affect all thalli, and that lichen populations have a statistically normal distribution of thallus sizes (Innes 1985; McCarthy 1999). This approach allows for the use of parametric tests and regression analysis in growth curve construction, the ability to date diachronous surfaces, and the ability to apply statistically defined confidence intervals around surface ages (Matthews 1974; McCarroll 1994; Bull and Brandon 1998; McCarthy 1999).

However, McCarthy (1999) shows that various statistical approaches are based on questionable assumptions and, therefore, cautions against placing "great trust" on age predictions from statistical techniques that are at best weakly supported by biological data. The fact that lichenometry is used to date moraines of different ages indicates in itself changing environmental conditions over time. McCarthy (1999) also notes that normality in lichen size is seldom met. Such apparent violations of the underlying assumptions, combined with the statistical requirement of large data sets and a random sampling design, limits the use of this approach.

More commonly used are traditional lichenometric methods based on the sampling of the largest lichen, or average of the largest lichens, growing on a substrate.

With this method, a number of growth indices are used to characterize the size of a lichen thallus; the two most common methods are the largest inscribed circle (LIC), and the longest axis of the thallus (Figure 4.8) (Innes 1985).

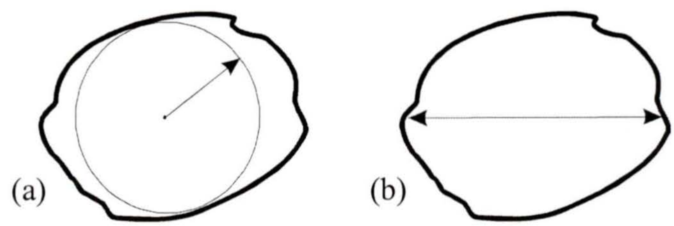


Figure 4.8. Two common lichen measuring methods: (a) largest inscribed circle (LIC), and (b) longest axis.

There are two main sampling approaches regarding the number of thalli used in dating a deposit: (*i*) the single largest thallus, and (*ii*) the average of several of the largest thalli (Innes 1988). The single largest lichen method is often used in areas where there is considerable variation in the substrate, whereas the averaging method is used where there is little substrate variability (Gordon and Sharp 1983).

Some studies suggest that only the single largest thallus on a surface is representative of the age of the deposit, as there are frequently only a few lichen that are substantially larger than others growing on the same surface (Webber and Andrews 1973; Benedict 1993). These thalli are believed to have inhabited the most favourable microsites and represent the closest to optimal growth. However, several studies argue that the single largest thallus is not truly representative of the sampled area and suggest adopting a sampling strategy based on an averaging procedure (Matthews 1974; Locke *et al.* 1979; Matthews and Shakesby 1984; Innes 1985, 1988). The assumption that the largest lichen is the first coloniser growing under optimal conditions is replaced with the assumption that the mean size of the largest few lichens is a function of the substrate age (Matthews 1974).

Obtaining a reliable surface date is highly dependent on the size of the area sampled, with the probability of finding the largest lichen on a substrate surface increasing with the area searched (Innes 1984). It is, therefore, important to search a sufficiently large area on each surface. If time permits, the optimal sampling strategy is to search the entire deposit (Innes 1985, 1988). If it is not possible to search the entire area, an alternative sampling strategy can be adopted. Sampling methods range from the 'random walk' in which the largest lichen are measured during a fixed time interval, to an area-based quadrat method (Innes 1988). Anywhere from 5 to 10 quadrats with an area ranging from 25 m^2 to 600 m^2 have been used in previous studies (Webber and Andrews 1973; Locke *et al.* 1979; Gordon and Sharp 1983; Innes 1985).

4.6.3 *Rhizocarpon* Dating Curve

As a relative-age dating technique lichenometry is based on the assumption that a given index of lichen size increases with time. Based on this assumption, a surface with larger diameter lichen is considered older than a corresponding surface with smaller diameter lichen. However, as a calibrated age-dating technique, lichenometry is based on the additional assumption that an empirical relationship can be established between lichen size and time since lichen colonisation (*i.e.*, surface stabilisation)

In order to utilise lichen growth as a dating method for surfaces of unknown age, a local growth curve, or dating curve, relating lichen size to substrate age is first required. A line fitted to the largest thalli measured from deposits of different known ages will provide an estimate of the maximum historical growth rates of a given species. This curve can then be used to determine minimum estimates of substrate age within regions that have similar environmental conditions (Grove 1988).

Lichen growth curves can be established through two methods: direct measurements of individual thalli over time, or indirectly, by measuring the diameter of thalli growing on surfaces of known age. In the direct approach, repeat measurements of the same thalli using photographic techniques provide the most accurate means of determining the growth rate over a specific time period (Innes 1985; McCarthy 1999). However, this method requires repeat measurements of the same lichen over extended periods of time, assumes that the growth rate determined during the measurement interval is the same as it has been in the past, and that environment conditions have remained constant over time (Innes 1988).

The majority of dating curves have been established using the indirect method. This involves measuring the largest lichen diameters on substrates of known age, natural or anthropogenic in origin, and plotting size as a function of the surface age. A 'best-fit' curve is then fit to the data, or an envelope curve is placed around the data, which provides a smoothed dating curve relating lichen size to substrate age. Unlike direct measurements, this technique can account for fluctuations in lichen growth associated with changing climate over time (Innes 1985).

Dating control is provided by measuring lichen from historically dated surfaces such as gravestones, buildings, and dams, as well as surfaces that have been dated by dendrochronological methods (*e.g.*, McCarthy and Smith 1995). Greater confidence can be placed in a dating curve that has been constructed from a larger number of control points. Ideally, the same sampling strategy that was used to date the substrate of unknown age should be used to construct the dating curve. This may not be possible where small surfaces such as tombstones are used as control points; as tombstones generally have a small area, a growth rate derived from these surfaces may result in an underestimation of the maximum growth rate. However, if a sufficiently large graveyard is sampled and the dating curve is verified with independent data not used in the curve construction, a reasonable estimate of the growth rate can be established (Innes 1988). As lichen growth rates can vary greatly from one site to another, the growth rate for a given species should ideally be derived from local surfaces of known age in a similar environment as the undated landforms.

4.6.4 Lichenometric Methods in this Study

The moraine complexes at each site were small enough to be searched in their entirety and thus, subplots were not required. The longest axis sampling method was used, and the 'A' and 'B' axes of notable lichen were measured to the nearest ±0.01 mm using digital callipers (Figure 4.9). Sampling was limited to near circular lichen to avoid anomalously large thalli as a result of bird perches or coalescence. Lichen identification was limited to the subgenus level in the field and location of the lichen were mapped on local site maps.

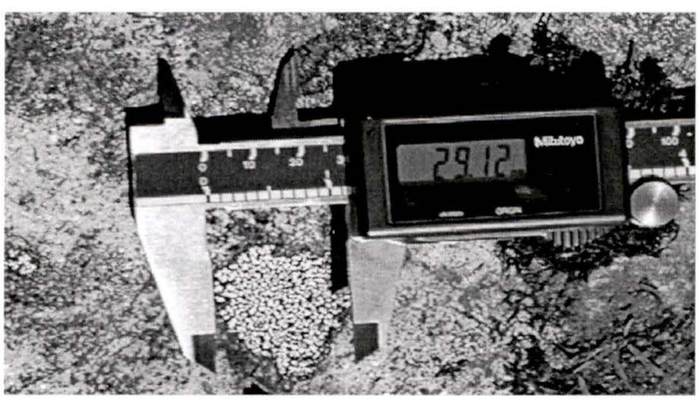


Figure 4.9. Lichen measurement using digital callipers.

A local growth curve was constructed from *Rhizocarpon sp*. measured on gravestones from the Cumberland Community Cemetery (49° 37' N, 125° 02' W; 160 m asl). Additional control points were provided by lichen growing on tree-ring dated moraines at the two glacier study sites. Like the moraine lichen, the primary and secondary axes of the largest lichen from each gravestone were measured to the nearest ± 0.01 mm and the date of the surface recorded.

CHAPTER 5. DATING CONTROLS

5.1 Introduction

This chapter presents the results of investigations needed to develop the LIA history of Colonel Foster and Septimus glaciers: (*i*) mountain hemlock growth rate and sampling height correction; (*ii*) ecesis interval; (*iii*) development of the master ring-width chronology; (*iv*) determining the climate-tree relationship; and (*v*) construction of the local lichen dating curve.

5.2 Dendrochronological Controls

5.2.1 Growth Rate/Sampling Height-Age Correction

The mountain hemlock coring height-age correction factor was established from a sample of ten young mountain hemlocks from the Colonel Foster Glacier forefield. The height of ten saplings and seedlings (*ca.* 1 m tall) was measured and the trees were cored at the root crown in order to estimate a general age-height relationship (Table 5.1).

Table 5.1. Mountain hemlock apical growth rates

Sample	Height (cm)	Age (yrs)	Growth Rate (cm/yr)
CFGR-1	94.0	30	3.1
CFGR-2	125.0	42	3.0
CFGR-3	112.5	29	3.9
CFGR-4	75.0	31	2.4
CFGR-5	105.0	33	2.8
CFGR-6	90.0	41	2.1
CFGR-7	101.0	33	3.1
CFGR-8	104.0	44	2.3
CFGR-9	98.5	46	2.1
CFGR-10	96.5	39	2.5
Mean	**100.2**	**37.8**	**2.7**

From the trees sampled, it was shown that mountain hemlock at this site required between 29 to 46 years to reach to *ca.* 100 cm in height, with a mean of 38 years. Growth rates ranged from 2.1 to 3.9 cm/yr, with a mean of 2.7 cm/yr (0.37 yrs/cm). These values are similar to mountain hemlock growth rates determined 80 km to the southeast of Strathcona Park in a previous alpine study by Laroque *et al.* (2001).

For trees cored above the root crown, a sampling height correction factor of 0.37 yrs/cm was applied to the age of the tree corresponding to the height at which the core was extracted. As most cores were taken in the bottom 5 cm of the stem with the increment borer angled down towards the root crown, the sampling height correction factor was limited to only a small number of trees.

5.2.2 Ecesis Interval

A local mountain hemlock ecesis interval was determined from 13 mountain hemlock growing on the bedrock surface of the 1946 landslide scar, 100 m below the Landslide Lake (Table 5.2).

Table 5.2. Mountain hemlock ecesis samples from lower Landslide Lake

Sample	Ring count	Pith Date	Ecesis Interval (Yrs)
CFEC-01	36	1962	16
CFEC-02	37	1961	15
CFEC-03	33	1965	19
CFEC-04	***48***	***1950***	***4***
CFEC-05	43	1955	9
CFEC-06	41	1957	11
CFEC-07	43	1955	9
CFEC-08	40	1958	12
CFEC-09	33	1965	19
CFEC-10	44	1954	8
CFEC-11	38	1960	14
CFEC-12[1]	46	1952	6
CFEC-13[1]	42	1956	10

bold = ecesis value used in this study; [1] denotes increment core

The oldest tree (CFEC-04) sampled from this site had a ring count of 48 years; a correction for sampling height was not needed as the tree was sampled close enough to the ground (<1cm). The basal pith date of 1950 provides an ecesis interval of 4 years for mountain hemlock in this area. After tree ages were corrected for sampling height and pith distance, a four-year ecesis interval was added to the age of all trees used in dating moraine surfaces at Colonel Foster and Septimus glaciers.

5.2.3 Mountain Hemlock Master Chronology Development

Three mountain hemlock ring-width chronologies were developed in this study; two individual site chronologies from Colonel Foster and Septimus glaciers, and a regional Strathcona Park master chronology consisting of tree-ring series from both sites (Figure 5.1). At Colonel Foster Glacier, 44 increment cores were extracted from 22 mountain hemlock trees in the adjacent forest 10-50 m north of the moraine complex. At Mount Septimus, 62 cores from 31 mountain hemlock were sampled above the north side of Cream Lake in 1995. A summary of the chronology statistics is given in Table 5.3.

Table 5.3. Summary statistics for Strathcona Park tree-ring chronologies

	Col. Foster	**Septimus**	**Strathcona Master**
Number of trees	20	21	41
Number of cores	40	35	75
Interval (yrs)	1673-1998 (326)	1412-1995 (584)	1412-1998 (587)
Mean core length (yrs)	251	359	301
Series Intercorrelation	0.671	0.638	0.618
Mean Sensitivity	0.253	0.268	0.261

The high series intercorrelation and mean sensitivity indicate that: (*i*) mountain hemlock from Strathcona Park are responding homogeneously to a similar environmental forcing; (*ii*) trees from both sites can be combined into a single, regional master chronology; and (*iii*) the trees should have good dendroclimatic utility.

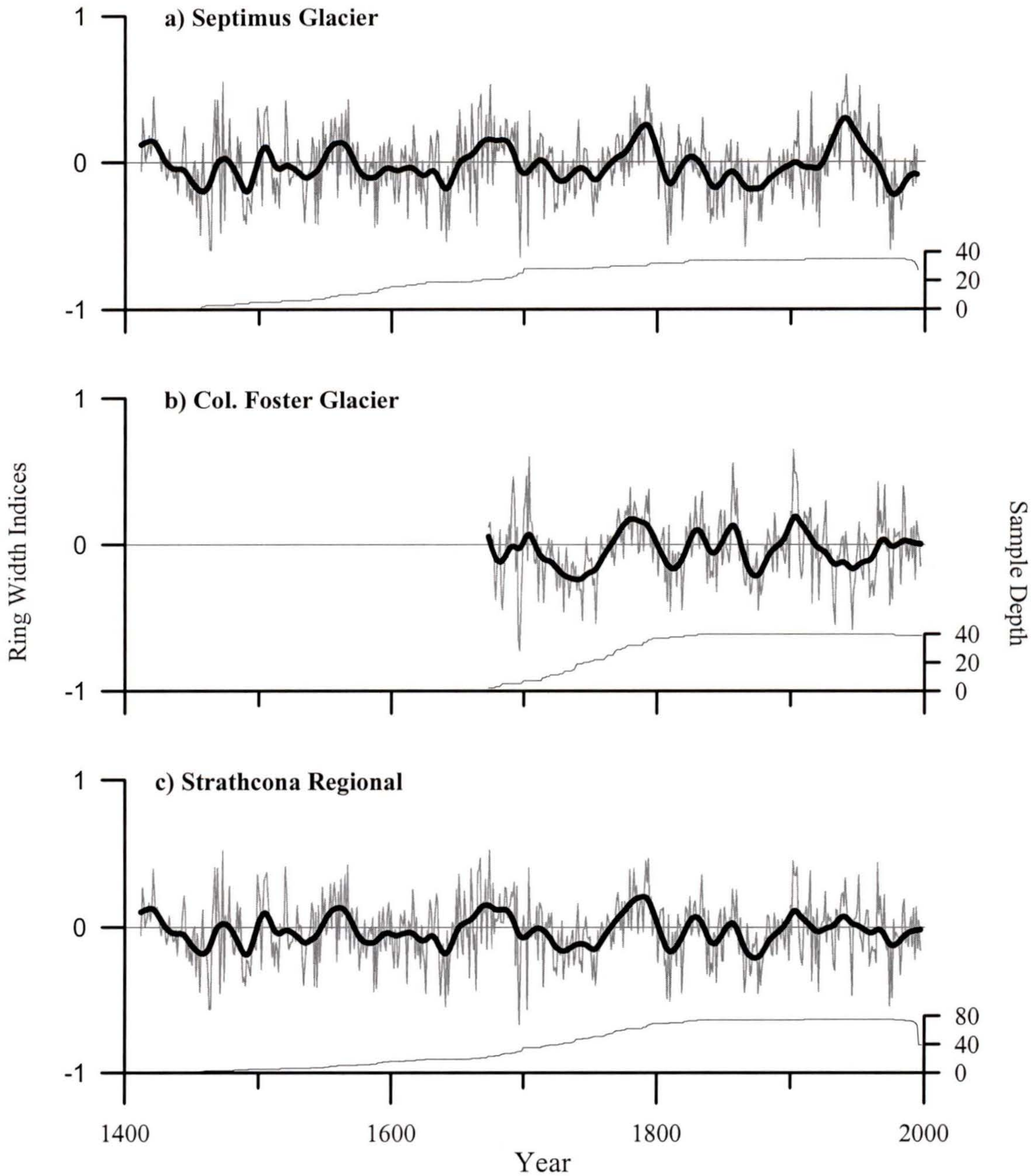


Figure 5.1. Indexed mountain hemlock chronologies for Strathcona Park: (a) Septimus Glacier, (b) Colonel Foster Glacier, and (c) Strathcona Regional master chronology. A 25-year cubic smoothing spline (bold line) is fit to the data to emphasize the long-term trends. The sample depth, or number of cores contributing to the annual index, is also given for all chronologies.

Cores contributing to the Strathcona regional chronology span the interval from 1412 to 1998 AD. The last 250 years of the chronology are represented by ≥50 cores. Prior to this period, the number of samples contributing to the chronology progressively decreases from 16 cores at 1600 AD to ≤8 cores from 1550 to 1412 AD (Figure 5.1c).

Growth trends in the early part of the Strathcona regional chronology (1412 to 1550 AD) are quite variable and likely an artifact of the limited sample depth. Significant intervals of reduced growth rates occur from 1575 to 1650, 1690 to 1765, 1800 to 1820, and 1835 to the 1860's AD. Less significant episodes of reduced growth occur in the 20th century, from 1915 to 1930 and again in the 1970's. Notable intervals of above average growth occurred in the late 1600s, late 1700's, and early 1900's AD (Figure 5.1c).

