# Dendroclimatological and dendroglaciological investigations at Confederation and Franklin glaciers, central Coast Mountains, British Columbia, Canada.

by

Bethany L. Coulthard
B.A., Mount Allison University, 2007

A Thesis Submitted in Partial Fulfillment
of the Requirements for the Degree of

MASTER OF SCIENCE

in the Department of Geography

# Supervisory Committee

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# Abstract

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It has become increasingly clear that climate fluctuations during the Holocene interval were unusually frequent and rapid, and that our current understanding of the temporal and spatial distribution of these oscillations is incomplete. Little paleoenvironmental research has been undertaken on the windward side of the central Coast Mountains of British Columbia, Canada. Very high annual orographic precipitation totals, moderate annual temperatures regulated by the Pacific Ocean, and extreme topographic features result in a complex suite of microclimate conditions in this largely unstudied area.

Dendroclimatological investigations conducted on a steep south-facing slope near Confederation and Franklin glaciers suggest that both mountain hemlock (*Tsuga mertensiana*) and subalpine fir (*Abies lasiocarpa*) trees at the site are limited by previous year mean and maximum summer temperatures. A regional subalpine fir chronology for the central and southern Coast Mountains indicates that subalpine fir trees at the study site experience physiological stress with warm summer temperatures, despite the high annual precipitation totals experienced there. This response is likely a result of the extreme gradient and the aspect of the slope at the sampling location, underscoring the importance of site characteristics on annual radial tree growth. Local (AD 1820-2008) and regional (AD 1700-2008) tree ring width chronologies were used to reconstruct previous July mean and maximum temperatures, explaining between 13% and 36% of the variance in climate. The proxy record features cool intervals that are comparable to other paleoenvironmental research from the region, and cyclical oscillations in temperature commonly associated with the El Niño Southern Oscillation and Pacific Decadal Oscillation. Century-scale fluctuations may be connected to changes in solar irradiance.

Dendroglaciological investigations were undertaken at the confluence of the Confederation and Franklin glaciers with the intention of exploring the Holocene behaviour of low-elevation maritime glaciers in this region. These glaciers are suspected to be sensitive to variations in the mean position of winter freezing level heights and warm winter temperatures, and may respond differently to changes in climate than more continental glaciers. Buried wood samples were radiocarbon-dated and cross-dated to construct three floating chronologies. Float A ($r$ = 0.467) suggests an early Little Ice Age advance of the two glaciers, and Float B ($r$ = 0.466) suggests an early Tiedemann advance of Confederation Glacier. Float C ($r$ = 0.519) is dated to the Garibaldi Phase of glacier expansion, but may not have been killed by glacial activity. The temporal synchronicity of these findings with glacial events documented throughout the region suggests a spatially coherent response of maritime and continental glaciers to the dominant climate–forcing mechanisms operating in Pacific North America throughout the late Holocene.

The dendroclimatological and dendroglaciological findings of this study help to fill a spatial research gap in the current understanding of Holocene climate variations in British Columbia. Because of the complex and at times topographically-controlled

response of conifers to climate in the study area, this region may provide a particular challenge in terms of reconstructing Holocene climate variability.

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# Chapter 1: Introduction

## 1.1 Introduction

It has become clear that climate conditions during the Holocene interval have been unusually variable (Mayewski *et al*. 2004) and that an understanding of the complexity of these changes has not yet been reached (Walker and Pellatt 2003). While a variety of paleoenvironmental indicators have been employed in an effort to reconstruct climate histories, few of these have the temporal resolution of annual tree ring width records. Sensitive tree ring series have been used to create proxy records of specific climate variables throughout Pacific North America (PNA), many of which are comparable to other paleoenvironmental reconstructions from the region (Wiles *et al*. 1998; Gedalof and Smith 2001a; Larocque and Smith 2005). Dendrochronological techniques have also informed our understanding of past environments through dendroglaciological investigations, which use glacially-killed trees to date periods of glacier advance and retreat, and corresponding warm and cool climate intervals. In PNA dendroglaciological findings are in broad agreement with comparable climate proxy records (Walker and Pellatt 2003; Menounos *et al*. 2009).

Dendroclimatological and dendroglaciological research has provided valuable insight into past temperature and precipitation regimes in many high-altitude environments in the southern Canadian Cordillera (Luckman *et al*. 1993; Luckman 2000; Gedalof and Smith 2001b; Wilson and Luckman 2003; Menounos *et al*. 2004; Wood and Smith 2004; Reyes *et al*. 2006; Koch *et al*. 2007a, 2007b; Osborn *et al*. 2007). However, virtually no dendroclimatological investigations have been conducted on the windward slopes of the central Coast Mountains of British

Columbia (B.C.), where very high annual precipitation totals, moderate coastal temperatures, and steep mountain topography results in a complex range of microclimate environments (Bancroft 1931; Tuller 2001). Similarly, no dendroglaciological investigations have been undertaken on low-elevation maritime glaciers in B.C., a common glacier type in the coastal mountains. Previous research has shown that low-elevation maritime glaciers may be more sensitive to warm winter temperatures and fluctuating winter freezing level heights than continental glaciers (Arendt *et al*. 2009).

## 1.2 Research goals and objectives

Two primary goals underlie this research: 1) to increase the spatial resolution of dendroclimatological studies in British Columbia through an investigation of climate-tree ring radial growth responses on the windward side of the Coast Mountains; and, 2) to document the Holocene behaviour of two low-elevation maritime glaciers in coastal British Columbia. Specific objectives were to:

1. determine the climatic variables influencing the radial growth of mountain hemlock (*Tsuga mertensiana*) and subalpine fir (*Abies lasiocarpa*) trees, and to explore the physiological influence of these factors on the study species;
2. create proxy climate reconstructions from tree ring width records collected at the study site;
3. create regional proxy climate reconstructions from a spatial network of tree ring width records; and,
4. use radiocarbon-dating and dendrochronological cross-dating techniques to reconstruct the Holocene behaviour of Confederation and Franklin glaciers.

## 1.3 Thesis format

This thesis consists of five chapters. Chapter One provides an introduction to the research and reviews the goals and objectives of this project. Chapter Two presents a review of dendroglaciological and dendroclimatological research that has been conducted in Pacific North America. Chapter Three presents dendroclimatic reconstructions of previous July temperatures using temperature-stressed trees from the windward central Coast Mountains. Chapter Four presents dendroglaciological findings from the Confederation and Franklin glaciers confluence area. Chapter Three and Four are formatted as manuscripts prepared for submission to refereed journals. Chapter Five summarizes the findings of the research, provides some concluding comments, and identifies study limitations and suggestions for further research.

# Chapter 2: Review of Dendroglaciological and Dendroclimatological Research in Pacific North America

## 4.1. Dendrochronology

The science of dendrochronology uses the annual growth rings of trees to determine the age and chronological order of past events (Fritts 1976). Because annual tree growth is largely a function of the limits to growth associated with climate variables, the successive annual growth rings contained in trees can also be used as a proxy for past climate conditions (Fritts 1976). Trees rings are important in that, with the exception of those limited by site-specific non climate-related variables, they can provide a temporally extensive (beyond the instrumental record) annually resolved and physically stationary record of past climate (Fritts 1976).

### *2.1.1 Principles of Dendrochronology*

Dendrochronological studies are based upon the principle of uniformitarianism which holds that the natural processes and interactions operating in the present are the same as those that were operating in the past (Hutton 1785). In the context of dendrochronology it is assumed that the physical and biological relationships between tree rings and climate have been consistent over time (Fritts 1976). Dendrochronology is also based upon the principle of limiting factors, according to which biological processes such as growth are regulated by the most limiting environmental variable (Fritts 1976). Dendrochronological studies are frequently designed to sample trees that are limited by a common climate variable. This can be achieved by sampling at the margins of a species' ecological amplitude, or range (Fritts, 1976).

### *2.1.2 Tree Ring Formation*

The radial growth of most trees at higher latitudes in the northern hemisphere is more rapid in the spring and early summer and generally declines by August (Kramer and Kozlowski 1960). As a rule growth ceases earlier at high altitudes (Daubenmire 1945). Diameter growth in the tree stem is initiated in the cambium, a lateral meristem that forms the divide between the bark, composed of a dead outer layer and a living inner layer, and the wood, which is composed of an outer layer of sapwood and dead inner heartwood (Bannan 1962). When frost leaves the ground and water becomes available in the spring cambium cell division is initiated, giving rise to cell division of xylem on the sapwood side and phloem on the inner bark side (Kramer and Kozlowski 1960; Bannan 1962).

The annual increment of xylem cells produced early in the season is referred to as earlywood, while the cells produced later in the season are referred to as latewood. Because the rate of radial growth is reduced at the end of the growing season, latewood is generally comprised of a larger number of cells per unit of area and appears darker than earlywood (Kramer and Kozlowski 1960). Earlywood cells grade into latewood cells, which end abruptly where they abut the earlywood of the following year (Kramer and Kozlowski 1960). These annual increments of earlywood and latewood form tree “rings”.

### *2.1.3 Dendroclimatology*

Outside of the equatorial regions the annual radial growth increment of trees is primarily a function of the limits to growth imposed by temperature and/or precipitation (Fritts 1976). As such, variations in the width of annual growth rings can be used as a proxy for past climate conditions (Fritts 1976). Dendroclimatological research is based upon, and affirms the existence of, large-scale geographic patterns of common annual tree-ring width variabilities that are related to climate (Hughes 2002). Understanding these relationships provides insight not only into the dynamics of existing forest communities, but also into how tree species distributions may change in response to changing climate (Laroque and Smith 2003; Larocque and Smith 2005; McKenney *et al*. 2007), insect (Swetnam and Lynch 1989; Mason *et al*. 1997), fire (Larsen 1996; Stephens *et al*. 2003), and other disturbances (Briffa *et al*. 1998). The discipline relies upon the continued development of extensive networks of tree-ring chronologies that meet common standards (Hughes 2002).

#### *2.1.4 Dendroglaciology*

Dendroglaciology is the study of glacial processes and history through the dendrochronological dating of glacier landforms and advances (Smith and Lewis 2007). The approximate date of surface stabilization of a relatively young (hundreds of years) glacial deposit can be ascertained by establishing the age of a tree growing on its surface and by taking into account an estimated period of ecesis, i.e. the time interval between the point of surface stabilization and when plants begin to colonize the surface (McCarthy and Luckman 1993).

Dendroglaciological techniques can also be used to assign relative dates to much older glacial deposits through the dendrochronological and/or radiocarbon-dating of trees that have been overrun, scarred, buried and/or killed directly by an advancing glacier or by the construction of a glacial landform (i.e. Luckman 1998). If a tree remains in growth position (i.e. rooted in a paleosol) the age of the outermost tree ring may delineate the position of the ice margin at the time of death (i.e. Luckman 1995; Wood and Smith

2004). The position of *in situ* samples — that is samples recovered from the position where they were deposited by glacial activity — can also shed light on past glacier dynamics. Scattered tree boles and detrital wood fragments located on glacial deposits and forefields can be used to build floating tree ring-width chronologies that, when either cross-dated or radiocarbon-dated, provide an estimate of the timing of surface stabilization and corresponding age of maximum glacial advance (i.e. Jackson *et al*. 2008). A floating chronology can be linked to a particular location, landform, and/or time period if cross-dated to an *in situ* sample or sample in growth position.

Many dendroglaciological samples originate from high altitude or latitude environments where climate strongly limits the radial growth of trees (Tranquillini 1979). These temperature or precipitation sensitive tree-ring chronologies can be used to reconstruct past climates and provide useful proxies for the reconstruction of summer and winter glacier mass balance (Larocque and Smith 2005; Smith and Lewis 2007). Such investigations have been employed to highlight the long-term relationships between glacier mass balance dynamics and various climate-forcing mechanisms (Luckman 1986; Larocque and Smith 2005; Allen and Smith 2007; Koch *et al*. 2007a; Barclay *et al*. 2009).

## 2.2 Regional Climate-Forcing Mechanisms

In Pacific North America (PNA) radial tree growth and regional climates are influenced by a suite of climate-forcing mechanisms (Gedalof and Smith 2001a). Bonsal *et al*. (2001) indicate that the El Niño Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) explain a significant portion of climatic variance in western

Canada, with the latter having slightly more influence, especially in extreme western regions.

### *2.2.1 The El Niño Southern Oscillation (ENSO)*

ENSO is a coupled ocean-atmospheric process that, while centered on the equatorial Pacific region, constitutes the largest single source of interannual climatic variability around the globe (Diaz and Markgraf 1992). Prior to understanding the dual nature of the phenomenon, the ocean circulation component of ENSO was referred to as El Niño (opposite phase La Niña), while the atmospheric component was referred to as the Southern Oscillation (Rasmussen and Wallace 1983). It is now recognized that these processes interact to produce ENSO events: recurrent weather and climate anomalies defined by changes in atmospheric circulation and sea surface temperatures (SST) in the equatorial Pacific (Diaz and Markgraf 1992). These events occur at varying temporal and spatial scales, but usually take place every 7 to 10 years (Rasmussen and Wallace 1983).

ENSO events are broadly defined by changes in the equatorial trade wind systems resulting in a reduction of the "normal" cross-Pacific sea level pressure gradient. A see-saw effect occurs between the southeast Pacific subtropical high and the region of low pressure usually centered on the Indian Ocean (Rasmussen and Wallace 1983). Simultaneously, warm water in the western equatorial Pacific is displaced eastward where it suppresses cold-water upwelling in the eastern equatorial Pacific, resulting in increased SSTs off the west coast of South America (Rasmussen and Wallace 1983). These changes in tropical circulation are expressed in the extratropics due to large-scale Rossby-wave patterns that result in characteristic rainfall and temperature responses in specific "ENSO areas" around the world. In PNA, ENSO events are usually expressed as a negative center over the North Pacific, a positive centre over western Canada, and a negative centre over the southeastern United States (Rasmussen and Wallace 1983). In western Canada, ENSO events are generally correlated to warmer- and drier-than-normal winter (January — February —March) conditions (Shabbar and Khandekar 1996; Bonsal *et al*. 2001).

### *2.2.2 The Pacific Decadal Oscillation (PDO)*

The PDO is a sea surface temperature (SST) anomaly the influence of which is predominantly felt during winter months in the extratropics, particularly in the north Pacific and in North America (Mantua and Hare 2002). Although the phenomenon is thought to originate in the tropics (Evans *et al*. 2001), the specific cause(s) and mechanism(s) of the PDO are not yet fully understood. This said, some temporal and spatial characteristics of the PDO are generally recognized.

Temporally the PDO is characterized by two phases, a warm (positive) phase and a cool (negative) phase. It has been shown to operate at a range of timescales including interannual (< 8 years), decadal (10-20 years), and interdecadal (30-70 years) modes (Gedalof and Smith 2001a). PDO regimes are thought to shift abruptly, as evidenced by the 1976/77 shift. Other historic regime shifts have been identified in the 1920s, 1940s, and 1970s (Mantua and Hare 2002). Gedalof and Smith (2001a) used tree ring records to identify six shifts between 1650 and 1850. Their reconstruction suggests that PDO phases have an average period length of 23 years, and indicate that a quiescent period in the interdecadal mode of the PDO may have occurred between 1840 and 1930.

Warm phases of the PDO are characterized by anomalously cool SSTs in the central north Pacific and abnormally warm SSTs along the Pacific coast of North

America. Warm phases are also characterized by low sea level pressure over the north Pacific, resulting in increased wind moving counter clockwise toward the North American coast. Cool phases of the PDO result in the opposite effects (Mantua and Hare 2002). The climatic outcomes of these SST and wind regimes are complex and exert varying influences throughout the entire Pacific region. During the warm PDO phase, dry conditions persist throughout interior Alaska and much of PNA. Wet conditions dominate the Gulf of Alaska and the southwestern United States (Mantua and Hare 2002). In terms of temperature, anomalously warm conditions are experienced in northwestern North America and the southeastern United States during warm phases (Mantua and Hare 2002). Reversed temperature and precipitation regimes affect these areas in cool phases (Mantua and Hare 2002).

### *2.2.3 Summary*

It is difficult to discern the impact of these mechanisms on PNA climates due to the interactions between different oscillations and variations in their spatial and temporal extent (Bonsal *et al*. 2001; Papineau 2001; Bond and Harrison 2006). Bonsal *et al*. (2001) suggest that when El Niño (La Niña) events occur in association with a positive (negative) PDO, a significantly stronger winter temperature response is observed over western Canada then if they had been operating independently. Other unstable modes of climatic variability influencing climate in PNA include the Arctic Oscillation, Pacific North American Pattern, and the Aleutian Low (Zhang *et al*. 1997; Overland *et al*. 1999; Bonsal *et al*. 2001; Schneider and Cornuelle 2005; Bond and Harrison 2006). Despite the complex nature of these climate phenomena, PDO (D’Arrigo *et al*. 2001; Larocque and

Smith, 2005) and ENSO (Larocque and Smith, 2005; Watson *et al*. 2006) signals have been shown to correlate to tree-ring data in PNA.

## 2.3 Climate-Radial Growth Responses

### *2.3.1 Mountain hemlock* (Tsuga mertensiana)

The most pervasive mountain hemlock climate-growth response reported in PNA is a positive correlation between ring width and current-year summer temperature. This relationship is documented in the central Coast Mountains of B.C. (Larocque and Smith 2005), the Cascade and Olympic ranges in Washington and Oregon (Graumlich and Brubaker 1986; Peterson and Peterson 2001), and at coastal sites ranging from northern California to Alaska (Wiles *et al*. 1998; Gedalof and Smith 2001b). Warm summers increase radial growth by regulating soil temperatures, rates of respiration and photosynthesis, metabolic processes, and consequent carbohydrate production (Gedalof and Smith 2001b). They also mitigate the reduced radial growth that occurs during cone crop years and increase snowmelt and water availability in high-elevation environments (Woodward *et al.*1994; Peterson and Peterson 2001; Larocque and Smith 2005). Mountain hemlock trees demonstrate a ubiquitous negative growth response to spring snowpack depth in the Coast Mountains (Larocque and Smith 2005), the northwestern United States (Graumlich and Brubaker 1986; Peterson and Peterson 2001) and on Vancouver Island (Laroque *et al*. 2001). Smith and Laroque (1998) report that when seasonal snow packs exceed 4 m in depth, mountain hemlock radial growth is significantly reduced, regardless of the growing season temperature. Graumlich and Brubaker (1986) also found that deep spring snow packs overwhelm the positive influence of warm spring temperatures on radial growth trends in the Cascade Ranges of


Figure 2.1: Map of major mountain regions in Pacific North America.

Washington state. The pronounced negative response to spring snowpack depth is commonly explained by the reduced duration of seasonal cambial activity and photosynthesis that occurs as a result of burial under snowpack (Gedalof and Smith 2001b; Peterson and Peterson 2001; Laroque 2002). Laroque and Smith (2001) found that trees on Vancouver Island were positively correlated with mean April temperature and suggest that when mountain hemlock trees are not buried under spring snow packs radial growth is enhanced by earlier bud burst (Owens 1984) and activation of earlywood growth.

A negative response to previous year spring/summer temperature was found at high-elevation sites in the Cascade and Olympic ranges in Washington and Oregon state (Graumlich and Brubaker 1986; Peterson and Peterson 2001), Vancouver Island (Laroque and Smith 2001), and at coastal sites ranging from northern California to Alaska (Gedalof and Smith 2001b). This relationship was attributed by Peterson and Peterson (1994) to summer drought effects that limit photosynthetic production and initiate higher respiration rates. Larocque and Smith (2005) report the opposite response on the eastern slopes of the Coast Mountains, however, which they attribute to the alleviation of summer moisture deficits and a resultant extended growing season (Peterson and Peterson 1994).

Less common mountain hemlock climate-growth responses reported in PNA include a positive correlation with current August precipitation at high-elevation sites on Vancouver Island (Laroque and Smith 2001) and a positive correlation with previous January air temperatures in the Mount Waddington area (Larocque and Smith 2005). Current-year August precipitation reduces the effects of moisture deficit (Peterson and

Peterson 1994), while warmer than normal air temperatures in January are thought to enhance the overwintering capacity of vegetative buds, thus increasing photosynthetic potential (Larocque and Smith 2005). In general, spring snowpack and summer temperature are reported as the primary climatic limits to the radial growth of mountain hemlock throughout much of its geographic range (Peterson and Peterson 2001).

#### *2.3.2 Subalpine fir* (Abies lasiocarpa)

Reported subalpine fir climate radial-growth relationships in PNA are similar to those observed in mountain hemlock trees. A positive response to current-year spring and/or summer air temperature is well documented and appears to be the most pervasive climate-growth signal (Villalba *et al*. 1994; Parish *et al*. 1999; Spelchtna *et al*. 2000; Luckman *et al*. 2002; Larocque and Smith 2005). This relationship was reported in the Prince William Sound region of Alaska (Barclay *et al*. 1999), the Olympic Peninsula (Ettl and Peterson 1995; Peterson *et al*. 2002), in the southern Canadian Rocky Mountains (Wig and Smith 1994), and in the Cascade Ranges of Washington state (Heikkinen 1985; Peterson *et al*. 2002). Like mountain hemlock trees, this relationship is assumed to be related to an extension of the growing season associated with the seasonal melting of lingering snow packs (Peterson and Peterson 1994; Ettl and Peterson 1995). The benefits of warm summer air temperatures are particularly important during periods of reduced growth associated with cone crop years (Woodward *et al*. 1994; Ettl and Peterson 1995). Negative correlations to winter and spring precipitation are reported at the same sites and reflect a response to snowpack similar to that of mountain hemlock (Peterson and Peterson 2001). A negative response of subalpine fir radial growth to previous summer temperature is reported in the central Coast Mountains (Larocque and Smith 2005), on the Olympic Peninsula (Ettl and Peterson 1995; Peterson *et al*. 2002), in the southern Canadian Rocky Mountains (Wig and Smith 1994) and in the Cascades (Peterson *et al.* 2002), and is likely related to moisture stress, which has been shown to limit radial growth in the following year (Peterson *et al*. 2002). Interestingly, Larocque and Smith (2005) found that subalpine fir trees correlate positively with previous fall air temperature in the central Coast Mountains and suggest that this reflects an extended period of nutrient storage, leading to enhanced radial growth in the following growing season (Peterson and Peterson, 1994; Ettl and Peterson, 1995). In general, snowpack and summer temperature serve as the primary limits to the growth of subalpine fir trees at wet, high-elevation sites (Peterson and Peterson 1994; Ettl and Peterson 1995).

## 2.4 Holocene Glaciation in the Pacific Northwest

The Pleistocene epoch extended from 2.5 million to 12 000 cal. yrs BP and encompassed several ice ages, the most recent of which resulted in the advance of the Cordilleran Ice Sheet over much of the western Canadian Cordillera including the Coast Mountains south of $60^{o}$ N (Ryder *et al*. 1991). Subsequent climate warming and ice decay led to glacial conditions in B.C. similar to those at present by 11 500 cal. years BP (Menounos *et al.* 2009).

