

Late Holocene Glacial Activity of Bridge Glacier, British Columbia Coast Mountains

By

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B.A., McGill University, 1999

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of the Requirements for the Degree of

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Abstract

Bridge Glacier is a prominent eastward-flowing valley glacier located on the east side of the Pacific Ranges within the southern British Columbia Coast Mountains. The terminus of Bridge Glacier has retreated at rates ranging from 0 to 125 m/year over the last 50 years and currently calves into proglacial Bridge Lake. Field investigations of the recently deglaciated terrain at Bridge Glacier in 2002 and 2003 led to the discovery of detrital boles and glacially-sheared stumps. Dendroglaciological analyses of this subfossil wood allowed for the construction of five radiocarbon-controlled floating tree-ring chronologies. The relative age and stratigraphic location of these samples revealed that Bridge Glacier experienced at least four periods of significant advance during the late Holocene: a Tiedemann-aged advance at ca. 3000 ¹⁴C years BP, an unattributed advance at ca. 1900 ¹⁴C years BP, a First Millennial Advance at ca. 1500 ¹⁴C years BP, and an early Little Ice Age advance at ca. 700 ¹⁴C years BP. Lichenometric investigations at eight terminal and lateral moraine complexes led to the recognition of early Little Ice Age moraine-building events during the late 13th to early 14th centuries, with subsequent Little Ice Age episodes in the mid 15th, early 16th, mid-late 17th, early 18th, mid-late 19th, and early 20th centuries. These interpretations provide an exceptional long-term perspective on the extent and character of a glacier within this region during the late Holocene.

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Chapter One: Introduction

Glaciers have long been recognized as an exceptional source of proxy paleoclimatic information (Grove 1988). Alpine glaciers, in particular, have been shown to be especially sensitive to both air temperature and precipitation variations on annual to decadal scales (Grudd 1990; Lawby et al. 1994). Consequently, investigations designed to discern the long-term mass balance response of a glacier can reveal much about past climate conditions (Watson and Luckman 2004, 2006).

In Pacific North America, the dating of Holocene-age terminal and lateral moraine deposits provides spatial insights into the response of individual glaciers to regional-scale climate forcing events. Detailed site-specific chronologies have been developed primarily through the application of dendroglaciology and lichenometry.

The purpose of this thesis is to reconstruct the late Holocene history of Bridge Glacier in the British Columbia Coast Mountains. Although prior research at Bridge Glacier had detailed a framework for its late-Holocene behaviour, reconnaissance fieldwork carried in July 2002 at the site resulted in the discovery of freshly exposed, *in situ* and detrital subfossil wood on the glacier's forefield. Prompted by these discoveries and the results of preliminary radiocarbon dating, I returned to the area in July 2003 to complete a dendroglaciological and lichenometric assessment of the site to establish the late Holocene dynamics of Bridge Glacier.

The objectives of this project are:

- I. To establish the late Holocene glacial history of Bridge Glacier, by:
 - using dendroglaciology to identify glacier advance events, extents, and rates of advance,

- using lichenometry to reconstruct the Little Ice Age (LIA) sequence of glacial activity at Bridge Glacier, and
- using photogrammetry to determine Bridge Glacier's 20th century rates of recession.

II. To place Bridge Glacier's late Holocene activity in the framework of the existing proxy late Holocene glacier records for the cordillera of northwestern North America.

This thesis consists of four chapters and five appendices. Following this introduction, Chapter Two is a discussion of the geobotanical dating techniques applied in this project, followed by a review of the current state and findings of dendroglaciological and lichenometric investigations in the cordillera of northwestern North America. Chapter Three is a manuscript prepared for submission to a refereed journal, describing the Bridge Glacier study site, the methods applied in this research, and the findings of my investigations. Chapter Four presents a synthesis of my research activities and offers suggestions for further research. Appendix A contains the lichenometric control points and growth curve used in this study. Appendices B and C provide supplementary information on dendrochronological methods and results, and illustrations of the tree-ring chronologies developed from the Bridge Glacier site. Appendix D describes the dendroclimatological investigations attempted at the site and Appendix E presents my efforts to reconstruct a proxy record of glacier mass balance history at Bridge Glacier. Appendix F provides details of the subfossil wood samples collected at the site.

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Chapter Two: Research Background

This chapter reviews what is known about Holocene glacier fluctuations in the northwestern cordillera of North America, revealed using geobotanical dating techniques. Two main techniques have been applied within this region: dendroglaciology and lichenometry. Dendroglaciology has the potential to provide an absolute calendar date for a specific glacial event (e.g., Luckman 1985), whereas lichenometric dating provides only a relative age of the termination of glacier activity (e.g., Andrews and Webber 1969).

2.1 Dendroglaciology

Dendroglaciology involves the use of dated tree rings to reconstruct past glacial activity. Dendroglaciological research techniques have been employed to reconstruct local histories of glacier fluctuations by:

- Establishing minimum ages for glacial landforms by determining the age of the oldest tree growing on the surface of the deposits. Dating is achieved by counting the annual growth rings and adding an appropriate ecesis interval (i.e. the length of time necessary for germinants to establish on a recently formed or deglaciated surface).
- Dating scars or compression wood in a tree that has been pushed over by a glacier. Compression wood is the denser, more brittle reaction wood that forms on the lower side of a leaning coniferous tree.
- Determining the kill date of trees that were overridden by a glacier by crossdating the subfossil wood sample with a local living tree ring chronology. Crossdating is the process of matching variations in ring width or other ring characteristics among

several tree-ring series. Subfossil wood contains non-petrified cambium tissue that has been preserved due to anoxic and/or cold conditions.

- Radiocarbon (^{14}C) dating of wood samples can provide relative ages for wood samples that cannot be crossdated with absolutely dated chronologies. Unsuccessful crossdating can occur if the time since death exceeds the length of the site living chronology, or if the tree rings are affected more by the approaching glacier than by a common factor such as climate. Chronologies formed by crossdating undated tree ring series with each other are called floating chronologies.