5.2.3.1 Mountain Hemlock-Climate Relationship

The Strathcona regional master chronology was first analysed to determine if the climatic variables responsible for limiting tree growth at the two sites were consistent with other mountain hemlock dendroclimatological studies in the PNW. The software program PRECON 5.17c (Fritts 1976; Fritts and Wu 1986; Fritts *et al.* 1991; Fritts 1998) was used to identify relationships between the standardised growth index (regional master chronology) and records of monthly average temperature and monthly total precipitation. PRECON recalculates matrices of climatic data using principal components analysis (PCA) to form new orthoganalized variables that maximize the variance in the climatic factors influencing tree growth (Fritts *et al.* 1971; Blasing *et al.* 1984). The new variables are entered into a stepwise multiple regression procedure where the regression coefficients are multiplied by the principle components of climate to obtain a new set of regression coefficients related to the original monthly precipitation and temperature variables. The new coefficients express the relative importance of each monthly climate variable to the tree-ring chronology (Fritts *et al.* 1971; Fritts *et al.* 1991). The output from this analysis is graphically represented as a response function demonstrating the relationship between variations in annual tree growth and the limiting climate variables (Fritts 1976; Cook and Kairiukstis 1990). Climate data (1945-1994) are from the Meteorological Service of Canada (MSC) meteorological Station 'A' at Comox, B.C. (49°43' N, 124°54' W, 24 m asl), *ca.* 35 km east of Strathcona Park (Figure 3.1).

Response function analysis identified climate variables that have significant associations with radial growth of the Strathcona regional mountain hemlock chronology. Figure 5.2 shows the growth response of the mountain hemlock chronology to temperature and precipitation data from the MSC Comox 'A' climate station.

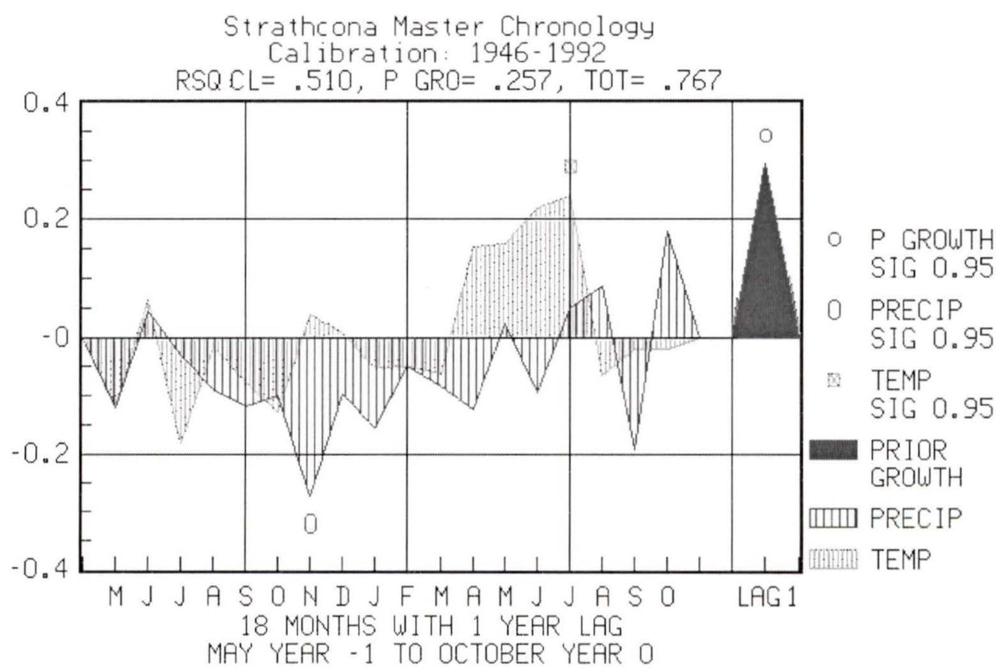


Figure 5.2. Results of the response function analysis for the Strathcona regional chronology.

Figure 5.2 illustrates the amount of variation in ring-width explained by temperature and precipitation over an 18-month interval, from May of the previous year to October of the growth year. An 18-month growth period was used to capture the annual growth signal, as high-elevation trees are often influenced by growth in the preceding year (Colenutt and Luckman 1991).

The response function analysis shows that mountain hemlock growth at these sites is most strongly correlated to mean July temperature of the growing season, and to precipitation in the preceding November (Figure 5.2). These results show that 77 percent of the ring-width variance detected in the Strathcona mountain hemlock chronology (1946-1992) is attributed to climate variables in the present year (51 %), and tree growth (21%) in the previous year.

Previous studies by Graumlich and Brubaker (1986) and Smith and Laroque (1998b), show that the effect of monthly temperature and precipitation on mountain hemlock growth is not a simple linear relationship, and that annual spring snowpack depths play a significant role in governing mountain hemlock growth. Such a non-linear relationship makes climatic reconstruction from mountain hemlock a difficult and complex procedure. Nevertheless, a simplified relationship between climate and tree-growth was used to put the glacier record in the context of a qualitative proxy climate record. Based on previous dendroclimatological studies on Vancouver Island and the surrounding Pacific Northwest (Brubaker 1986; Peterson and Peterson 1994; Smith and Laroque 1998b; Gedalof and Smith 2001a; Peterson and Peterson *in press*), in conjunction with the mountain hemlock-climate relationship determined in this study, periods of below average mountain hemlock growth are considered to reflect years with shorter, cooler summers and increased winter precipitation (*i.e.*, greater spring snowpack depths). Conversely, periods of increased mountain hemlock growth is considered to reflect years with warmer, drier summers and drier winter conditions (*i.e.*, decreased spring snow packs).

5.3 Lichenometric Controls

5.3.1 Strathcona Lichen Growth Curve

Although a number of *Rhizocarpon* growth-curves are available for dating landforms, each one is a product of local environmental factors and, therefore, their character will vary between locations (Figure 5.3). Only two of the existing growth

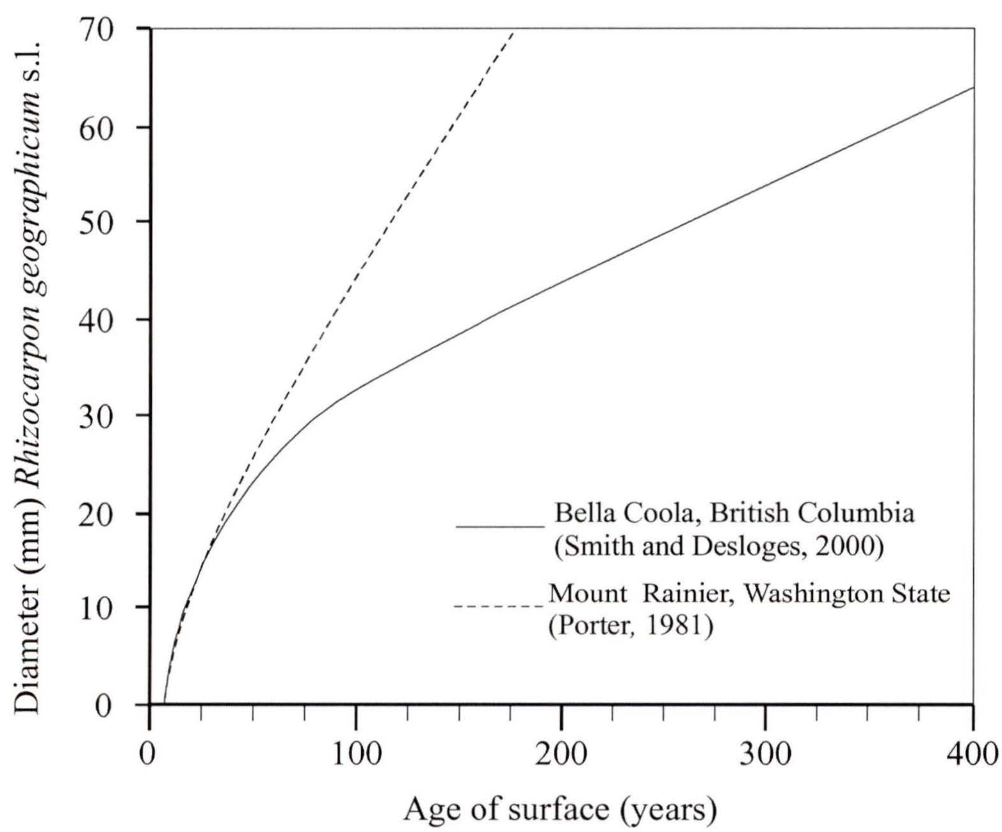


Figure 5.3. An example of two different *Rhizocarpon geographicum* growth curves from the Pacific Northwest.

curves in the Pacific Northwest, Bella Coola (Smith and Desloges 2000) and Mount Rainier (Porter 1981b), may have been applicable to studies on Vancouver Island (Figure 5.3). However, as the two growth curves display markedly different growth characteristics, a local growth curve was constructed that would better represent lichen growth trends in Strathcona Park.

The local *Rhizocarpon* growth curve was developed primarily from gravestones in the Cumberland Cemetery, with four additional control points added from tree-ring dated moraines at Colonel Foster and Septimus glaciers (Table 5.4).

Table 5.4. Control points for Strathcona lichen dating curve

Sample #	Thalli Diameter (mm)	Date[1]	Sample #	Thalli Diameter (mm)	Date[1]
1	11.8	1988	14	29.7	1943
2	10.1	1987	15	31.0	1934
3	11.9	1983	16	36.5	1932
4	13.4	1982	17	32.7	1931
5	15.7	1981	18	34.1	1929
6	18.7	1978	19	34.0	1926
7	22.2	1976	20	36.0	1916
8	23.9	1971	21	36.5	1915
9	22.6	1969	22	40.2	1909
10	22.2	1967	23	42.7	<1898[2]
11	20.9	1965	24	78.0	<1775[2]
12	25.6	1958	25	96.6	<1708[2]
13	27.1	1953	26	97.4	<1710[2]

[1] All 20th century dates are taken from gravestones at the Cumberland Cemetery
[2] 18th and 19th century dates are minimum dates from tree-ring dated moraines at Colonel Foster and Septimus glaciers

Two sets of independent control data used to check the growth curve are from the 1958 Strathcona Dam (100 thalli) (50° 00' N, 125° 32' W; 201 m asl), and the 1946 landslide scar below Landslide Lake (50 thalli) (49° 29' N, 125° 32' W; 905 m asl) (Table 5.5). The resulting *Rhizocarpon* growth curve spans 290 years from 1708 to 1998 AD

(Figure 5.4). Lichen samples were identified down to the species level as *Rhizocarpon geographicum* with assistance from the University of Victoria Biology Department.

Table 5.5. Independent control data for Strathcona lichen growth curve

Site	Surface date	Elevation (m asl)	# of lichen sampled	Average of 5 largest (mm)	Single Largest (mm)
Strathcona Dam	1958	201	100	26.0	26.6
North facing			*50*	*25.6*	*26.6*
South facing			*50*	*25.1*	*25.9*
Landslide Scar	1946	880 – 905	50	-	-
Bedrock			*12*	*32.0*	*32.2*
Boulders			*38*	*37.0*	*38.2*

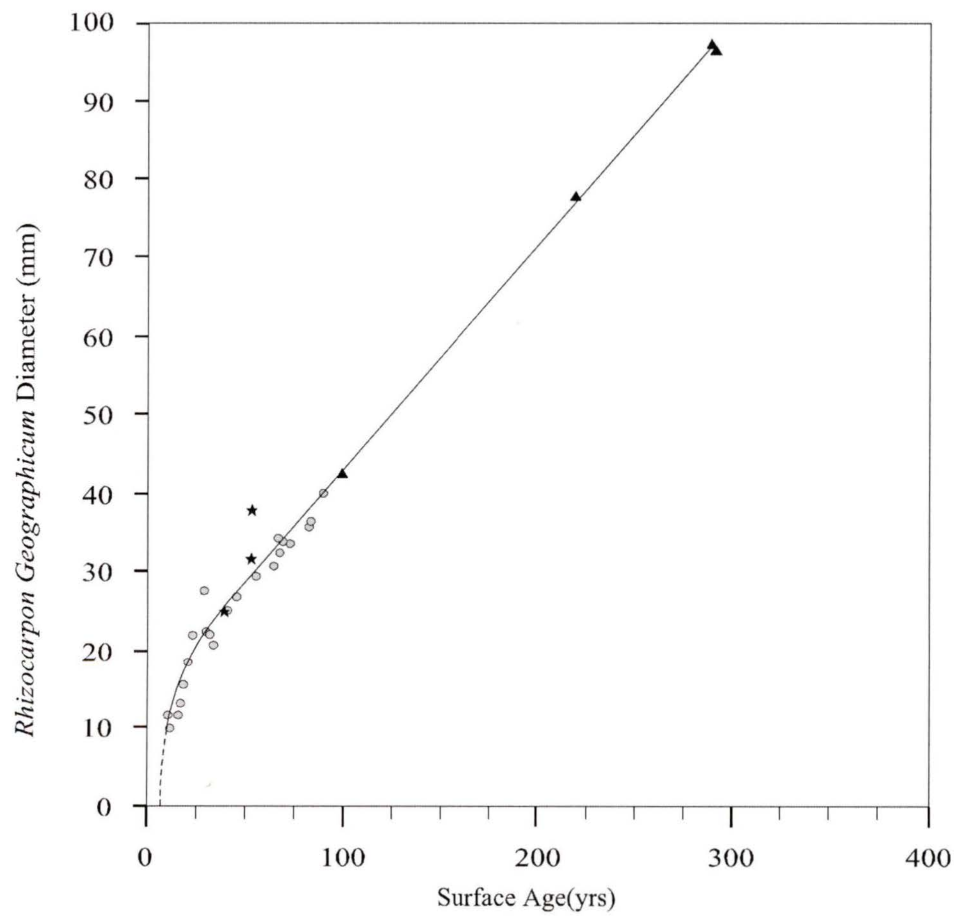


Figure 5.4. *Rhizocarpon geographicum* dating curve for Strathcona Park. Circles are tombstone dates, triangles are tree-ring dated surfaces, and stars are independent control points from the 1958 Strathcona Dam and the 1946 landslide scar.

All lichen sampled from the Strathcona Dam are growing on boulders *ca*. 30 – 150 cm in diameter. Lichen sampled from the north side of the dam have a slightly faster growth rate than those sampled from the south side. A slight difference in growth between lichen from the north and south side of the dam is likely due to the presence of a large water source to the north (Buttle Lake). However, the difference in lichen size between the north and south sides of the dam is minimal, and the largest lichen from entire dam was taken as being representative of the 1958 surface.

Of the 50 thalli sampled along the landslide scar, only 12 were recorded from the bedrock surface. The majority (76%) of the landslide lichen were sampled from various sized boulders (25 - 120 cm diameter) sitting directly on the bedrock. A marked difference in lichen size, and thus growth rates, between the two surfaces is evident. The largest lichen diameter from the bedrock (32.2 mm) was significantly smaller (6.0 mm) than the largest lichen (38.2 mm) from the boulders of the same age (52 years). Lichen from both surfaces were retained as control points in the lichen growth curve.

CHAPTER 6. LIA RESULTS FROM STRATHCONA PARK

6.1 Introduction

This chapter provides moraine-dating results from Colonel Foster and Septimus glaciers, and puts them in the context of previous LIA studies on Vancouver Island, the Pacific Northwest, and Canadian Rocky Mountains. Eighty-three living trees were sampled on moraines in the forefields of both Colonel Foster (68 trees) and Septimus Glacier (15 trees). An additional 337 lichen were measured; 73 from the western moraines at Colonel Foster Glacier, and 264 from moraines in the Septimus Glacier forefield. Although dendroglaciological methods can provide calendar-year precision, limitations associated with lichenometry and tree-ring dates result in uncertainties of *ca.* ± 5 years for moraine surfaces in optimal circumstances. Dating uncertainties, however, are often closer to ± 10 years and may even be greater on older surfaces (Luckman, 2000).

6.2 Colonel Foster Glacier

Fifty-four trees were sampled from eight moraines (Moraines 1 - 8) in front of Colonel Foster Glacier. Moraines 1 to 5 of the inner suite were dated primarily through the sampling of live trees growing on their surface, with supplementary dating provided by lichenometry. Additional dating control for Moraine 5, as well as the limiting date for Moraine 6, was provided by an *in situ* stump from Moraine 6 that had been partially buried by debris from the emplacement of Moraine 5. Moraines 6, 7 and 8 of the outer moraine suite were otherwise unable to be dated adequately due to relatively youthful trees growing on their surface, and an understory growth that limited lichen colonisation.

6.2.1 Moraine dates derived from live trees

Table 6.1 summarises the oldest trees from each moraine that were used to develop the moraine chronology. The relative location of each tree on the moraine is given (*i.e.*, crest, proximal or distal face), as well as the corrected tree age and the corresponding limiting age of the surface upon which they were growing. Complete results from all tree samples collected at this site are presented in Appendix 1.