Following this latest ice age, the Holocene or post glacial epoch began in ca. 11 500 cal. yrs BP and extends to present (Roberts 1998). The Holocene has been characterized by relatively significant and rapid climate changes (Mayewski *et al*. 2004). Based on palynological reconstructions of vegetation and climate, Hebda (1995) indicates that three phases characterize Holocene climates in B.C.: a warm, dry “xerothermic” interval from 9500 to 7000 cal. yrs BP; a warm, moist "mesothermic" interval from 7000 to 4500 BP; and, a moderate and moist interval from 4500 cal. yrs BP to the present (Hebda 1995). These phases follow a general trend of increasing moisture from 8500 to 6000 cal. yrss BP and increased cooling from 4500 to 3000 cal. yrs BP (Hebda 1995). The magnitude and timing of these climate trends differ to varying extents throughout the region (Hebda 1995).

The Holocene glacial history discussed here focuses on the regions and events most relevant to this research, namely those that occurred in the Coast Mountains, in Alaska, in the Cascade and Olympic ranges in Washington, and in the Canadian Rocky Mountains.

### *2.4.1 The Early Holocene*

The period from 11 500 to 7800 cal. yrs BP is considered the early Holocene (Figure 2.2), with the “xerothermic” interval extending from 9500 to 7000 cal. yrs BP. During the xerothermic interval climate conditions were warmer and drier than they are at present in both coastal and interior B.C. (Mathewes 1985). Pollen records suggest that xerothermic conditions in B.C. peaked around 7500 to 7000 yrs BP (Mathewes 1985; Hebda 1995).

Glacial history reconstructions from the Sierra Nevada in California (Clark and Gillespie 1997), and Mount Rainier (Heine 1998) and Mount Baker (Thomas *et al*. 2000) in Washington State, as well as lake sediment records and detrital wood evidence from glacier forefields in the southern Coast Mountains (Menounos *et al*. 2004), provide evidence for an *early Holocene glacial advance* in western North America at ca. 8000 cal. yrs BP (Reasoner *et al*. 2001; Menounos *et al*. 2004). It has been suggested that this event was much smaller in both temporal and spatial magnitude than better-known late Holocene advances (Menounos *et al*. 2004).

Radiocarbon-dated lake sediments from Mount Baker and Mount Rainier suggest that this event may comprise two distinct periods of advance, one between ca. 8400 and 9450 cal. yrs BP and a second between 10 000 and 10 900 cal. yrs BP (Heine 1998; Thomas *et al*. 2000). Menounos *et al*. (2004) notes the concurrence of this early Holocene glacial advance with the well-documented 8200-year cold event recorded in the North Atlantic region (Alley *et al*. 1997). This relationship would suggest that large-scale climatic linkages existed between the North Atlantic and North Pacific regions at this time (Menounos *et al*. 2004).

Despite existing glacial and lacustrine evidence, a period of positive glacial mass balance during the early Holocene is discordant with contemporaneous reconstructions of summer air temperature (Clague and Mathewes 1989; Hebda 1995; Pellatt and Mathewes 1997; Palmer *et al*. 2002) and calculated summer insolation values (Berger 1978). These reconstructions suggest that the climate in PNA was significantly warmer than it is at present, with estimates ranging between 1 and 4°C (Reasoner *et al*. 2001).

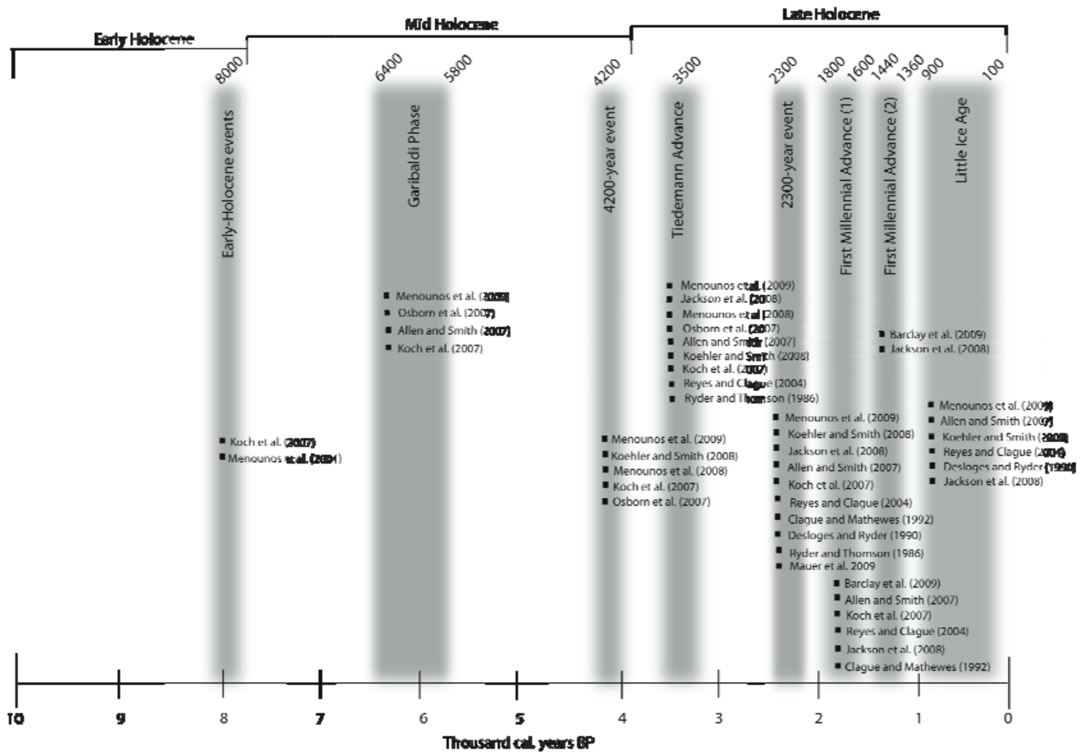


Figure 2.2: Schematic of Holocene glacial history in the B.C. Coast Mountains. Bulleted references indicate research that has contributed dendroglaciological evidence supporting the associated advance.

### *2.4.2 The Mid Holocene*

The interval from 7000 to 4500 cal. yrs BP is considered the mid Holocene or mesothermic period (Hebda 1995). Rising lake levels, bog expansion, forest encroachment, and expanded ranges of moisture-adapted trees suggest that this was a period of transition during which xerothermic conditions gradually gave way to a cooler and wetter climate (Mathewes 1985). Hebda (1995) defines the onset of the mesothermic as the time when annual precipitation rose to present day levels, while temperatures remained warmer than present. He suggests that the end of the mesothermic was reached when mean annual temperatures reached those of the present day.

Following early Holocene ice advances glaciers likely receded before readvancing during the *Garibaldi Phase* ca. 6400 – 5800 cal. yrs BP (Ryder and Thomson 1986). Dates from glacially overridden tree stumps in growth position in Garibaldi Provincial Park as well as roots on a nearby nunatak provided the initial evidence for a phase of glacial expansion at this time (Lowden and Blake 1975; Ryder and Thomson 1985). Dendroglaciological evidence of this event is found throughout the Coast Mountains (Allen and Smith 2007; Koch *et al*. 2007b; Osborn *et al*. 2007; Menounos *et al*. 2008), in the Cascade Range of Washington (Miller 1969), and in mountain ranges in Europe, the Himalaya, New Zealand, and the Andes (Thompson *et al*. 2006). It is unclear whether Garibaldi glacial expansion was slow and continuous or the result of two separate advances in the Coast Mountains (Koch 2006; Koch *et al*. 2007b).

Garibaldi ice expansion was followed by a period of recession before a readvance ca. 4200 cal. yrs BP (Menounos *et al*. 2008). Termed the *4200-Year Event*, dendroglaciological evidence of this advance comes from dendroglaciological discoveries in the Canadian Rocky Mountains (Gardner and Jones 1985; Luckman 1995; Wood and Smith 2004) and in the Coast Mountains (Osborn *et al*. 2007; Koch *et al*. 2007b; Koehler 2009; Menounos *et al*. 2008; Menounos *et al*. 2009).

This mid Holocene advance is thought to have been less extensive than subsequent late Holocene advances, as evidenced by *in situ* radiocarbon-dated wood samples from glacier forefields throughout western Canada (Menounos *et al*. 2008). Positive mass balance conditions are thought to have persisted for only several decades to a century (Osborn *et al*. 2007; Koch *et al*. 2007b; Menounos *et al*. 2008; Koehler 2009; Menounos *et al*. 2009). Proxy climate records corroborate the existence of colder and wetter climate conditions in the region during this interval, with notable changes in both coastal and interior B.C. vegetation at ca. 4000 cal. yrs BP (Hebda 1995; Viau *et al*. 2002; Mayewski *et al*. 2004; Booth *et al*. 2005; Zhang and Hebda 2005; Menounos *et al*. 2008).

### *2.4.3 The Late Holocene*

An interval of relative cooling, termed the Neoglacial or late Holocene period (Figure 2.2), was initiated in the Coast Mountains ca. 3800 cal. yrs BP and has persisted until the present (Mathewes 1985; Hebda 1995). The mid Neoglacial *Tiedemann Advance* was initially documented at sites in the Coast Mountains and was proposed to have taken place between 3000 to 1900 cal. yrs BP (Ryder and Thomson 1986). Recent discoveries suggest that it consists of an early and late phase (Allen and Smith 2007; Koch *et al*. 2007b; Koehler 2009; Menounos *et al*. 2009). Distinct climate episodes during the Tiedemann period are highlighted in a bog near Tiedemann Glacier where three distinct pollen zones from before 2600 $^{14}$C yrs BP, from 2500 to 2300 $^{14}$C yrs BP, and after 1900 $^{14}$C yrs BP are reported (Arsenault *et al*. 2003).

Evidence of an *early Tiedemann* glacier advance at ca. 3500 cal. yrs BP is reported from sites located in the Coast Mountains at Beare River Glacier (Haspel *et al*. 2005; Spooner *et al*. 2005), Surprise Glacier (Jackson *et al*. 2008), and Manatee Glacier (Koehler 2009). Palynological evidence from Marion Lake (Mathewes and Heusser 1981) and chronomid assemblages in North Crater Lake and Lake of the Woods in southern B.C. (Palmer *et al*. 2002) suggest cold and wet environments at this time. Subalpine ponds on the Queen Charlotte Islands (Pellatt and Mathewes 1997), and radiocarbon dates from fossil plant material in a proglacial lake at Berendon Glacier (Clague and Matthewes 1996) suggest similar conditions in northern B.C. A decrease in the fire frequency in mountain hemlock forests in southwestern B.C. at ca. 3500 cal. yrs BP highlights a shift to Neoglacial climate conditions at this time (Hallett *et al*. 2003).

Early Tiedemann age glacial activity has also been documented in the mountain ranges of the western United States (Burke and Birkeland 1983; Clark and Gillespie 1996) and in Alaska (Wiles and Calkin 1990; Calkin *et al*. 2001; Wiles *et al*. 2002). In the Canadian Rocky Mountains contemporaneous glacial activity is recorded at Peyto, Saskatchewan, Robson, and Yoho glaciers (Luckman *et al*. 1993; Wood and Smith 2004)

At sites in the Coast Mountains, Tiedemann-aged deposits distal to Little Ice Age (LIA) moraines suggest that some glaciers reached their maximum Holocene extent during this interval (Ryder and Thompson 1986; Larocque and Smith 2003). Significantly increased clastic sedimentation rates in proglacial lakes in the southern Coast Mountains around 3000 $^{14}$C yrs BP are also comparable to rates associated with the LIA (Osborn

2007; Minkus 2006). Menounos *et al*. (2009) highlight the possibility of multiple early Tiedemann ice advances.

The early Tiedemann advance appears to have been followed by an interval of ice retreat that persisted for several centuries, before the late Tiedemann advances began. Palynological and lake sediment evidence from the southern interior of B.C. suggest cool and moist conditions at this time (Pellatt *et al*. 2000; Spooner *et al*. 2003; Lamoureux and Cockburn 2005).

Evidence for a *late Tiedemann* (*2300-year*) *Advance* comes from investigations at Manatee Glacier in the southern Coast Mountains (Koehler 2009), where a 250-year-old *in situ* glacially-killed tree growing on a glacial forefield was radiocarbon-dated to 2340 +/- 70 $^{14}$C yrs BP (Koehler 2009). Buried trees and sheared stumps of similar age have been found in growth position above early Tiedemann-age paleosols on glacial forefields and in lateral moraines at sites throughout the Coast Mountains, notably at Tiedemann Glacier (Ryder and Thomson 1985), Tide Lake (Clague and Matthewes 1992), Beare Glacier (Johnson *et al*. 1997), Lillooet Glacier (Reyes and Clague 2004), Bridge Glacier (Allen and Smith 2007), at various glaciers in Garibaldi Provincial Park (Koch *et al*. 2007b), and at Castle Creek Glacier (Mauer *et al*. 2009). Analogous late Tiedemann advances are documented in the Canadian Rocky Mountains at Peyto and Robson glaciers (Luckman *et al.* 1993), Stutfield Glacier (Osborn *et al*. 2001), and Peyto Glacier (Luckman 2006).

Following the late Tiedemann advance, radiocarbon and lichenometric evidence from glaciers in coastal British Columbia and Alaska provide widespread evidence of a *First Millennial Advance* (FMA) from ca. 1800 to 1300 cal. yrs BP (Reyes *et al*. 2006).

Glacier expansion into standing forests during this interval is reported from sites in the southern Coast Mountains (Reyes and Clague 2004; Allen and Smith 2007; Koch *et al*. 2007b), the northern Coast Mountains (Laxton 2005; Jackson *et al*. 2008), in coastal Alaska (Barclay *et al*. 2009), and in the Cariboo Mountains (Mauer *et al*. 2009). Glaciers may have attained ice-front positions comparable to those later reached during the LIA (Reyes *et al*. 2006; Jackson *et al*. 2008; Mauer *et al*. 2009). Dendrochonological and stratigraphic investigations in Alaska and the northern Coast Mountains suggest two distinct FMA phases, one from ca. 1360 to 1440 cal. yrs BP and a second from ca. 1600 to 1800 cal. yrs BP (Jackson *et al*. 2008; Barclay *et al*. 2009).

Synchronous FMA advances are reported from sites in the Canadian Rocky Mountains (Luckman 1996), Iceland (Gudmundsson 1997), the European Alps (Holzhauser *et al*. 2005), and New Zealand (Gellatly *et al*. 1988). A variety of proxy climate reconstructions suggest generally cool temperatures in PNA and elsewhere around the globe during this time (Bond *et al*. 1997; Hu *et al*. 2001; Pierce *et al*. 2004; Moberg *et al*. 2005). The timing of the FMA advance is consistent with the previously-suggested notion of millennial-scale climate cycles in the Pacific Northwest (Denton and Karlén, 1973; Reyes *et al*. 2006).

With the onset of the *Little Ice Age* (LIA) ca. 1000 cal. yrs BP, glaciers began to expand and advance worldwide (Grove 1988). Dendroclimatological reconstructions suggest that temperatures in southern British Columbia at this time were more than 1 ºC cooler than at present (Graumlich and Brubaker 1986; Luckman 2000). In the Coast Mountains three distinct intervals of LIA ice expansion are recognized, from ca. AD 1100 to 1200, from AD 1600 to 1700, and from AD 1800 to 1900 (Ryder and Thomson

1985; Desloges and Ryder 1990; Smith and Desloges 2001; Larocque and Smith 2003; Lewis and Smith 2004a; Reyes and Clague 2004; Koch *et al*. 2007a; Jackson *et al*. 2008; Menounos *et al*. 2008; Koehler 2009). LIA advances culminated between AD 1600 to 1900 (Menounos *et al*. 2008), at which point terminal and lateral moraines representing maximum Holocene glacial extent were built at most sites in the Coast Mountains (Menounos *et al*. 2009).

Dendroglaciological and lake sediment analyses from northern B.C. (Spooner *et al*. 2005; Jackson *et al*. 2008) and dendroglacialogical evidence from central B.C. (Allen and Smith 2007) suggest an earlier onset of the LIA in the Coast Mountains than elsewhere in PNA. Allen and Smith (2007) report evidence of moraine stabilization at Bridge Glacier at AD 1367. It is thought that the distinct periods of expansion and retreat during the LIA occurred in response to climate fluctuations related to variations in solar insolation over the last millennia (Wiles *et al*. 1999b; Larocque and Smith 2003; Barclay *et al*. 2009).

Comparable LIA activity is described from sites throughout PNA, notably at Mount Baker and Mount Rainier in the Cascades (Miller 1969; Sigafoos and Hendricks 1972; Fuller 1980; Burbank 1981; Heikkinen 1984), in Alaska (Wiles *et al*. 1999a,1999b; Calkin *et al*. 2001; Barclay *et al*. 2009), and in the Canadian Rocky Mountains (Luckman 1986; Smith *et al*. 1995; Luckman 2000).

## 2.5 Summary

An increasingly complex understanding of historical climate and mass balance oscillations over the past 10 000 years emphasizes the importance of continued inquiry into the nature of Holocene climate and ecosystem changes. Dendroclimatological and dendroglaciological investigations have contributed to this growing body of research, providing some of the few annually resolved records of past climate. In concert with other paleoenvironmental evidence, the ongoing collection of tree ring records and dendroglaciological samples throughout PNA will provide further insight into the dynamic Holocene environmental history of this region.

# Chapter 3: Dendroclimatological reconstruction of Mean and Maximum July Temperatures in the central Coast Mountains of British Columbia, Canada

## 3.1 Introduction

A growing recognition of the long-term consequences of ongoing global climate changes has underscored the importance of understanding past environmental dynamics (Pauli *et al*. 1996; Walther *et al*. 2002). This is especially true for vulnerable mountain ecosystems where climate changes are already having significant impacts (Diaz *et al*. 2003). In the absence of long-term instrumental climate records in many high-elevation areas, tree rings have been used to provide robust annually resolved proxy records of past climate with both temporal and spatial resolution (e.g. LaMarche 1974; Briffa *et al*. 1992; Luckman 1994; Cook *et al*. 2003; Frank and Esper 2005).

In the western Canadian Cordillera, dendroclimatological methods have been used to reconstruct detailed temperature and precipitation histories over the past 400 to 500 years. In many settings, these tree ring-based proxy climate histories are corroborated by other paleoclimate proxies (Walker and Pellatt 2003). Combined, these records provide increasingly detailed histories of Holocene climate fluctuations and their impacts on ecosystem changes in the Canadian Rocky Mountains (Walker and Mathewes 1989; Leonard and Reasoner 1999; Pellatt *et al*. 2000; Wilson and Luckman 2003; Wood and Smith 2004), the northern Coast Mountains (Gilbert *et al*. 1997; Penrose 2007; Jackson *et al*. 2008; Laxton 2008), the southern Coast Mountains (Lowe *et al*. 1997; Osborn *et al*. 2007; Gedalof and Smith 2001a; Palmer *et al*. 2002; Menounos *et al*. 2004; Menounos *et al*. 2005; Reyes *et al*. 2006; Allen and Smith 2007), and in the Vancouver Island Insular

Mountains (Laroque and Smith 2003; Laroque and Smith 1999; Gedalof and Smith 2001a; Brown and Hebda 2002; Lewis and Smith 2004a, 2004b).

In the central Coast Mountains of British Columbia (B.C.) most prior dendroclimatological research has been completed at sites located along the eastern slopes (i.e. Larocque and Smith 2003, 2004, 2005). These slopes are comparatively dry and warm in the summer and dry and cold during the winter due to rain shadow effects and the influence of continental air masses. By contrast, only a limited amount of dendroclimatological research has been undertaken at sites located along the western windward slopes of the Coast Mountains (i.e. Penrose 2007), where the proximity to the Pacific Ocean results in a distinct maritime climate (Tuller, 2001). Characterized by substantial orographic precipitation, winter snow packs often reach over 5.4 m in depth at higher elevations (avg. max May 1$^{st}$ snowpack at Bella Coola, Toba River, and Mt. Seymore, B.C. Ministry of Environment Historic Snow Survey Data 1960-1991). The only regional insights into the radial growth response of conifer trees in comparable high-elevation settings comes from research completed in the Olympic Mountains of Washington state (Woodward *et al*. 1994; Peterson *et al*. 2002; Shu *et al*. 2004), along the windward slopes of mountains on Vancouver Island (Laroque 2002), and in the state of Alaska (Wiles *et al*. 1996, 1998).

The purpose of the research presented in this paper was to characterize the radial growth response of conifer trees to high-elevation maritime climates at sites in the central Coast Mountain region. The study initially focused on climate-growth relationships at a single site in order to reconstruct a local history of climate fluctuations. Following preliminary analyses of the data collected, the research was expanded to characterize the regional character of climate-radial growth relationships at high-elevation sites located throughout the central Coast Mountains.

## 3.2 Physical Setting

The Coast Mountain ranges span almost the entire western perimeter of B.C., and in many settings form a 160 km wide barrier between the Pacific Ocean and the interior plateau. They are characterized by some of the highest peaks, largest icefields, and most variable climate conditions in the province (Tuller 2001). The bedrock geology of the area consists of plutonic (igneous intrusive) rocks, as well as metamorphized early Palaeozoic to middle Triassic sedimentary and volcanic rocks (Bancroft 1931). During the Pleistocene, the Cordilleran Ice Sheets repeatedly overwhelmed the region carving the exceptionally deep fjords and large u-shaped valleys that now characterize the Coast Mountain topography (Clague *et al*. 1980). Subalpine forests along the windward slopes of the Coast Mountains are located within the maritime mountain hemlock zone (Pojar *et al*. 1987).

In the winter months, incoming maritime polar air masses from the north Pacific transport cold, moist air to the windward side of the range where orographic effects release abundant precipitation (Tuller 2001). Much of this moisture is generated by mid-latitude cyclonic storms that generally peak in January (Koeppe 1931). At lower elevations this moisture and the moderating effects of warm maritime air masses result in mild and wet conditions (Tuller 2001). At higher elevations temperatures are significantly cooler resulting in the accumulation of deep seasonal snow packs (Kendrew and Kerr 1955). Summer months are generally characterized by distinctly drier conditions as a result of the northern migration of the Aleutian Low system (Kendrew and Kerr 1955).

Cool summer conditions arise due to the influence of dominant northwesterly winds originating in the Pacific Ocean (Tuller 2001).

Year-to-year and decadal-scale climate variability in this region is largely an expression of elevation and climate-forcing from ocean-atmospheric processes originating in the Pacific Ocean. The dominant climate modes are related to the El Niño-Southern Oscillation (ENSO) and the Pacific Decadal Oscillation (PDO) (Rasmussen and Wallace 1983; Ropelewski and Halpert 1987; Mantua *et al*. 1997; Minobe 1997; Zhang *et al*. 1997; Bond and Harrison 2000; Biondi *et al*. 2001). These systems interact and influence each other (Bonsal *et al*. 2001).

ENSO is an ocean-atmospheric process that is often described as a see-saw effect between the southeast Pacific subtropical high and the region of low pressure usually centered on the Indian Ocean (Rasmussen and Wallace 1983; Diaz and Markgraf 1992). Simultaneously the eastward displacement of warm water in the western equatorial Pacific results in increased sea surface temperatures (SSTs) off the coast of South America (Rasmussen and Wallace 1983).

ENSO events are the recurrent weather and climate anomalies that result from these changes in atmospheric circulation and SSTs (Diaz and Markgraf 1992). They occur at varying temporal and spatial scales but usually take place every 7 to 10 years (Rasmussen and Wallace 1983). In the Coast Mountains ENSO events are generally correlated to warmer- and drier-than-normal winter (January - March) conditions (Shabbar and Khandekar 1996; Bonsal *et al*. 2001).