Dendroglaciology has been applied globally for decades and has been the subject of methodological review and refinement (*c.f.* Holzhauser 1984 *in* Schweingruber 1988; Villalba et al. 1990; Bräuning; 1995; Luckman 1998*a*). Although widely accepted as an accurate and relatively inexpensive method for dating glacier advances and moraine establishment, dendroglaciology has its limitations. Because it is a branch of the larger discipline of dendrochronology, dendroglaciology is subject to the same problems inherent to all tree-ring studies. It also has challenges unique to the marriage of tree-ring analysis and interpretation of glaciological events. These challenges include: a) sampling height-age errors that arise from collecting tree-ring data from living trees and subfossil wood at positions above the ground surface (McCarthy et al. 1991), b) the difficulties encountered when a sample is missing the pith and/or perimeter wood or bark (Luckman 1986), and c) determining tree ecesis, i.e. the interval of time elapsed between ice retreat and the germination of trees upon the deglaciated surface. These challenges can lead to the underestimation of sample ages and, in the case of missing perimeter wood, an inaccurate estimation of the kill-date.

The rarity of good, intact samples for dendroglaciological analyses can be a limitation. *In situ* samples are rare and the provenance of detrital wood is often uncertain, as wood fragments can be transported by water, avalanche, mass wasting, and/or glacier activity. For samples of unknown origin, estimates of the initial source of the wood can be made based on the sample's position in moraines, on outwash plains, or within the glacier (Ryder and Thomson 1986). During transportation, the wood can become masticated and fragmented, which can remove bark and/or pith material, and compromise the sample's utility for dendrochronological analysis (Luckman 1988).

Crossdating of subfossil wood samples with both living and dead samples can be difficult even when the wood has not been reworked by an external mechanism. Trees that have been scarred or killed by a glacial advance will often experience several years of physical and/or microclimatological stress induced by the encroaching glacier (e.g., Holzhauser 1984 in Schweingruber 1988; Luckman 1996). If the effects are persistent, the resulting tree-ring pattern will be difficult to crossdate with local chronologies.

When subfossil samples cannot be crossdated to a living tree-ring chronology, a relative age can be assigned by radiocarbon dating. Radiocarbon dating is based on the rate of radioactive decay of ^{14}C in wood tissue younger than approximately 50,000 years (Arnold and Libby 1949). To compensate for long-term variations in atmospheric ^{14}C , calibration curves have been constructed (Stuiver 1982), which for some time periods result in more than one calibrated date being assigned to a single sample (Luckman 2000).

2.2 Lichenometry

Lichenometry is the study of lichen growth rates to date surface rock features. It is based on the assumption that the largest lichen growing on a substrate is the oldest individual. If the growth rate for a given species is known, the maximum lichen size will provide a minimum age for the substrate (Grove 1988). Lichen age-size relationships are determined by measuring lichen thalli sizes growing on surfaces of known age. Once a sufficient number of reliable calibrations points have been collected, a regression model is developed to describe the size-age relationship. This equation can then be applied to lichen thalli measurements collected from glacial landforms to determine an age.

Species used in lichenometry in Pacific North America include *Rhizocarpon* sp. (particularly *R. geographicum*) on calcium-poor substrates (Denton and Karlén 1977; O'Neal and Shoenenberger 2003; Larocque and Smith 2004), and *Xanthoria elegans* (Osborn and Taylor 1975; McCarthy 1993) and *Aspicilia candida* (McCarthy and Smith 1995) on calcium-rich substrates. Lichenometry has been used to determine the dates of abandonment of glacier landforms in the Washington Cascades (Porter 1981; O'Neal and Shoenenberger 2003), the British Columbia (B.C.) Coast Mountains (Smith and Desloges 2000; Lewis and Smith 2004; Larocque and Smith 2004), the Canadian Rocky Mountains (Osborn and Taylor 1975; Luckman 1977; Smith et al. 1995), and within the Alaskan cordillera (Denton and Karlén 1977; Wiles et al. 2002; Solomina and Calkin 2003).

A significant concern in lichenometry is the calibration methods used to construct lichen growth curves (Jochimsen 1973; McCarthy 1999; Karlén and Black 2002). There are several important considerations:

- Human-made structures and monuments, such as gravestones, cairns, and monuments, are often used as known-age surfaces for lichen growth curve calibration (Osborn and Taylor 1975; McCarthy and Smith 1995). Photogrammetry and historical accounts can be applied to the studied surface to obtain local calibration points (Osborn and Taylor 1975; Smith and Larocque 2004). Lichen thalli growing on moraines dated using dendroglaciological methods have also been used to calibrate growth curves, but these calibration points are then subject to the same limitations inherent to dendroglaciology. For example, curves calibrated using radiocarbon-dated surfaces carry the errors associated with radiocarbon dating (Larocque and Smith 2004; Lewis and Smith 2004). Similarly, curves calibrated using the age of trees growing on a substrate require accurate determination of both tree and lichen ecesis intervals (Luckman 1977).
- Lichen growth curves require substrates of known age. Before an age can be assigned to the largest lichen found on these dated surfaces, it is essential that the local ecesis interval for lichen establishment be determined (McCarthy and Luckman 1993). Ecesis intervals are often regionally variable and are always species-specific (McCarthy and Smith 1995). For instance, while Larocque and Smith (2004) report that *R. geographicum* spp. in the Mt Waddington area have a 9-year ecesis interval, Smith and Desloges (2000) estimate that *R. geographicum* spp. in the Bella Coola area typically have a 20-year ecesis interval.
- The growth conditions in the region from which the calibration points are collected must be representative of the conditions where the growth curve will be applied. Variations in microclimate, substrate lithology, biotic influences, and abiotic

disturbances can produce significant differences in lichen growth rates that can introduce errors in substrate age estimates (Beschel 1961; Osborn and Taylor 1975; Gellatly 1982).

- The effectiveness of regression models in explaining lichen growth has also been questioned (e.g., Boucher 1997), as lichen thallus age-size relationships are not always normally distributed (McCarthy 1999). This limitation is partially overcome by introducing a non-linear component to lichen growth curves in the initial stages of lichen thallus development (e.g., O'Neal and Schoenenberger 2003; Larocque and Smith 2004).