Table 6.1. Tree ages and minimum dates for moraines at Colonel Foster Glacier

Moraine and Sample ID	Location on the Moraine[1]	Ring Count	Pith Error	Estimated Tree Age	Minimum Surface Date[3]
1. Inner Suite					
Moraine 1					
98CFWM-46	PF	59	0	59	1935
Moraine 2					
98CFWM-35	PF	96	0	96	1898
Moraine 3					
98CFWM-33	DF	206	10	216	1778
Moraine 4					
98CFWM-23	PF	214	10	224	1770
Moraine 5					
98CFWM-09	DF	286	0	286	1708
2. Outer Suite					
Moraine 6					
98CFWM-08	DF	358	10	403[2]	1591
98CFLOG-01	DF	315[4]	0	> 350[2]	> 1396
Moraine 7					
-	-	-	-	-	-
Moraine 8					
-	-	-	-	-	-

[1] PF = Proximal Face, CR = Crest, DF = Distal Face
[2] Tree age includes a coring-height correction of 35 years
[3] Surface date includes 4 year ecesis interval
[4] Missing outer rings

6.2.1.1 Ring-width anomalies

Figure 6.1 illustrates variations in the ring-width pattern of tree CFWM-07, growing on the crest of Moraine 6 (2 m outside of Moraine 5).

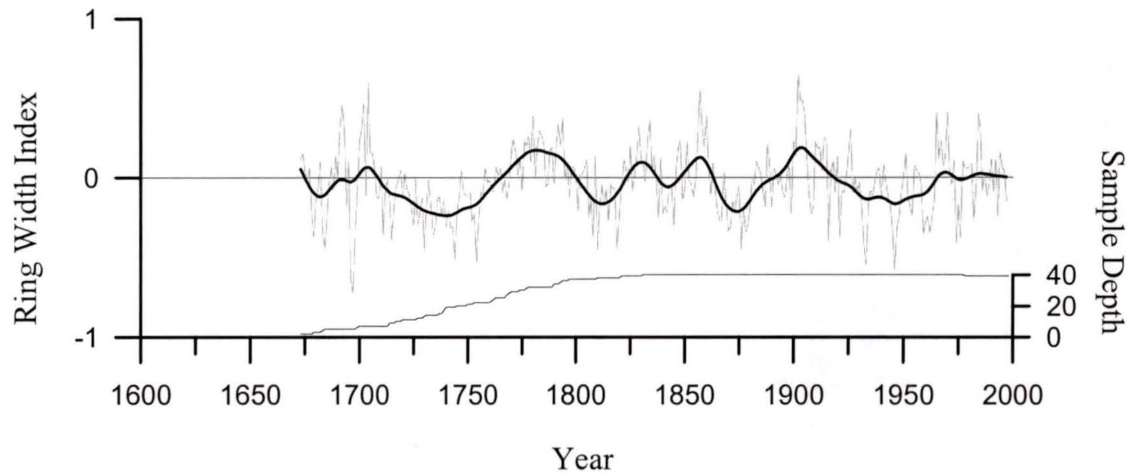


Figure 6.1. Mountain hemlock ring-width chronology from Colonel Foster Glacier. Thin lines are annual values and thick line is a 15-year cubic smoothing spline to emphasise longer-term growth trends. Sample depth is the number of cores contributing to the annual index.

Ring-width patterns of the Colonel Foster mountain hemlock (MH) chronology show a reduction in radial growth in the late 1690's, early 1710's to early 1770s, early 1800's to early 1820's, mid 1830's to mid 1840's, early 1860's to late 1880's, 1930's, mid 1940's, mid 1960s and mid 1970's (Figure 6.1). The low growth rates of the late 1940's are a result of the 1946 earthquake, which is also evident as a scar in many of the tree cores. Two distinct intervals of increased radial growth occur in the late 1700's and early 1900's, bracketing two minor periods of increased growth in the late 1820's and 1850's. A general trend of 'average' growth occurs from mid 1970's on to the present (Figure 6.1).

6.2.2 Moraine dates based on sub-fossil wood

Given the brevity of the local (Colonel Foster) master ring-width chronology, the sub-fossil stump from Moraines 5 and 6 (Figure 6.2) was crossdated using the Strathcona regional master chronology. Of the 315 rings that could be counted and measured, only the outer 122 years of the stump were able to be crossdated with any certainty. The difficulty in crossdating the earlier ring-widths was due to numerous cracks in the sample and distorted ring-width patterns resulting from repeated episodes of reaction wood formation. Crossdating attempts for the earlier years of the stump were also limited by a lack of sample depth in the master chronology prior to 1550 AD.

Figure 6.2. Photograph of the *in situ* stump (CFLOG-01) from the base of Moraine 5 at Colonel Foster Glacier. Arrow in upper right of the picture indicates the general glacier direction during glacier advances.

Table 6.2 summarises the COFECHA crossdating verification results for the last 122 years of the stump cross-section. A strong crossdate was obtained at 1750 AD for the outermost ring ($r = 0.59$, $r > 0.3281$ is significant at the 99% confidence level), providing a germination date of 1435 AD. Corrections for sampling height and ecesis interval suggest a minimum surface date of 1396 AD for Moraine 6 (Table 6.1). In addition to pith and kill dates, a number of distinctive event indicators (*i.e.*, scars and reaction wood episodes) in the stump cross-section were observed and dated (Table 6.3; Figure 6.3).

Table 6.2. Crossdating results of CFLOG-01 for the last 122 years

	50-yr segment (AD)	Correlation
	1628-1677	0.54
	1653-1702	0.65
	1678-1727	0.62
	1701-1750	0.55
Total interval:	**1628-1750**	**0.59**

Table 6.3. Ring-width anomalies and event indicators for CFLOG-01

Event Indicator	**Event Date(s) AD**
Concentric Growth	Pith to 1511
Reaction Wood	1512-1523, 1620-35, 1681-1750
Scar	1619, 1631, 1681, 1696, 1736, 1750[1]

[1]kill date

An initial period of concentric growth occurred from germination until 1511 AD, at which time there was a distinct change to eccentric growth patterns. Periods of pronounced reaction wood occur in the intervals 1513 to 1523, 1620 to 1635, and 1681 to 1750 AD, whereas major scars are dated to 1619, 1681, 1696, and 1736 AD.

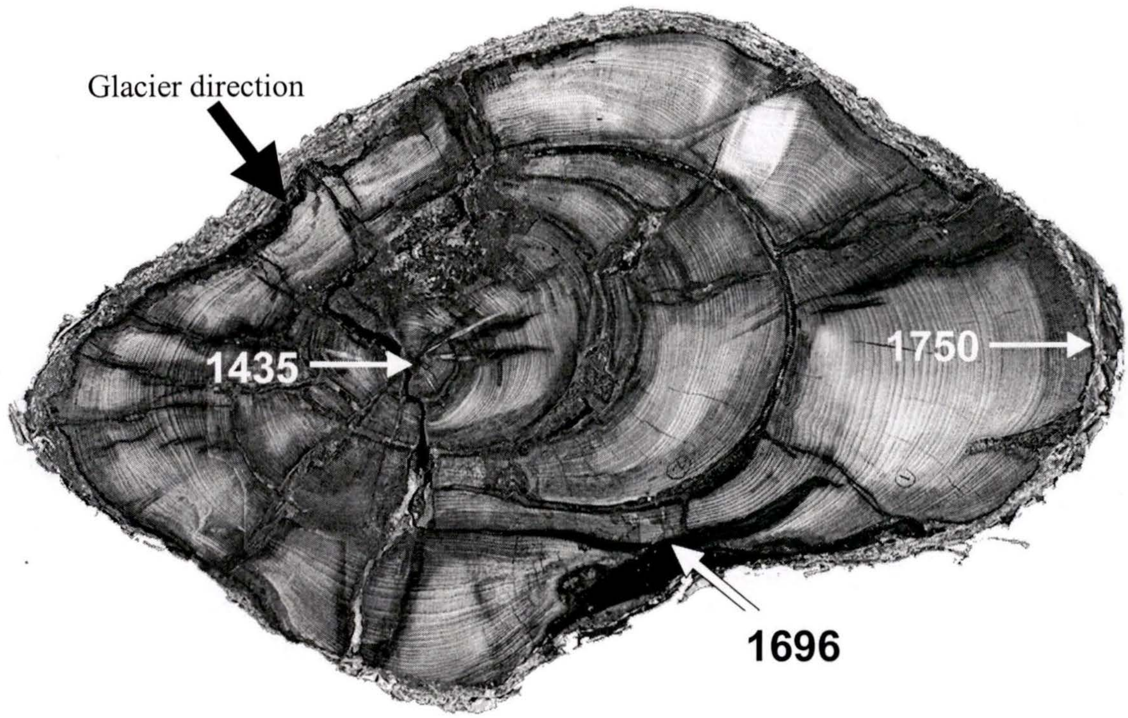


Figure 6.3. Cross-section of the stump (CFLOG-01) located at the base of Moraine 5 with key event dates and general glacier direction highlighted.

6.2.3 Lichenometric moraine dates from Colonel Foster Glacier

Additional moraine stabilisation dates were determined from measurements of 143 *R. geographicum sp.* Table 6.4 summarises the results of the lichen measurements, including the number of lichen sampled on each moraine, the size of both the single largest and the average of the five largest thalli sampled, and the corresponding minimum surface date interpolated from the Strathcona dating curve. Lichen measurements were not collected on Moraines 6 – 8 of the outer moraine suite due to the dense forest cover and lack of lichen due to competition from understory growth.

Table 6.4. Lichen dating results for the moraines at Colonel Foster Glacier

Moraine	# Thalli Measured	Largest Single Lichen (mm)	Minimum Surface Date[1]	Average Largest 5 Lichen (mm)	Minimum Surface Date[2]
1	17	39.3	**1912**	37.1	1920
2	8	33.5	**1933**	30. 9	1940
3	48	78.0	**1775**	75.5	1784
4	7	57.4	**1847**	53.1	1863
5	63	96.6	**1708**	92.3	1753

[1] Minimum surface date based on single largest lichen diameter
[2] Minimum surface date based on average of 5 largest lichen diameters
BOLD indicates dates used for moraine chronology

The mean diameter of the five largest thalli were between 3.2 and 7.7 percent smaller than the single largest diameter, with an average difference of -5.7 percent. The small difference between the two measurement methods indicates that the single largest lichen from each moraine is not anomalously large (*i.e.*, bird perch or coalesced) and, therefore, can be used confidently to provide minimum moraine stabilisation dates. Considering the relationship between the two methods, and that the growth curve was developed from single largest thalli measurements, lichenometric moraine dates from Colonel Foster were based on measurement of the single largest lichen.

6.2.4 Interpretation of LIA History at Colonel Foster Glacier

The morphology of the moraine suite and moraine dates provided by both live and dead trees, in conjunction with the ring-width suppression record the Colonel Foster mountain hemlock chronology, suggests three main intervals of glacier activity relating to three episodes of LIA moraine formation at Colonel Foster Glacier (Figures 6.4 and 6.5).

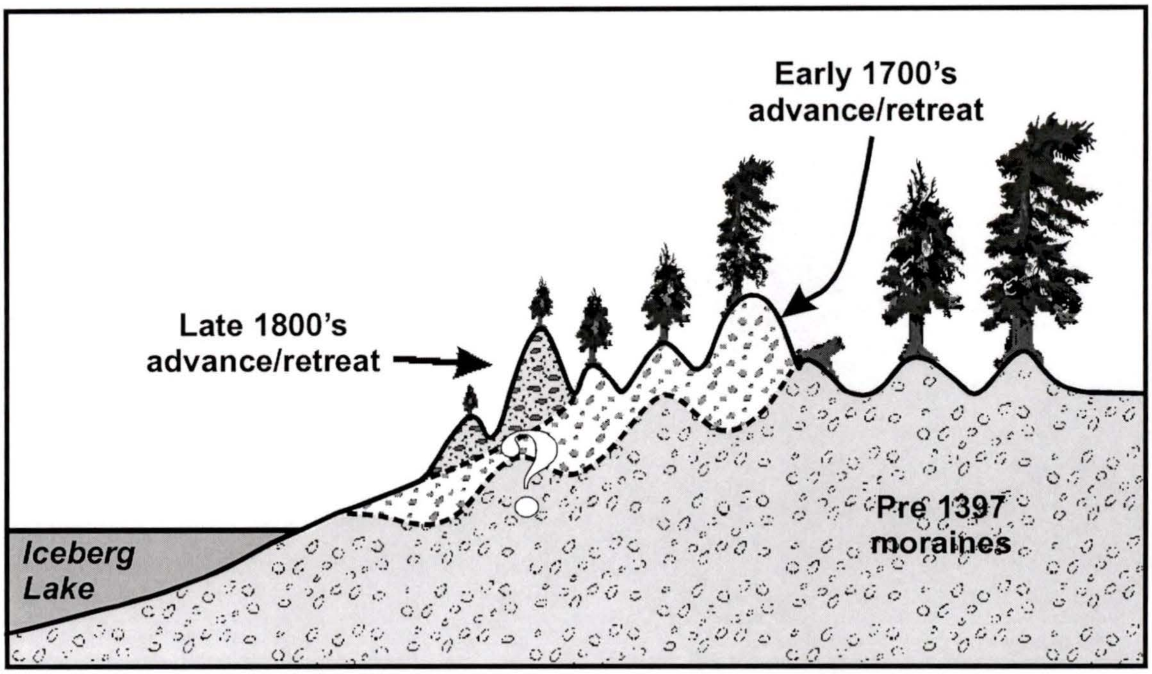


Figure 6.4. Schematic cross-section of geomorphic evidence for three LIA events at Colonel Foster Glacier. Shading and textural differences represent different events, and the question mark indicates that previous surfaces (dashed lines) are unknown.

Moraine dates provided by both dating techniques are given in Table 6.5. Tree-ring derived moraine dates follow the expected order of progressively younger moraines with closer proximity to the glacier front. Lichenometric moraine dates, however, do not follow a similar chronological trend. The lichen-derived date for Moraine 2 is 21 years younger than that of Moraine 1 (the youngest moraine), and the date for Moraine 4 is 72 years younger than the more proximal Moraine 3, and 77 years younger than the tree-ring derived date.

Table 6.5. Summary of tree-ring and lichen dating results at Colonel Foster Glacier

Moraine	Tree-ring derived moraine date (AD)	Lichen derived moraine date (AD)
1	1935	1912
2	1898	1933
3	1778	1775
4	1770	1847
5	1708	1708

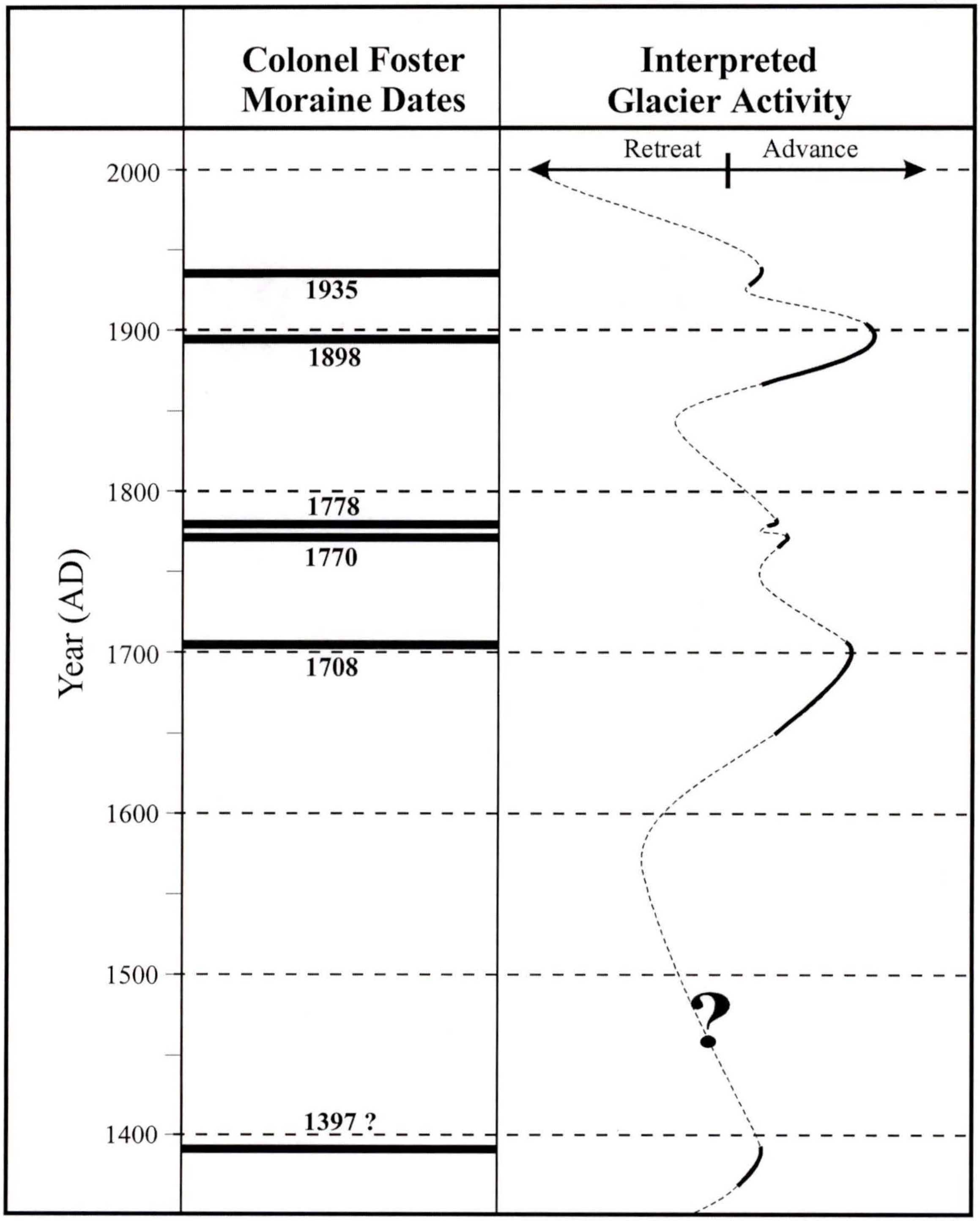


Figure 6.5. Interpreted episodes of LIA glacier activity at Colonel Foster Glacier. Horizontal bars on the left column represent dated moraines; solid curves in the right column represent glacier activity associated with the dated moraines, whereas the dashed line is suggested glacier activity between the dated events.