The Pacific Decadal Oscillation is a SST anomaly the influence of which is dominant during winter months in the extratropics, particularly in the north Pacific and in western North America (Mantua and Hare 2002). While the specific cause(s) and mechanism(s) of the PDO are not yet fully understood, the phenomenon is characterized by two recognizable phases, a warm (positive) phase and a cool (negative) phase.

The PDO operates at a range of timescales, including interannual (<8 years), decadal (10-20 years), and interdecadal (30-70 years) modes (Gedalof and Smith 2001a). Warm phases of the PDO are characterized by anomalously cool SSTs in the central north Pacific and abnormally warm SSTs in coastal PNA. Warm phases are also characterized by low sea level pressure over the north Pacific, resulting in increased numbers of warm and wet weather systems moving counter clockwise to impact the North American coast. Cool phases of the PDO result in the opposite effects (Mantua and Hare 2002). During the warm PDO phase, anomalously warm and dry conditions persist in the Coast Mountains (Mantua and Hare 2002). Reversed temperature and precipitation regimes affect these areas in cool phases of the PDO (Mantua and Hare 2002).

## 3.3 Field Methods

Mountain hemlock (*Tsuga mertensiana*) and subalpine fir (*Abies lasiocarpa*) trees were sampled on a steep forested site at 1300 m asl located near the confluence of Franklin and Confederation glaciers, approximately 17 km west of the summit of Mount Waddington in the central Coast Mountains (Lat 51°16’11” N, Long 125° 26’17” W; Figures 3.1 and 4.2). The site is characterized by a mature even-aged forest stand dominated by mountain hemlock trees. Interpolated B.C. Climate data indicate an average annual air temperature of 1.1 °C and precipitation totals of approximately 1695 mm/yr at the site (Wang *et al*. 2005).


Figure 3.1: Map showing the location of the study area.

Increment core samples were collected from at least 20 trees of each species at standard breast height with a 5 mm borer. Two cores were taken from each tree at 180º from one another. The healthiest, most undisturbed, and oldest-looking trees were preferentially sampled in order to minimize the influence of local environmental variables on tree ring width and to temporally extend the growth record as far as possible (Fritts 1976). Care was taken to avoid trees with broken masts, as well as growth irregularities near roots and branches (Stokes and Smiley 1968; Schweingruber *et al*. 1990). The samples were stored in plastic straws, labelled, and transported back to the University of Victoria Tree-Ring Laboratory (UVTRL) for analysis.

## 3.4 Data Analysis

After being allowed to air dry for several weeks, the cores were glued into slotted mounting boards. The exposed face of each core was then sanded using a belt sander equipped with progressively finer grits of sandpaper (maximum 600 grit) in order to ensure that all tree rings were visible (Stokes and Smiley 1968). Each core was then measured to an accuracy of 0.001 mm using a Velmex measuring stage and a Wild M3B stereomicroscope equipped with a Sony 3CCD video camera. Ring-width measurements were taken along the centre of each core and captured by J2X software (v.3.2.1, 1994). Measurements were converted to standard decadal (Tucson) format using the program FMT from the Dendrochronology Program Library (DPL) (Holmes 1994).

### *3.4.1 Cross-dating*

The cores were visually cross-dated and quality-checked using the International Tree Ring Data Bank (ITRDB) software program COFECHA (Holmes 1983). The program was used to evaluate the relationship between individual tree ring width series by calculating the Pearson's correlation coefficients between successive 50-year ring-width segments with a 25-year lag (Grissino-Mayer 2001). Statistical significance was determined at the 99% confidence level. Low-frequency variance was removed by fitting the ring-width series with a smoothing spline with a standard 50% frequency cutoff at a wavelength of 32 years, and persistence was removed through autoregressive modeling (Grissino-Mayer 2001).

Visual and COFECHA-assisted cross-dating was carried out between the chronologies constructed from trees in the Confederation and Franklin glaciers confluence (local) area and a regional network of chronologies previously developed by UVTRL researchers on Vancouver Island (Smith and Laroque 1998), the eastern slopes of the Waddington Range (Larocque and Smith 2005; Hart 2009), in the Lillooet Icefield area of the southern Coast Mountains (Allen and Smith 2007; Koehler 2009), and from the Todd Icefield area of the northern Coast Mountains (Jackson *et al*. 2008). The field, sample preparation, and cross-dating methods employed in the development of these chronologies were consistent with those used in this study. Similar methods were also used by Parish (2006) in developing a Vancouver Island mountain hemlock chronology that was also included in subsequent analyses.

### *3.4.2 Standardization*

Tree ring series generally exhibit a negative age-related trend as rapid juvenile radial growth is replaced by smaller increments of annual growth as a tree matures. This results in larger growth rings around the pith and narrower rings near the bark. Ring width increments are also influenced by local endogenous and stand-wide exogenous disturbances unrelated to climate. In attempting to isolate climate-related growth trends for the purposes of dendroclimatological research, the influence of these growth trends must be reduced, or standardized. The ITRDB program ARSTAN (Cook and Krusic 2005) was used to fit a negative exponential curve, a linear regression line of negative slope, or a horizontal line through the mean of each tree ring series. Ring width measurements from each series were then divided by the expected value of the curve, resulting in relative tree ring indices with a defined mean of 1.0 and stabilized variance (Cook 1985). Because single detrending primarily succeeds in removing the age-related growth trend, a double detrending approach was employed whereby a smoothing spline was fit to the data to further reduce the influence of endogenous and exogenous disturbance (Cook 1985; Cook and Krusic 2005). The spline was applied with a 67% frequency response cutoff, thereby preserving 50% of the variance in the ring-width series at a frequency equal to two-thirds of the length of the series (Cook 1985). Auto Regressive Moving Average (ARMA) modeling was also applied to the ring width series in order to remove autocorrelation. The order of the ARMA model was defined using the Akaike Information Criterion (Cook and Krusic 2005). Finally, a master ring-width chronology was created for each species by averaging individual ring-width series using a biweight robust mean (Cook *et al*. 1990). ARSTAN was also used to calculate Express Population Signal (EPS) values at 20-year moving windows with 10-year overlap so as to control chronology quality with decreasing sample size over time. Tree ring width chronologies were truncated when running EPS values fell below the standard value of 0.85 (Wigley *et al*. 1984; Briffa and Jones 1990).

Regional subalpine fir and mountain hemlock chronologies that successfully cross-dated to the chronologies developed in this study were added to the local species chronologies. The resulting regional subalpine fir and mountain hemlock chronologies were converted into master chronologies via the standardizing methods described above.

### *3.4.3 Correlation and Response Function Analysis*

Bootstrapped Pearson's correlation and response function analyses were conducted in order to determine the relationship(s) between the master chronologies and seasonal and monthly climatic variables using the software program DENDROCLIM (Biondi and Waikul 2004). Partial correlation analysis was employed to control for correlations between two predictor variables. Climate data were acquired from Environment Canada's (1996) Adjusted Historical Homogeneous Canadian Climate Dataset (AHCCD) (Vincent and Gullet 1999) and from ClimateBC (Daly *et al*. 2002; Mitchell and Jones 2005). The latter downscales PRISM climate data and incorporates historical climate data to interpolate monthly and annual temperature and precipitation variables based on a site's geographical coordinates and altitude (Wang *et al*. 2005). Climate variables that explained the highest statistically significant percentage of the variance in tree ring width for both local and regional subalpine fir and mountain hemlock chronologies were reconstructed using regression analysis. The program DENDROCLIM was also used to conduct bootstrapped moving interval response function analyses in order to evaluate the temporal stability of the relationship between tree ring width indices and monthly climate variables.

Regression models for reconstruction were calibrated and verified using split period cross-validation technique, according to which a transfer function is developed using only half of the available climate data, and then verified through comparison with the remaining half of the data (Fritts 1976; Blasing *et al*. 1981; Gordon 1982). The ability of the model to reconstruct climatic variability was assessed using Pearson's correlation coefficients. Wavelet analyses were employed in order to highlight temporal trends and modes of variability in the data.

## 3.5 Observations

### *3.5.1 Chronology characteristics*

Thirty-one cores from 16 of the 30 mountain hemlock trees sampled at the study site were included in a 409-year long (AD 1599-2009) local mountain hemlock chronology with a series intercorrelation value of $r = 0.493$ (Table 3.1). The regional master mountain hemlock chronology includes the local site chronology collected as part of this study and chronologies from two additional locations, Mount Caine on northern Vancouver Island (Parish 1996; Laroque and Smith 2003) (Table 3.1) and in the Bella Coola area (Gedalof and Smith 2001b) (Table 3.1). In total the 734 year-long master mountain hemlock chronology contains 119 series from 71 trees and spans the interval from AD 1274 to 2007 ($r$ =0.493) (Table 3.1). Several series that did not cross-date once combined into the master chronology were omitted.

The local subalpine fir chronology was created using 28 cores from 15 of the 26 subalpine fir trees sampled at the site, and extends over a 315-year period (AD 1693-2008, Table 3.2). The chronology has a series intercorrelation value of $r = 0.573$ (Table 3.2). The regional master subalpine fir chronology was compiled from cross-dated chronologies developed from 7 sites in the Coast Mountains. Two hundred and eighty four series from 160 trees were included in the chronology which spans the period from 1572 to 2007 and has an intercorrelation value of $r = 0.509$ (Table 3.2). Four of the site chronologies included in the master chronology were collected by Larocque and Smith

Table 3.1: Chronology statistics for local and regional mountain hemlock (MH) chronologies, and for included regional chronologies.

| Chrono.[a] | Source | *r* value | # Yrs | Interval (yrs. AD) | # Trees (cores) | Mean sens.[b] | Std. dev.[c] | Auto-corr.[d] | EPS[e] |
|---|---|---|---|---|---|---|---|---|---|
| MH Local | This study | 0.493 | 421 | 1586-2006 | 16 (30) | 0.267 | 0.377 | -0.023 | 1750 |
| Caine (1) | Parish (1996) | 0.572 | 606 | 1394-1999 | 36 (64) | 0.248 | 0.362 | -0.026 | 1480 |
| Caine (2) | Laroque (1996) | 0.468 | 831 | 1166-1996 | 15 (23) | 0.273 | 0.360 | -0.014 | 1650 |
| Bella Coola | Gedalof and Smith (2001b) | 0.485 | 285 | 1712-1996 | 20 (33) | 0.217 | 0.359 | -0.028 | 1740 |
| MH Regional | This study | 0.493 | 734 | 1274-2007 | 71 (119) | 0.250 | 0.366 | -0.024 | 1470 |

[a] Chronology name
[b] Mean sensitivity, with ~0.650 being very sensitive and ~0.150 being complacent
[c] Standard deviation of the ring-width measurements (in millimeters) for the given series
[d] Autocorrelation, or the correlation between the ring width in year n and the ring width in year n-1, with 0.300 being very low and 0.900.
[e] The Express Population Signal (EPS) year signifies the point at which EPS > 0.85 due to decreasing sample size over time.

Table 3.2: Chronology statistics for local and regional subalpine (SAF) chronologies, and for included regional chronologies.

| Chrono.[a] | Source | *r* value | # Yrs | Interval (yrs. AD) | # Trees (cores) | Mean sens.[b] | Std. dev.[c] | Auto-Corr.[d] | EPS[e] |
|---|---|---|---|---|---|---|---|---|---|
| SAF Local | This study | 0.573 | 317 | 1691-2007 | 15 (28) | 0.206 | 0.366 | -0.024 | 1470 |
| Hope | Larocque and Smith (2005) | 0.612 | 302 | 1700-2001 | 20 (37) | 0.241 | 0.367 | -0.017 | 1760 |
| Escape | Larocque and Smith (2005) | 0.583 | 223 | 1778-2000 | 19 (36) | 0.188 | 0.379 | -0.014 | 1830 |
| Liberty | Larocque and Smith (2005) | 0.555 | 296 | 1705-2000 | 20 (38) | 0.189 | 0.364 | -0.013 | 1730 |
| Cathedral | Larocque and Smith (2005) | 0.573 | 359 | 1642-2000 | 21 (38) | 0.237 | 0.370 | -0.025 | 1710 |
| Nostetuko | Hart (2009) | 0.447 | 279 | 1729-2007 | 19 (29) | 0.214 | 0.378 | -0.009 | 1930 |
| Bridge | Allen and Smith (2007) | 0.617 | 322 | 1681-2002 | 20 (40) | 0.203 | 0.381 | -0.019 | 1800 |
| Manatee (1) | Koehler (2009) | 0.614 | 435 | 435 | 6 (12) | 0.233 | 0.345 | -0.008 | 1820 |
| Manatee (2) | Koehler (2009) | 0.541 | 325 | 1681-2005 | 12 (18) | 0.206 | 0.388 | -0.020 | 1890 |
| SAF Regional | This study | 0.509 | 436 | 1572-2007 | 160 (284) | 0.214 | 0.372 | -0.107 | 1700 |

[a] Chronology name
[b] Mean sensitivity, with ~0.650 being very sensitive and ~0.150 being complacent
[c] Standard deviation of the ring-width measurements (in millimeters) for the given series
[d] Autocorrelation, or the correlation between the ring width in year n and the ring width in year n-1, with 0.300 being very low and 0.900.
[e] The Express Population Signal (EPS) year signifies the point at which EPS > 0.85 due to decreasing sample size over time.

(2005) along the eastern perimeter of the Mount Waddington area (Table 3.2). Also included were individual site chronologies collected by Hart (2009) in the nearby Homathko Icefield area, and Allen and Smith (2007) and Koehler (2009) in the vicinity of the Lillooet Icefield (Table 3.2).

### *3.5.2 Evaluating climate-growth relationships*

According to interpolated climate data (Wang *et al.* 2005) the majority of the mountain hemlock sites included in the regional chronology are characterized by annual climates that are both warmer and wetter than those characterizing the subalpine fir regional chronology sites (Table 3.3). Cooler mean annual temperatures at subalpine fir sites result from significantly colder conditions during winter months. In contrast, summer temperatures at the mountain hemlock and subalpine fir sites are more similar, ranging between 13.0ºC and 19.4ºC (Table 3.3). The significantly higher annual precipitation totals that characterize the mountain hemlock sites are largely a function of the heavy precipitation regimes that persist during the winter months (Table 3.3).

The statistical relationship between the local and regional chronologies and various climate variables were calculated for all AHCCD climate stations (Vincent and Gullet 1999) in the central and southern Coast Mountain area, as well as for a suite of site-specific monthly temperature and precipitation variables generated by ClimateBC software (Wang *et al*. 2005) (Table 3.4). Climate data from the station located in Comox, Vancouver Island (Station # 1021830; Lat 49°42'N; Long 124°53W) was the most strongly correlated to ring width patterns exhibited by trees at the study site and was used for the remaining analysis (Table 3.4).

Table 3.3: Locations and climate characteristics of sites included in the subalpine fir and mountain hemlock regional chronologies.

| | Location (Lat. Long). | Elev. (m asl) | Annual mean temp. (ºC) | Min. mean monthly temp. (ºC) | Max mean monthly temp. (ºC) | Mean annual precip. (mm) | Min mean monthly precip. (mm) | Max mean monthly precip. (mm) |
|---|---|---|---|---|---|---|---|---|
| Study Site | 51º16'N 125º26'W | 1300 | 1.1 | -12.1 | 17 | 1695 | 48 | 248 |
| Caine (1) | 50º13'N 126º19'W | 270 | 9.0 | 3.1 | 14.2 | 3936 | 54.3 | 225.9 |
| Caine (2) | 50º13'N 126º19'W | 270 | 9.0 | 3.1 | 14.2 | 3936 | 54.3 | 225.9 |
| Bella Coola | 52º13'N, 126 º19'W | 1035 | 2.1 | -1.9 | 16.6 | 2568 | 54.3 | 225.9 |
| Hope | 51º31'N, 124º53'W | 1830 | -0.8 | -11.6 | 13.0 | 1461.0 | 52.0 | 213.0 |
| Escape | 51º37'N, 125º07'W | 1760 | 0.1 | -11.3 | 14.3 | 1294 | 50.0 | 181.0 |
| Liberty | 51º35'N, 124º50'W | 1525 | 0.3 | -11.3 | 14.4 | 1338 | 50 | 192 |
| Cathedral | 51º14'N, 124º52'W | 1600 | 2.0 | -12.1 | 19.4 | 1523 | 37 | 236 |
| Nostetuko | 51º17'N 124º30'W | 1400 | 6.7 | -3.9 | 13.8 | 898 | 11.03 | 56.3 |
| Bridge | 50º49'N 123º35'W | 1544 | 1.0 | -11.8 | 16.5 | 2044 | 70 | 291 |
| Manatee (1) | 50º37'N 123º41'W | 1600 | 0.6 | -11.8 | 16.1 | 2639 | 77 | 379 |
| Manatee (2) | 50º37'N 123º41'W | 1600 | 0.6 | -11.8 | 16.1 | 2639 | 77 | 379 |

Table 3.4: Highest bootstrapped response function values calculated between local/regional chronologies and annual climate variables recorded at AHCCD climate stations and interpolated by Climate BC software, using the program DENDROCLIM.

| Chronology | Climate Station | | | | | |
|---|---|---|---|---|---|---|
| | Tatloyoko Lake (Stn # 1088010) | Bella Coola (Stn # 1060840) | Port Hardy (Stn # 1026270) | Comox (Stn # 1021830) | Nanaimo (Stn # 1024483) | BC Climate data |
| Mean annual temperature | | | | | | |
| MH Local | 0.240 | -0.225 | -0.200 | -0.380 | -0.251 | -0.222 |
| MH Regional | 0.299 | -0.286 | -0.279 | -0.317 | -0.262 | - |
| SAF Local | 0.318 | -0.227 | -0.270 | -0.330 | -0.238 | -0.302 |
| SAF Regional | 0.370 | 0.289 | -0.275 | -0.408 | -0.264 | - |
| Mean annual precipitation | | | | | | |
| MH Local | - | 0 | 0.292 | -0.301 | -0.259 | -0.300 |
| MH Regional | - | -0.210 | 0.309 | -0.308 | -0.311 | - |
| SAF Local | - | 0 | 0.305 | -0.323 | -0.268 | -0.214 |
| SAF Regional | - | 0.190 | 0.363 | -0.251 | -0.290 | - |

Bootstrapped Pearson’s correlation and response function relationships between the master chronologies and seasonal and monthly climatic variables indicate that current July temperature was the only variable correlated (negatively) to mountain hemlock growth at the study site (Table 3.5). This response is opposite to all of the other sites included in the regional mountain hemlock chronology, which generated positive responses to the same variable (Table 3.5). Both local and regional subalpine fir chronologies developed in this study demonstrate a consistent, dominant negative response to previous July mean (Figure 3.2) and maximum (Figure 3.3) temperatures (Table 3.6).

Moving interval analyses highlight the temporal instability of the relationships between radial tree growth and climate. All four subalpine fir response function analyses are characterized by increasingly weak negative correlations between previous July temperature and ring width from past to present (Figures 3.4 and 3.5). These findings indicate that the strength of relationships between climate and radial growth and the influence of limiting factors may not be static over time.

Table 3.5: Statistically significant (difference between the 97.5 and 2.5 percentile) bootstrapped moving response function values of individual mountain hemlock chronologies included in regional chronology to Comox monthly (previous June to current August) mean and maximum temperature values.

| Site | Month | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | Previous Year | | | | | Current Year | | | | |
| | J | J | A | S | O | N-J | F | M-M | J | J | A |
| *(A) Comox monthly mean temperature* | | | | | | | | | | | |
| Local MH | | | | | | | | | | -0.361 | |
| Caine (Parish) | | -0.300 | | | | | | | | | |
| Caine (Laroque) | | -0.204 | | | -0.300 | | | | 0.230 | 0.400 | |
| Bella Coola | | | 0.263 | | | | | | 0.320 | | |
| *(B) Comox monthly maximum temperature* | | | | | | | | | | | |
| Local MH | | | | | | | 0.217 | | | -0.400 | |
| Caine (Parish) | | 0.280 | | | | | | | | | |
| Caine (Laroque) | | -0.188 | | | -0.280 | | | | 0.280 | 0.340 | |
| Bella Coola | | | | | | | | | | 0.340 | |

Table 3.6: Statistically significant (difference between the 97.5 and 2.5 percentile) bootstrapped moving response function values of individual subalpine fir chronologies included in regional chronology to Comox monthly (previous June to current August) mean and maximum temperature values.

| Site | Month | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | Previous Year | | | | Current Year | | | | | |
| | J | J | A | S-F | M | A | M | J | J | A |
| *(A) Comox monthly mean temperature* | | | | | | | | | | |
| Local SAF | | -0.330 | | | | | | | 0.214 | |
| Hope | | -0.317 | | | | | | | 0.280 | |
| Escape | | -0.370 | | | | | | | 0.248 | |
| Liberty | | -0.342 | | | | | | 0.185 | .266 | |
| Cathedral | | -0.351 | | | | | | 0.194 | 0.378 | |
| Nostetuko | | -0.249 | | | | | | | 0.201 | |
| Bridge | | -0.364 | | | | | | | 0.269 | |
| Manatee (1) | | -0.384 | | | | | | | 0.280 | |
| Manatee (2) | | -0.416 | | | | | -0.200 | | 0.282 | |
| *B) Comox monthly maximum temperature* | | | | | | | | | | |
| Local SAF | | -0.300 | | | | | | 0.231 | 0.211 | |
| Hope | | -0.294 | | | | | | | 0.261 | |
| Escape | | -0.400 | -0.220 | | | | | | 0.240 | |
| Liberty | -0.192 | -0.324 | | | | | | | 0.220 | |
| Cathedral | | -0.332 | | | 0.197 | | | 0.240 | 0.350 | |
| Nostetuko | -0.203 | -0.217 | | | | | | | | |
| Bridge | | -0.400 | | | 0.231 | | | | 0.240 | |
| Manatee (1) | | -0.372 | -0.216 | | | | -0.222 | | 0.250 | |
| Manatee (2) | | -0.390 | -0.230 | | | | -0.244 | | 0.270 | |

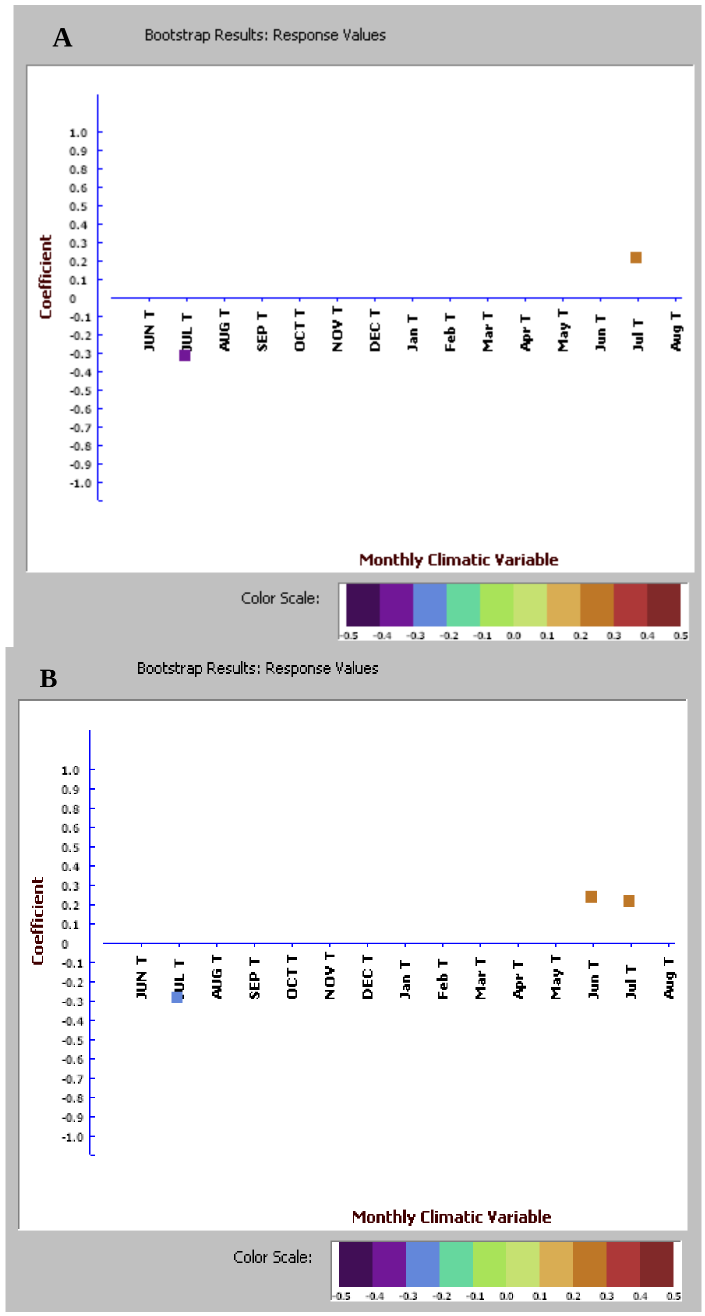


Figure 3.2: Significant bootstrapped response function relationships between the local subalpine fir chronology and monthly (previous June to current August) mean temperature (panel A) and maximum temperature (panel B) (Comox). Months of the previous year are identified with capital letters. Significant positive relationships are highlighted with boxes in shades of red while negative correlations are highlighted with boxes in shades of blue.