2.3 Late Holocene glacial fluctuations in western North America

At least three regionally identifiable glacial advances are reported to have occurred in the B.C. Coast Mountains during the late Holocene: the Tiedemann Advance from ca. 3300 to 2200 ^{14}C years BP; the First Millennium Advance (FMA) (referred to locally as the Bridge Advance) from ca. 1700 to 1350 ^{14}C years BP; and the Little Ice Age (LIA) advance from ca. 900 to 100 years A.D. Coeval advances have been reported from the Canadian Rocky Mountains, the Alaskan cordillera, and the Cascade Mountains.

2.3.1 Tiedemann Advance

Unlike most glaciers in the Coast Mountains that achieved their late Holocene maximum extent during the LIA, Tiedemann Glacier in the Mount Waddington area attained its maximum extent ca. 2940 to 2250 \pm 130 ^{14}C years BP (Fulton 1971) during an advance that began prior to 3345 \pm 115 ^{14}C years BP (Ryder and Thomson 1986). Palynological studies on the Queen Charlotte Islands (Pellat and Mathewes 1994, 1997) and in the Iskut region of British Columbia (Spooner et al. 2002) provide evidence for

substantial environmental cooling at the time of the Tiedemann advance. Corresponding glacial advances in the B.C. Coast Mountains are referred to as the Tiedemann Advance and are assumed to be coeval with the ca. 3100-2500 years BP Peyto Advance in the Canadian Rockies (Luckman et al. 1993, discussed below).

Evidence for Tiedemann-age glacial activity is widespread. In the southern Coast Mountains radiocarbon dating of subfossil wood led to the recognition of Tiedemann events in the Lillooet Icefield (Lillooet Glacier [Reyes and Clague 2004], and Bridge Glacier [this study]) and within Garibaldi Provincial Park (Decker Glacier [Koch 2006]). In the Bella Coola area, Desloges and Ryder (1990) found evidence supporting Tiedemann-age glacial advances in the Monarch Icefield area (Jacobsen Glacier).

Recent findings in the northwestern Coast Mountains support a Tiedemann Advance in the Stewart-Cassiar area (Todd Glacier [Laxton 2005], Surprise Glacier [Jackson and Smith 2005], Forrest Kerr Glacier [Lewis and Smith 2005], and Bear River Glacier [Haspel et al. 2005]). Clague and Mathews (1992) also found evidence for a Tiedemann-age advance of Frank Mackie Glacier, while Clague and Mathews (1996) reported similar evidence for nearby Berendon Glacier.

Coeval advances in western North America

Detrital and *in situ* wood from the forefields of several glaciers in the Canadian Rocky Mountains have provided evidence for a widespread advance between 3100 and 2400 ^{14}C years BP termed the Peyto Advance (Luckman et al. 1993; Osborn et al. 2001). There is convincing evidence for a major period of expansion of Peyto Glacier, where radiocarbon dated detrital wood and *in situ* samples from the upper forefield indicate that the glacier advanced between ca. 3000 and 2800 ^{14}C years BP (Lowdon et al. 1971;

Luckman et al. 1993). Evidence for correlative advances has been found at Robson Glacier (four *in situ* glacially-sheared stumps dated from 3130 ± 70 to 3360 ± 60 ^{14}C years BP; Luckman 1993), Saskatchewan Glacier (*in situ* forest of glacially-sheared stumps dated at 2910 ± 60 to 2730 ± 60 ^{14}C years BP; Wood and Smith 2004), Yoho Glacier (detrital log radiocarbon dated to 2830 ± 80 ^{14}C years BP; Luckman et al. 1996), and Stutfield Glacier (detrital wood fragment radiocarbon dated to ca. 2400 ^{14}C years BP; Osborn et al. 2001).

In the St. Elias and Wrangell Mountains of Alaska and Yukon Territory, Denton and Karlén (1977) identified a major expansion of several glaciers between 2900 and 2100 ^{14}C years BP; the maximum extent occurred ca. 2600 and 2800 ^{14}C years BP. This advance is supported by evidence from glaciers in the Alaskan Wrangell Mountains (detrital wood samples radiocarbon dated to 2700 - 2200 ^{14}C years BP; Denton and Karlén 1977), and the Kenai Mountains (glacial expansions ca. 3300 ^{14}C years BP [Wiles and Calkin 1994]).

In the Cascade Range of Washington State, the only evidence for glacial activity contemporary with the Tiedemann/Peyto Advance comes from glaciers located on Mount Rainier. Crandell and Miller (1964, 1974) report a moraine-building event ca. 3500-2000 ^{14}C years BP. Burbank (1981) also reports a moraine stabilization date of >2200 years ago at Mount Rainier (identified through tephrochronology).

2.3.2 Bridge Advance

Recent research has revealed evidence for a pre-LIA advance in the Coast Mountains at ca. 1500 ± 50 ^{14}C years BP, or ca. 500 A.D. (Allen and Smith 2003; Laxton and Smith 2004; Reyes and Clague 2004; Reyes et al. 2006). First reported by Allen and

Smith (2003), this advance is substantiated by an *in situ* whitebark pine (*Pinus albicaulis* Engelman) stump discovered in an abandoned meltwater channel in the forefield of Bridge Glacier in 2002. Radiocarbon dating identified a kill date of 1500 ± 50 ^{14}C years BP, with the position and shape of the stump strongly indicating that the tree was killed by an advance of Bridge Glacier. This advance was locally referred to as the “Bridge Advance” (Allen and Smith 2003; Reyes et al. 2004b), and has since been termed the “First Millennium A.D.” advance (FMA; Reyes et al. 2006).