The non-sequential lichen dates could be a function of substrate size and late-lying snow cover, which combine to provide differential lichen growth rates (Innes 1985, 1988; Benedict 1993). Several large boulders and blocks on Moraines 3 and 5 provide elevated surfaces that become exposed above the snowpack earlier in the year. As it is not uncommon to have snow cover last into July and August, as well as have snowfall begin by September, the presence of larger host boulders allows for a prolonged lichen growth season. For less prominent moraines, or for those moraines with few large boulders, this equates to a shortened growth season and smaller lichen, resulting in an underestimation of the surface age.

Although late-lying snow cover can be a problem for lichen growth rates, it is less of a problem for tree growth. Persistent snow cover can inhibit seedling establishment and hinder growth for small, young saplings, however, once trees are established and become large enough to emerge from the snowpack, photosynthesis can start and annual growth can occur even though snow cover is still on the ground. This appears to be the case at Colonel Foster Glacier, where the moraines sit in a heavily shadowed cirque shaded by both the tall mountains and the adjacent forest, allowing snow to linger well into the summer months.

Except for a few larger boulders, most moraines lacked adequate lichen cover for dating purposes. Of the 143 lichen sampled, 78 percent (111 thalli) were from the only two moraines (Moraines 3 and 5) with large boulders. Nine of the ten largest lichen on Moraine 3 were from a single large block (Lichen Rock), and all seven lichen from Moraine 4 were sampled from a single boulder.

Differential lichen growth is also evident along the landslide scar, where 50 thalli were measured to provide independent dating control for the lichen growth curve (Table 5.5). Both the growth rate and the number of thalli were lower for lichen sampled from the bedrock surface. The majority of the lichen measured (76%) were from large boulders (0.5 m – 1.5 m diameter), where the growth rate of the largest lichen was 16 percent greater than the largest bedrock lichen.

Given the relationship between climate and mountain hemlock growth on Vancouver Island, the ring-width record from the Colonel Foster MH chronology (Figure 6.1) is interpreted as a combined response to climate conditions (as outlined in Section 5.2.3.1) and effects of glacier proximity. The collective effect of these two parameters is considered a proxy indicator for glacier mass balance conditions. Radial growth values of 1 are considered average growth, whereas values > 1 are indicative of climate conditions associated with a negative mass balance regime (warmer/drier) and with glacier retreat. An exception to this would be increased radial growth rates in conjunction with reaction wood formation. The presence of reaction wood suggests that the increased growth is a result of tilting due to ice proximal debris or increased snow-loading; both associated with of positive mass balance climate conditions and an advanced state of the glacier. Conversely, below average radial growth-rates (<1) reflect shorter cooler summers, and snowier winters; climate conditions that favour a positive mass balance regime (cooler/wetter) and glacier advances.

Termination of suppressed mountain hemlock growth intervals coincide with a number of minimum moraines dates determined from trees growing on the moraine surfaces and from the sub-fossil stump on Moraine 6 (Figure 6.6). The 1898, 1778, 1770

AD moraines all occur shortly after intervals of suppressed radial growth. The 1708 moraine appears to follow an interval of suppressed radial growth, but the ring-width record is too short to be certain. The timing of the 1935 recessional moraine appears to fall within a period of suppressed radial growth. However, the prolonged suppressed growth pattern is due to the effect of the 1946 earthquake, which resulted in a period of sharply reduced growth at this site during the mid 1940's.

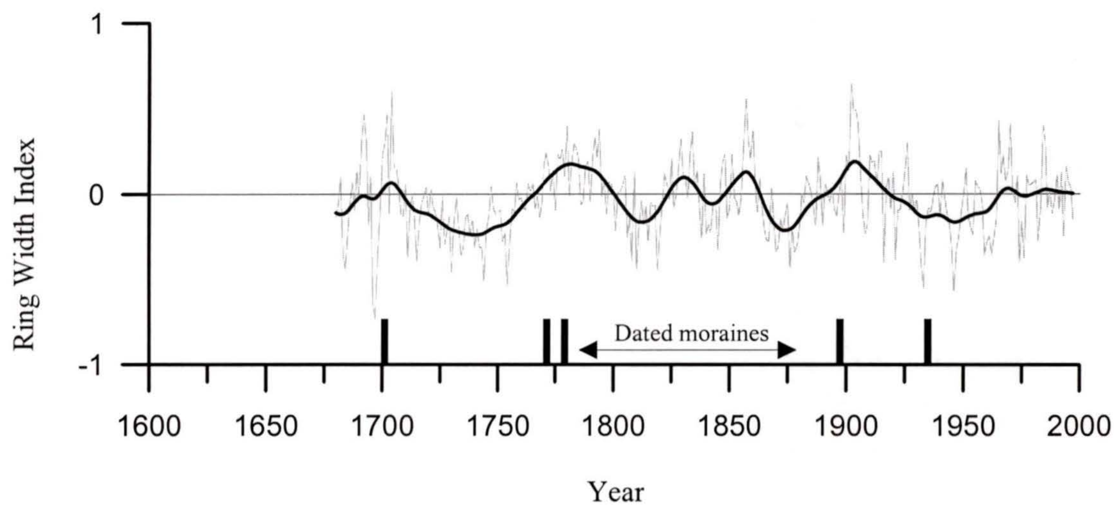


Figure 6.6. A comparison of dated moraines at Colonel Foster Glacier with the Colonel Foster mountain hemlock chronology. The thin grey line is annual values, the thick black line is a 25-year cubic smoothing spline to emphasize long-term variability, and the short black vertical bars are dated moraines.

The age of the *in situ* stump (CFLOG-01) from Moraine 6 indicates that Colonel Foster Glacier had not advanced past this point since at least 1396 AD. Periods of reaction wood formation in the late 1600's, in conjunction with scars dated to 1681 and 1696 AD, suggest increased snow loading and (or) debris mobilization from the front of the advancing glacier. The 1681 scar and subsequent reaction wood indicate an impact event from mobilized debris near the glacier front, with a subsequent tilting of the tree as the glacier advanced downvalley. The large 1696 scar on the north side of the tree marks the date when the tree was pushed over onto boulders below, scarred, and partially buried by debris from the deposition of Moraine 5. These data provide a maximum emplacement date of 1696 AD for Moraine 5.

The surface date of Moraine 5 determined from tree 98CFWM-09 indicates that the glacier had receded from this moraine prior to 1708, and provides a minimum date for moraine stabilisation. The evidence provided by tree 98CFWM-09 and the stump (CFLOG-01) suggest that Colonel Foster Glacier was advancing in the late 1600's and had reached its greatest down valley extent since at least 1396 sometime between 1696 and 1708, when it had begun retreating from Moraine 5.

Ring-width patterns and moraine morphology suggest Moraines 4 and 3 were deposited during stillstands or minor re-advances of the glacier after retreat from its 1708 maximum position. The ring-width suppression record the Colonel Foster MH chronology suggests two relatively strong intervals of positive mass balance conditions in the 1800's; however, no evidence of moraine formation is preserved. A problem with the preservation of many mid to late LIA moraines is that latter advances often override and rework moraines from earlier depositions. This appears to be the case at Colonel Foster Glacier, as the smaller size and discontinuous morphology of Moraines 4 and 3 compared to the larger, distinct, continuous nature of the late 1800's moraine, (Moraine 2), suggests a more significant late 1800's advance that most likely incorporated previous 19th century moraine deposits into the formation and deposition of Moraine 2.

Following the transition from suppressed to above average radial growth (Figure 6.6), interpreted as a change to ameliorating climate conditions at the beginning of the

1900's, a minimum stabilisation date of 1898 for Moraine 2 was determined from tree CFWM-35 growing on the proximal face of the moraine. Bracketing dates for the emplacement of this moraine were not possible as no trees were available for sampling from the distal slope of the moraine.

The surface age provided by tree CFWM-44 for the inter-moraine area between Moraines 2 and 1 indicates that this area was ice-free by at least 1926. The youngest and innermost moraine (Moraine 1) was deposited sometime after 1926 during a stillstand or minor re-advance that culminated by 1935 based on the age of tree CFWM-46. This is also reflected in the mountain hemlock ring-width record, with a return to suppressed growth by the 1920's, suggesting a short-term return to a positive mass balance climate regime (Figure 6.6).

6.2.5 Colonel Foster LIA Summary

Dendrochronological evidence suggests three main intervals of glacier activity at Colonel Foster Glacier to account for the LIA moraine formation at this site. The first notable episode of glacier activity occurred sometime prior to 1396 (early LIA?), and was likely responsible for the deposition of the outer moraine suite (Moraine 6, 7 and 8). However, due to the lack of dating control for the outer moraine suite, there is a possibility these moraines are from an earlier advance.

The second episode of moraine formation (Moraines 5, 4 and 3 of the inner suite) follows a late 1600's advance that culminated with the deposition of Moraine 5 sometime between 1696 and 1708 AD. The smaller, less developed ridges of Moraines 4 and 3 were deposited during stillstands or minor re-advances in the late 1700's as the glacier was retreating from its LIA maximum extent of the early 1700's.

A distinct change in morphology with Moraine 2 delineates the third and final episode of advanced LIA activity of Colonel Foster Glacier. The anomalous lack of evidence in the moraine record between the mid 1700's and late 1800's, however, is contrasted by periods of suppressed radial-growth in adjacent mountain hemlock; indicative of either close glacier proximity and (or) below average climate conditions. This suggests there may have been additional periods of glacial activity not recorded in the moraine record at this site. Two possible interpretations are that either the glacier remained in an advanced state during this interval, or that a late 1800's advance obliterated or reworked any previous moraine deposits and culminated with the deposition of Moraine 2 prior to 1898. Finally, Moraine 1 was deposited as recessional moraine during a minor re-advance or stillstand sometime between 1926 and 1935 as the glacier was retreating towards its present state.

6.3 Septimus Glacier

Dendrochronological evidence at Septimus Glacier was limited to 15 trees from two small stands of trees growing on prominent bedrock outcrops along the west side of the forefield. A large number of the trees in both stands show evidence of snow damage, or are very young and, therefore, unsuitable for dating purposes.

Table 6.6 lists the oldest trees sampled on each moraine, the location on the moraine surface, the corrected tree-age, and estimated moraine date. Complete results from all tree samples collected at Septimus Glacier are given in Appendix 2. At the upper tree site (Site A), corrected tree ages from eight trees (80 to 120 years old) provided minimal surface stabilisation dates ranging from 1912 – 1874, with two main moraine stabilisation episodes dated to 1898 and 1874 (Table 6.6).

Table 6.6. Tree ages and minimum surface dates for moraines at Septimus Glacier

Moraine and Sample ID	Location on Moraine[1]	Ring Count	Pith Error	Corrected Tree Age	Minimum Surface Date[4]
Site A					
Moraine 2					
98Sept-A5	Cr	86	10	86	1898
Moraine 3					
98Sept-A7	PF	120	0	120	1874
Site B					
Moraine 4					
98Sept-B7	DF	254	15	284[3]	1710
98Sept-B4	DF	345	5	360[2]	1634

[1] PF = Proximal Face, CR = Crest, DF = Distal Face
[2] Tree age includes a coring-height correction of 10 years
[3] Tree age includes a coring-height correction of 15 years
[4] Surface date includes 4 year ecesis interval

All trees sampled at the lower site (Site B) were collected from the distal slope of Moraine 4, with all but one (98Sept-B4) germinating in the 18th century (Table 6.6). 98Sept-B4, the oldest tree, was located near the base of the moraine on the distal slope. Coring height and pith corrections for this tree result in a corrected tree-age of 360 years and a germination date of 1638 AD. The ring-width pattern and J-shaped morphology of this tree suggest it was established prior to the formation of Moraine 4. Closer to the crest of the distal slope, the next oldest tree (98Sept-B7) has a corrected age of 284 years, indicating a surface stabilisation date of 1710 AD for Moraine 4.

6.3.1 Live Tree Dating Results

6.3.1.1 Ring-width Anomalies

Figure 6.7 illustrates variations in the ring-width pattern of the Septimus mountain hemlock (MH) chronology. Radial growth patterns of the chronology show periods of reduced growth in the 1430's, 1470's to late 1490's, early 1500's, late 1500's to mid 1650's, 1690's to 1760's, early 1800's, 1830's to 1850's, 1860's to 1900, and a minor episodes in the 1930's and 1970's. Notable intervals of above average mountain hemlock growth occur in mid 1500s, mid to late 1600s, late 1700's and early 1900's AD.

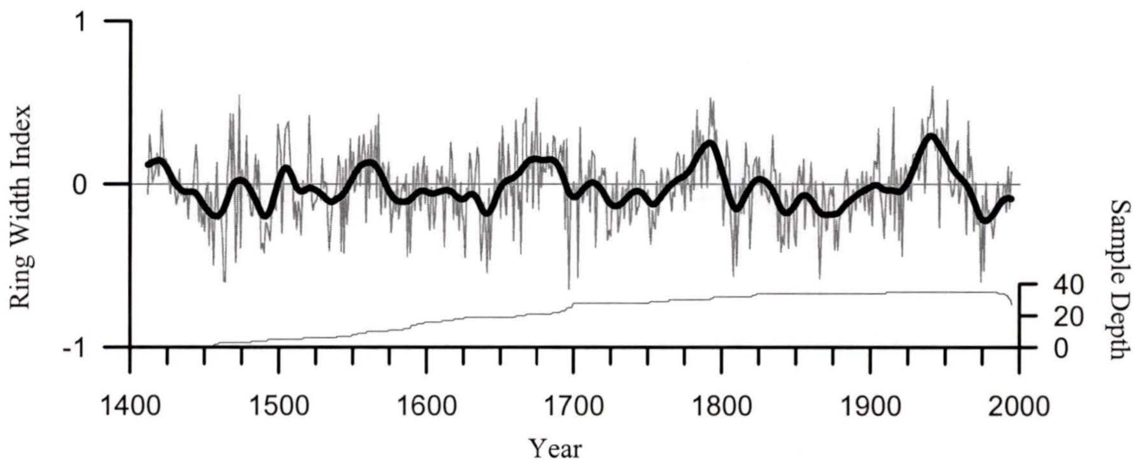


Figure 6.7. Mountain hemlock ring-width chronology from Septimus Glacier. Thin lines are annual values and thick line is a 15-year cubic smoothing spline to emphasise longer-term growth trends. Sample depth is the number of cores contributing to the annual index.

6.3.2 Lichenometric moraine dates from Septimus Glacier

Stabilization dates for the Septimus Glacier moraines were also determined from 264 *R. geographicum sp.* measurements taken from four moraines on both the western and eastern sides of the glacier forefield. Table 6.7 summarizes the results of the lichen measurements, including the number of lichen sampled on each moraine, the size of both the single largest and the average of the five largest thalli, and the corresponding minimum surface date interpolated from the Strathcona dating curve.

Table 6.7. Lichen dating results for the moraines at Septimus Glacier

Moraine	# Lichen Sampled	Single Largest (mm)	Minimum Surface Date	Average Largest 5 (mm)	Minimum Surface Date
Moraine 1	31	33.2	**1934**	31.9	**1938**
-western	*19*	*33.2*	*1934*	*31.9*	*1938*
-eastern	*12*	*28.3*	*1951*	*24.6*	*1961*
Moraine 2	60	42.7	**1898**	41.5	**1902**
-western	*19*	*42.7*	*1898*	*41.2*	*1902*
-eastern	*41*	*40.2*	*1908*	*38.2*	*1916*
Moraine 3	78	60.8	**1836**	58.7	**1842**
-western	*19*	*59.6*	*1838*	*56.9*	*1849*
-eastern	*59*	*60.8*	*1836*	*55.5*	*1855*
Moraine 4	95	97.4	**1706**	93.6	**1718**
-western	*24*	*52.2*	*1858*	*50.4*	*1863*
-eastern	*71*	*97.4*	*1706*	*93.6*	*1718*

Italics are minimum surface dates from the eastern and western moraine components
Bold are minimum surface dates for the entire moraine based on eastern and western component data

With the exception of Moraine 1, there is a noticeable difference in thalli distribution at Septimus Glacier. Lichen were more abundant on the larger and more prominent eastern moraines (183 thalli) than on the western moraines (81 thalli). The single largest thalli from each moraine was only 2.8 to 3.9 percent larger than the mean of the five largest diameters from the same moraine, indicating that the single largest lichen were not anomalously large, and therefore, suitable for providing minimum stabilization dates for the moraines.

6.3.3 **Interpretation of Dating Methods at Septimus Glacier**

Moraine dates provided by both dating methods are summarised in Table 6.8. Both series of moraine dates follow the expected order of decreasing age with an increased proximity to the glacier front. Tree-ring dating methods, however, did not provide a date for Moraine 1 due to a lack of adequate tree cover on the eastern moraines, and on the western side where tree growth was limited to the larger outer moraines (Moraines 2, 3 and 4). Conversely, lichen dates were recorded from all moraines on both sides of the valley.