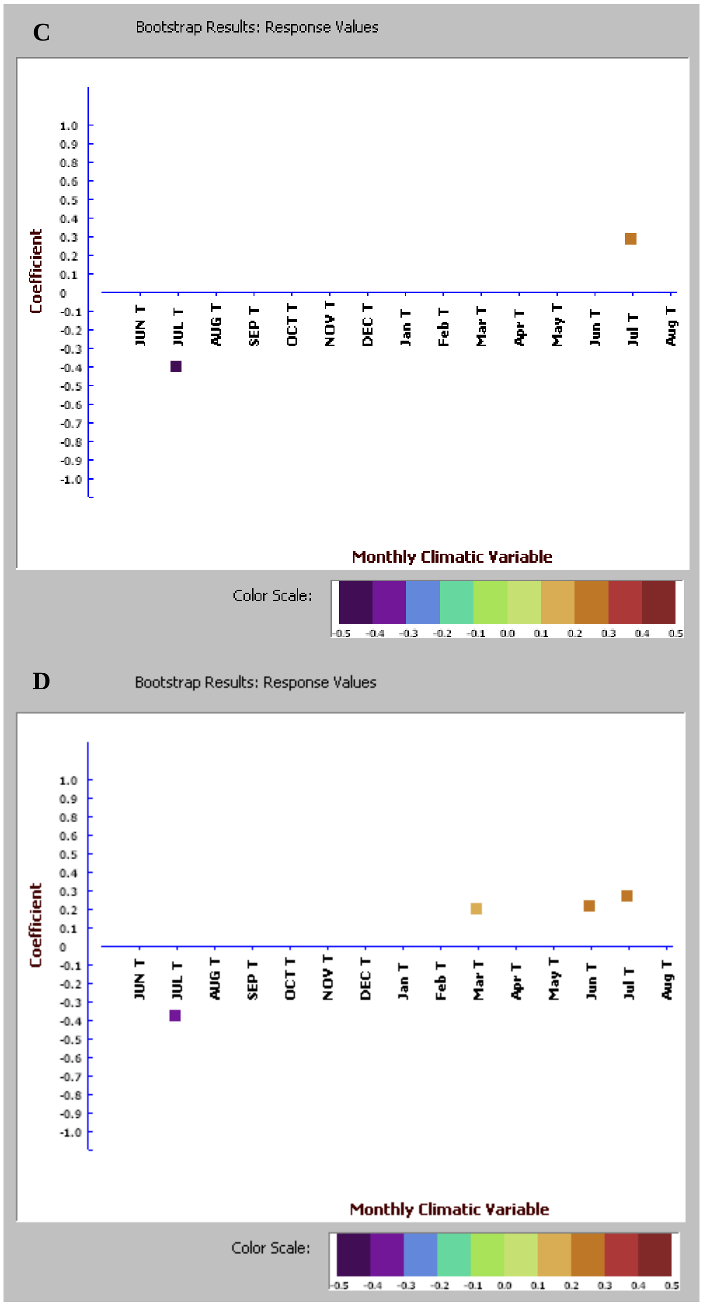


Figure 3.3: Significant bootstrapped response function relationships between regional subalpine fir chronology and monthly (previous June to current August) mean temperature (panel C) and maximum temperature (panel D) (Comox). Months of the previous year are identified with capital letters. Significant positive relationships are highlighted with boxes in shades of red while negative correlations are highlighted with boxes in shades blue.

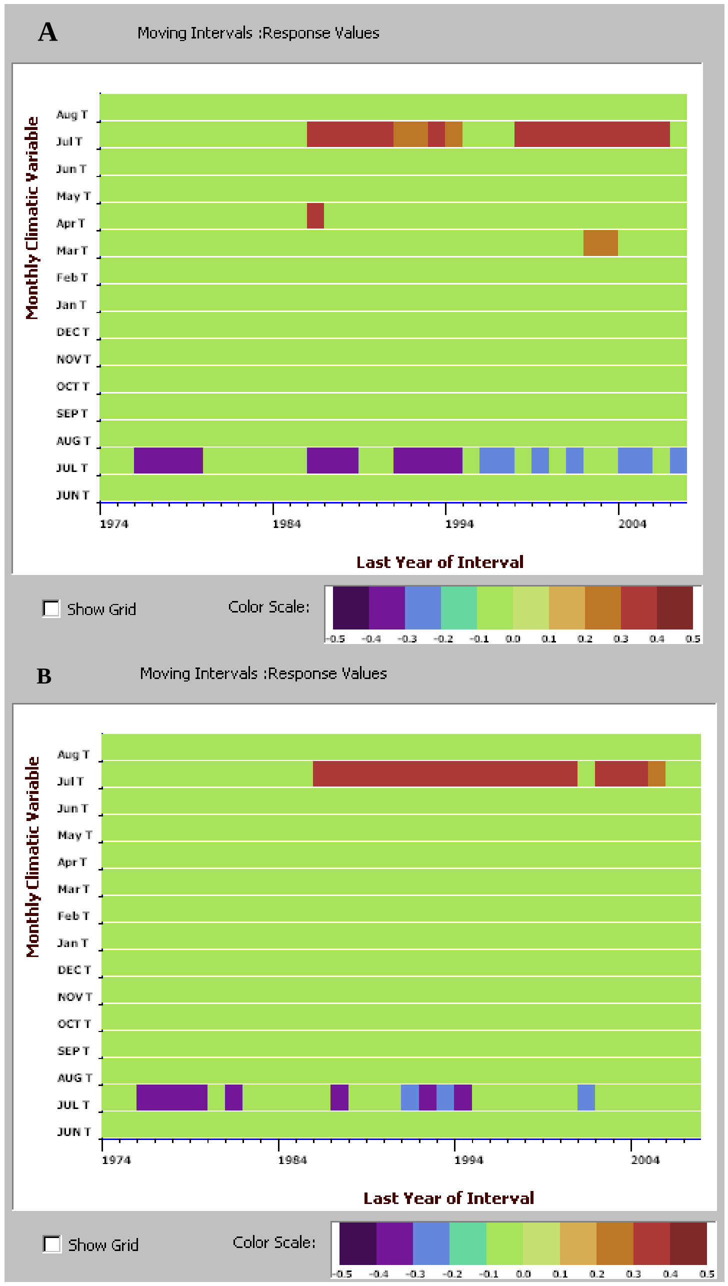


Figure 3.4: This figure illustrates bootstrapped moving interval response function analyses calculated using the program DENDROCLIM. Thirty-year intervals were employed to test the strength of relationships between tree ring indices and monthly June to August air temperature values over time (Comox). The panels depict the relationships between the local subalpine fir chronology and mean (A) and maximum (B) temperature values. Statistically significant positive relationships are highlighted in shades of red and negative relationships in shades of blue. Months of the previous year are identified with capital letters.

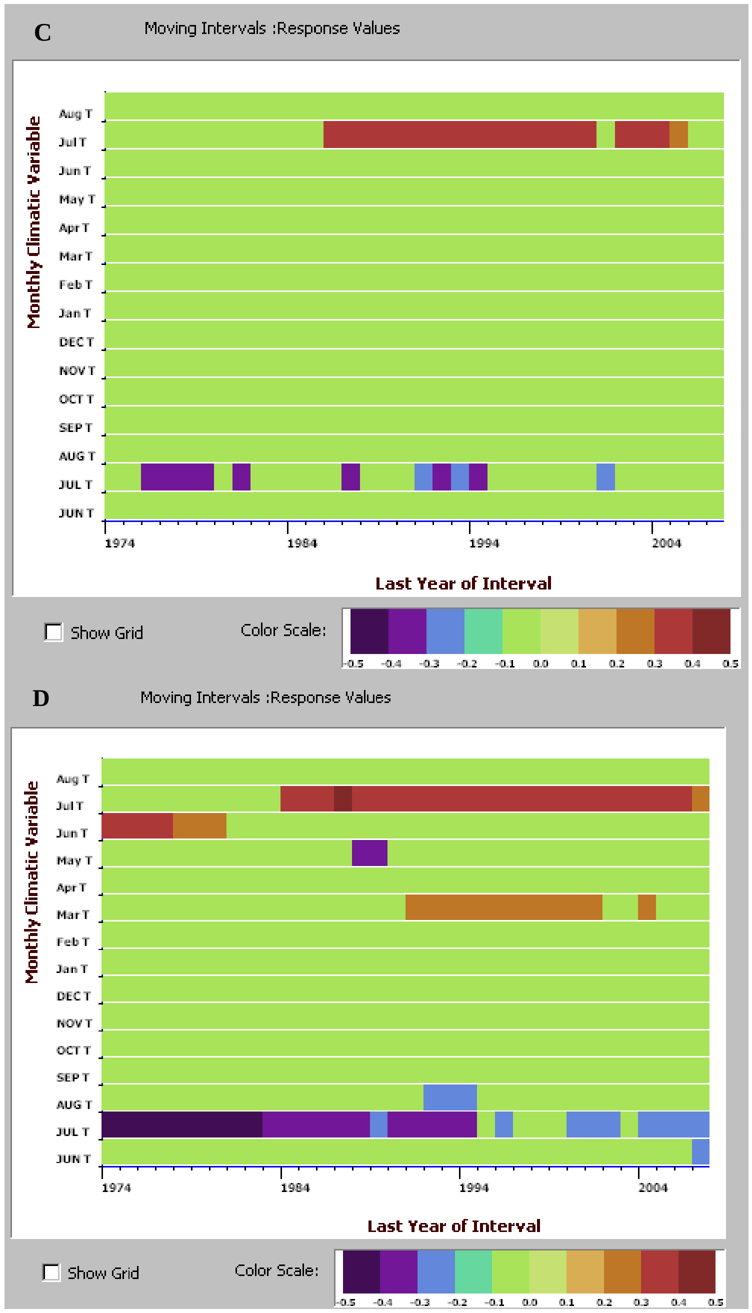


Figure 3.5: This figure illustrates bootstrapped moving interval response function analyses calculated using the program DENDROCLIM. Thirty-year intervals were employed to test the strength of relationships between tree ring indices and monthly June to August air temperature values over time (Comox). The panels depict the relationships between the regional subalpine fir chronology and mean (C) and maximum (D) temperature values. Statistically significant positive relationships are highlighted in shades of red and negative relationships in shades of blue. Months of the previous year are identified with capital letters.

## 3.6 Results

### *3.6.1 Mountain hemlock climate response*

The unstable response of mountain hemlock trees to current July temperature in the study region may result from a complex species-specific response to climate (Gedalof and Smith 2001b; Laroque *et al*. 2001; Penrose 2007). It may be, however, that the local chronology was sampled at too low an elevation, thereby weakening the influence of a single limiting growth factor. Sampling was conducted close to the lower altitudinal limit of the mountain hemlock trees at the study site. Furthermore while the local chronology appears to be sensitive to growing-season summer-temperature induced stress, likely related to factors such as slope gradient, runoff, and slope aspect, the regional mountain hemlock chronologies respond positively to growing season temperature. Because the mountain hemlock chronologies behave individualistically, they were omitted from the remaining analysis.

### *3.6.2 Subalpine fir climate response*

A positive radial growth response of subalpine fir trees to current summer temperatures is the most pervasive relationship documented in related studies across PNA (Heikkinen 1985; Ettl and Peterson 1995; Parish *et al*. 1999; Spelchtna *et al*. 2000; Peterson *et al*. 2002; Larocque and Smith 2005). In contrast, the response function analyses completed for both the local and regional subalpine fir chronologies demonstrate a consistent, negative response to previous July mean and maximum temperatures. While a comparable lagged response has been reported elsewhere, the influence of previous summer temperatures has generally been of secondary importance to the role played by temperature during the growing season (i.e. Wig and Smith 1994; Peterson *et al*. 2002; Larocque and Smith 2005).

The negative response to previous summer temperature likely results from the combined effects of deep spring snow packs and the influence of warm summer temperatures. Peterson *et al*. (2002) found that previous year winter precipitation, snowpack depth, and soil moisture limited radial growth in subalpine fir trees at high elevation sites in Washington state. Peterson and Peterson (1994) and Smith and Laroque (1998) report that warm summer temperatures only favour subalpine fir growth in years with below average snowpack. Snowpack-induced growth setbacks are thought to predispose trees to soil moisture sensitivity later in the summer (Fonda and Bliss 1969; Brooke *et al*. 1970; Evans and Fonda 1990; Peterson and Peterson 1994), resulting in reduced respiration and net photosynthesis and subsequent stomatal closure, reduced transpiration, and increased leaf temperature (Fritts 1976). Warm summer temperatures can also reduce frost hardiness in the following winter leading to death and damage of new shoots (Tranquillini 1979). Ultimately, the combination of deep, persistent spring snowpacks and warmer–than-average summer temperatures reduces the length of the growth season, thereby limiting the accumulation and storage of resources that can be allocated to radial growth in the following season.

Subalpine fir trees in PNA may be susceptible to a temperature threshold effect wherein warmer temperatures enhance radial growth up to a certain threshold, beyond which continued warming limits radial growth. In environments where the early-season activation of photosynthetic and metabolic processes is delayed by the presence of snow pack, this point may be reached earlier or at lower temperatures than in environments where snow packs are shallower; optimum temperatures for photosynthesis in alpine environments can be as low as 10 – 15°C (Fritts 1976; Tranquillini 1979).

The coherent response of the subalpine fir tree ring widths to previous July temperature is likely a function of the distinct climate regimes characterizing the sites where the chronologies included in this analysis were sampled. With the exception of the study site, subalpine fir trees were sampled in environments that typically experience warm and very dry summers (Table 3.3). The local study site, although located in a windward environment with high summer precipitation and moderate summer temperatures, is located on an extremely steep south-facing slope, the impact of which was noted earlier in summer-temperature-stressed mountain hemlock trees. The correlation of trees in this environment to trees growing in warm and dry-summer climate regimes emphasizes the impact of slope gradient and aspect on the susceptibility of conifers to summer-temperature induced physiological stress, even in environments with high summer precipitation totals (Table 3.3).

The Comox climate station, located in the rain shadow of the Vancouver Island insular mountains, appears to most closely describe the July moisture regimes thought to characterize the various sampling sites. As Table 3.7 illustrates, the climate records from Comox indicate that this area experiences both the warmest and driest July conditions of all of the station records examined.

#### *3.6.3 July temperature reconstructions*

Based on the results of the response function analyses, linear regression equations were calculated to represent the relationship between the subalpine fir local and regional chronologies, and the historical mean and maximum temperature data recorded at

Table 3.7 July mean precipitation and maximum temperatures as recorded at the various subalpine fir and mountain hemlock study sites and at the Comox climate station.

| Location | Mean July precipitation (mm) | Max July temperature (ºC) |
|---|---|---|
| Study Site | 66 | 17.5 |
| Caine | 121 | 18.8 |
| Bella Coola | 81 | 17.3 |
| Waddington Area | 70 | 15.6 |
| Bridge | 70 | 16.2 |
| Manatee | 77 | 15.8 |
| Comox | 33 | 21.1 |

the Comox climate station. Regression statistics for the local and regional subalpine fir chronology reconstructions are presented in Tables 3.8 and 3.9. The subalpine fir climate reconstructions are plotted along with graphs of the common interval between each reconstruction and the instrumental climate data from which it was generated. Ten-year running means highlight the cyclical trends inherent in the reconstructions (Figures 3.6, 3.8, 3.10, and 3.12).

The reconstructed climate variables were found to explain between 13% and 36 % of the variation in climate over time. These proxy records are comparable to similar climate reconstructions from the region (Table 3.10). Although the significance values determined in this analysis are low, the common interval plots depict a strong directional relationship between the datasets (Figures 3.7, 3.9, 3.11, and 3.13) and indicate the reconstruction closely follows the high amplitude trends.

Table 3.8: Climate model statistics for reconstructions of previous July mean and maximum temperature using the local subalpine fir chronology.

| Climate Variable | Statistical measure | | |
|---|---|---|---|
| *(A) Regression Analysis* | $r^2$ | *Period* | *Equation* |
| Previous July mean temp. | 0.162 | 1470-2007 | 20.840+(-3.291*x*) |
| Previous July max. temp. | 0.185 | 1700-2007 | 27.558+(-5.009*x*) |
| (B) *Calibration* | $r^2$ | *Period* | |
| Previous July mean temp. | 0.227 | 1944-1975 | |
| Previous July max. temp. | 0.255 | 1944-1975 | |
| (C) *Validation* | $r^2$ | *Period* | |
| Previous July mean temp. | 0.156 | 1975-2007 | |
| Previous July max. temp. | 0.131 | 1975-2007 | |

Table 3.9: Climate model statistics for reconstructions of previous July mean and maximum temperature using the regional subalpine fir chronology.

| Climate Variable | Statistical measure | | |
|---|---|---|---|
| *(A) Regression Analysis* | $r^2$ | *Period* | *Equation* |
| Previous July mean temp. | 0.244 | 1700-2007 | 21.687+(-4122*x*) |
| Previous July max. temp. | 0.301 | 1700-2007 | 29.084+(-6.514*x*) |
| (B) *Calibration* | $r^2$ | *Period* | |
| Previous July mean temp. | 0.273 | 1944-1975 | |
| Previous July max. temp. | 0.362 | 1944-1975 | |
| (C) *Validation* | $r^2$ | *Period* | |
| Previous July mean temp. | 0.265 | 1975-2007 | |
| Previous July max. temp. | 0.257 | 1975-2007 | |

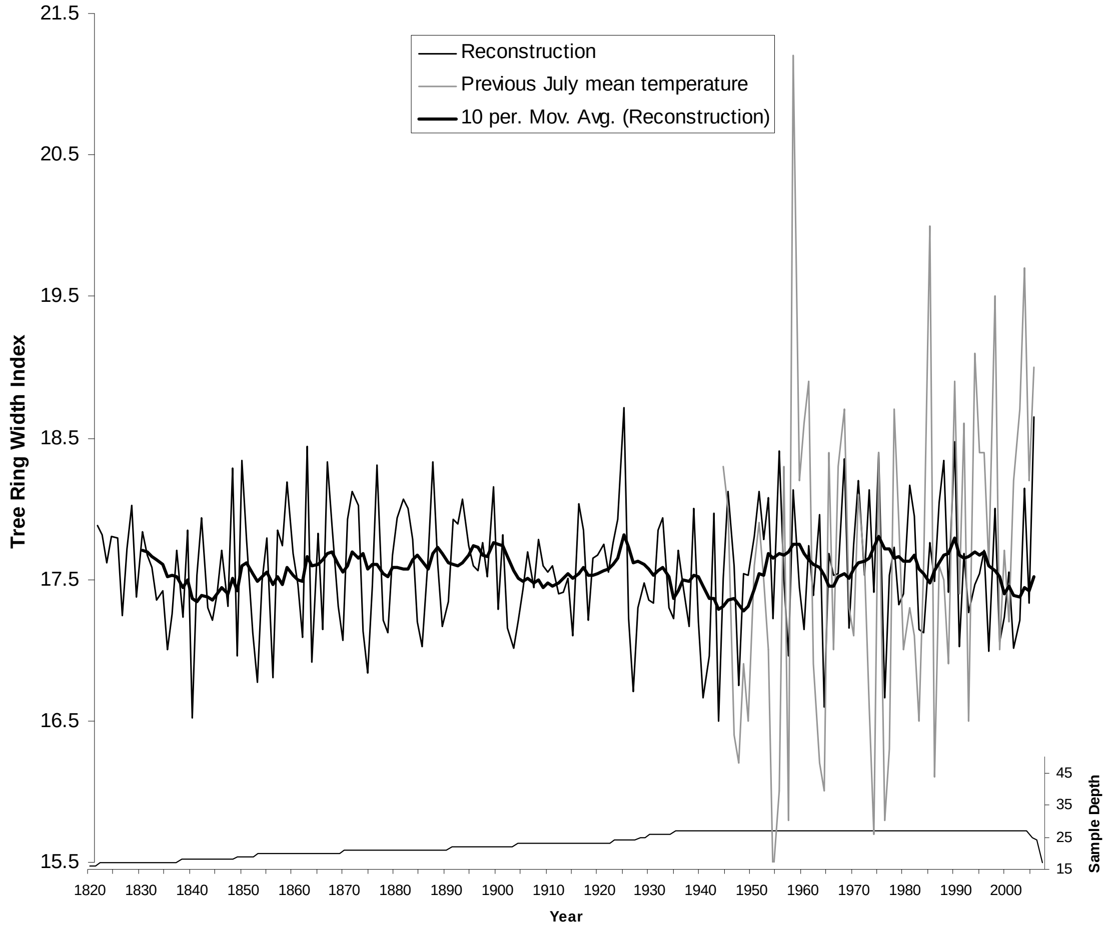


Figure 3.6: A graph of instrumental previous July mean temperature data from the Comox climate station (grey) and a reconstruction of previous July mean temperature from the local subalpine fir chronology developed at the study site (black). A ten-year running mean has been plotted through the data, and sample depth over time is plotted at the bottom of the graph.