Koch et al. (2004a, 2004b) and Koch (2006) reported the southernmost evidence for a Bridge-age glacial advance in Garibaldi Provincial Park in the southern Coast Mountains. A glacially-killed detrital snag at Sphinx Glacier was radiocarbon dated to 1570 ± 40 ^{14}C years BP (Koch 2006). At Lillooet Glacier, Reyes (2003) and Reyes and Clague (2004) dated wood fragments in the uppermost layer of a paleosol overlain by till: six of these detrital samples yielded radiocarbon ages from 1390 ± 50 ^{14}C years BP to 1720 ± 42 ^{14}C years BP, indicating that the paleosol was buried by a Bridge-aged advance (Reyes et al. 2006).

In the Mount Monmouth area of the central Coast Mountains, lichenometric evidence from Miserable Glacier in the Tchaikazan Valley indicates that a moraine-building event took place ca. 1269 years BP (733 A.D.; Smith 2003). In the Mount Waddington area, Larocque and Smith (2003) found lichenometric evidence for a Bridge Advance at Tiedemann Glacier dated to 1380 years BP (620 A.D.) that supports previous reports of logs buried by an advance at this time (1330 ± 65 ^{14}C ; Ryder and Thomson 1986).

Several sites in the vicinity of Stewart, B.C. in the northern Coast Mountains provide additional evidence for a Bridge-age advance. At Todd Glacier, two logs buried by an advance date to 1540 and 1690 ± 60 ^{14}C years BP (Laxton and Smith 2004; Laxton 2005). At Surprise Glacier, numerous glacially-transported boles dating to ca. 1600 ^{14}C years BP were found buried within a prominent lateral moraine (Jackson and Smith 2005). At Berendon Glacier, Clague and Mathewes (1996) found subfossil samples buried beneath mud layers deposited during a glacier advance ca. 1500 ^{14}C years BP. Nearby, Clague and Mathewes (1992) report that Frank Mackie Glacier advanced and impounded a glacial lake 1600 ± 40 ^{14}C years BP. Further to the north in the vicinity of the Iskut River, Lewis and Smith (2005) found till-covered Bridge-age wood samples at Forrest Kerr Glacier that date to ca. 1500 ^{14}C years BP.

Coeval advances in western North America

Evidence for an FMA advance exists in the Canadian Rocky Mountains. At Peyto Glacier a stump found buried within a lateral moraine indicates that the glacier was advancing just prior to 1550 ± 60 ^{14}C years BP (Luckman et al. 1993). Subfossil logs found in outwash sediments at Robson (1140 ± 80 ^{14}C years BP) and Saskatchewan glaciers (1590 ± 70 ^{14}C years BP) suggest both glaciers were also advancing at this time (Luckman et al. 1993). Detrital wood flushed from the snout of Cavell Glacier dates to 1600-1900 ^{14}C years BP, suggesting that the glacier was advancing into a standing forest at this time (Luckman et al. 1996)

In the Kenai Mountains of Alaska, Wiles and Calkin (1994) describe an advance at ca. 1400 ^{14}C years BP that occurred simultaneously in land-terminating and calving tidewater glaciers. At Bering Glacier in the Chugach Range of southern Alaska, Wiles et

al. (1999*b*) found dendroglaciological evidence for an advance at ca. 1500 ¹⁴C years BP. In the St. Elias and Wrangell Mountains of Alaska and Yukon Territory, Denton and Karlén (1977) identified a short period of glacier expansion between 1230 and 1050 ¹⁴C years BP. Fjord glaciers in Icy Bay south of the St. Elias Mountains also experienced an advance 1200 ± 160 ¹⁴C years BP (Porter 1989), at about the same time as Tebenkoff and other nearby glaciers were advancing (Barclay et al. 1999; Wiles et al. 1999*a*).

2.3.3 Little Ice Age (LIA) Advances

In Garibaldi Provincial Park in the southern Coast Mountains, Koch et al. (2003, 2004*a*, 2004*b*) report dendroglaciological and lichenometric evidence for LIA initiation around 1050 A.D. LIA glacier activity in this area peaked in the early 18th century with moraine-building events culminating in 1725 A.D. at Lava Glacier (Mathews 1951) and 1735 A.D. at Overlord Glacier (Koch et al. 2003). Whereas Overlord Glacier experienced a subsequent moraine-building event in 1835 A.D. (Koch et al. 2003), dendrochronological evidence at Helm Glacier indicates that it was still advancing at this time (Mathews 1951). Koch et al. (2003) dated additional moraines at Overlord Glacier to 1890 and 1920 A.D.

On Vancouver Island, Smith and Laroque (1996) identified a post 1718 A.D. advance at Moving Glacier and noted that the outermost moraine at this site stabilized in ca. 1818 A.D. Lewis and Smith (2004) employed lichenometry and dendrochronology to assign dates to moraines at Colonel Foster and Septimus glaciers. Their findings indicate that LIA moraines were deposited prior to 1397 A.D., in the late 17th/early 18th centuries, in the mid-late 19th century, and in the early-mid 20th century.

In the Silverthrone Icefield area of the central Coast Mountains, Ryder and Thomson (1986) discovered *in situ* glacially-killed wood at Klinaklini Glacier that indicates the glacier was advancing during the late 11th/early 12th century (900 ± 40 ¹⁴C years BP) until at least the late 16th century (400 ± 45 ¹⁴C years BP). Similar evidence recovered within the Mount Waddington area shows that Franklin Glacier was advancing downvalley during the mid 12th century (835 ± 45 ¹⁴C years BP; *ibid.*). Lichenometric investigations at glaciers east of Mount Waddington suggest the LIA was initiated in the early 13th century, followed by at least 10 distinct moraine-building events spanning the mid-15th to mid-20th centuries (Larocque and Smith 2003). Further to the north in the Bella Coola region, Desloges and Ryder (1990) used dendrochronology to show that LIA glacier activity commenced prior to the 13th century and culminated in the late 19th century.