Table 6.8. Summary of tree-ring and lichen dating results at Septimus Glacier

Moraine	Tree-ring date	Lichen date
1	-	1934
2	1898	1898
3	1874	1836
4	1710	1706

Although both methods provide a series of dates for moraine deposition, the two moraine chronologies are not entirely synchronous. Deposition dates determined by the two methods are roughly equivalent for Moraines 2 and 4; however, there is a 38-year difference between dates for Moraine 3. Lichenometry indicates that the moraine had stabilized by 1836, whereas tree-ring dates indicate Moraine 3 stabilized by 1874 (Table 6.8). A possible explanation for the difference in moraine dates may be recorded in the radial growth record of the Septimus MH chronology (Figure 6.7). Suppressed ring-widths during the mid to late 1800's suggest climate conditions favouring a positive mass balance regime, which may have resulted in a minor glacier re-advance or stillstand. Considering that tree-ring data for this moraine is taken from Site A, which is protected by a bedrock outcropping, it is possible that the tree-ring date is from a remnant recessional moraine that wasn't reworked by the deposition of the 1898 moraine (Moraine 2), or that the pioneering trees of this moraine may have been killed during an minor re-advanced phase of the glacier and that the trees sampled are second-generation trees. If it is a remnant moraine, a time-lag for tree establishment, or an extended ecesis interval may have resulted from periods of enhanced snow avalanching, or deeper than average snow cover. Increased snow avalanching off the margin of the glacier may have killed trees that had originally established on the 1836 moraine, and the trees that were sampled in 1998 are second-generation trees, thus, underestimating the age of the moraine.

Similar lichen dates of 1836 and 1838 for Moraine 3 were obtained from both east and west moraine suites respectively. Unfortunately, it is difficult to compare the two dating methods directly due to lack of adequate lichen growth at the two tree locations. Sites A and B on the western moraine suite lack sufficient lichen growth as the trees have out-competed the lichen at these two locations.

6.3.4 Interpretation of LIA History at Septimus Glacier

A compilation of moraine dates provided by the largest lichen diameters and living trees growing on the moraine surfaces, in conjunction with the Septimus mountain hemlock ring-width record, suggests three intervals of glacier activity relating to three episodes of LIA moraine formation at Septimus Glacier. LIA moraine dates are centred on 1706, 1836, and 1898, and possibly an additional minor advance in the 1870's (Figure 6.8 and 6.9).

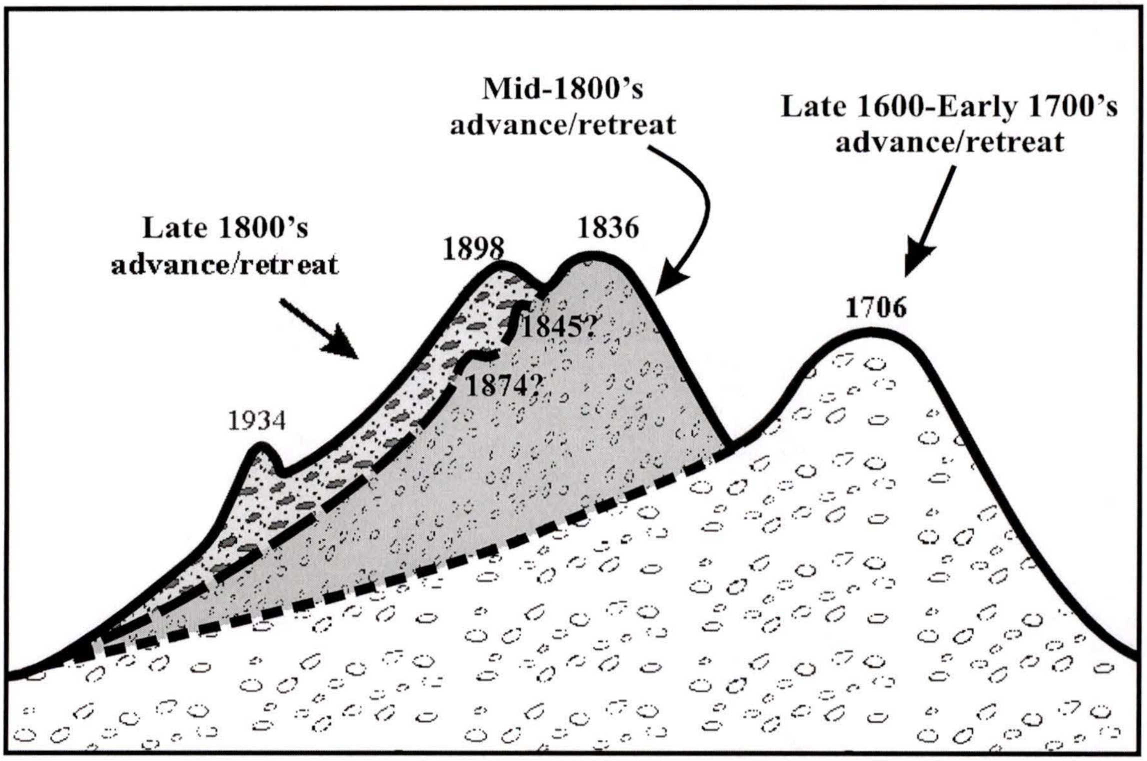


Figure 6.8. Schematic cross-section of the eastern moraines at Septimus Glacier illustrating the timing of three LIA events. Shading and textural differences represent different events, and dashed lines indicate that previous surfaces are unknown.

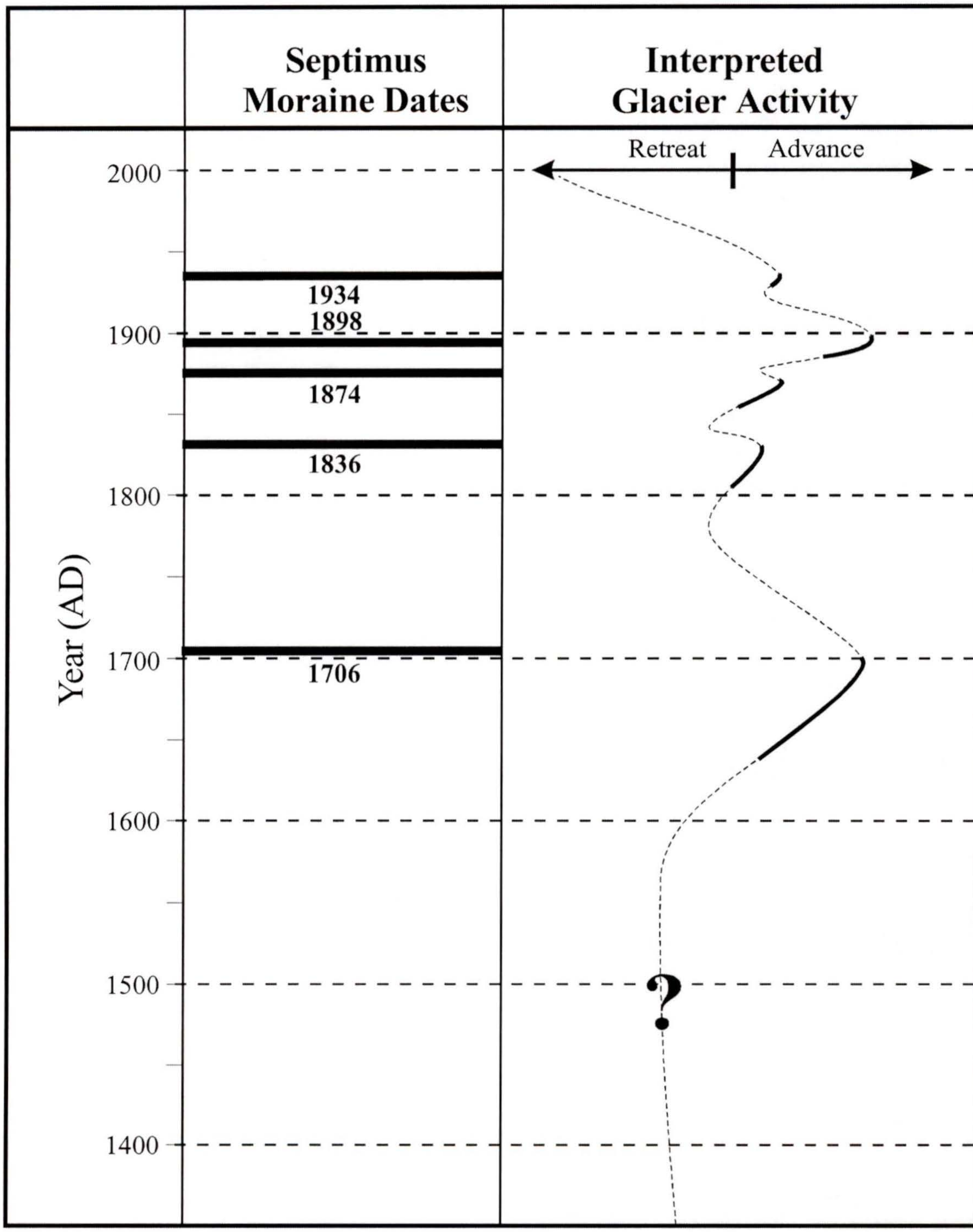


Figure 6.9. Interpreted episodes of LIA glacier activity at Septimus Glacier. Horizontal bars on the left column represent dated moraines; solid curves in the right column represent glacier activity associated with the dated moraines, whereas the dashed line is suggested glacier activity between the dated events.

The "J-shaped" morphology of 98SEPT-B4 indicates that the tree had germinated prior to the deposition of the outermost moraine (Moraine 4). In the Septimus MH chronology, suppressed radial growth patterns in the mid 1600's and late 1600's to early 1700's is interpreted as climate conditions favouring a positive glacier mass balance regime that led to the late 1600's advance of Septimus Glacier. The appearance of reaction wood during the 1680's and early 1690's in a number of tree cores taken from Site B, in conjunction with the trunk morphology indicate that the tree was tipped over by the advancing glacier during this interval, or by enhanced snow creep due to above average snowpacks. A sharp decrease in mountain hemlock growth between 1693 and 1697 suggests the occurrence of a relatively short-term traumatic event. This event is also recorded in the regional chronology, indicating it was not site specific, but instead a response to a regional climatic event.

Moraine dates provided by lichen measurements (1706 AD) and tree-rings (1710 AD) indicate that the glacier had retreated from Moraine 4 by the early 1700's. The presence and age of tree 98SEPT-B4 at the base of Moraine 4 suggests that it may have been growing on a previously deposited moraine, and by *ca.* 1695 the glacier had reached its maximum down-valley extent since at least 1638, before retreating from Moraine 4 by 1706 – 1710.

A second phase of glacier activity involving at least two separate advances is dated to the early and mid 1800's. However, the Septimus MH chronology illustrates a prolonged period of below average tree-growth in the early and mid 1700's. In conjunction with a lack of moraine evidence from this time, it suggests either an advanced state of Septimus Glacier, or reworking of any mid to late 1700's moraines by subsequent advances. Lichen dates from moraines on both sides of the moraine complex indicate that the glacier had retreated from Moraine 3 by 1836. Two further episodes of suppressed tree-growth in the MH chronology suggest positive mass balance conditions occurred at least twice in the 1800's. However, no moraine evidence is preserved from either of these events, unless the 1874 tree-ring date is from a remnant moraine fragment not reworked by the 1898 advance.

A return to below average mountain hemlock growth in the late 1800's coincides with the last major advance of Septimus Glacier. Lichen and tree-ring dates show that the late 19th century advance culminated with the deposition of Moraine 2 prior to 1898. The size of Moraine 2 and its close proximity to Moraine 3 suggests that the 1898 advance was extensive enough to override and incorporate lesser, earlier advances of the early and mid 1800's. An increase in mountain hemlock radial growth rates suggest ameliorating climate conditions in the early 1900's, followed by a minor drop in radial growth the late 1910's to 1920's indicating a minor deterioration in climate conditions. Finally, Moraine 1 was deposited during a minor re-advance or stillstand that culminated by 1934, corresponding to the early 1900's period of reduced radial growth.

6.3.5 Septimus LIA Summary

The moraine chronology provided by lichen and tree-ring dated moraine surfaces, in conjunction with the Septimus MH chronology, provides evidence of three main episodes of LIA moraine formation centred around 1706, 1836, and 1898, and possibly an additional minor advance in the 1870's (Figures 6.7, 6.8 and 6.9).

The first episode of moraine formation follows a late 1600's advance, which culminated with the maximum LIA glacier extent in *ca.* 1695, and by 1706 to 1710 the glacier had begun retreating from Moraine 4. A second advance in the mid- to late 1700's culminated with the deposition of Moraine 3 by 1836. Although the tree-ring record suggests conditions favouring a positive mass balance regime and possible re-advances in the 1820's, 1840's or 1870's, any moraines associated with these advances are not preserved and may have been reworked and incorporated into the deposition of Moraine 2 (1898). The youngest, and innermost moraine (Moraine 1) was deposited during a stillstand or minor re-advance that ended prior to 1934 (Figures 6.7, 6.8 and 6.9).

6.4 LIA in Strathcona Provincial Park

A comparison of the moraine chronologies from the two study sites is given in Table 6.9. The Colonel Foster moraine chronology is based solely on tree-ring methods, whereas the Septimus moraine chronology utilizes dendrochronological and lichenometric dating methods.

Table 6.9. Summary of moraine dates from Colonel Foster and Septimus glaciers

Colonel Foster Glacier	Septimus Glacier
1935	1934
1898	1898
1778	**1874**
1770	**1836**
1708	1706

Bold dates denote asynchronous moraine dates

The moraine records at Colonel Foster and Septimus Glaciers indicate that glacier response to climate conditions during the LIA was generally synchronous with larger and longer-term climatic events, yet asynchronous at lesser, shorter time-scales (*i.e.*, annual to decadal) events. Both glaciers appear to have responded synchronously to three major climatic events, the first in the late 1600's to early 1700's, the second in the late 1800's, and a third event in the mid 1930's. Based on geomorphic and dendrochronological evidence, however, glacier response to climate conditions between the larger events appears to be asynchronous. This suggests that the glaciers were responding to some sort of environmental threshold, below which glacier activity is governed by local environmental conditions such as glacier shape, bed slope, topography, and local climate, but above this threshold the glaciers respond to a greater, regional signal.

The asynchronous response of the glaciers between the early 1700's and late 1900's advances may be explained in terms of differences in the local site conditions. Colonel Foster Glacier is wider, and relatively unconfined compared to Septimus Glacier, which is longer, thinner and constrained between steep bedrock walls on its eastern side and bedrock outcrops on its western side. Colonel Foster is also in a more protected and shaded cirque than Septimus Glacier. Septimus Glacier is also a land-terminating glacier, whereas Colonel Foster Glacier presently calves into Iceberg Lake, which has definitely altered its dynamic response over time.

The LIA chronology of Septimus and Colonel Foster glaciers' is in agreement with previous LIA research at Moving Glacier where dendroglaciological evidence provided dates for LIA glacier advances in the early 1700's and after 1818 (Smith and Laroque 1996). Since Moving Glacier terminated in Milla Lake throughout the LIA, however, the flow dynamics of this glacier are likely quite different from that of Colonel Foster and Septimus Glaciers. Along with Moving Glacier, both Colonel Foster and Septimus Glaciers reveal a similar magnitude of retreat and down-wasting since reaching their maximum LIA positions (Table 6.10)

Although LIA moraine evidence is limited to three glaciers, the findings from this study and that of Smith and Laroque (1996) indicate that glaciers in Strathcona Park have responded synchronously to large-scale climatic events during the mid and late stages of the LIA. In the context of the larger regional LIA picture, the Strathcona Park LIA moraine record corresponds well with LIA moraine record the Canadian Rocky Mountains and the combined Coast Mountain LIA record from Alaska, British Columbia, and Washington. Two main pulses of LIA activity are dated to the early 1700's and late 1800's, followed by a minor advance in the 1930's (Figure 6.10).

Table 6.10. Comparison of retreat and areal ice losses of Colonel Foster, Septimus and Moving glaciers.

Glacier	Retreat from LIA maximum (m)	% of LIA maximum area remaining
Col. Foster	300	22
Septimus	600	18
Moving[a]	834[a]	10[a]

[a] from Smith and Laroque (1996)

Dendroclimatic temperature reconstructions in both Alaska and the Canadian Rocky Mountains have shown that the mid to late 1690's were characterized by sustained cold summers, and that following this 'cold snap' many glaciers deposited their outermost LIA moraines (Wiles *et al.* 1996b; Luckman *et al.* 1997; Wiles 1997). Similarly, both temperature reconstructions indicate the late 1800's were also one of the coldest decades, and that following this interval most glaciers deposited a late 1800's moraine marking the end of the LIA.