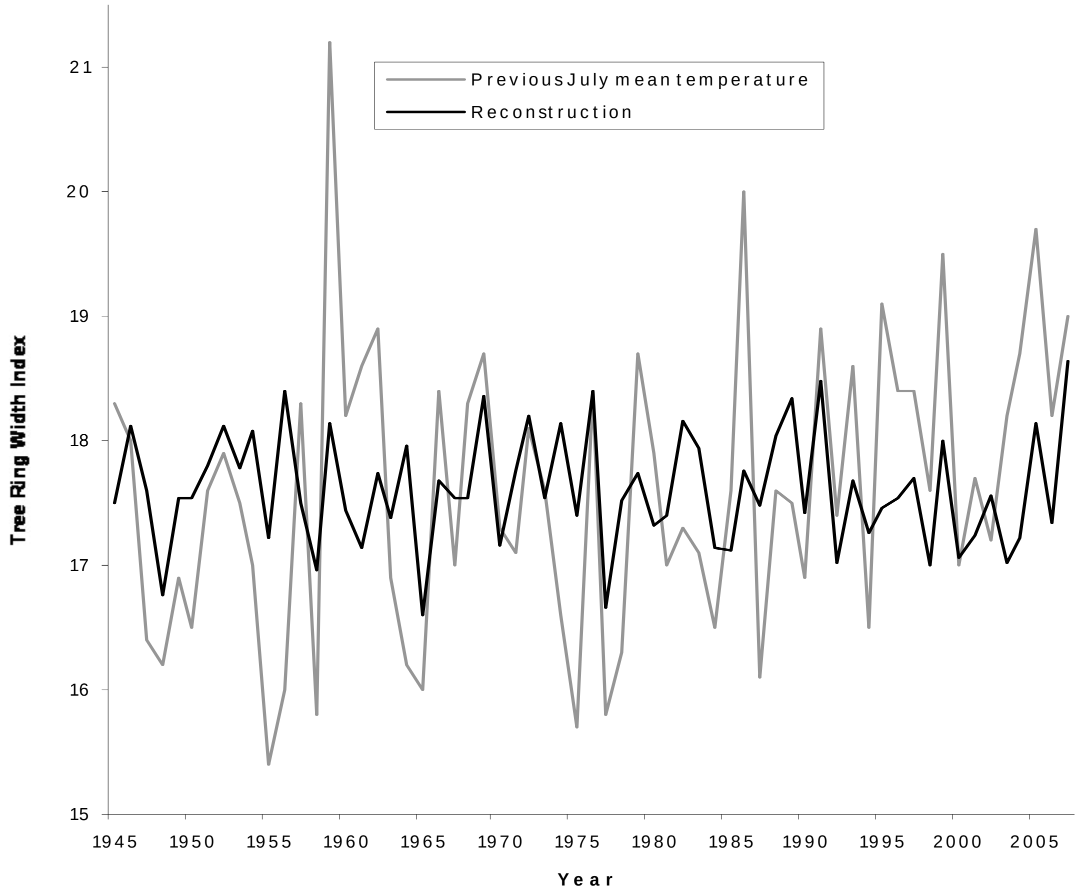


Figure 3.7: A graph of the common interval of data between instrumental previous July mean temperature data from the Comox climate station (grey) and a reconstruction of previous July mean temperature from the local subalpine fir chronology developed at the study site (black).

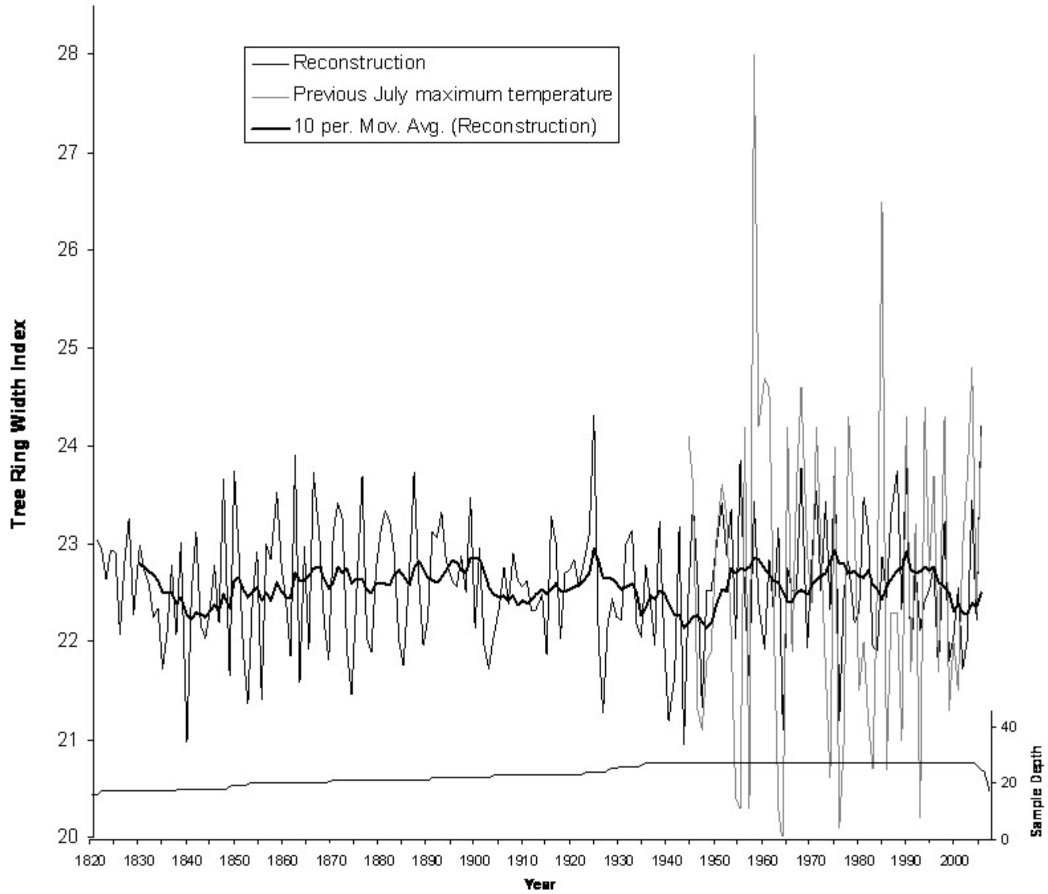


Figure 3.8: A graph of instrumental previous July maximum temperature data from the Comox climate station (grey) and a reconstruction of previous July maximum temperature from the local subalpine fir chronology developed at the study site (black). A ten-year running mean has been plotted through the data, and sample depth over time is plotted at the bottom of the graph.

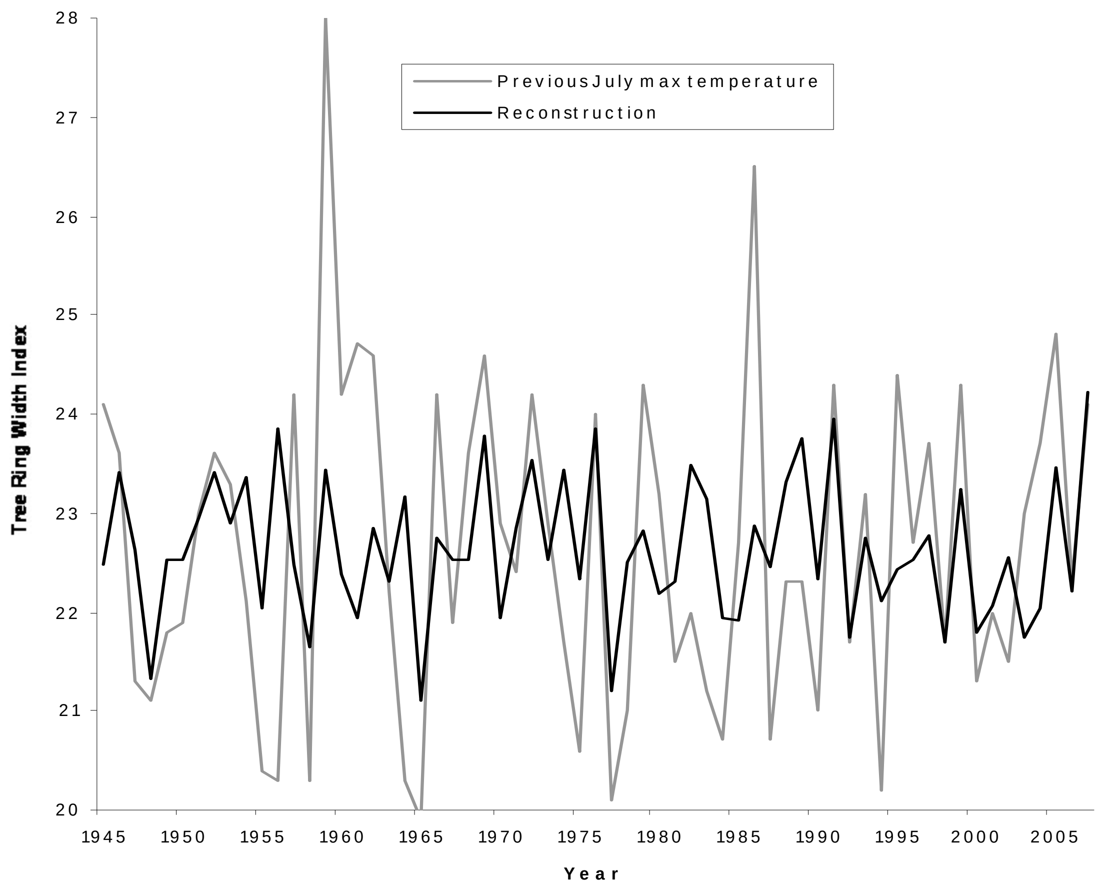


Figure 3.9: A graph of the common interval of data between instrumental previous July maximum temperature data from the Comox climate station (grey) and a reconstruction of previous July maximum temperature from the local subalpine fir chronology developed at the study site (black).

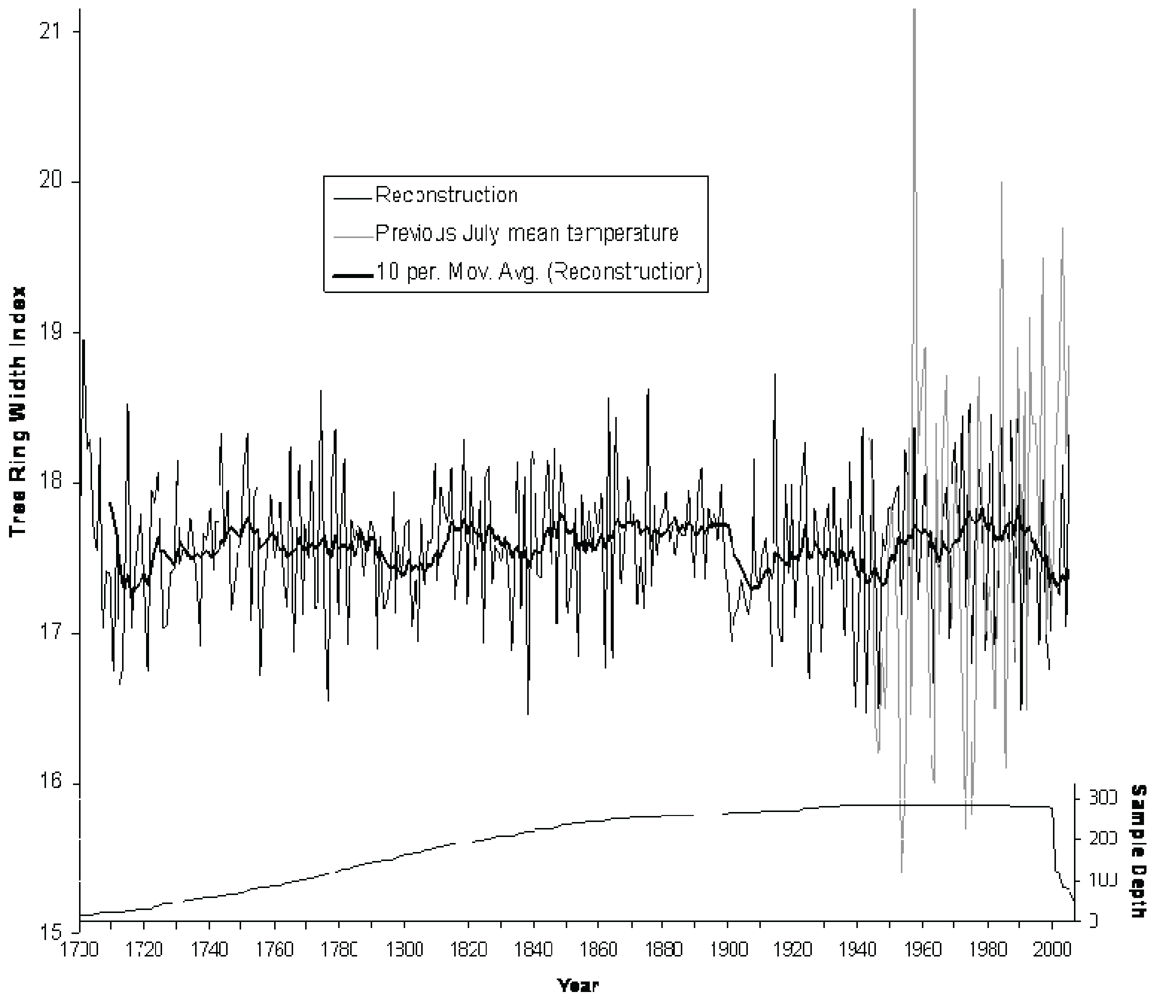


Figure 3.10: A graph of instrumental previous July mean temperature data from the Comox climate station (grey) and a reconstruction of previous July mean temperature from the regional subalpine fir chronology (black). A ten-year running mean has been plotted through the data, and sample depth over time is plotted at the bottom of the graph.

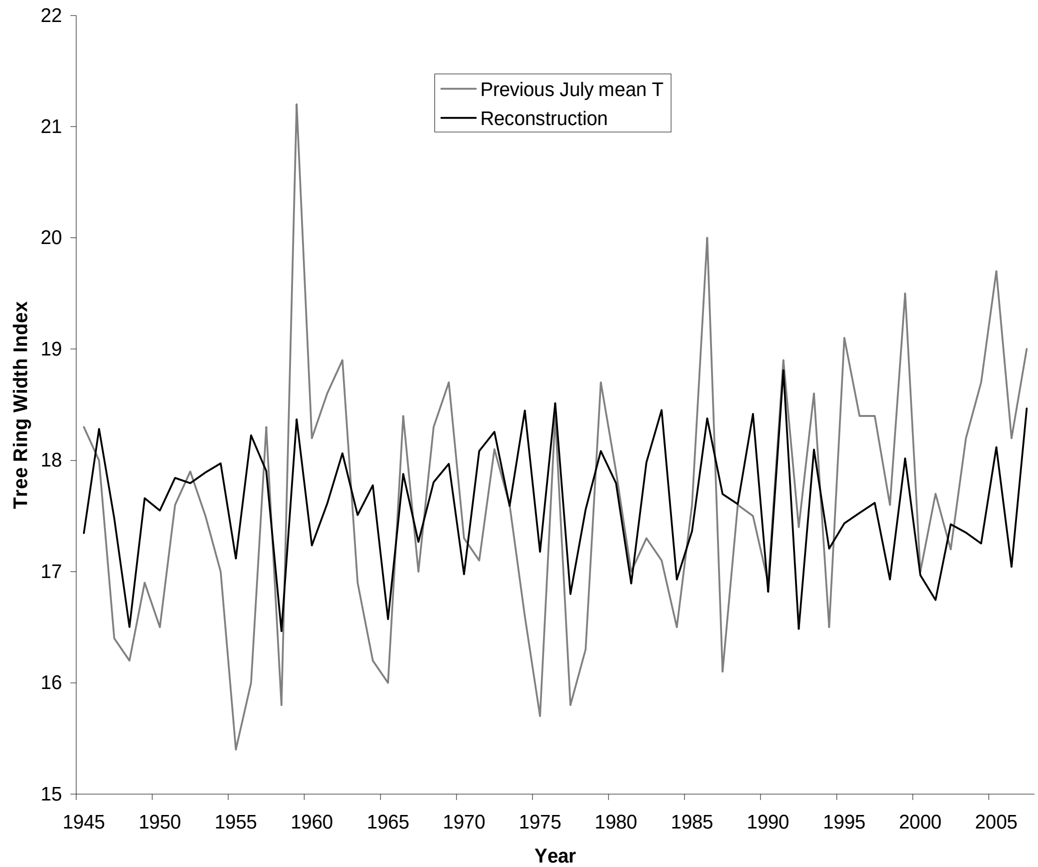


Figure 3.11: A graph of the common interval of data between instrumental previous July mean temperature data from the Comox climate station (grey) and a reconstruction of previous July mean temperature from the regional subalpine fir chronology (black).

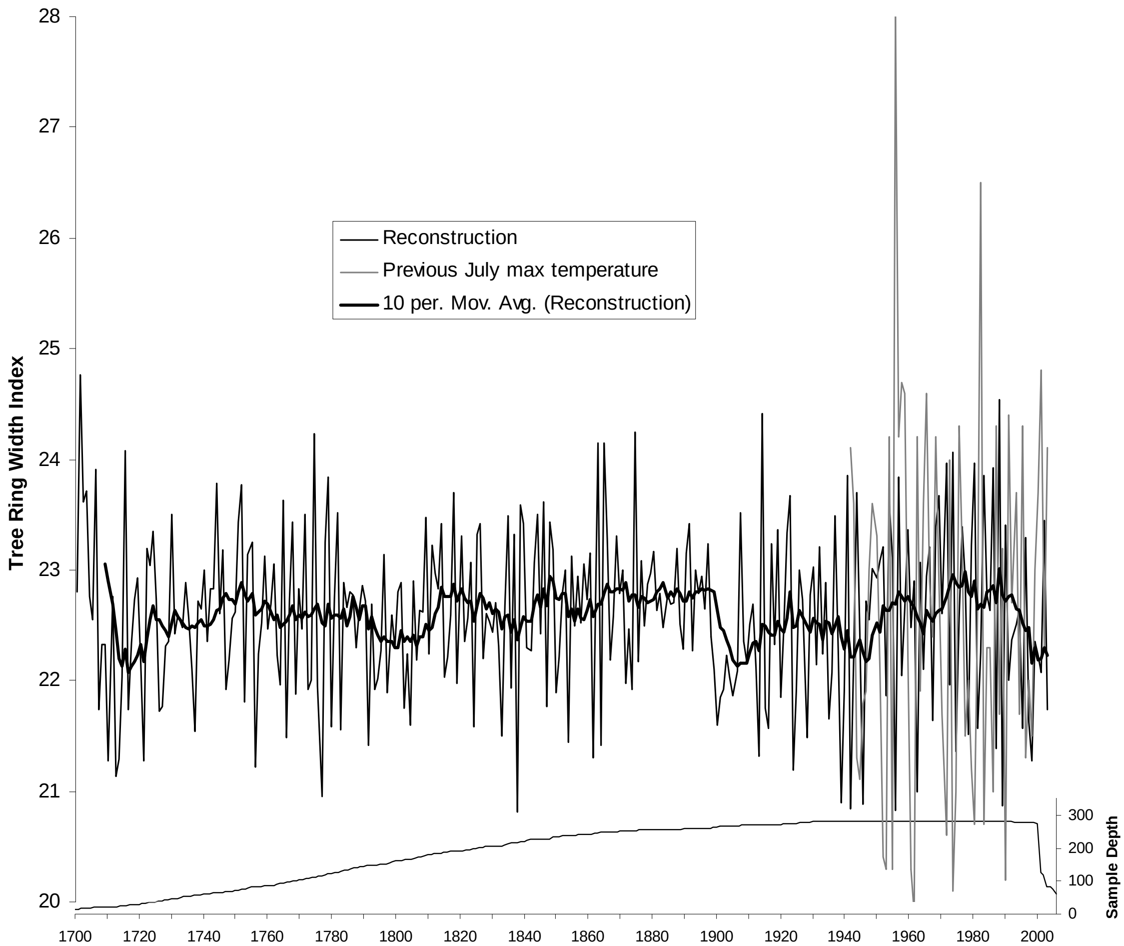


Figure 3.12: A graph of instrumental previous July maximum temperature data from the Comox climate station (grey) and a reconstruction of previous July maximum temperature from the regional subalpine fir chronology (black). A ten-year running mean has been plotted through the data, and sample depth over time is plotted at the bottom of the graph.

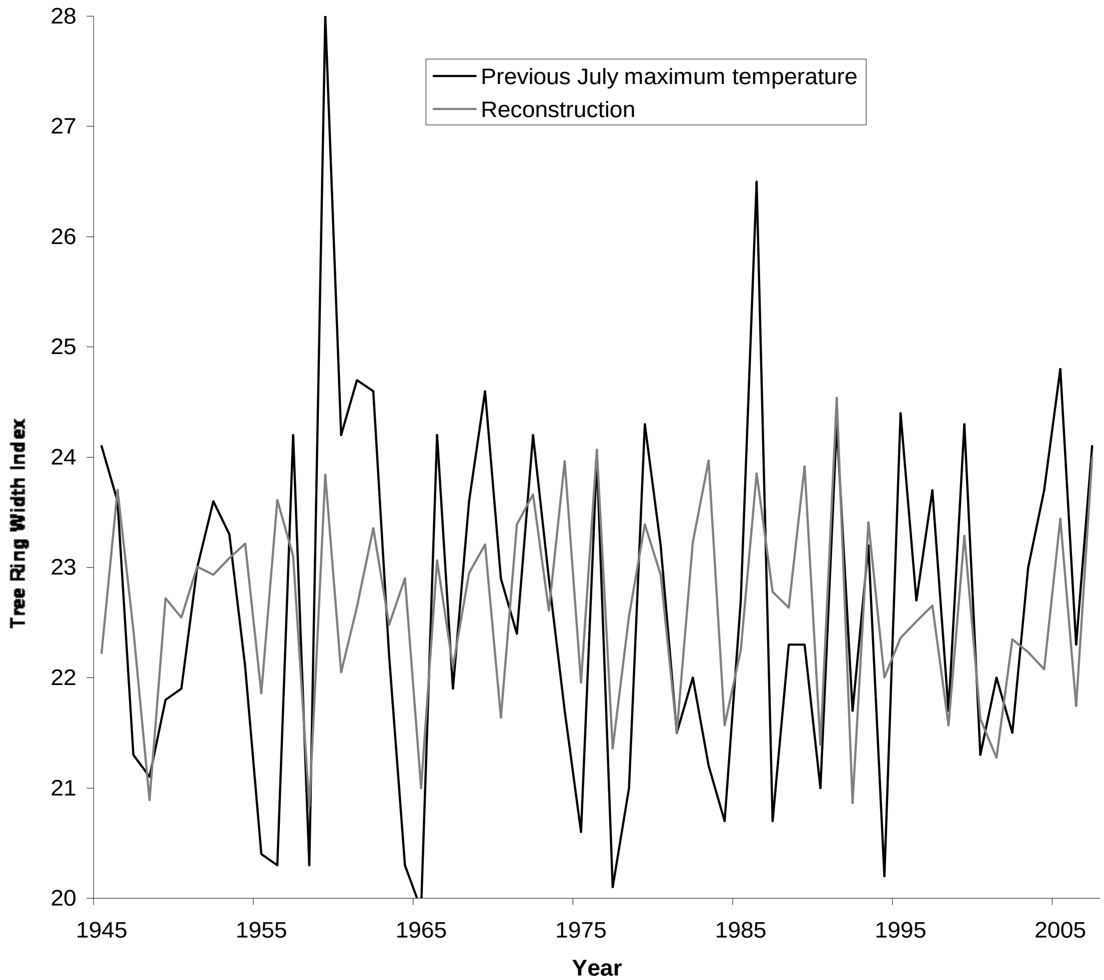


Figure 3.13: A graph of the common interval of data between instrumental previous July maximum temperature data from the Comox climate station (grey) and a reconstruction of previous July maximum temperature from the regional subalpine fir chronology (black).