At Berendon and Frank Mackie Glaciers in the northern Coast Mountains, Clague and Mathewes (1996) and Clague and Mathewes (1992) report that the LIA initiated prior to 500 ¹⁴C years BP, and culminated in the early 17th century. Subsequent readvances may have occurred in the 18th or early 19th centuries. In the Stewart area, dendroglaciological evidence revealed that Surprise Glacier has a complex LIA history. Evidence for at least four intervals of glacier expansion and retreat dating to the 14th, 17th, 18th, and 19th centuries were inferred from a composite lateral moraine at this site (Jackson and Smith 2005). Investigations at nearby Todd and Bear glaciers reveal a period of significant glacier expansion in the late 13th century (660 ¹⁴C years BP), followed by advances in the 14th, 16th, 17th and 19th centuries (Haspel et al. 2005; Laxton 2005). Further to the north in the Andrei Icefield area, Lewis and Smith (2005) suggest,

on the basis of lichenometric evidence, that the LIA spans the period between the early 13th and late 19th centuries.

Coeval advances in western North America

In the Canadian Rocky Mountains, early-LIA glacier advances are reported to have occurred in the 12th and 13th centuries at Peyto Glacier (Luckman 1993, 1996) and Robson Glacier (Luckman 1993, 1995). This period of glacier expansion persisted until the 14th century at Robson Glacier (Luckman 1995). Mid-LIA glacier advances occurred in the 15th (Manitoba Glacier; Robinson 1998) and 16th centuries (Kananaskis region [Smith et al. 1995] and Turret Glacier [Luckman and Osborn 1979]). Several glaciers in the Canadian Rocky Mountains achieved their LIA maxima in the early 18th century (South Cirque Glacier [Bednarski 1979], Yoho Glacier [Bray and Struik 1963], Maligne Icefield [Kearney 1981], and Angel Glacier [Luckman 1977]). At other sites, LIA advances are reported to have occurred in the late 18th and 19th centuries (at Robson Glacier [Heusser 1956; Luckman 1995], Dome Glacier [Luckman 1998b], glaciers in Peter Lougheed and Elk Lakes Provincial Parks [Smith et al. 1995], Stutfield Glacier [Osborn et al. 2001], Manitoba Glacier [Robinson 1998], and the Premier Range [Watson 1986]). Some glaciers readvanced in the mid 20th century (Premier Range; Luckman et al. 1987).

The LIA in Alaska began as early as the 10th to 12th century A.D. (Kenai Mountains [Wiles and Calkin 1994] and Prince William Sound [Wiles et al. 1999a]). Thirteenth-century LIA activity is widely reported (Brooks Range [Evison et al. 1996], Wrangell Mountains [Wiles et al. 2002], western Prince William Sound [Wiles et al. 1999a], Bering Glacier [Wiles et al. 1999b], and southern Kenai Mountains [Wiles and

Calkin 1994]). Little Ice Age glacial maxima at many sites occurred during the 15th and 16th centuries (Wrangell Mountains [Denton and Karlén 1977], Brooks Range [Evison et al. 1996], and Kenai Mountains [Wiles and Calkin 1994]). Seventeenth to 18th century LIA advances are evident in several areas (Kigluaik Mountains [Calkin et al. 1998], Wrangell Mountains [Wiles et al. 2002], Prince William Sound [Wiles et al. 1999a]), and Kenai Mountains [Wiles and Calkin 1994]) followed by mid-late 19th century LIA advances (*ibid.*, Donjek Glacier [Reyes et al. 2004a], and Lowell Glacier [Clague et al. 1982; Reyes and Smith 2001]).

The earliest LIA glacial activity reported for the Cascade Mountains occurred during the 13th and early 14th centuries (Dome Peak [Miller 1969], North Nowich Glacier [Burbank 1981], and Cowlitz Glacier [Crandell and Miller 1964]). Glaciers in the Dome Peak area reached their LIA maxima in the 16th century (Miller 1969). Sites at Mount Rainier have moraines that date to the early-mid 16th century (Sigafos and Hendricks 1972; Burbank 1981), the early-mid 17th century (Sigafos and Hendricks 1972), and the late 17th century (Burbank 1981). In the Mount Baker area there is evidence for mid 16th century (Boulder and Deming Glaciers; Easterbrook and Burke 1971, 1972) and early 17th century moraine stabilization (Deming Glacier [Easterbrook and Burke 1971; Fuller 1980]). Eighteenth century events are evident at Mount Rainier (Crandell and Miller 1964; Sigafos and Hendricks 1972). Late LIA activity has been reported at nearly all reviewed sites in the Cascades in the 19th century (Nisqually Glacier [Sigafos and Hendricks 1961], Van Trump Glacier [Crandell and Miller 1964], Mount Rainier [Burbank 1981], Dome Peak area [Miller 1969], and Mount Baker [Easterbrook and Burke 1971, 1972; Fuller 1980; Heikkinen 1984]), and at several sites in the 20th century

(Dome Peak area [Miller 1969], Mount Baker area [Easterbrook and Burke 1971, 1972; Fuller 1980; Heikkinen 1984]).

2.4 Summary

Dendroglaciology and lichenometry are geobotanical dating techniques that are commonly used to provide the dates for the stabilization of glacial landforms, and to assist in the reconstruction of the advance/retreat histories of individual glaciers.

Application of these methods at glacier sites throughout the northwestern North American cordillera has led to the description of three regionally synchronous late Holocene glacial advances:

1. The Tiedemann Advance (termed "Peyto Advance" in the Canadian Rocky Mountains), ca. 3000 ^{14}C years BP, for which there is evidence in the B.C. Coast Mountains, the Canadian Rocky Mountains, the Alaskan cordillera, and the Washington Cascades.
2. The First Millennial Advance (termed "Bridge Advance" in the southern B.C. Coast Mountains), ca. 1500 ^{14}C years BP, for which there is evidence in the B.C. Coast Mountains, the Canadian Rocky Mountains, and the Alaskan cordillera.
3. The Little Ice Age, ca. 10th-12th centuries to the 20th century, for which there is evidence in the B.C. Coast Mountains, the Canadian Rocky Mountains, the Alaskan cordillera, and the Washington Cascades.

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Chapter Three: Late Holocene glacial activity at Bridge Glacier, Coast Mountains, British Columbia.