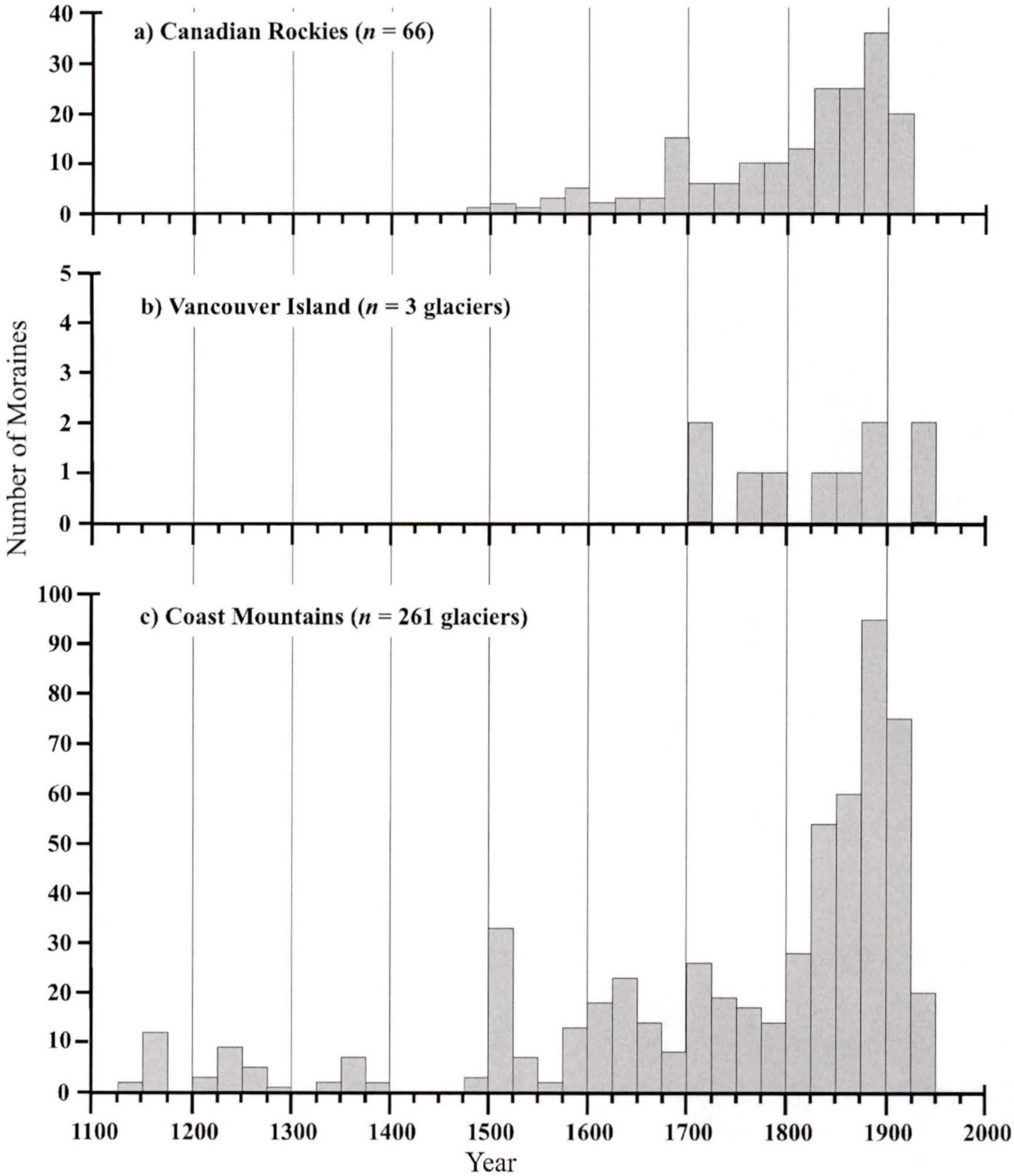


Figure 6.10. A comparison of LIA glacier activity on Vancouver Island to regional LIA records of the Coast Mountain ranges and the Canadian Rocky Mountains.

CHAPTER 7. MASS BALANCE RECONSTRUCTION

7.1 Glacier Mass Balance

Variations in energy and mass fluxes at glacier surfaces result in changes in accumulation and ablation that ultimately lead to glacier mass balance fluctuations. Mass balance is defined as the difference between the mass input (accumulation) and the mass output (ablation) for a glacier, measured over a particular time interval and expressed in terms of water equivalence (w.eq.) (Paterson 1995).

Changes in glacier mass balance alter the dynamics of glacier flow and velocity, and are often manifested as fluctuations in glacier area. During intervals when accumulation exceeds ablation (positive net balance), glaciers often expand in area, advance and deposit moraines that mark the former ice extent following glacial retreat (negative net balance).

For most glaciers outside of the polar regions, winter accumulation (snowfall) and summer ablation (temperature) are the primary controls on total annual mass gain and loss, respectively (Tangborn 1980; Letréguilly 1988; Walters and Meier 1989; Harper 1993; McCabe and Fountain 1995; Hodge *et al.* 1998). The degree to which the mass balance of alpine glaciers responds to changes in summer temperature and winter precipitation is a function of their continentality (Letréguilly and Reynaud 1989; Oerlemans 1998). For continental glaciers such as Peyto Glacier in the Canadian Rocky Mountains (Figure 7.1), the net mass balance is primarily a response to fluctuations in summer temperature (Demuth and Keller *in press*). Maritime glaciers in the Pacific Northwest, such as South Cascade Glacier in Washington State, and Place, Helm and Sentinel Glaciers in the Coast Mountains of British Columbia, are more sensitive to variations in winter precipitation (Figure 7.1). For more southerly maritime glaciers (*i.e.*, South Cascade), however, the effect of summer temperature has an increasingly more significant role in mass balance (Letréguilly 1988; Letréguilly and Reynaud 1989; Walters and Meier 1989; Brugman 1992; Demuth and Keller *in press*).

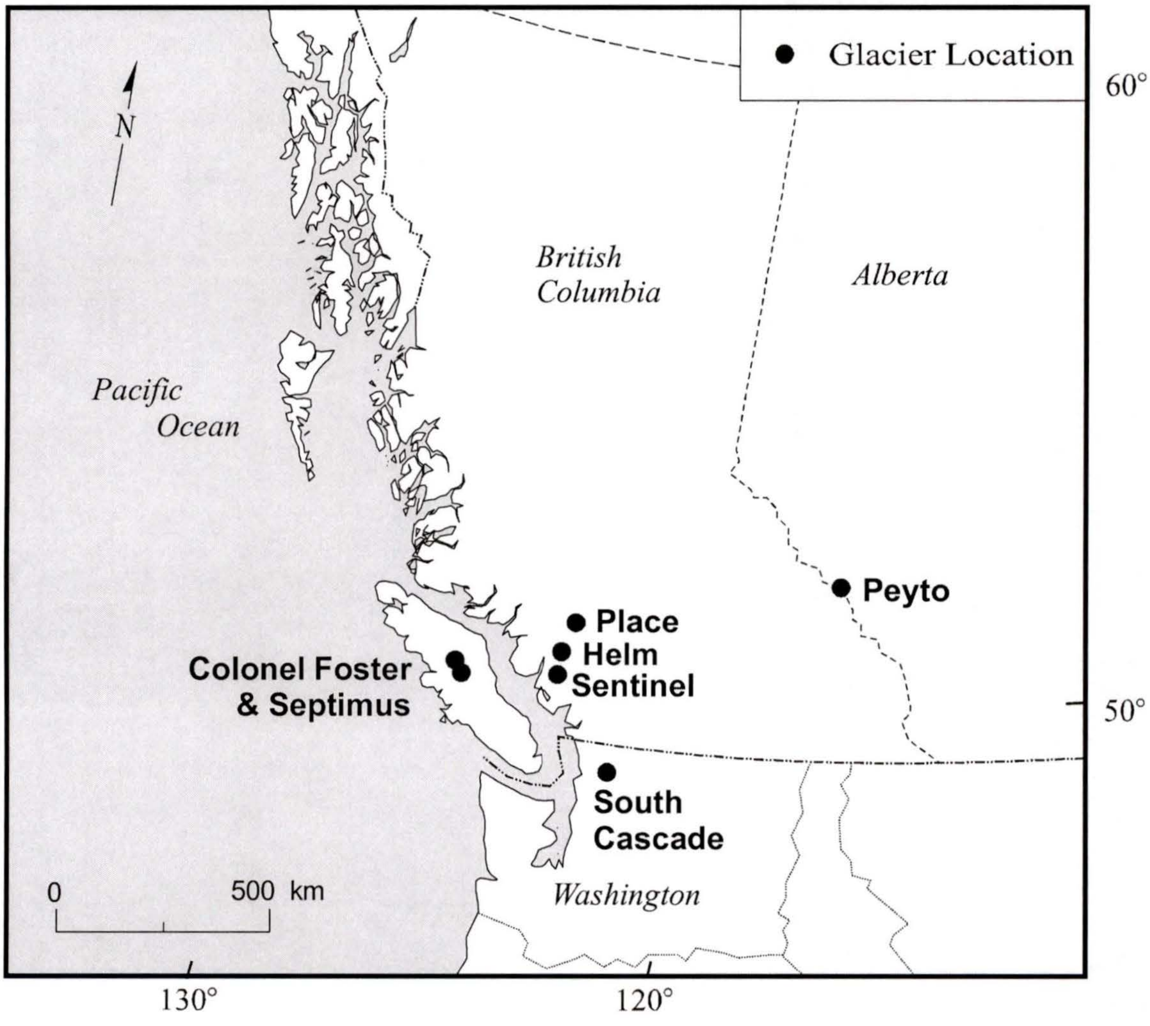


Figure 7.1. Map illustrating the locations of some maritime (Septimus, Col. Foster, Place, Sentinel, Helm and S. Cascade) and continental (Peyto) glaciers in Pacific North America.

Since the temporal and spatial variations in surface weather variables, such as precipitation and temperature, are controlling factors of glacier mass balance, and as they are a function of upper level atmospheric circulation, glacier mass balance measurements can be an effective proxy indicator of regional climate change (McCabe and Fountain 1995; McCabe *et al.* 2000). Recent glacier-climate studies in the Pacific Northwest show that significant relationships exist between variations in glacier mass balance and upper atmospheric circulation (Yarnal 1984; Walters and Meier 1989; McCabe and Fountain 1995; Moore 1996; Hodge *et al.* 1998; Bitz and Battisti 1999; McCabe *et al.* 2000). This relationship suggests that a homogeneous glacial response to regional climate patterns exists among glaciers in southwestern British Columbia and northwestern Washington State (Letréguilly and Reynaud 1989; Walters and Meier 1989; Brugman 1992; McCabe and Fountain 1995; Bitz and Battisti 1999). Mass balance variations have been statistically correlated for distances up to 500 km (Letréguilly 1988; Letréguilly and Reynaud 1989, 1990), and indicate that a common mass balance signal is applicable to glaciers within that area (Letréguilly and Reynaud 1989). Based on this relationship, it is assumed that the response of glaciers on Vancouver Island will be similar to those found in the nearby Coast Mountains of British Columbia and Cascade Mountains of northern Washington State. Consequently, a regional mass balance record constructed from glaciers in this region should be representative of Vancouver Island glaciers, for which there are no mass balance records.

7.2 Mass Balance Reconstruction

Numerous authors have shown that variations in high elevation tree-ring chronologies correlate well with changes in glacier mass balance, oscillations in glacier length, and periods of glacial advance (Matthews 1977; Karlén 1984; Scuderi 1987;

Serebryanny and Solomina 1989; Kaiser 1993; Luckman 1993). For instance Lamarche and Fritts (1971) found that annual radial growth variations within high elevation trees in the Austrian Alps were significantly correlated to local glacier activity. These findings are supported by other studies which show that conifers in alpine environments are particularly sensitive to climatic conditions that induce glacial activity, and can, therefore, be successfully used as proxy indicators of glacial variation (Villalba *et al.* 1990; Bhattacharyya and Yadav 1996; Nicolussi and Patzelt 1996).

Regional studies in southwestern British Columbia and northwestern Washington State have shown that the radial growth of mountain hemlock in the Pacific Northwest is highly sensitive to summer temperatures and winter precipitation (represented by April 1 snowpack depth) (Heikkinen 1985; Graumlich and Brubaker 1986; Smith and Laroque 1998b; Gedalof and Smith 2001a; Peterson and Peterson *in press*). The climate variables are the same ones that govern glacier mass balance fluctuations in the same region, as these glaciers are also shown to be highly sensitive to summer temperature and winter precipitation (Tangborn 1980; Burbank 1982; Letréguilly 1988; Brugman 1992; McClung and Armstrong 1993; Moore and McKendry 1996; Bitz and Battisti 1999).

Although the relationship between climate and glacier mass balance is complex, it is known that high winter precipitation and short, cool, cloudy summers favour net accumulation (positive mass balance) and lead to glacial advances (Paterson 1995). Similar climate conditions in the Pacific Northwest have been shown to result in a shortened growing season and below average (narrow) mountain hemlock radial growth on Vancouver Island (Smith and Laroque 1998a, 1998b; Gedalof and Smith 2001a). Conversely, climate conditions that promote above average radial growth (wide rings) in mountain hemlock, relatively warm/dry winters and moderately warm summers are conditions that favour glacial ablation (negative mass balance) and glacial retreat (See Bray and Struik 1963). Such relationships, between periods of reduced ring-width and glacial expansion, suggest that in combination with other proxy indicators such as moraine records, dendroclimatic data may provide a continuous proxy record of the climatic conditions that have governed LIA glacier activity.

For such an approach to be robust, a quantitative relationship needs to be established between climate parameters in the form of local meteorological data (*i.e.*, temperature and precipitation) and glacier mass balance variation. A similar relationship between climate and ring-width variation of climatically sensitive trees must also be established. Once the relationships have been defined and an accurately dated moraine chronosequence has been developed, the relationship between climate induced ring-width depressions and episodes of moraine emplacement can be shown if both records show a quantitative response to similar climate conditions (Villalba *et al.* 1990). Moraine emplacement dates derived through dendrogeomorphological, dendroglaciological, and lichenometric dating techniques can then be used to verify the reconstructed glacier or mass balance record from the continuous tree-ring series. Similar approaches were used by Nicolussi and Patzelt (1996) to reconstruct a continuous glacial mass balance record for Hintereisferner Glacier in the Tyrol back to 1600 AD, and by Matthews (1977) and Villalba *et al.* (1990) to reconstruct glacial histories in Norway and Argentina.

7.2.1 Climate-Tree Relationship

The mountain hemlock-climate relationships established earlier (Section 5.2.3.1) indicate that the climatic variables responsible for limiting radial growth of mountain hemlock in Strathcona Provincial Park are summer temperature and winter precipitation. These results corroborate findings of previous mountain hemlock-climate studies on Vancouver Island. The reconstructed mass balance index is limited to 1600 AD, as the Subsample Signal Strength (SSS) values for the Strathcona mountain hemlock chronology (>0.85 acceptable) indicate that the common signal within the trees prior to 1600 AD is unreliable (Briffa and Jones 1990).

7.2.2 Mass balance record

As there are no mass balance records for glaciers on Vancouver Island, a regional mass balance anomaly record was constructed from the mass balance records of four mainland glaciers in close proximity to Vancouver Island (Table 7.1; Figures 7.1 and 7.2; Appendix 2).

Table 7.1. Mass balance records from PNW glaciers that were used in the construction of the regional mass balance record

Glacier	Period used	Location (Lat., Long.)
Place	1965-1994[1]	50º 26'N, 123º 36'W
Helm	1975-1994[1]	49º 58'N, 123º 00'W
Sentinel	1966-1990[2]	49º 53'N, 122º 59'W
South Cascade	1959-1994[3]	48º 22'N, 121º 03'W
Regional	**1966-1994**	

[1] Provided by Mike Demuth, GSC, Glaciology Division
[2] WGMS archive data
[3] Provided by Roy Krimmel, USGS, Water Resources

To calculate a regional mass balance anomaly record, each observation was first standardized as follows:

$$\mathrm{SMB}_t = \frac{(\mathrm{MB}_t - m)}{S} \qquad (7.1)$$

where SMB_t is the standardized mass balance anomaly for year t; MB_t is the observed annual net mass balance at year t (in metres of water equivalence (m w.eq.)); and m and s

are the mean and standard deviation of the mass balance time series, respectively (in m w.eq.). By standardizing the mass balance records, the data become dimensionless indices that can be used in regression relations with standardized tree-ring chronologies for the reconstruction of historical mass balance anomalies. The standardized mass balance records were then averaged using an arithmetic mean into a single regional time series of mass balance deviations relative to 1966-1994 mean. The regional mass balance record is restricted to the 1966-1994 interval in order to incorporate the longest possible record of mass balance data, as well as to coincide with the last full year of growth in the Septimus Glacier tree-ring chronology. For the reconstruction, the mass balance records for Sentinel (1966-1990) and Helm (1975-1997) glaciers were combined through regression analysis to create a sufficiently long composite record (1966-1994) that would otherwise be limited to the common interval of 1975-1990 for the two sites. Mass balance is no longer recorded at Sentinel Glacier due to the close proximity to, and high correlation ($r = 0.90$) with Helm Glacier.

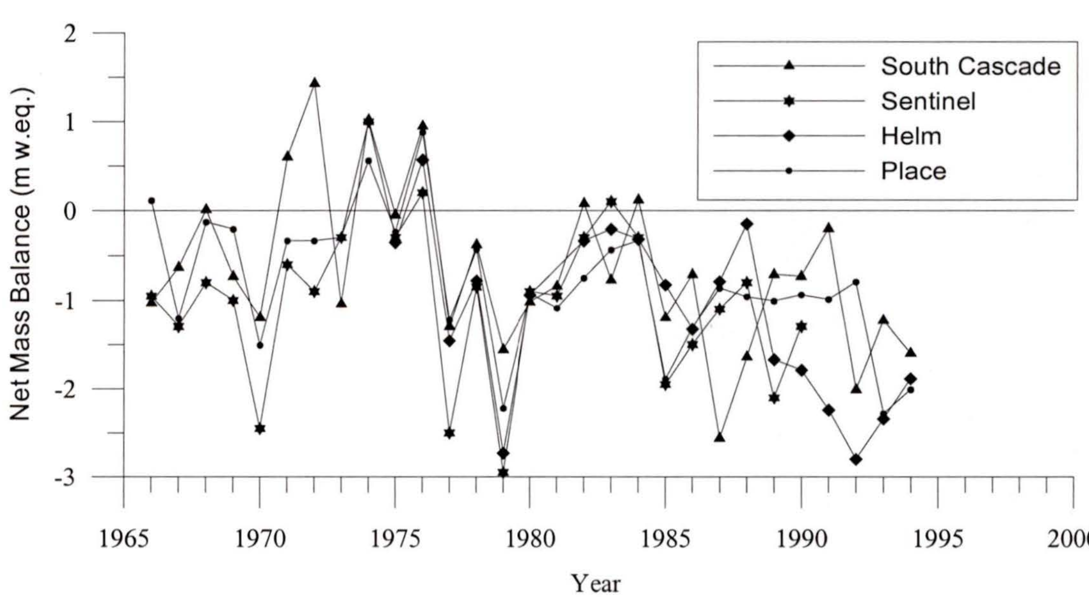


Figure 7.2. Annual net balance variations for South Cascade, Sentinel, Helm and Place glaciers from 1966 to 1994.