## 3.7 Discussion

The proxy records of mean and maximum July temperature indicate that periods of cooler than average temperatures occurred from 1705 to 1717, 1769 to 1811, 1816 to 1834, 1890 to 1912, 1922 to 1950, in the 1970s, and from 1990 to 2006 (Figure 3.14). Synchronous episodes of positive mass balance conditions recorded in proxy records developed for glaciers located in the Coast Mountains, the Canadian Rocky Mountains and the Cascade Mountains (Lewis and Smith 2004a; Larocque and Smith, 2005; Osborn *et al*. 2007; Malcomb and Wiles 2009) (Table 3.10) highlight the regional coherence and glaciological significance of these cold episodes.

Cyclicity in the datasets, highlighted by ten-year running means, shows that a strong 15 to 20 year pattern extends from the 1950s to present in both the local and regional reconstructions of mean and maximum July temperature (Figures 3.6, 3.8, 3.10, and 3.12). Examination of temporal trends in the previous 250 years suggests that they are better characterized by 3 to 8 year cyclical fluctuations in temperature. The two regional chronologies exhibit underlying oscillations in frequency at 15 to 25 year intervals.

These cyclical tendencies were highlighted using wavelet analyses, which draws attention to the dominant modes of variability. As shown in Figure 3.15, the regional reconstructions of mean and maximum July temperatures are characterized by high frequency modes of variability with a period length of around 20 years and medium frequency variability with a period length of around 120 years. While the 20 year mode of variability is commonly associated with radial growth trends related to PDO fluctuations (Wiles *et al*. 1998; Gedalof and Smith 1999; Biondi 2000; Peterson *et al*.

2002; Larocque and Smith 2005), the 120-year mode of variability may be related to century-scale climate cycles related to variations in solar irradiance that are known to induce wide-ranging climatic responses (Yu and Ito 1999; Domack *et al*. 2001; Shindell *et al*. 2001; Neff *et al*. 2001; Steinhilber *et al*. 2009). Similar radial growth relationships have been observed in the dendroclimatological reconstructions presented by Larocque and Smith (2005) and in tree ring chronologies from Vancouver Island (Laroque 2002).

## 3.8 Conclusion

Mean and maximum July temperatures were reconstructed using both local (1820-2008) and regional (1700-2008) subalpine fir chronologies. The reconstructions explain between 13% and 36% of the overall climate variance over time which, while relatively low, was expected given the climatic and topographic complexity of the central Coast Mountains.

The reconstructions depict significant relationships between instrumental and reconstructed climate data in terms of direction and amplitude. The warm and cool intervals predicted by the reconstructions are consistent with other proxy environmental records from the surrounding region. Temporal trends within the reconstructions highlight the likely influence of PDO and ENSO climate oscillations on regional climate and are suggestive of the potential influence of changes in solar activity on the regional climate regime.

This research draws a connection between previous year warm summer temperatures and the radial growth response of subalpine fir trees in the central Coast Mountains. At the local scale, these conditions negatively influence the radial growth of subalpine fir trees found growing on a steep south-facing slope, even in a setting characterized by high summer and annual precipitation totals. At the regional scale, previous year warm summer temperatures appear to negatively influence the radial growth of subalpine fir trees at sites with very low summer precipitation totals. The similarity of the subalpine fir response at dry-summer regional sites to that of subalpine trees at the local site emphasizes the role of slope gradient and aspect in terms of the susceptibility of conifers to summer-temperature induced physiological stress. In conclusion, it appears that warm summer temperatures are the dominant climate parameters influencing the radial growth of subalpine fir trees at Coast Mountain sites with either low summer precipitation, or with higher precipitation combined with steep and/or south-facing slopes.

Table 3.10: Comparison of cool climate intervals recorded in PNA through tree ring climate reconstructions[a], lichenometric[b] and dendroglaciological[c] moraine dating, and lake sediment analysis[d].

| Source | Location | Cool Intervals | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| *This study* [a] | | *1705-1717* | *1748* | *1769-1811* | *1816-1834* | *1865* | *1890-1912* | *1922-1950* | *1975* | *1980-1985* |
| Ware and Thompson 2000[a] | PNA (USA and S. Canada) | | 1746-1754 | 1761-1781 | 1826-1854 | | 1879-1922 | | | |
| Briffa *et al*. 2001[a] | | 1605-1705 | | | 1810s, 1830s | | | | | 1980s |
| Larocque and Smith 2003[a] | Central Coast Mtns. | | | 1760-1785 | 1820-1840 | | 1870-1900 | 1915-1930 | | |
| Luckman and Wilson 2005[a] | Canadian Rocky Mtns. | 1685-1704 | 1727-1746 | 1799-1818 | 1819-1838 | | | | | |
| Allen and Smith 2007[b] | S. Coast Mtns. | 1706 | 1745, 1756 | 1780 | 1817, 1831 1836 | 1869 | 1891, 1911/ 12/ 14 /16 | 1925, 1930, 1949 | | |
| Heikkinen 1984[c] | Cascades | | 1740 | | 1820-1890 | | 1820-1890 | 1920 | | |
| Luckman 2000[c] | Canadian Rocky Mtns. | 1700-1725 | | | 1825-1850 | | 1850-1920 | | | |
| Moore and Demuth 2002[c] | S. Coast Mtns. | | | | | | Pre-1922 | | 1976 | |
| Lewis and Smith 2004a[c] | Vancouver Island | 1690-1710 | | | 1840 | | 1890 | 1930 | | |
| Osborn *et a*l. 2007[c,d] | S. Coast Mtns. | 1700s | | | 1830s | | 1890s | 1920s | | |

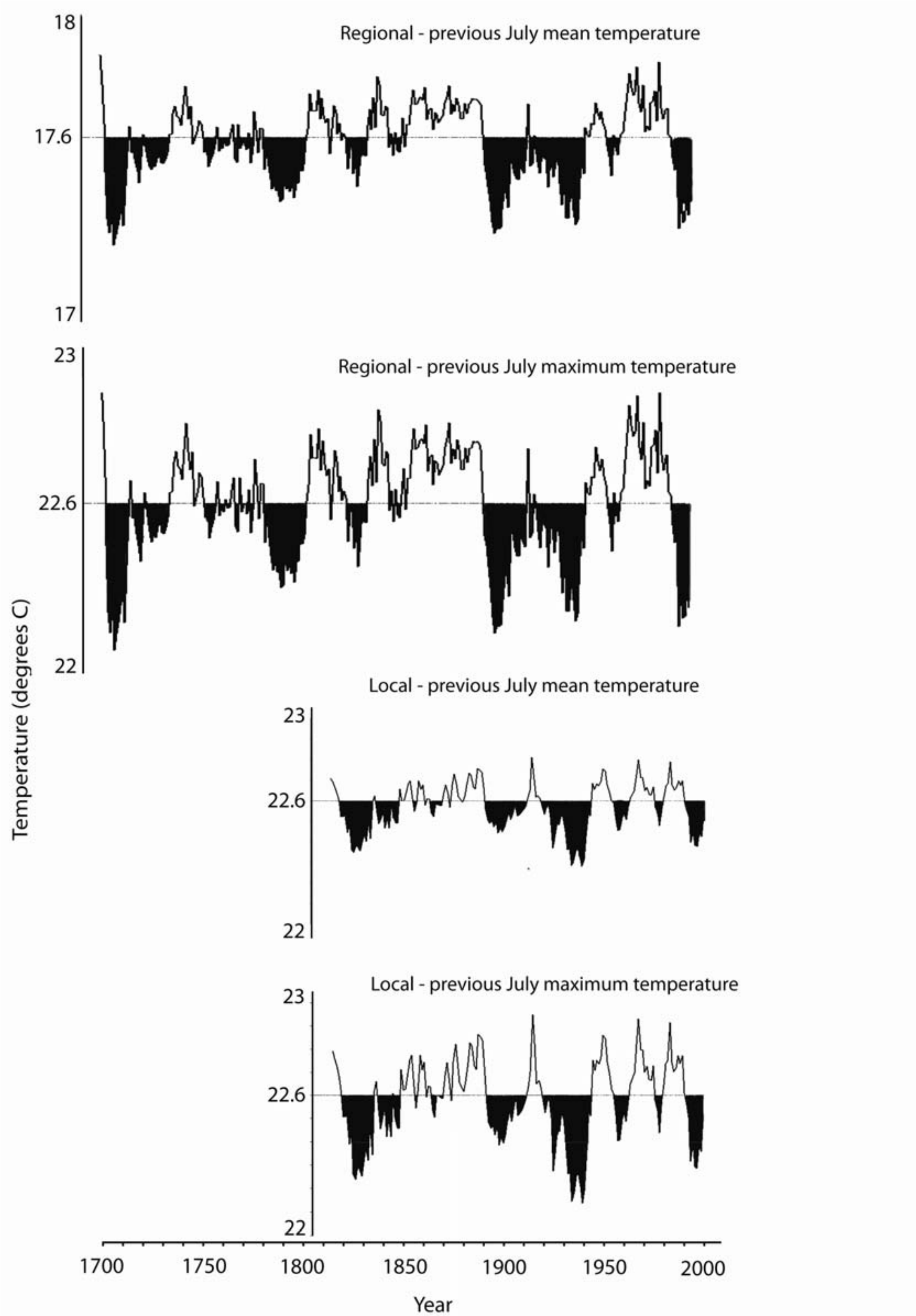


Figure 3.14: Local and regional proxy climate reconstructions of mean and maximum July temperature. White portions identify intervals of above average temperature and black portions identify intervals below average temperature.

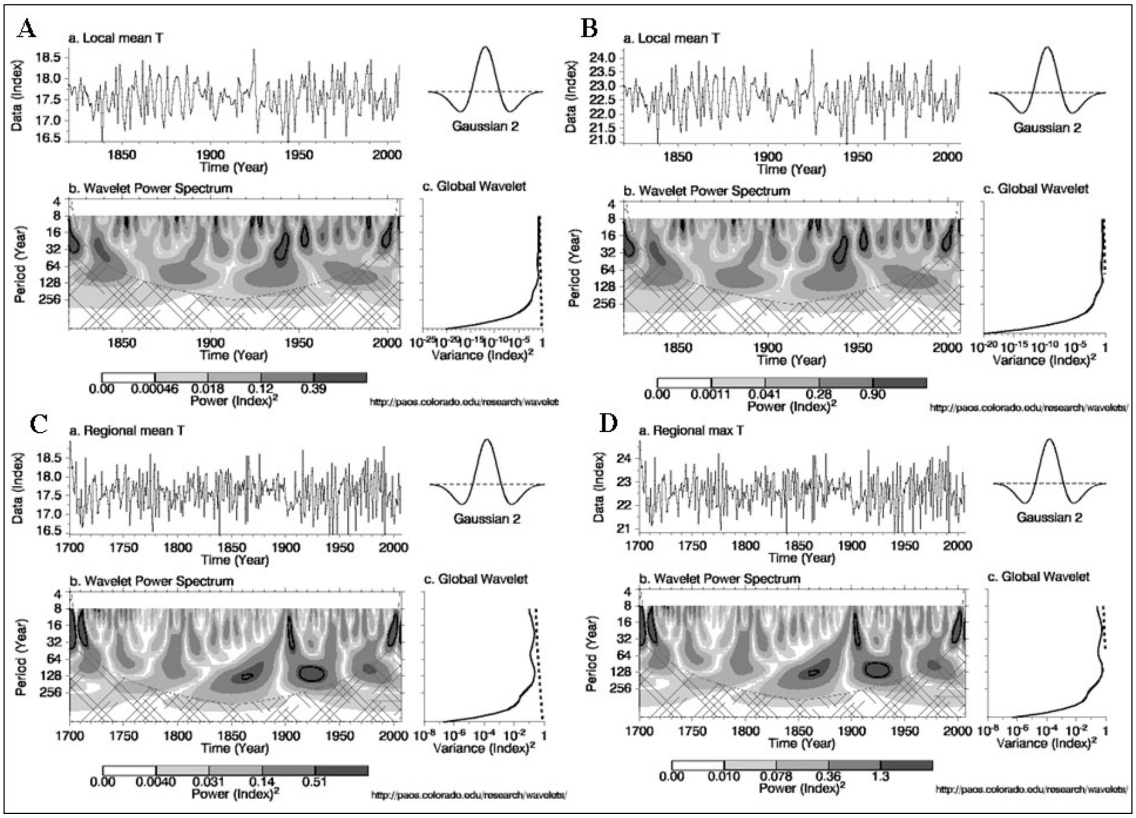


Figure 3.15: (a) Previous July maximum temperature reconstruction from the regional subalpine fir chronology. (b) Wavelet power spectrum. The contour levels represent 75%, 50%, 25%, and 5% of wavelet power. Cross-hatching represents the cone of influence where zero padding has reduced the variance. The black contours represent 5% significance levels, using a white-noise background spectrum. (c) Global wavelet spectrum (black line) and significance (dashed line) assuming the same background and significance level as in (b) (Torrence and Campo 1998).

# Chapter 4: Dendroglaciological investigations at Confederation and Franklin Glaciers, central Coast Mountains, British Columbia, Canada.

## 4.1 Introduction

Most glaciers in the Coast Mountains of British Columbia (B.C.) reached their maximum Holocene extent in recent centuries during the Little Ice Age (LIA) (Ryder and Thompson 1985). Since this event, the majority of glaciers in the region, and around the globe, have experienced significant periods of general retreat and downwasting (Grove 1988; Larocque and Smith 2003; Arendt *et al*. 2009). Today, most Coast Mountain glaciers are surrounded by recently exposed forefields and extensive unweathered LIA moraines. These features often overlie and bury older glacial deposits, hampering efforts to reconstruct records of pre-LIA glacier activity (Ryder and Thompson 1985; Larocque and Smith 2003; Reyes *et al.* 2006).

Our current understanding of Holocene glacial history in Pacific North America (PNA) is incomplete, with initial interpretations of glacial activity proving to be too simplistic as retreating glaciers expose more dendroglaciological evidence of repeated advances over the last 10 000 years (Menounos *et al*. 2009). The goal of this study was to contribute to a better understanding of Holocene glacial history in the Coast Mountains through dendroglaciological investigations at a low-elevation maritime valley glacier site where enhanced precipitation regimes create a distinct and little studied glaciological setting in B.C. Recent studies suggest that warm winter temperatures and fluctuating winter freezing level heights may affect the synchronicity and timing of maritime glacier responses to climate changes differently than continental glaciers (Arendt *et al*. 2009).

## 4.2 Previous Research

### *4.2.1 Holocene Glacial Activity in the British Columbia Coast Mountains*

During the Pleistocene epoch, the Coast Mountains of B.C. were mantled by the Cordilleran Ice Sheet (Ryder *et al*. 1991). After reaching its maximum extent ca. 14 500 to 14 000 cal. yrs BP, the ice sheet decayed rapidly and by ca. 11 500 cal. yrs BP glacial conditions in PNA were similar to those at present (Menounos *et al.* 2009). The ensuing interval from ca. 11 500 cal. yrs BP to present is referred to as the Holocene or post glacial epoch (Roberts 1998), and is characterized by considerably extensive and rapid climate fluctuations (Mayewski *et al*. 2004).

Paleoecological research in coastal B.C. identifies three distinct climate phases during the Holocene (Hebda 1995). The early Holocene or xerothermic period from 11 500 to 7800 cal. yrs BP is characterized by relatively warm and dry conditions (Mathewes 1985; Walker and Pellatt 2003). The mid Holocene or mesothermic interval from 7800 to 3800 cal. yrs BP is regarded as an interval of transition during which xerothermic conditions gradually gave way to cooler and wetter climates (Mathewes 1985; Walker and Pellatt 2003). The Neoglacial or late Holocene interval begins ca. 3800 cal. yrs BP and is an interval of relative cooling that has persisted until the present (Mathewes 1985; Walker and Pellatt 2003).

The terminal positions and thickness of glaciers in the Coast Mountains appear to have fluctuated in concert with these Holocene climate oscillations. In the southern Coast Mountains lake sediment records and subfossil wood evidence indicate an early Holocene glacier advance occurred ca. 8000 cal. yrs BP (Menounos *et al*. 2004). Correlative evidence of the regional character of this early Holocene advance comes from subfossil wood collected at Castle Glacier in the B.C. Cariboo Mountains (Mauer *et al*. 2009), and from buried subfossil wood located within glacial forefields on Mounts Baker and Rainier in Washington state (Clark and Gillespie 1997; Heine 1998; Thomas *et al*. 2000). Although Menounos *et al*. (2004) suggest that this advance may be associated with the well-documented 8200-year cold event in the North Atlantic region (Alley *et al*. 1997), paleobotanical evidence and calculated summer insolation values indicate that this period was 1 to 4 °C warmer than present in PNA (Reasoner *et al*. 2001).

Following this early Holocene advance, glaciers in the Coast Mountains receded before readvancing during the mid Holocene "Garibaldi Phase" (Ryder and Thompson 1986). Dendroglaciological investigations at glaciers in Garibaldi Provincial Park and elsewhere in the southern Coast Mountains indicate that between 6400 and 5800 cal. yrs BP ice-fronts advanced to positions ca. 800 m further down valley than they are at present (Stuiver and Deevey 1961; Lowden and Blake 1968, 1973; Ryder and Thomson 1986; Allen and Smith 2007; Koch *et al*. 2007b; Osborn *et al*. 2007; Menounos *et al*. 2008). Analogous glacier activity is reported in the Cascade Range of Washington state (Miller 1969) and in mountain ranges around the world (Thompson *et al*. 2006). An ice-building episode at this time is consistent with the transition from warmer, drier xerothermic climates to cooler, wetter conditions in the mesothermic period (Mathewes 1985; Hebda 1995). It is unknown whether ice accumulation during this period was slow and continuous, or associated with distinct advance episodes (Koch *et al*. 2007b).

The Garibaldi Phase was followed by an interval of glacier recession and a subsequent advance ca. 4200 cal. yrs BP (Menounos *et al*. 2008). Dendroglaciological and glaciolacustrine investigations indicate that glaciers in the Coast Mountains experienced positive mass balance conditions for several decades to a century (Osborn *et al*. 2007; Koch *et al*. 2007b; Koehler 2009; Menounos *et al*. 2008; Menounos *et al*. 2009). Proxy records confirm that climate conditions in the region were colder and wetter during this time (Menounos *et al*. 2008). Radiocarbon-dated wood samples from glacial forefields throughout the southern Canadian Cordillera suggest that this “4.2 ka” event was not as extensive as subsequent late Holocene advances (Menounos *et al*. 2008).

Following the early and mid Holocene intervals of ice expansion, glaciers throughout the Coast Mountains retreated before the initiation of regionally extensive late Holocene advances ca. 3500 cal. yrs BP (Ryder and Thomson 1986; Desloges and Ryder 1990; Larocque and Smith 2003, Koch *et al*. 2003b, 2007b; Reyes and Clague 2004; Allen and Smith 2007; Osborn *et al*. 2007; Jackson *et al*. 2008; Koehler, 2009; Menounos *et al*. 2009). Accumulating evidence from sites in PNA make it increasingly clear that the late Holocene was characterized by multiple episodes of glacier expansion and retreat, culminating in most cases with the LIA glacier advances.

The Tiedemann Advance is recognized as the earliest of the late Holocene advances. Originally described by Ryder and Thompson (1986) as having occurred from 3000 to 1800 cal. yrs BP, recent investigations indicate that the Tiedemann Advance is distinguished by an early and a late phase (Allen and Smith 2007; Koch *et al*. 2007b; Koehler 2009; Menounos *et al*. 2009). In southern B.C., paleoclimate records have consistently reported a gradual shift from warmer, drier mesothermic conditions toward cooler, wetter conditions coeval with the onset of the Tiedemann Advance (Mathewes and Heusser 1981; Mathewes 1985; Clague and Matthewes 1996; Pellatt and Mathewes 1997; Palmer *et al*. 2002; Hallett *et al*. 2003). At sites in the Coast Mountains

Tiedemann-aged deposits distal to LIA moraines suggest that some glaciers reached their maximum Holocene extent during this interval (Ryder and Thompson 1986; Larocque and Smith 2003).

The earliest evidence of Tiedemann-age expansion in the Coast Mountains indicates that many glaciers were advancing down valley by at least 3500 cal. yrs BP (Haspel *et al*. 2005; Spooner *et al*. 2005; Jackson *et al*. 2008; Koehler 2009). Menounos *et al*. (2009) suggest that the early Tiedemann phase may have been characterized by multiple pulses of ice advance, similar to those that occurred during the LIA in this region. Analogous glacier expansion was underway throughout PNA (Clark and Gillespie 1996; Burke and Birkeland 1983; Wiles and Calkin 1990; Calkin *et al*. 2001; Barclay *et al*. 2009), including in the Canadian Rocky Mountains where this interval of ice expansion is referred to as the Peyto Advance (Luckman *et al*. 1993; Wood and Smith 2004).

The early Tiedemann Advance was followed by an interval of glacier retreat and downwasting of sufficient duration for trees to colonize and grow to be more than 250 years old at some sites (Koehler 2009). Glacier expansion during the subsequent late Tiedemann Advance ca. 2300 cal. yrs BP is recorded by buried trees and sheared stumps found in growth position above early Tiedemann-age paleosols (Desloges and Ryder 1990; Koehler 2009). Correlative evidence of ice expansion has been found on glacier forefields and within lateral moraine deposits at numerous Coast Mountain sites (Ryder and Thomson 1985; Clague and Matthewes 1992, 1996; Johnson *et al*. 1997; Reyes and Clague 2004; Allen and Smith 2007; Koch *et al*. 2007b; Menounos *et al*. 2009; Mauer *et al*. 2009). There is also limited evidence of a similar late Tiedemann episode of ice expansion in the Canadian Rocky Mountains (Luckman *et al*. 1993; Osborn *et al*. 2001; Luckman 2006). As with the early Tiedemann interval, paleoclimatic investigations suggest that the Coast Mountains were characterized by cool and moist conditions at this time (Pellatt *et al*. 2000; Spooner *et al*. 2003; Lamoureux and Cockburn 2005). The existence of two distinct climate episodes during the Tiedemann period is highlighted at Moraine Bog near Tiedemann Glacier where three distinct pollen zones — before 2600 $^{14}$C yrs BP, from 2500 to 2300 $^{14}$C yrs BP, and after 1900 $^{14}$C yrs BP — have been identified (Arsenault *et al*. 2003).

Following the Tiedemann-age advances, glaciers throughout the region retreated and downwasted before expanding during the First Millennial Advance (FMA) from ca. 1800 to 1300 cal. yrs BP (Reyes *et al*. 2006). Dendrochonological and stratigraphic investigations suggest there were two distinct FMA phases, an initial interval of ice expansion between ca. 1360 and 1440 cal. yrs BP and a second interval between ca. 1600 and 1800 cal. yrs BP (Jackson *et al*. 2008; Barclay *et al*. 2009). The FMA corresponds to an interval of generally cool temperatures (Bond *et al*. 1997; Hu *et al*. 2001; Pierce *et al*. 2004; Moberg *et al*. 2005) that resulted in glacier expansion into standing forests throughout PNA (Reyes and Clague 2004; Laxton 2005; Allen and Smith 2007; Koch *et al*. 2007b; Jackson *et al*. 2008). Advances in the northern Coast Mountains and in the Cariboo Mountains suggest that some glaciers expanded to attain ice-front positions comparable to those later reached during the LIA (Reyes *et al*. 2006; Jackson *et al*. 2008; Mauer *et al*. 2009).