3.1 Introduction

The glaciological response of temperate alpine glaciers to year-to-year climate fluctuations is often marked by rapid changes in glacier mass balance state (Ostrem and Brugman 1991). Over the historical period glaciers in the British Columbia (B.C.) Coast Mountains have experienced recurring decades of negative annual mass balance (Mathews 1951; Ryder and Thomson 1986; Hodge et al. 1998; Larocque and Smith 2003) in response to the effects of climate variability and change (Moore and Demuth 2001). The resulting glaciological response has been a reduction in glacier area, downwasting of glacier surfaces, and accelerated frontal retreat (Mathews 1951; Osborn and Luckman 1988). In the mountains of coastal B.C., this recession and downwasting is exposing land surfaces and features buried by glacial advances that in many locations predate the recent Little Ice Age (LIA) glacial episode (Smith and Laroque 1996; Smith and Desloges 2000).

This paper presents the results of field investigations conducted at Bridge Glacier in the B.C. Coast Mountains. Although prior research by Ryder and Thomson (1986) and Ryder (1991) detailed a chronological framework for the late Holocene behaviour of Bridge Glacier, reconnaissance fieldwork carried in July 2002 within recently deglaciated terrain resulted in the discovery of freshly exposed *in situ* and detrital subfossil wood. Prompted by these discoveries and the results of preliminary radiocarbon dating, I returned to the site in July 2003 to complete a dendroglaciological and lichenometric study of the late-Holocene dynamics of Bridge Glacier.

3.2 The Holocene behaviour of glaciers in the B.C. Coast Mountains

The Holocene Epoch in the B.C. Coast Mountains is characterized by at least five episodes of glacier advance that include: an early Holocene event (ca. 8000 ^{14}C years BP); the Garibaldi Phase (ca. 6000 ^{14}C years BP); the Tiedemann Advance (ca. 3000 ^{14}C years BP); a First Millennium A.D. advance (ca. 1500 ^{14}C years BP); and the Little Ice Age advances (ca. 800 ^{14}C years BP).

Within the B.C. Coast Mountains the earliest record of Holocene glacier activity comes from lake sediments and detrital wood samples collected within two contemporary glacier forefields in Garibaldi Provincial Park (Menounos et al. 2004). Radiocarbon dated to the interval between 7720 to 7380 ^{14}C years BP (8630 to 8020 cal years BP), this early Holocene advance occurred during a time when corollary paleoenvironmental data describe a climate that was drier and warmer than present in southern and coastal B.C. (Clague and Mathewes 1989; Hebda 1995; Pellatt et al. 1998; Pisaric et al. 2003; Walker and Pellatt 2003). Supporting evidence for comparable advances of alpine glaciers in the cordillera of Pacific North America (PNA) comes from the southern Canadian Rocky Mountains (Luckman 1988), the Washington Cascade Mountains (Heine 1998; Thomas et al. 2000), and the northern B.C. Coast Mountains (Smith et al. 2005).

A mid-Holocene B.C. Coast Mountain glacial event was first identified in Garibaldi Provincial Park, where glacially-killed *in situ* subfossil stumps were radiocarbon dated to ca. 5300 to 6000 ^{14}C years BP (Stuiver et al. 1960; Lowdon and Blake 1968, 1975). Presently recognized as a distinct period of glacier expansion between ca. 6000 to 5000 ^{14}C years BP (Souch 1994; Ryder and Thomson 1986; Osborn and Luckman 1988; Cashman et al. 2002; Davis et al. 2006), this event is referred to as

the 'Garibaldi Phase' due to the lack of evidence for any ensuing retreat of these glaciers to upvalley pre-advance positions until the 20th century (Ryder and Thomson 1986).

Recent studies in Garibaldi Provincial Park have sought to broaden the Garibaldi Phase to include evidence for glacier advances from 7300 to 5800 ¹⁴C years BP (Koch et al. 2004*a*, 2004*b*; Koch and Clague 2005; Koch 2006). Nonetheless, there is mounting evidence from ongoing investigations in the Coast Mountains at Tchaikazan (Smith 2003*b*), Fyles (Laxton et al. 2003), and Icemaker (Smith et al. 2006) glaciers that points to a distinct regional episode of frontal advance into mature forest stands from 6000 to 5000 ¹⁴C years BP. Paleoenvironmental data from this time indicate a climate characterized by increased precipitation and cooler summer temperatures (Spooner et al. 2002; Walker and Pellatt 2003), conditions recognized as conducive to positive glacier mass balance states and glacier advances (Bitz and Battisti 1999).

Glacier advances comparable in age to those associated with the Garibaldi Phase have been reported from several locations in PNA. In Alaska, glaciers in the Brooks and Alaska Ranges were advancing ca. 5800 years BP (revealed through lichenometry; Calkin 1988), and Hubbard Glacier (Seward Peninsula, Alaska) expanded ca. 6850 cal years BP (Calkin et al. 2001). In the Cascade Mountains of Washington state, South Cascade Glacier (Dome Peak) experienced an advance at ca. 4900 ¹⁴C years BP (Miller 1969). No evidence currently exists for coeval advances in the Canadian Rockies (Luckman 1998*b*, Smith 2003*a*).

The Tiedemann Advance ca. 3000 ¹⁴C years BP is recognized at glacier sites throughout the B.C. Coast Mountains (Fulton 1971; Ryder and Thomson 1986). Evidence supporting this event has been reported from sites in the southern B.C. Coast Mountains

(Lillooet Glacier [Reyes and Clague 2004]; Decker Glacier [Koch et al. 2003]), the mid B.C. Coast Mountains (Jacobsen Glacier [Desloges and Ryder 1990]), and the northern B.C. Coast Mountains (Frank Mackie Glacier [Clague and Mathews 1992]; Berendon Glacier [Clague and Mathews 1996]; Bear River Glacier [Haspel et al. 2005]; Surprise Glacier [Jackson and Smith 2005]; Todd Glacier [Laxton 2005]; and Forrest Kerr Glacier [Lewis and Smith 2005]). Palynological studies on the Queen Charlotte Islands (Pellat and Mathews 1994, 1997) and in the Iskut region of northwestern B.C. (Spooner et al. 2002) provide evidence for substantial cooling at the time of the Tiedemann advance.