7.3 Results of the Strathcona Mass Balance Reconstruction

Correlation between the regional ring-width chronology and the regional mass balance record show that the mass balance record has a strong, negative correlation to the mountain hemlock chronology (R = -0.71 at the 99% confidence level). Equation 7.2 is the result of a simple linear regression of the standardized mass balance regional mass balance record onto the Strathcona regional ring-width chronology and provides an R^2 of 0.50 for the period 1966-1994 (Figure 7.3).

$$\text{SMB} = 4.37(SRW) + 3.74 \tag{7.2}$$

where SMB is the standardized mass balance deviation, and *SRW* is the standardized mountain hemlock ring-width index. Due to the brevity of the regional mass balance record, the reconstructed anomalies (1966-1994) were not divided into separate calibration and verification subsets. Based on the reasonable reproduction of measured record, the Strathcona chronology was used to develop the paleo-mass balance record back to 1600 AD (Figure 7.4).

Reconstructed intervals of positive mass balance (MB) anomalies occur from 1622 to 1668, 1696 to 1702, 1721 to 1762, 1802 to 1820, 1839 to 1847, 1864 to 1886 AD, and a minor positive anomaly in the mid 1970s. To assess the accuracy of the reconstruction, the reconstructed mass balance record was compared to the dated moraine record from Pacific North America (Figure 7.5). Periods of synchronous moraine dates were used for verification, as they are believed to represent intervals of regional climate forcing.

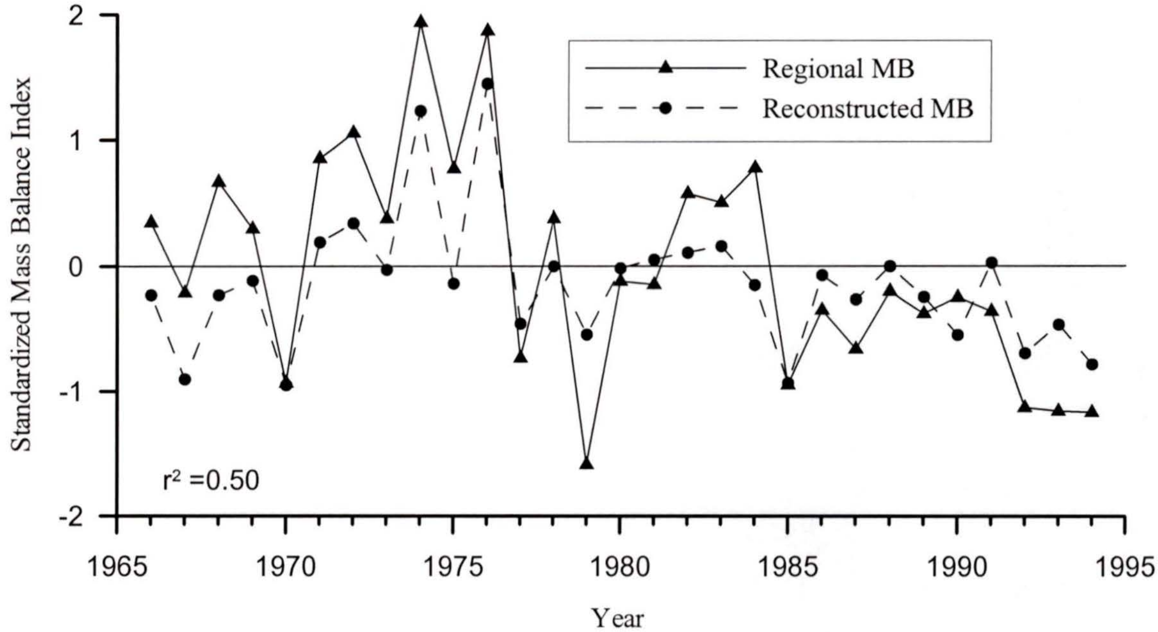


Figure 7.3. Reconstructed and regional standardized mass balance anomalies for the period 1966 to 1994. Solid line is the regional record and dashed line is the reconstructed record.

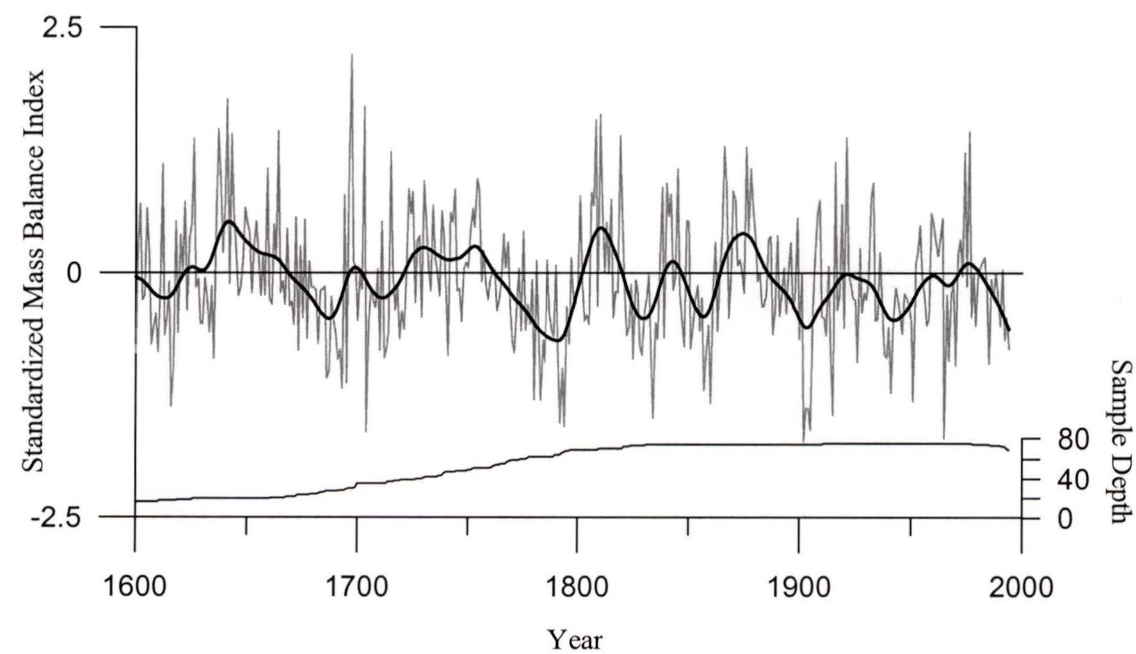


Figure 7.4. Reconstructed mass balance anomalies from 1600 to 1994 AD. Thin line is the annual values and bold line is a 25-year cubic smoothing spline to emphasise long-term trends. Sample depth is the number of cores contributing to the annual value.

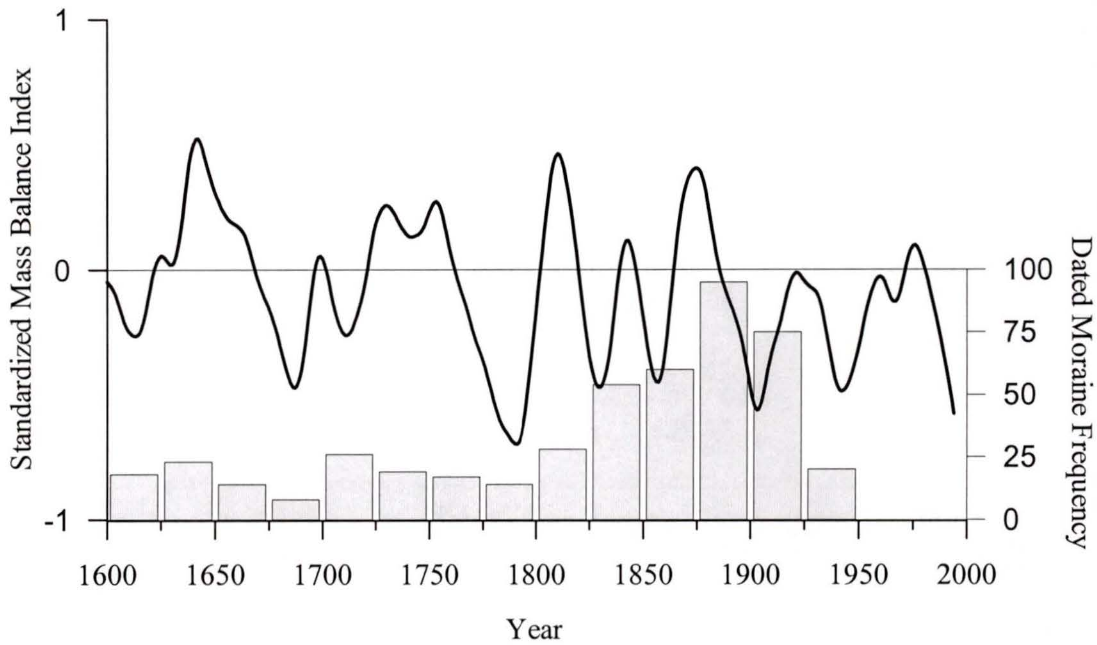


Figure 7.5. Relationship between dated LIA moraines in Pacific North America and reconstructed mass balance anomalies from 1600 to 1994 AD. Standardized mass balance anomalies are represented by a 25-yr cubic spline (thick black line) to emphasize long-term trends; dated moraines are from Alaska, British Columbia and Washington, and are grouped in 25-year intervals.

In comparing the Pacific North American moraine record to the reconstructed mass balance anomaly record, it is evident that although there are intervals with a reasonable relationship between the two records, the relationship between the two weakens prior to the 1800's. The trend of increasing MB anomalies prior to the deposition of the 1935 moraine agrees with the geomorphic interpretation that the glacier came to a stillstand at this time, depositing a recessional moraine. Deposition of late 1800's moraines also correspond reasonably well with reconstructed positive mass balance anomalies, as peak moraine formation lags behind the 1875 peak in positive mass balance anomalies by *ca.* 12 years. It is possible that the lag prior the late 1800's moraine is due to glaciers remaining in an advanced state until the early 1890's.

The other set of 'classic' LIA moraines (early 1700's) also follows a 10 to 12-year lag behind a short, but positive trend of mass balance anomalies. A similar relationship is also noted for mid 1600's moraines; however, the peak in moraine dates appears to lead the mass balance anomalies by *ca.* 5 years. Conversely, large positive mass balance anomalies in the mid 1700's and early 1800's do not correlate with any peaks in LIA moraine dates.

Possible explanations for the indirect correlation between the mass balance and LIA moraines include the fact that both records are composites representing large areas, which may result in diluted signals. Other considerations are variations in methods used to date moraines and their differing degrees of dating accuracy, and the combined lack of knowledge regarding response times of these particular glaciers and the length of time required for moraines to stabilize once the ice retreats. Also, there is limited moraine evidence prior to the 1700's. Early 18th century moraines may very well be reworked 17th century moraines, similar to those found in both Alaska and Washington State. Similarly, the large 1898 moraine could have reworked earlier LIA moraines deposits. At a number of sites, the 19th century moraines are often the largest and represent the greatest downvalley extent of many glaciers during the LIA.

Although there are intervals where the LIA moraine chronology is in general agreement with the mass balance reconstruction, the lack of a direct correlation for the majority of the record may be also attributed to significant differences in synoptic scale circulation patterns during the mass balance calibration period and the LIA. Mass balance anomalies are calculated relative to the 1966 – 1994 average net mass balance, an interval that includes 16 of the warmest and 14 of the driest years on record, which is reflected in increasingly negative glacier mass balance over this interval (Figure 7.6).

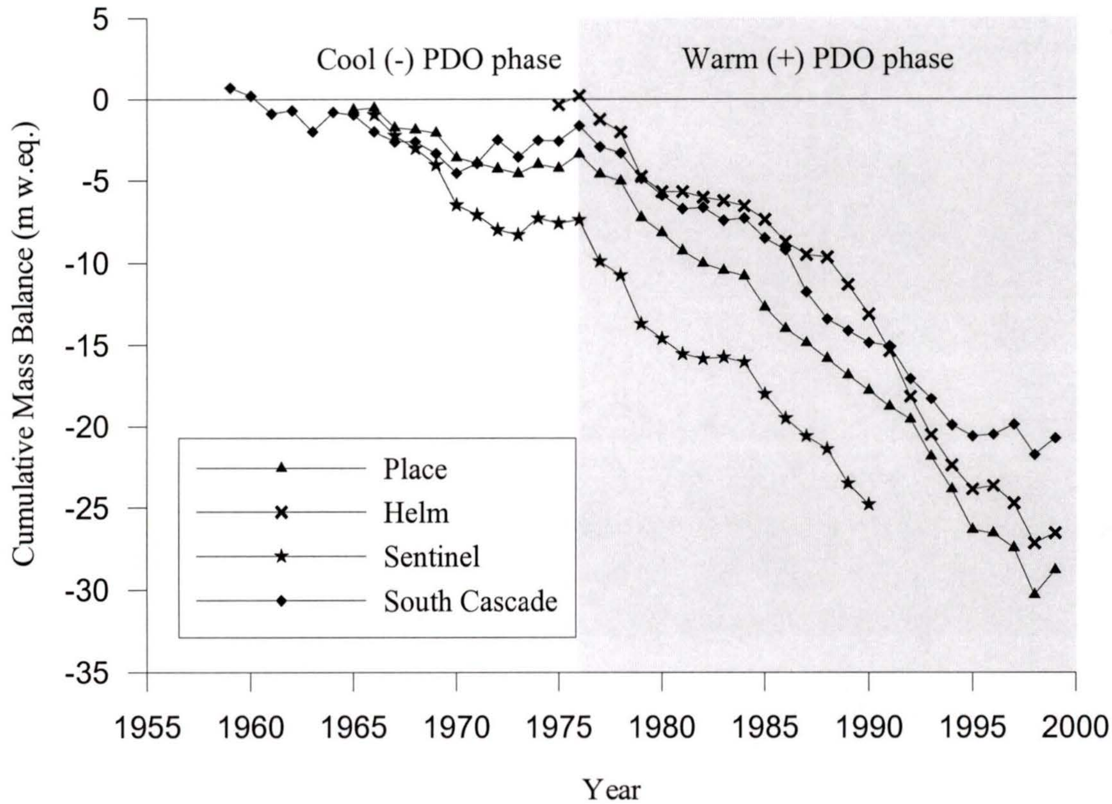


Figure 7.6. The effect of the PDO on cumulative mass balance of four glaciers in the Pacific Northwest (1966 – 1999). Grey area on the right side of the graph emphasises the 1976 shift to a warm (+) phase of the PDO.

The distinct climate shift of 1976 that resulted in warmer air temperatures, reduced winter precipitation (snow), and an overall reduction in winter storminess, is manifested in a change to strongly negative mass balance conditions throughout southern British Columbia and northwestern Washington State (Walters and Meier 1989; McCabe and Fountain 1995; McCabe and Legates 1995; Mantua *et al.* 1997; Cayan *et al.* 1998; Hodge *et al.* 1998; Bitz and Battisti 1999; McCabe *et al.* 2000). Similar patterns in climate variability have been recorded at least twice in the last century, and are regarded as symptomatic of the Pacific Decadal Oscillation (PDO) (Hare 1996; Mantua *et al.* 1997; Gedalof and Smith 2001b; Laroque and Smith 2001).

The PDO is a long-lived El Niño-like pattern of Pacific climate variability characterised by alternating regimes of warm and cool sea surface temperatures in North Pacific (Zhang *et al.* 1997). The positive phase of the PDO is characterized by an enhanced Aleutian Low and reduced storminess in the PNW. Storm tracks tend to be diverted away from the PNW towards Alaska, resulting in warmer, drier winters with below average snowpacks and increasingly negative winter mass balance. Conversely, the negative phase of the PDO is associated with a diminished Aleutian Low and increased winter storminess in the PNW, as the storm tracks are diverted away from Alaska, resulting in cooler air temperatures, increased precipitation, and greater snowpack depths in the PNW.

Recent mass balance-climate studies have found that interdecadal climate variability associated with the PDO is negatively correlated with the net winter balance of maritime glaciers in the PNW, accounting for 56 to 60 percent of winter mass balance variability (McCabe and Fountain 1995; Bitz and Battisti 1999; McCabe *et al.* 2000). The significant relationship between the PDO and winter mass balance of PNW glaciers is a result of greater variability in winter atmospheric circulation compared to summer circulation patterns, and as a result of the high sensitivity of maritime glaciers to changes in mesoscale atmospheric circulation (*i.e.*, winter storminess).