By ca. 1000 cal. yrs BP, glaciers throughout PNA were advancing down valley in response to the global LIA climate interval (Grove 1988). In the Coast Mountains, three distinct LIA advances are broadly recognized: from ca. AD 1100 to 1200, from ca. AD 1600 to 1700, and ca. AD 1850 to 1900 (Ryder and Thomson 1985; Desloges and Ryder 1990; Smith and Desloges 2001; Larocque and Smith 2003; Lewis and Smith 2004a; Reyes and Clague 2004; Koch *et al*. 2007a; Jackson *et al*. 2008; Menounos *et al*. 2008; Koehler 2009). At most sites in the Coast Mountains the LIA limits mark the maximum Holocene glacial extent (Menounos *et al*. 2009), with advances culminating for the most part in the period between AD 1600 and 1900 (Menounos *et al*. 2008). Comparable LIA activity is described from sites throughout PNA (Miller 1969; Sigafoos and Hendricks 1972; Fuller 1980; Burbank 1981; Heikkinen 1984; Luckman 1986; Smith *et al*. 1995; Wiles *et al*. 1999a, 1999b; Luckman 2000; Calkin *et al*. 2001; Barclay *et al*. 2009), although there is evidence for an earlier onset of LIA activity in the Coast Mountains (Spooner *et al*. 2005; Allen and Smith 2007; Jackson *et al*. 2008). Recent research suggests that these distinct periods of expansion and retreat occurred in response to climate fluctuations related to variations in solar insolation over the last millennium (Wiles *et al*. 1999b; Larocque and Smith 2003; Barclay *et al*. 2009).

## 4.3 Study Area

Fieldwork was completed on recently deglaciated terrain in the vicinity of the Confederation and Franklin glaciers confluence in the central Coast Mountain region (Figure 4.1). The study area is located approximately 17 km west of the summit of Mount Waddington and 18 km from the head of Knight Inlet (Figure 4.2). This high relief landscape contains some of the tallest peaks in B.C. and is characterized by granitic bedrock, rugged terrain, glaciers, and glacial landforms (Clarke and Holdsworth 2002). Both glaciers descend to terminal positions located well below the local treeline at approximately 1650 m asl. Runoff from the Franklin Glacier drains into the Franklin River and eventually to the Pacific Ocean.

Figure 4.1: The Franklin Glacier (right) and Confederation Valley (bottom) confluence in the central Coast Mountain region. Mount Waddington is the prominent peak shown in the background (© Scurlock, J. 2007).

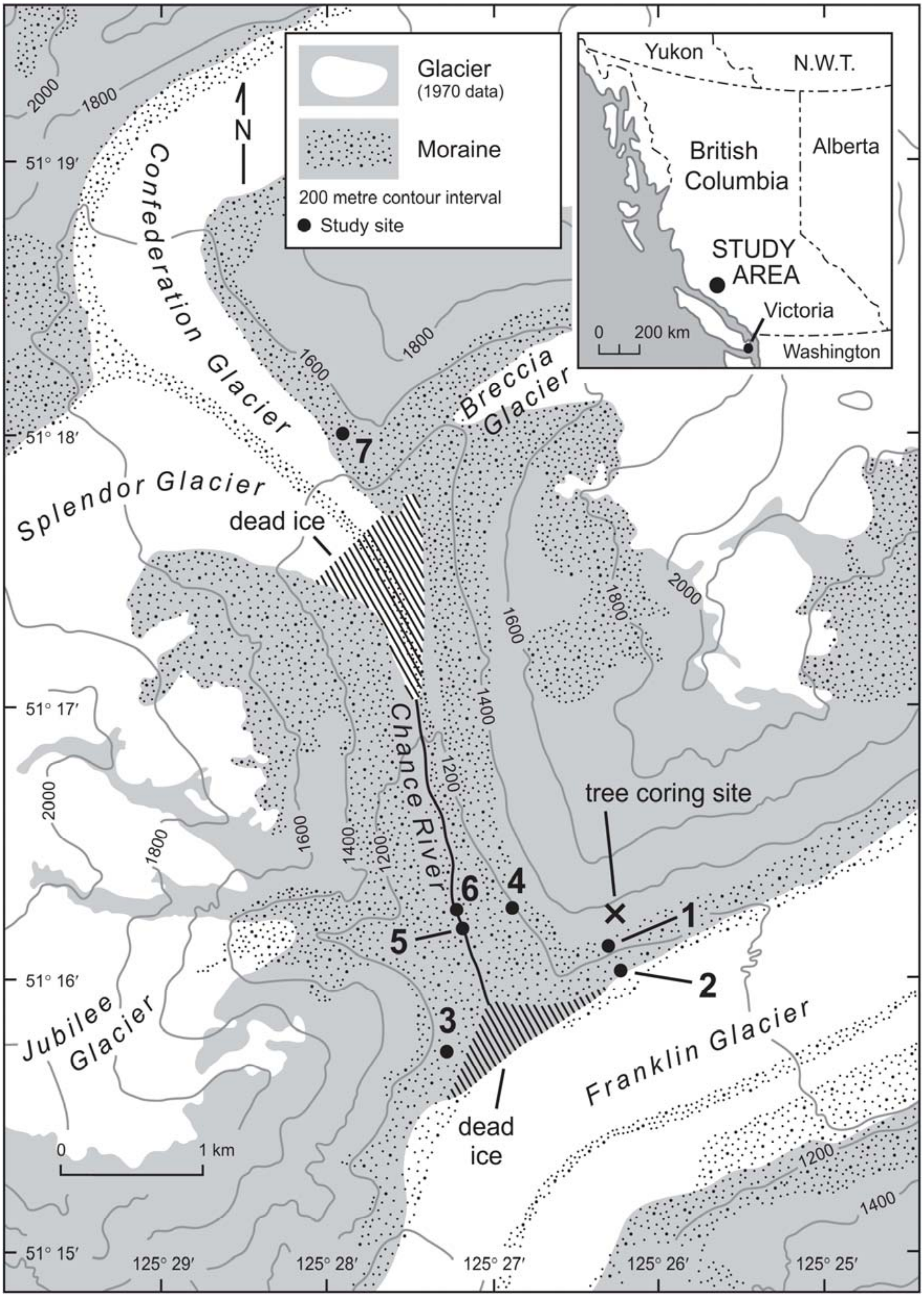


Figure 4.2: A map showing the Confederation and Franklin glaciers study area, central Coast Mountain region.

The climate of the central Coast Mountain region is strongly influenced by weather systems originating in the Pacific Ocean, characterized by average annual air temperatures of 1.1 ºC and average annual orographic precipitation totals around 1700 mm (Wang *et al*. 2005). The surrounding montane forests are composed of mountain hemlock (*Tsuga mertensiana*) and subalpine fir (*Abies lasiocarpa*) trees, with undisturbed valley bottom forests composed of western red cedar (*Thuja plicata*) and western hemlock (*Tsuga heterophylla*) trees (Meidinger and Pojar 1991).

Based on their proximity to the Pacific Ocean and the prevalence of high annual precipitation and moderate annual temperature regimes, Confederation and Franklin glaciers are classified as maritime glaciers (Letréguilly and Reynaud 1989; Lewis and Smith 2004a). These glaciers were first explored by mountaineers Don and Phyllis Munday in the summer of 1927 (Watson and King 1935; Munday 1948). At that time there was evidence of a recent major advance in the region. Munday (1939) describes moraine features overrun and buried by large-scale composite lateral moraines bordering Franklin Glacier. Trees growing on the Franklin Glacier terminal moraine were estimated to be no more than 60 years old, indicating that the glacier had only recently reached its maximum Holocene down valley extent. Buried and drowned wood deposits examined at nearby Klinaklini, Scimitar, and Waddington glaciers lend further support to Munday's (1939) contention that glaciers throughout the Mt. Waddington area had recently advanced. Munday (1939) estimated that the terminus of the Franklin Glacier was retreating at 'no less' than 90 m/yr, and that recently deposited push moraines indicated that this ice retreat was periodically interrupted by minor readvances.

Following these early observations, Ryder and Thomson (1985) reported on buried tree roots and wood fragments found 25 m below the summit of lateral moraines located near the confluence of Confederation and Franklin glaciers. An *in situ* subalpine fir tree root in the associated paleosol was interpreted to suggest that the Franklin Glacier was advancing down valley at 835 ± 45 $^{14}$C yrs BP (Sample S-1568, Table 4.3)

Fieldwork completed on the eastern flank of the Waddington Range by Larocque and Smith (2003, 2004) allowed for the documentation of moraine stabilization intervals in AD 620, 925–933, 1443–1458, 1506–1524, 1562–1575, 1597–1621, 1657–1660, 1767–1784, 1821–1837, 1871–1900, 1915–1928, and 1942–1946. Analysis of historical glacier-climate relationships revealed that El Niño Southern Oscillations (ENSOs), Pacific Decedal Oscillations (PDOs), and perturbations in solar insolation activity interact in this region to cause a repetitive 65-year cycle of positive mass balance conditions (Larocque and Smith 2005).

VanLooy and Forster (2008) employed remote sensing techniques to determine that the Franklin Glacier retreated ca. 4100 m in the period between 1926 and 1999 (mean 56 m/yr), noting several minor intervals of advance and retreat occurring between 1926 and 2001. They also report that glacial responses to climate in the area are complex and that the most rapid ice thinning occurred between 1927 and 1974 when winter temperatures were lower and snowfall amounts higher than present. They suggest that increases in winter temperatures, as well as increases in winter rainfall and a decrease in winter snowfall totals since the mid-1980s have jointly contributed to recent ice-thinning.

### *4.3.1 Confederation Glacier*

Confederation Glacier flows approximately 4 km southwest from the 2700 m peaks of Mounts Myrtle and Bezel down a bedrock-confined valley to its present-day snout position at 1400 m asl, close to the point of confluence with tributary Splendor Glacier (Figure 4.2). At its maximum Holocene extent Confederation Glacier filled the 8 km long Confederation Valley (*unofficial nam*e) and was confluent with the Franklin Glacier at approximately 900 m asl (Figure 4.2).

In the 1920s Confederation Glacier was fed by tributary flows from Breccia Glacier from the east and Jubilee Glacier from the west. The snout of Breccia Glacier is currently located in a tributary valley approximately 1.5 km from Confederation Valley. Jubilee Glacier has undergone extensive vertical retreat and downwasting and its snout now calves from a steep bedrock wall approximately 1 km from the margin of the Confederation Valley (Figure 4.2). The retreat of all four glaciers has exposed a 4 km long expanse of glacial forefield. The ice front margins of both Confederation and Splendor glaciers are now characterized by extensive tongues of stagnant, sediment-covered ice.

The recently deglaciated slopes of Confederation valley are heavily eroded and extensive bedrock outcrops characterize much of the valley margin. Sections of the valley are demarcated by a single high-relief, sharp-crested lateral moraine deposit. An oblique photograph taken by Munday (1939) shows that approximately 80 years ago Confederation Valley was ice-filled to within a few metres of the present-day lateral moraine crest (Figure 4.4). At that time the glacier surface was characterized by multiple medial moraines as a result of ice and debris inputs from Splendor, Breccia, and Jubilee glaciers.


Figure 4.3: Illustration of historical ice margins of Confederation Glacier shown on a Google Earth image from 2007. Delineated historical ice front positions from vertical aerial photographs taken in 1965, 1978, and 1996 (B.C. Air Photo Library, 1:40,000).

Figure 4.4: A photograph taken by Don Munday in 1927 of the confluence of Franklin Glacier (foreground) and Confederation Glacier (centre). Jubilee Glacier (left) and Breccia Glacier (right) flow into Confederation Glacier.

Prominent lateral moraines flank Confederation Valley. The moraine crests consist of a series of smaller unvegetated nested moraines that are interpreted as being linked to periods of LIA glacier expansion. Standing approximately 200 m above the valley floor, the eroded proximal face of the lateral moraines contain laterally homogeneous till units that are separated at intervals by subfossil wood mats (Figure 4.1). Many of the steep gullies eroded into this deposit contain scattered subfossil wood fragments, as do the surfaces of many of the underlying colluvial fans.

Chance River (*unofficial name*) flows from a subglacial portal in the stagnating snout of Confederation Glacier before disappearing below an extensive field of dead ice at the confluence of Confederation Valley and Franklin Glacier (Figure 4.2). Near the confluence of Jubilee and Confederation valleys, a 200 m long section of the river is characterized by a massive unit of sorted sand. The deposit exceeds 6 m in depth and is incised by terrace surfaces left behind as Chance River has eroded through them.

### *4.3.2 Franklin Glacier*

Franklin Glacier is a large valley glacier approximately 12 km long and up to 2 km wide. The glacier flows southwest from Mount Waddington to a snout position at 610 m asl, approximately 4 km down valley from its point of confluence with Confederation valley (Figures 4.1 and 4.2). In the immediate vicinity of Confederation Valley, the hanging valleys of Chasm, Splinter, Marvel, Yataghan, and several other smaller once-tributary glaciers stand above Franklin Glacier. Further up-valley, the larger Dauntless, Repose, and Whitetip glaciers contribute directly to the flow of the Franklin (Munday 1939).

Like Confederation Glacier, Franklin Glacier has retreated and downwasted significantly in recent decades (VanLooy and Forster 2008). The photograph taken by Munday in 1927 shows that the Franklin Valley was ice-filled at that time (Figure 4.4). It also shows that a sizeable medial moraine extended down valley from the confluence of Confederation and Franklin glaciers. Lateral moraine crests along the perimeter of Franklin Glacier currently stand an estimated 300 m above the ice surface. A sizeable field of dead ice originating either from Franklin Glacier, Confederation Glacier, or a combination of both, extends almost 1 km from the perimeter of Franklin Glacier into Confederation Valley. In the gullies of the upper northeastern Franklin lateral moraine, near its confluence with Confederation Glacier, two distinct horizontal till units are visible. Subfossil boles and tree roots separate these units and were previously examined by Ryder and Thompson (1985).

## 4.4 Research Methods

Field investigations were carried out near the confluence of Confederation and Franklin glaciers in July 2008. Subfossil wood samples were collected at various locations to build a record of Holocene glacier activity. The kill date of a subfossil sample buried by an advancing glacier or buried by moraine building activity is interpreted to provide a minimum date for an individual glacial event.

### *4.4.1 Sample Collection*

Standard dendroglaciological research methods rely on local living tree chronologies in order to situate floating tree-ring series in time (Luckman 1998; Smith and Lewis 2007). Living tree core samples were collected from a subalpine forest positioned above the ice limit in order to assign absolute dates to the floating tree-ring chronologies developed from subfossil wood samples (Stokes and Smiley 1968).

The trees were sampled using a standard 5-mm increment borer. Older trees were selectively sampled to provide the longest possible record of tree ring-width variability (Fritts 1976). Two core samples were removed from each tree at 180º from one another, with care taken to avoid growth irregularities, stem rot, and other physical damage (Stokes and Smiley 1968; Schweingruber *et al*. 1990). Cores were removed at standard breast height and stored in labelled plastic straws.

Rooted samples in growth position, *in situ* samples located within glacier deposits, and scattered detrital samples were collected from locations on the valley floor and from lateral moraine faces. Detrital samples were also collected from talus deposits below the eroding proximal faces of lateral moraines. Sampling was conducted by cutting cross sectional disks from the wood samples with a chainsaw. The samples were wrapped in duct tape, labelled, and transported back to the University of Victoria Tree-Ring Laboratory for analysis.

### *4.4.2 Sample Preparation and Ring-width Measurement*

Tree core samples were dried and glued to slotted mounting boards. The subfossil samples were glued where necessary to preserve their structural integrity. Both sample sets were polished with progressively finer sandpaper to either a 600-grit or 1200-grit finish using a belt sander. Once the ring boundaries were clearly visible, the width of each annual ring was measured to 0.001 mm using a Velmex measuring stage system and a Wild M3B stereomicroscope equipped with a Sony 3CCD video camera. Ring-width measurements were recorded along the centre of each core sample. Two or more paths from perimeter to pith were measured on each cookie sample. Reaction wood and other anomalous growth were avoided in order to ensure that all rings were measured. Once captured by J2X software (v3.2.1, 1994) the raw tree ring data were converted to standard decadal format using the program FMT from the Dendrochronology Program Library (DPL) (Holmes 1994).

Ring-width measurement data was cross-dated using the program COFECHA 3.0 (Holmes 1983). This quality-control program uses segmented time series correlation techniques to assess the quality of cross-dating in measurement series (Grissino-Mayer 2001). The program applies a Pearson’s product moment correlation at the 99% significance level in order to check the overall quality of tree-ring chronologies, and assists in evaluating measurement accuracy and identifying outliers (Grissino-Mayer 2001). COFECHA also removes the low-frequency growth trends that are inherent in raw tree ring data via a combination of spline fitting, autoregressive modeling, and log transformations (Grissino-Mayer 2001). All reported series correlation values (*r*) were calculated by the program COFECHA and are significant at the 99% confidence level.

Each of the time series pairs were internally cross-dated to ensure that no rings were missed during measuring. The internally cross-dated tree-ring series from the living cores were subsequently cross-dated with the remaining cores from each sampling site to produce species-specific master chronologies.

Tree ring radial growth is more rapid when the trees are young (Fritts 1976). This age-related growth trend was removed by fitting a negative exponential curve, a linear regression, or a horizontal line passing through the mean using the software program ARSTAN (Cook and Krusic 2005). Individual ring-widths were then divided by the value of the fitted curve for that year to calculate a dimensionless tree ring width index (Stokes and Smiley 1968; Fritts 1976). These indices were combined to create a standardized master chronology for each species (Stokes and Smiley 1968; Fritts 1976). Standardization was followed by a second detrending to reduce any non-climatic variability associated with endogenous or exogenous disturbance events or other stand-related dynamics (Cook and Peters 1981; Cook and Holmes 1986). In this instance, a spline with a 50% response frequency over 67% of the series length was employed.

Floating site-specific chronologies were constructed from the subfossil wood samples following established cross-dating protocols (Fritts 1976; Stokes and Smiley 1968; Grissino-Mayer 2001). An attempt was made to determine the absolute age of each floating chronology by cross-dating to the standardized living chronology. Where this was unsuccessful, efforts were made to cross-date floating chronologies to both living and radiocarbon-dated floating chronologies developed from the surrounding region (Smith and Laroque 1998; Larocque and Smith 2005; Parish 2006; Allen and Smith 2007; Jackson *et al*. 2008; Koehler 2009; Hart 2009). Where cross-dating failed, the outer rings from selected samples contributing to the floating chronologies were radiocarbon-dated by Beta Analytic Inc. Calendar ages were ascribed to the assigned $^{14}C$ ages by using a cubic spline fit (Talma and Vogel 1993). Midpoint dates were used to calibrate $^{14}C$ ages. Dates are reported in calendar years BP (cal. yr BP), with the exception of late LIA dates, which are reported in years AD. In the case of multiple age intercepts, all possible dates are reported.

Microscopic wood identification was used to determine the species identity of select subfossil samples. Radial, tangential, and transverse thin sections (20-30 microns) were cut from the subfossil samples using a sliding microtome, washed in an ethyl alcohol bath, and mounted on slides. Some samples were fixed to slides with a small amount of permou nt and left on a hotplate at 50ºC for 12 hours, after which time they were cleaned again with xylene (Hoadley 1990). Sample species were then identified with the aid of anatomical keys (Panshin and Zeeuw 1980), using a Motic BA310 microscope.

## 4.5 Observations

### *4.5.1 Dendrochronology*

Thirty-one cores from 16 of the 30 mountain hemlock trees sampled were included in the chronology (Table 4.1). The mountain hemlock chronology extends over 409 years (AD 1599-2009) and has a mean series correlation of $r$ = 0.521. The subalpine fir chronology was constructed from 28 cores from 15 of the 26 subalpine fir trees sampled. The 315-year chronology (AD 1693-2008) and has a mean series correlation of *r* = 0.573 (Table 4.1). Almost all of the trees at the site exhibited severely hockey stick shaped trunks, with some individuals showing evidence of fungus growing on their trunks and on exposed root systems.

### *4.5.1 Dendroglaciology*

Subfossil boles and wood fragments were collected at seven locations in the study area (Table 4.2) (Figure 4.2). The majority of the sites were located within or below the lateral moraines near the historical confluence of the Confederation and Franklin glaciers (Sites 1, 2, 3, and 4, Table 4.2) (Figures 4.2 and 4.5), or on the recently deglaciated floor of Confederation Valley (Sites 5 and 6, Table 4.2) (Figures 4.2 and 4.5). Additional samples were collected within a west-facing lateral moraine located 1.5 km from the 2008 snout position of Confederation Glacier (Site 7, Table 4.2) (Figures 4.2 and 4.5).

*Site 1*

Site 1 is located within deep gullies on the upper slopes of the most distal northeastern Franklin lateral moraine at approximately 1200 m asl (Table 4.2) (Figures 4.2 and 4.5). Boles with mean diameters of 100 cm were found protruding laterally from an organic layer located approximately 10 m below the moraine crest. A root with a radiocarbon age of 835 +/- 45 $^{14}$C yrs BP believed to be from this horizon was previously collected by Ryder and Thomson (1985) (Sample S-1568, Table 4.3).

When Site 1 was visited in 2008, the vertical proximal face of the lateral moraine rendered most boles inaccessible. Sampling was restricted to the upper portion of a single gully where a log had spilled from the proximal face (Figure 4.5). The gully headwall also provided access to a buried paleosol where a stump in growth position was sampled (Sample 01A, Tables 4.2 and 4.3) (Figure 4.5).

Table 4.1: Living and subfossil tree-ring chronology statistics

| Species | No. cores/subfossil samples | No. of series | Interval (years) | Total length (years) | Pearson's value | Mean sensitivity | $^{14}$C Samples |
|---|---|---|---|---|---|---|---|
| Mountain hemlock | 16 | 30 | AD 1586-2006 | 421 | 0.493 | 0.267 | - |
| Subalpine fir | 15 | 28 | AD 1691-2007 | 317 | 0.573 | 0.206 | - |
| Site Chrono. 1 | 2 | 4 | - | 193 | 0.610 | 0.251 | 01A |
| Site Chrono. 2 | 5 | 17 | - | 332 | 0.492 | 0.233 | - |
| Site Chrono. 3 | 10 | 17 | - | 332 | 0.492 | 0.231 | 05D |
| Site Chrono. 4 | 17 | 23 | - | 308 | 0.488 | 0.260 | - |
| Site Chrono. 5*a* | 1 | 2 | - | 175 | 0.466 | 0.253 | 01R |
| Site Chrono. 5*b* | 1 | 2 | - | 174 | 0.584 | 0.268 | 04R |
| Site Chrono. 6 | 1 | 2 | - | 169 | 0.588 | 0.267 | 06R |
| Site Chrono. 7 | 1 | 2 | - | 111 | 0.477 | 0.234 | 04I |
| Float A | 20 | 37 | AD 1037 - 1658 | 622 | 0.469 | 0.249 | 01A 05D |
| Float B | 2 | 4 | 3409 - 3571 cal. yrs BP | 163 | 0.448 | 0.251 | 01R 06R |
| Float C | 4 | 8 | - | 353 | 0.519 | 0.224 | 04R |

Tree rings from the two samples cross-date to produce a 193-year floating chronology ($r$ =0 .610) (Float 1, Table 4.1). Outer rings on the stump were radiocarbon dated to 600 +/- 50 14 C yrs BP (Sample 01A, Table 4.3) indicating that Float 1 spans the interval from AD 1355 to 1548 (Table 4.1).