The Peyto Advance in the Canadian Rocky Mountains is coeval with the Tiedemann Advance (Luckman et al. 1993; Osborn et al. 2001; Wood and Smith 2004). Glacier advances dating to this time have also been reported from Alaska (Denton and Karlén 1977; Wiles and Calkin 1994; Wiles et al. 1995) and from the Cascade Mountains of Washington state (Crandell and Miller 1964, 1974; Burbank 1981).

Research at sites throughout the cordillera of PNA has revealed evidence for a pre-LIA glacial advance culminating around ca. 1500 ¹⁴C years BP (Reyes et al. 2006). Referred to regionally as the 'First Millennium A.D. Advance' (FMA, Reyes et al. 2006), the locally equivalent 'Bridge Advance' is supported by dendroglaciological evidence found at Bridge Glacier (Allen and Smith 2003), Lillooet Glacier (Reyes and Clague 2004), and at several sites in Garibaldi Provincial Park (Koch et al. 2004a, 2004b). In the central B.C. Coast Mountains, lichenometric studies have also revealed an episode of moraine-building activity ca. 1300 years BP at both Tiedemann (Larocque and Smith 2004) and Miserable (Smith 2003a) glaciers. In the northern B.C. Coast Mountains, ongoing studies are providing further evidence of Bridge-age glacial advances at Surprise

(Jackson and Smith 2005), Todd (Laxton and Smith 2004), Forrest Kerr, and Spahler glaciers (Lewis and Smith 2005). Previously published data describing a glacier advance at this time from Frank Mackie (Clague and Mathews 1992) and Berendon glaciers (Clague and Mathews 1996) in the same area are now also attributed to this event. Corollary evidence for a FMA advance has been reported from sites in both Alaska (Barclay et al. 1999; Wiles et al. 1999*a*, 1999*b*) and the southern Canadian Rocky Mountains (Luckman et al. 1993; Luckman 1996). Palynological records from southern B.C. Coast Mountains indicate that the ca. 1500 years BP time period is marked by a transition from wetter, cooler conditions to a climate regime similar to that existing today (Pellatt et al. 1998; Heinrichs et al. 2002). Evidence from lake sediments in southern B.C. (Lowe et al. 1997; Hallett et al. 2003) indicates drier conditions during the FMA with a subsequent transition to conditions similar to today.

The LIA was characterized by some of the most extensive glacial advances of the Holocene. Widely reported from sites in all regions of the mountainous PNA, the LIA is characterized by numerous advance and retreat episodes (Luckman 2000). LIA glacier advances have been reported from sites in Alaska (Wiles and Calkin 1994; Wiles et al. 1999*a*; Calkin et al. 2001), the southern Canadian Rocky Mountains (Smith et al. 1995; Luckman 2000), and the Washington Cascades (Sigafos and Hendricks 1972; Burbank 1981). Data from sites in the southern B.C. Coast Mountains show that LIA glacier advances were underway by the 11th to 12th centuries (Ryder and Thomson 1986; Koch et al. 2004*a*). Within the central B.C. Coast Mountains, equivalent LIA glacier advances appear to have initiated somewhat later, in the 12th to 13th centuries (Ryder and Thomson 1986; Desloges and Ryder 1990; Larocque and Smith 2003). In the northern B.C. Coast

Mountains coeval glacial advances appear to have culminated in the 13th to 14th centuries (Jackson and Smith 2005; Laxton 2005; Lewis and Smith 2005).

Following these early-LIA glacier advances, late-LIA moraine stabilization occurred in the 14th (Lewis and Smith 2004), 15th (Larocque and Smith 2004), 16th (*ibid.*), 17th (Clague and Mathews 1992; Clague and Mathewes 1996; Smith and Laroque 1996; Smith and Desloges 2000), 18th (Mathews 1951), and 19th (Desloges and Ryder 1990; Lewis and Smith 2004; Haspel et al. 2005) centuries, with an average periodicity of 65 years documented between advances in the Mount Waddington area (Larocque and Smith 2003). Proxy paleoclimate data from this time period indicate a probable relationship to variable air temperatures and precipitation totals (Barclay et al. 1999; Luckman 2000; Larocque and Smith 2004), as well as to variations in solar insolation (Larocque and Smith 2003, 2004).

3.3 Study area

Bridge Glacier is an outlet glacier of the Lillooet Icefield located in the Pacific Ranges of the southern B.C. Coast Mountains approximately 85 km northwest of Pemberton, B.C. (50° 49' 30" N, 123° 35' 30" W; Fig. 3.1). The mountain peaks surrounding Bridge Glacier range in elevation from 2000 to 2800 m above sea level (asl). They are composed of a northwest-to-southeast oriented belt of late Palaeozoic to Tertiary metamorphic and volcanic bedrock substantially modified by Tertiary erosion (Garver and Brandon 1994) and Pleistocene glaciation (Clague et al. 1989). Following the end of the Late Wisconsinan Fraser Glaciation and retreat of the Cordilleran Ice Sheet, most valleys and cirques within this region remained ice-free through the early Holocene (Ryder et al. 1991).

The climate of the study area is characterized by long, cold winters with persistent snowpacks and short, cool summers (Coupé et al. 1991). The average mean annual air temperature for the area is 6.7 °C, with a mean annual total precipitation of 898 mm (Canadian Institute for Climate Studies Gridded Climate Normals 1961-1990; Hopkinson 2000).

Glacier mass balance surveys were conducted at Bridge Glacier and at nearby Sykora and Zavisha glaciers from 1976 to 1984 (Mokievsky-Zubok 1985). The glaciers were monitored to aid with the management of the BC Hydro Downton Reservoir (Ommanney 2002) located 25 km downstream of the snout of Bridge Glacier.