If the majority of the calibration period for the mass balance reconstruction falls within one phase of the PDO (*i.e.*, positive/warm phase), reconstructing mass balance anomalies associated with the opposite phase (cold phase) will likely be weakened as a result of significantly different climate conditions being used for calibration and (or) verification.

Correlation analysis between the reconstructed mass balance record (1600 – 1994) and reconstructed Pacific Northwest PDO indices of Gedalof and Smith (2001b) and Laroque and Smith (2001) are given in Table 7.2. The Gedalof and Smith (2001b) PDO index (PDOI) is derived from mountain hemlock chronologies located at sites from Alaska to California, whereas the Laroque and Smith (2001) PDOI is derived from mountain hemlock and yellow-cedar chronologies from sites on Vancouver Island.

Table 7.2. Correlation results for the reconstructed mass balance and PDO time series

	PDO (G&S)	PDO (L)	MB (reconstruction)
PDO (G&S)	1	-	-
PDO (L)	0.52	1	-
MB (reconstruction)	-0.44	**-0.65**	1

all *p*-values are 0.00 at the 99% confidence interval

Figure 7.7 illustrates the comparison between the reconstructed mass balance record and the reconstructed mean spring (March – May) PDO record of Laroque and Smith (2001) for the interval 1600 to 1995 AD. Intervals of negative PDO (cool/wet) phases correspond well with intervals of positive mass balance anomalies. Positive PDO (warm/dry) phases also show a good correspondence with periods of negative mass balance anomalies. While there are some intervals where the connection between the two is not as strong (*i.e.*, late 1600's prior to the 1700 moraines, and the mid 1700's), the overall relationship between the PDO and the Strathcona mass balance reconstruction is good.

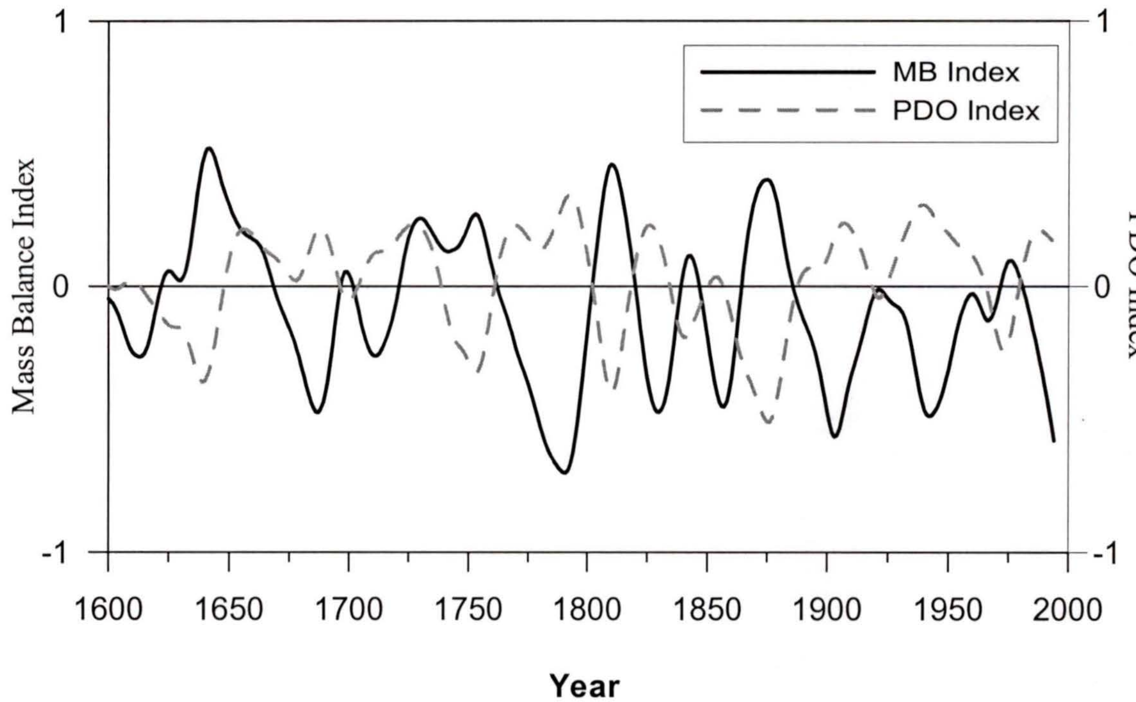


Figure 7.7. Reconstructed Strathcona mass balance index and mean spring (March – May) PDO Index. A 25-year cubic smoothing spline is fit to both records to emphasize the low frequency component. .

Finally, although both systems (glaciers and trees) are responding to the same climate forcing, albeit inversely, disparities between the mass balance response and the mountain hemlock growth response may also be related to the timing of the responses. Variability in the net mass balance of maritime glaciers in the PNW is strongly correlated to winter precipitation (snow). While this is also a limiting factor in mountain hemlock growth rates, the fact that the trees are not growing in the winter months means their response to winter precipitation is limited to spring snowpack depths and therefore, their response to winter precipitation is indirect whereas glacier response is direct. This response lag, therefore, may also contribute to a weakened relationship between the variability in glacier mass balance and the radial growth of mountain hemlock growing adjacent to the glaciers being studied.

CHAPTER 8. CONCLUSION

Little Ice Age glacier activity in Strathcona Provincial Park was investigated using both dendrochronological and lichenometric techniques. Moraine complexes from two small cirque glaciers within the park were dated and used to validate a regional mass balance reconstruction derived from a mountain hemlock ring-width chronology. Results show that glaciers on Vancouver Island are responding in a similar manner to glaciers in adjacent areas of coastal Pacific North America. Evidence from the two glacier forefields are in agreement with prior LIA research on Vancouver Island and indicate three main episodes of LIA glacier activity in Strathcona Provincial Park. The first episode is dated to sometime before 1397 AD, the second to the early 1700's (1706/08 AD), and the third to the late 1800's (1898 AD). The Vancouver Island LIA moraine chronology compares well with the coastal British Columbia LIA record, as well as both of the regional LIA records from maritime glaciers along the Washington-British Columbia-Alaska transect, and the continental glacier record of the Canadian Rocky Mountains

Lichenometric studies using *Rhizocarpon geographicum sp.* provide the first lichen-dating curve (1708 to 1998 AD) for this part of the PNW. The results show that the ecesis interval for *Rhizocarpon geographicum* in this area is only three years, and for mountain hemlock in the same environment is four years. The brevity of these ecesis intervals greatly reduces one of the main uncertainties in using geobotanical methods for dating late Neoglacial landforms. The moraine dates provided, therefore, give a good approximation of the shift in climate conditions that lead to the retreat of the glacier and stabilization of the moraines.

Dendrochronological and dendroglaciological methods proved to be more effective than lichenometry at Colonel Foster Glacier, which is the lower, and more protected of the two glacier sites. Situated below treeline, an abundant seed source ensures rapid colonisation of recently deglaciated surfaces. Conversely, lichenometry was more effective at the higher elevation site, Septimus Glacier, where the moraines are more exposed, and provide better conditions for optimal lichen colonisation. The upright growth form of trees is more susceptible to avalanche conditions at this site, resulting in limiting tree growth to two sites that are protected by bedrock outcrops. The Colonel Foster moraine chronosequence was derived solely from dendrochronological methods, whereas the Septimus moraine chronosequence utilised both lichenometry and dendrochronological methods.

Dating results from both sites suggest that the two glaciers are responding asynchronously to local variables, yet upon reaching a larger climatic threshold they begin to respond synchronously. Below the threshold, each glacier responds asynchronously to local factors such as the microclimate, topography, glacier geometry, and whether or not the glacier is terminating into a water body. Above the threshold, the glaciers respond synchronously to a stronger, regional climate signal.

Morphology of the moraine records also points to two levels of climatic forcing: (*i*) large-scale events that culminated with the synchronous deposition of the larger 1706 and 1898 AD moraines, and smaller 1935 AD moraine; and, (*ii*) small-scale events that likely instigate advances that are ultimately governed by local factors. These small-scale events are associated with the asynchronous deposition of moraines in the late 1700's at Colonel Foster and 1830's and 1870's moraines at Septimus Glacier.

Ring-width suppression records from trees growing adjacent to the glacier at each site suggests intervals with climatic conditions favouring positive mass balance regimes that are not preserved in the moraine record. Unfortunately, the large 1898 moraines at each site appear to have overridden and reworked morainic evidence of previous glacier advances, making direct comparisons to the moraine record somewhat difficult.

The reconstructed mass balance record (1966 - 1994) compares well with the combined instrumental record for PNW glaciers, explaining 50 percent of the variance. The success in reconstructing the recent record allowed for the reconstruction of a paleo-mass balance record from 1600 AD to 1994 AD. Verification of the mass balance reconstruction with the LIA moraine chronology from Pacific North America indicates that although the records don't correlate over the entire 1600 – 1994 AD interval, there are periods where the two records compares well. Reasons for the low correlation with parts of the moraine record may be: (*i*) a result of limited mass balance records for Canadian glaciers in southwestern British Columbia; (*ii*) the brevity of PNW mass balance records in general; (*iii*) a lack of dating constraints in the early part of the record; and (*iv*) the timing and abrupt changes in large scale atmospheric circulation patterns (*i.e.*, PDO) driving glacier mass balance in the PNW.

Finally, a comparison of the reconstructed Strathcona regional mass balance record with Laroque's (2001) reconstructed PDO index (1600 – 1994), reveals a significant relationship between the mass balance of Vancouver Island glaciers and large-scale climatic variability associated with the PDO. The promising nature of the preliminary results provides the groundwork for further studies in this area.

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Appendix 1

Tree ages and minimum surface dates for moraines at Colonel Foster Glacier

Moraine and Sample ID	Location on Moraine[1]	Ring Count	Pith Error	Estimated Tree Age	Estimated Moraine Age[3]
MORAINE 7W					
98CFWM-01a	Cr	200	15	250[2]	1744
98CFWM-01b	Cr	185	30	250[2]	1744
98CFWM-02a	Cr	240	15	290[2]	1704
98CFWM-03a	Cr	198	10	243[2]	1751
MORAINE 6W					
98CFWM-04a	Cr	221	5	261[2]	1733
98CFWM-04b	Cr	199	?	199	1795
98CFWM-05a	Cr	202	15	252[2]	1742
98CFWM-06b	Cr	213	20	268[2]	1726
98CFWM-07[4]	Cr	297	10	342[2]	1652
98CFWM-08	*DF*	*358*	*10*	*403*[2]	*1591*
MORAINE 5W					
98CFWM-09	*DF*	*286*	*0*	*286*	*1708*
98CFWM-10	DF	220	15	235	1759
98CFWM-11	PF	222	10	232	1762
98CFWM-12a	Cr	166	6	172	1822
98CFWM-12b	Cr	164	10	174	1820
98CFWM-13	Cr	-	-	-	-
98CFWM-14	Cr	204	15	219	1775
98CFWM-15	DF	131	8	139	1855
98CFWM-16a	DF	207	10	217	1777
98CFWM-16b	DF	213	4	217	1777
MORAINE 4W					
98CFWM-17	DF	202	3	205	1789
98CFWM-18	PF	138	8	146	1848
98CFWM-19	Cr	147	6	153	1841
98CFWM-20	PF	193	10	203	1791
98CFWM-21	PF	198	5	203	1791
98CFWM-22	PF	137	8	145	1849
98CFWM-23	*PF*	*214*	*10*	*224*	*1770*
MORAINE 3W					
98CFWM-24	PF	186	4	190	1804
98CFWM-25	PF	-	-	-	-
98CFWM-26	PF	165	5	170	1824
98CFWM-27	PF	134	5	139	1855
98CFWM-28	DF	206	0	206	1788

98CFWM-29	PF	201	10	211	1783
98CFWM-30a	DF	193	7	200	1794
98CFWM-30b	DF	185	15	200	1794
98CFWM-31	Cr	84	5	89	1905
98CFWM-32a	Cr	115	18	133	1861
98CFWM-32b	Cr	133	0	133	1861
98CFWM-33	*DF*	*206*	*10*	*216*	*1778*
98CFWM-34a	PF	118	5	123	1871
MORAINE 2W					
98CFWM-35	*PF*	*96*	*0*	*96*	*1898*
98CFWM-36	PF	79	5	84	1910
98CFWM-37	PF	82	0	82	1912
98CFWM-38	PF	52	10	62	1932
98CFWM-39	PF	84	0	84	1910
98CFWM-40	PF	54	5	59	1935
98CFWM-41a	PF	50	16	66	1928
98CFWM-41b	PF	56	10	66	1928
98CFWM-42	PF	76	3	79	1915
98CFWM-43a	PF	76	3	79	1915
98CFWM-43b	PF	64	15	79	1915
INTER-MORAINE AREA 1W/2W					
98CFWM-44a	BTWN	65	3	68	1926
98CFWM-44b	BTWN	65	3	68	1926
MORAINE 1W					
98CFWM-45	PF	39	3	42	1952
98CFWM-46	*PF*	*59*	*0*	*59*	*1935*
98CFWM-47	PF	42	0	42	1952
98CFWM-48a	PF	43	5	48	1946
98CFWM-48b	PF	38	10	48	1946
98CFWM-49a	PF	39	7	46	1948
98CFWM-50a	PF	42	4	46	1948
98CFWM-51	PF	38	0	38	1956
98CFWM-52a	PF	30	-	-	-
98CFWM-52b	PF	37	3	40	1954
98CFWM-53a	PF	41	3	44	1950

[1] PF = Proximal Face, CR = Crest, DF = Distal Face
[2] Tree age includes a coring-height correction of 35 years
[3] Surface date includes 4-year ecesis interval
[4] Yellow cedar, all other trees are mountain hemlock
Italics denotes limiting date

Tree ages and minimum surface dates for moraines at Septimus Glacier

Sample Site, and Tree ID	Tree Species	Location on Moraine[1]	Ring Count	Pith Correction	Estimated Tree Age	Estimated[4] Moraine Age
SITE A						
Moraine 2E						
98SEPT-A1a	MH	PF	77	5	82	1912
98SEPT-A1b	MH	PF	74	10	84	1910
98SEPT-A2	MH	PF	70	10	80	1914
98SEPT-A3	SAF	PF	77	10	87	1907
98SEPT-A5	*MH*	*PF*	*86*	*10*	*96*	*1898*
Moraine 3E						
98SEPT-A4a	SAF	PF	77	-	-	-
98SEPT-A4b	SAF	PF	103	10	113	1881
98SEPT-A6	*SAF*	*Cr*	*120*	*0*	*120*	*1874*
98SEPT-A7	*MH*	*Cr*	*120*	*0*	*120*	*1874*
98SEPT-A8	SAF	Cr	111	0	111	1883
SITE B						
Moraine 4E						
98SEPT-B1	SAF	DF	223	5	238[2]	1756
98SEPT-B2a	SAF	DF	231	10	256[3]	1738
98SEPT-B2b	SAF	DF	209	10	234[3]	1760
98SEPT-B3	YC	DF	-	-	0	-
98SEPT-B4a	MH	DF	329	15	359[3]	1635*
98SEPT-B4b	MH	DF	345	5	360[2]	1634*
98SEPT-B5	MH	DF	246	5	261[2]	1733
98SEPT-B6	MH	DF	230	15	260[3]	1734
98SEPT-B7a	MH	DF	209	-	-	-
98SEPT-B7b	*MH*	*DF*	*254*	*15*	*284[3]*	*1710*

[1] PF = Proximal Face, CR = Crest, DF = Distal Face
[2] Tree age includes a coring-height correction of 10 years
[3] Tree age includes a coring-height correction of 15 years
[4] Surface date includes 4-year ecesis interval
Italics denotes limiting date
* Located at base of moraine

Appendix 2

Net mass balance records from Place, Helm, Sentinel, and South Cascade glaciers

YEAR	Place	Helm	Sentinel	South Cascade
1994	-2010	-1885		-1600
1993	-2280	-2342		-1230
1992	-793	-2798		-2010
1991	-990	-2239		-200
1990	-938	-1790	-1300	-730
1989	-1010	-1670	-2100	-710
1988	-960	-150	-800	-1640
1987	-865	-790	-1100	-2560
1986	-1313	-1330	-1500	-710
1985	-1890	-825	-1950	-1200
1984	-340	-320	-300	120
1983	-440	-210	100	-770
1982	-750	-340	-300	80
1981	-1090		-950	-840
1980	-923	-940	-900	-1020
1979	-2221	-2730	-2950	-1560
1978	-433	-780	-850	-380
1977	-1227	-1460	-2500	-1300
1976	877	568	200	950
1975	-240	-357	-300	-50
1974	560		1000	1020
1973	-300		-300	-1040
1972	-340		-900	1430
1971	-340		-600	600
1970	-1510		-2450	-1200
1969	-210		-1000	-730
1968	-130		-800	10
1967	-1210		-1300	-630
1966	110		-950	-1030
Mean	**-800.2**	**-1178.3**	**-992.0**	**-652.8**
Std. Dev.	**755.0**	**953.9**	**903.8**	**916.3**

all values are listed in mm w.eq.

Vita

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WCAG Student Presentation Award (Hon. Mention)	1998
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Camosun College Chairman's Bursary	1991 and 1992

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Non-refereed contributions - specialized publications:

Lewis, D.H. and Smith, D.J. 1999. *Little Ice Age climate trends at treeline in Strathcona Provincial Park, Vancouver Island: insights from glaciers and trees*. Proceedings of the Workshop on Decoding Canada's Environmental Past: Adaptation Lessons Based on Changing Trends and Extremes in Climate and Biodiversity. Environment Canada. pp. 19-29.

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