Table 4.2: Subfossil and living tree study site names and their associated UTM coordinates and elevation values.

| Site Name | Lat. (N) | Long. (W) | Elevation (m asl) |
|---|---|---|---|
| Site 1 | 51º16’07“ | 125º26’20“ | 1200 |
| Site 2 | 51º15’59“ | 125º26’14“ | 1080 |
| Site 3 | 51º15’27“ | 125º27’09“ | 1200 |
| Site 4 | 51º16’43“ | 125º27’14“ | 1250 |
| Site 5 | 51º16’17“ | 125º27’24“ | 990 |
| Site 6 | 51º16’28“ | 125º27’32“ | 990 |
| Site 7 | 51º17’57“ | 125º28’03“ | 1475 |
| Living | 51º16’11“ | 125º26’17“ | 1320 |

*Site 2*

Site 2 is located on a colluvial slope at the base of the northeastern Franklin Glacier moraine, directly below Site 1 at approximately 1080 m asl (Figures 4.2 and 4.5). The detrial boles and branch fragments found spilled across this slope almost certainly originated from Site 1. Ten cookies were removed from boles at Site 2 and these cross-dated ($r$ = 0.492) to produce a 332-year floating chronology (Float 2, Table 4.1).

*Site 3*

Site 3 is located on a talus slope at the base of the northwestern lateral Franklin glacier moraine at approximately 1200 m asl (Figures 4.2 and 4.5). This site is characterized by large 2 to 3 m long detrital boles with diameters ranging from 60 to 120 cm. The woody detritus at the site originates from two horizontally oriented wood-rich horizons visible in the lateral moraine face at approximately 1180 m asl, about 10 m and 15 m from the upper crest of the moraine. Ten samples from Site 3 cross-date ($r$ = 0.492) to form a 332-year long floating chronology (Float 3, Table 4.1). The outer rings of sample 05D date to 550 +/- 50 $^{14}$C yrs BP and indicate that Float 3 spans the interval from AD 1370 to 1702 (Sample 05D, Tables 4.3 and 4.1).

*Site 4*

Site 4 is located on the upper slopes and gullies of the eastern Confederation Glacier lateral moraine at approximately 1250 m asl (Figures 4.2 and 4.5). Detrital boles found in the gullies are generally of a similar diameter and length to those found at sites 1, 2, and 3. Most of the boles are oriented down slope, having spilled out of the eroding gullies. Some remain partially buried in the gully walls.

Seventeen samples from Site 4 cross-date ($r$ = 0.488) to form a 308-year long floating chronology (Float 4, Table 4.1). Perimeter rings from a sample excavated from one of the gully walls date to 670 +/- 40 $^{14}$C yrs BP (Sample 07E, Table 4.3) but the sample did not cross-date to Float 4.

*Site 5*

Site 5 is located approximately 3 km downstream from the 2008 terminal position of Confederation Glacier at approximately 990 m asl (Figures 4.2 and 4.5), where several large masticated boles were discovered along the incised banks of the Chance River. Sample 01R was collected from a 5.5 m long bole that was 90 cm in diameter (Figure 4.5) and contained over 175 annual rings. Internal cross-dates of ring-widths measured on a cookie removed from the log resulted in a 175-year long floating chronology

Table 4.3: Summary of radiocarbon-dated dendroglaciological evidence recovered in the vicinity of the Franklin and Confederation glaciers confluence.

| Sample (lab ID) | Species | Site | Conventional age (no. rings dated) | $2\sigma$ Calib. (yrs BP) | Lat. (N) | Long. (W) | Alt. (m asl) | Description | Source |
|---|---|---|---|---|---|---|---|---|---|
| S-1568 | *Unknown* | 1 | 835 +/- 45 yrs BP (unknown) | — | 51°16’05” | 125°26’ 25” | 1170 | Tree root on paleosol | Ryder/ Thomson (1985) |
| 01A | *Tsuga mertensi-ana* | 1 | 600 +/- 50 yrs BP (47) | 670-520 | 51°16’07” | 125°26’20” | 1250 | Stump in growth position, W. Franklin lat. | This study |
| 07E | *Tsuga m.* | 4 | 670 +/- 40 yrs BP (47) | 680-620<br>610-560 | 51°16’25” | 125°27’09” | 1208 | Bole, east lat. Confed. moraine | This study |
| 05D | *Tsuga m.* | 3 | 550 +/- 50 yrs BP (38) | 650-510 | 51°15’43” | 125°27’17” | 1043 | Bole, base of NW Franklin lat. moraine | This study |
| 01R | *Tsuga m.* | 5 | 3480 +/- 70 yrs BP (56) | 3920-3570 | 51°16’28” | 125°27’38” | 990 | Bole, bank of C. River | This study |
| 04R | *Tsuga m.* | 5 | 4910 +/-50 yrs BP (74) | 5740-5590 | 51°16’39” | 125°27’40” | 990 | Bole, bank of C. River | This study |
| 06R | *Tsuga m.* | 6 | 3400 +/- 50 yrs BP (69) | 3820-3790<br>3770-3740<br>3730-3550<br>3520-3490 | 51°16’211” | 125°27’35” | 990 | Stump in growth position – E. river bank | This study |
| 04I | *Tsuga m.* | 7 | 3260 +/- 60 yrs BP (64) | 3630-3370 | 51°18’04” | 125°28’02” | 1475 | *In situ* bole, E Confed. lat. moraine | This study |

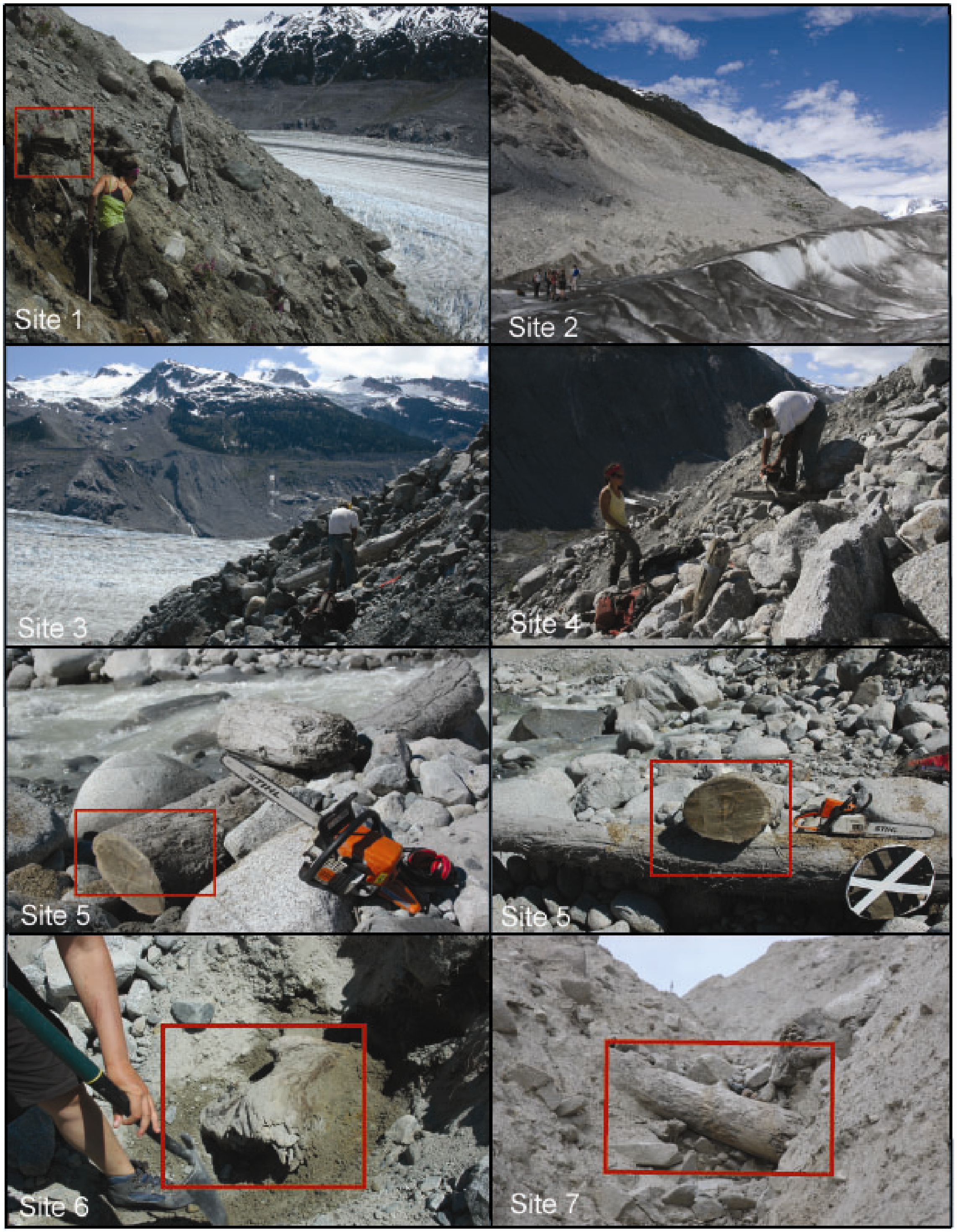


Figure 4.5: Dendroglaciological study sites and radiocarbon-dated samples in the vicinity of the Franklin and Confederation glaciers confluence. Radiocarbon-dated samples 01A (Site 1), 01R (Site 5, left), 04R (Site 5, right), 06R (Site 6), and 04I (Site 7) are identified with a red box.

($r = 0.466$) (Float 5*a*, Table 4.1). A perimeter wood sample was dated to 3480 +/- 70 $^{14}C$ yrs BP (Table 4.3).

Sample 04R was collected from a 5.46 m long bole with a diameter of 200 cm (Figure 4.5). Ring width series measured on a cookie removed from the log cross-dated to form a 174 long floating chronology ($r = 0.584$) (Float 5*b*, Table 4.1). The outermost tree-rings of sample 04R have a radiocarbon age of 4910 +/- 40 $^{14}C$ yrs BP (Sample 04R, Table 4.3).

*Site 6*

Site 6 is located 100 m upstream from the subfossil boles located at Site 5 at 990 m asl. Exposed within a 4 m section of the incised river bank is a paleosol covered by 2 to 3 m of sediment. Excavations revealed the remnants of a glacially-sheared stump rooted in growth position within the paleosol. Positioned behind a projecting boulder, a sample of wood from the stump internally cross-dated ($r = 0.588$) to form a 169-year long floating chronology (Float 6, Table 4.1). Radiocarbon dating of the outermost rings of the stump indicate the tree was killed in 3400 +/- 50 $^{14}C$ yrs BP (Sample 06R, Table 4.3).

*Site 7*

Site 7 is located approximately 1.5 km from the 2008 position of the Confederation Glacier ice front at 1775 m asl. Numerous pieces of detrital wood were discovered on the surface of a talus slope positioned below the west-facing lateral Confederation moraine (Figures 4.2 and 4.5). A 90 m traverse up steeply incised gullies above the slope led to the discovery of a 2.5 m *in situ* bole protruding across a channel (Figure 4.5). A small piece of perimeter wood contained 111 annual tree rings providing a floating chronology (*r* = 0.477) (Float 7, Table 4.1) dating to 3260 +/- 60 $^{14}$C yrs BP (Sample 041, Table 4.3).

## 4.6 Interpretation

### *4.6.1 Master Float A*

Floating chronologies 1, 2, 3, and 4 cross-date to form a 622-year long master floating chronology with a mean series intercorrelation of *r* =0.469 (Master Float A, Table 4.1). Microscopic wood identification indicates that the samples are mountain hemlock trees.

While absolute dates could not be assigned to Master Float A through cross-dating to living and floating subfossil chronologies (Smith and Laroque 1998; Larocque and Smith 2005; Parish 2006; Allen and Smith 2007; Jackson *et al*. 2008; Koehler 2009; Hart 2009), the two radiocarbon dates incorporated into the chronology indicate that it spans the interval from AD 1037 to 1658 (Table 4.1). The radiocarbon dates assigned to the chronology and to sample 07E from Site 4, and the radiocarbon date reported by Ryder and Thomson (1985), indicate that these sediments were deposited on the surface of forested lateral moraines during the early LIA.

### *4.6.2 Master Float B*

Master Float B includes site chronologies 5*a* and 6 (Master Float B, Table 4.1). When tree-ring segments containing extreme reaction wood were removed, the chronologies cross-dated to create a 163-year master floating chronology (*r* = 0.448). Two radiocarbon dates indicate that the floating chronology spans the interval between 3409 and 3571 cal. yrs BP (Table 4.1).

The rooted stump found in Confederation Valley indicates that Confederation Glacier was advancing into standing forests that were at least 170 years old ca. 3400 cal. yrs BP (Sample 06R, Table 4.3) (Figure 4.5). Supporting evidence for ice expansion at this time comes from sample 04I, collected further up valley at Site 7. Found buried *in situ* within the lateral moraine, this sample has a minimum age of 3260 +/- 60 $^{14}$C years BP and indicates that Confederation Glacier was expanding laterally as it moved down valley.

Microscopic wood identification showed that the two samples that make up Master Float B, as well as sample 06R, are from mountain hemlock trees. The chronology did not cross-date to other floating chronologies with comparable radiocarbon ages from the surrounding region (i.e. Smith and Laroque 1998; Parish 2006; Larocque and Smith 2005; Allen and Smith 2007; Jackson *et al*. 2008; Koehler 2009; Hart 2009).

### *4.6.3 Master Float C*

The radiocarbon date assigned to sample 04R (Table 4.3) from Floating Chronology 6 (Master Float C, Table 4.1) corresponds with — but may not be directly related to — a Garibaldi phase expansion of Confederation Glacier ca. 5000 years ago. Microscopic wood cell analysis identified the species of the bole as mountain hemlock. As the sample was not found in growth position or in proximity to other samples with comparable ages, no definitive explanation for its death can be offered. The sample did not cross-date to other subfossil chronologies from the surrounding region (Smith and Laroque 1998; Parish 2006; Larocque and Smith 2005; Allen and Smith 2007; Jackson *et al*. 2008; Koehler 2009; Hart 2009).

## 4.7 Discussion

### *4.7.1 Little Ice Age*

The early LIA moraine building event documented at this site is broadly synchronous with similar events recorded throughout the Coast Mountains (Smith and Desloges 2000; Larocque and Smith 2003; Laxton and Smith 2005; Allen and Smith 2007; Jackson *et al*. 2008), as well as on Vancouver Island (Smith and Laroque 1996), in Alaska (Wiles *et al*. 1999a, 1999b), and in the Canadian Rocky Mountains (Luckman 2000). This suggests that low-elevation maritime glaciers in British Columbia have responded similarly to the regional climate forcing mechanisms that have broadly influenced early LIA advances elsewhere in PNA.

### *4.7.2 Tiedemann Advance*

The radiocarbon dates assigned to the Tiedemann-aged chronologies recovered from Confederation Glacier forefield are broadly synchronous with the growing body of evidence for a regional period of ice expansion in ca. 3500 cal. yr BP. This includes evidence for an early Tiedemann advance in the Coast Mountains (Desloges and Ryder 1990; Clague and Matthews 1992, 1996; Koch *et al*. 2003b; Reyes and Clague 2004; Jackson *et al*. 2005; Lewis and Smith 2005; Allen and Smith 2007; Jackson *et al*. 2008) and in the B.C. Cariboo Mountains (Mauer *et al*. 2009). The broader regional significance of this period of ice expansion is evidenced by the documentation of contemporaneous advances in the Canadian Rocky Mountains (Gardner and Jones 1985; Luckman *et al*. 1993; Osborn *et al*. 2001; Wood and Smith 2004) and in Alaska (Porter and Denton 1967; Ellis and Calkin 1979; Barclay *et al*. 2009). The synchronicity of these events with the early Tiedemann moraine building episode documented at Confederation glacier suggests that mid Holocene glacial activity at maritime glaciers in British Columbia has been influenced by the same regional climate forcing mechanisms that govern glacier mass balance regimes throughout PNA.

#### *4.7.3 Garibaldi Phase: (6000-5000 $^{14}$C years BP)*

The radiocarbon date assigned to sample 04R falls within the Garibaldi Phase as recognized at glaciers throughout PNA (Stuiver *et al*. 1960; Lowden and Blake 1968; Miller 1969; Lowden and Blake 1973; Ryder and Thomson 1986; Allen and Smith 2007; Koch *et al*. 2007b; Osborn *et al*. 2007; Menounos *et al*. 2008) and around the world (Thompson *et al*. 2006). The growth of large mountain hemlock trees in the vicinity of the study site during this interval is consistent with a warmer, dryer late xerothermic climate (Mathewes 1985; Hebda 1995).

## 4.8 Summary

A floating chronology (Table 4.1) spanning the interval from AD 1037 to 1658 indicates that both Confederation and Franklin glaciers advanced to fill existing valleys during the early LIA. An early Tiedemann aged floating chronology (Table 4.1) spanning the period from 3409 to 3571 cal. yrs BP suggests that mature forests were growing in Confederation Valley prior to an advance of Confederation Glacier at this time. The size of the tree bole in Confederation Valley dated to ca. 4900 $^{14}$C yrs BP attests to the warm and dry late xerothermic growing conditions that preceded the Garibaldi Phase, but it is unknown whether or not the tree was killed as a result of an advance of Confederation Glacier.

These findings indicate that the mass balance regimes of low-elevation maritime and more continental glaciers in the central Coast Mountains have responded synchronously to two recognized phases of regional climate cooling over the past 3000 cal. yrs, despite the distinct maritime temperature and precipitation regimes. Although local factors such as warmer winters and lower freezing level heights may delay and/or stimulate maritime glacier responses to changing climates to a certain extent, the overall response of the study glaciers to at least two major late Holocene glacier advances is reflective of regional mass balance trends.

# Chapter 5: Conclusion

## 5.1 Summary

The dendroclimatological objectives of this research were to determine the climatic variables limiting the radial growth of *Tsuga mertensiana* and *Abies lasiocarpa* at the study site, to explore the physiological influence of these factors on the study species, and to develop both local and regional annually-resolved proxy records of recent climate changes in this region. My findings suggest that both study species are limited by previous year mean and maximum summer temperatures at the study site. The subalpine fir chronology developed from trees sampled at this site cross-dates to a network of fir chronologies collected elsewhere in the central and southern Coast Mountains. The chronologies were developed at sites that are characterized by low summer precipitation and warm temperatures from June through August. Response function analyses show that the radial growth of all of the subalpine fir chronologies examined in the study were sensitive to summer temperatures. In the case of trees sampled at the study site, this response is thought to be resultant of warm-temperature induced physiological stress that occurred as a result of the steep slope gradient and southern aspect of the study site and despite the high annual precipitation that characterizes this setting.

Local (AD 1820-2008) and regional (AD 1700-2008) subalpine fir tree ring width chronologies were constructed and employed to reconstruct previous July mean and maximum temperatures. The reconstructions explain between 13% and 36% of the variance in climate, and highlight warm and cool intervals that are comparable to those derived from other paleoenvironmental research in the region. Cyclical temperature oscillations commonly associated with the El Niño Southern Oscillation and Pacific

Decadal Oscillation are inherent in the proxy data, as well as century-scale fluctuations believed to be connected to long-term variations in solar irradiance.

The dendroglaciological objective of this study was to use radiocarbon dating and dendrochronological cross-dating techniques to reconstruct the Holocene behaviour of Confederation and Franklin glaciers. Glacially-buried and -killed detrital wood discovered near the confluence of the Confederation and Franklin glaciers allowed for the construction of three cross-dated floating chronologies that were interpreted to describe two late Holocene glacier advances. Radiocarbon dates indicate that two of the floating chronologies represent living trees that were killed during an early Tiedemann-age advance at ca. 3500 yrs BP and an early Little Ice Age advance at AD 1690. The dates ascribed to these advances of Franklin and Confederation glaciers are consistent with those recorded at glaciers throughout Pacific North America. A third floating chronology constructed from a single detrital log dates to the time of the lesser-known glacier advances during the Garibaldi Phase. No evidence was recovered, however, to indicate that this tree was killed as a result of an expansion of ice at that time. Based on the dendroglaciological evidence collected at the study site, it appears that these two low-elevation maritime glaciers in the central Coast Mountains have responded similarly to others in the region over the past 3000 years.

## 5.2 Conclusion

Little paleoenvironmental research has been undertaken on the windward slopes of the central Coast Mountains of B.C. The overarching goals of my research were to increase the spatial resolution of dendroclimatological studies in B.C. through an exploration of radial growth trends in this area and to use radiocarbon dating and dendrochronological cross-dating techniques to document the Holocene behavior of two nearby low-elevation maritime glaciers.

My dendroclimatological findings reinforce the importance of site characteristics on radial tree growth trends. In particular, these findings highlight the influence of slope gradient and aspect in alpine environments. In regions characterized by extreme topography such as the Coast Mountains, conifer trees may be more susceptible to warm-temperature induced physiological stress as a result of shallow soils and increased runoff, and despite high annual precipitation. The synchronicity of subalpine fir tree ring width reconstructions of previous mean and maximum temperatures to other paleoenvironmental indicators suggests that conifers growing on the windward side of the central Coast Mountains are broadly influenced by the same climate-forcing mechanisms that have been identified elsewhere in the region. The consistent generation of low reconstruction correlation values highlights the need for developing a more extensive network of high-elevation tree ring sites in coastal British Columbia.

My dendroglaciological findings suggest a temporal correlation between the behaviour of the study glaciers and glacial events that have been documented throughout the region. These results indicate a coherent mass balance response of both maritime and continental glaciers to the dominant climate fluctuations in the late Holocene. The low-elevation maritime glaciers that were studied do not appear to have been disproportionately affected by variations in the mean position of winter freezing level heights or warm winter temperatures during either the early Tiedemann or early Little Ice Age intervals. The absence of evidence for glacier advances at 2300 and 1500 cal. yrs BP in the central Coast Mountains may indicate that the mass of these glaciers has fluctuated very little following their extensions down valley during the Tiedemann Advances. It may be that only since the termination of the Little Ice Age in the early 20$^{th}$ Century have these glaciers downwasted and receded up valley to expose forests overrun 3000 years ago.

## 5.3 Limitations and Future Research

In terms of the dendroclimatic findings of this research, it is likely that more statistically significant tree ring based reconstructions of climate could have been created had the sample site not been located on such a steep, south-facing slope. Relationships between tree-ring width and climate may have been stronger if the sample site had been located at a higher elevation. Furthermore, particularly in the northern hemisphere where trees are most limited in either high- elevation or high latitude environments, dendroclimatic research is hindered by a lack of representative (high-elevation) climate data.

In terms of assessing differences in the timing of glacial activity between maritime and more continental glaciers, dendroglaciological investigations at a low-elevation maritime glacier with higher-resolution evidence of periods of glacial activity could be valuable. Small recessional LIA moraines would better attest to the influence of winter freezing level heights or warm winter temperatures on an annual or decadal scale.

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