The region surrounding Bridge Glacier is located within the ATunp (Alpine Tundra undefined parkland) and ESSFdv (Engelmann spruce/subalpine fir very cold dry) Biogeoclimatic Ecological Classification zones (Meidinger and Pojar 1991). The forest cover in the area is composed largely of stands of subalpine fir trees (*Abies lasiocarpa* [Hook.] Nutt), with whitebark pine (*Pinus albicaulis* Dougl. ex Hook) trees colonizing drier and disturbed sites (*ibid.*). At the Bridge Glacier study site, the forest consists mostly of whitebark pine on moraine crests and valley flanks, with subalpine fir established on the valley slopes up to treeline at 1600 m asl. Recently deglaciated surfaces in the Bridge Glacier valley predominantly host whitebark pine seedlings, except at higher, more exposed sites where whitebark pine and subalpine fir seedlings coexist.

3.3.1. Bridge Glacier study site

Bridge Glacier is a World Glacier Monitoring Service class 414 glacier (i.e. an outlet glacier with compound basins and a calving terminus; Haeberli et al. 1998). In 2002, the glacier was approximately 18 km in length from headwall to toe, and

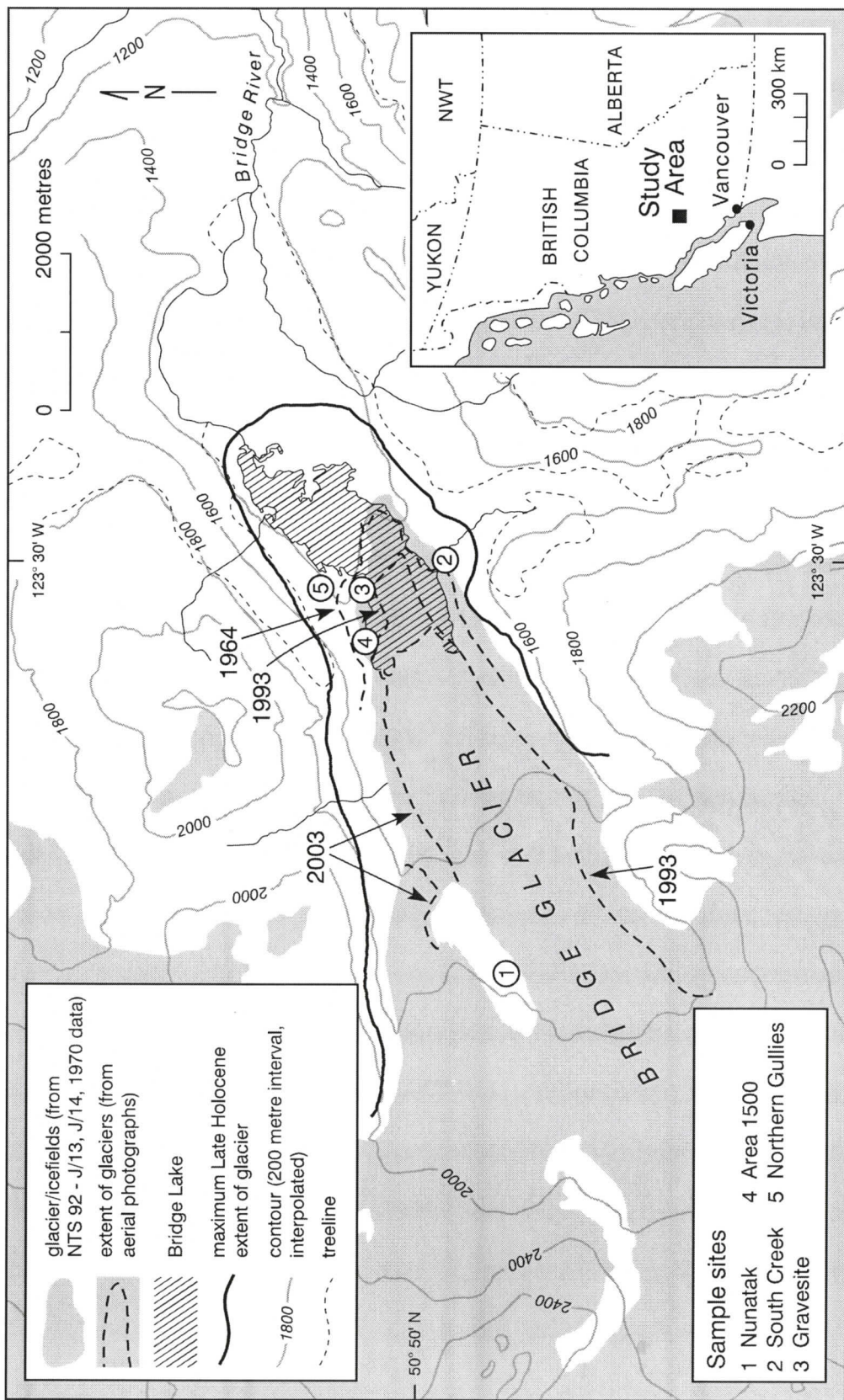


Fig. 3.1. Bridge Glacier site, located in the southern B.C. Coast Mountains, with locations of dendroglaciological sampling sites. Figure is based on 1970 National Topographic Series information; the 1964, 1993, and 2003 ice front positions are indicated by hatched lines.

covered an area of 88 km² (Dyurgerov 2002). Lateral moraines approximately 180 m in height flank Bridge Glacier and proglacial Bridge Lake (*unofficial name*). The lake presently abuts the glacier terminus at 1370 m asl and submerges much of the recently exposed glacial forefield. Between the north shore of Bridge Lake and the base of the north lateral moraine complex lies a former outwash surface incised by abandoned meltwater channels (Area 1500; Fig. 3.1).

Historical vertical aerial photographs show that prior to 1975 Bridge Glacier was retreating at a rate of 6 m/year; since 1979 the terminus has been receding upvalley at an average rate of 41 m/year (Table 3.2). Although accelerated recession over the last three decades may be related to climate variations that have seen glacier fronts receding rapidly throughout the Coast Mountains (Wheate et al. 2005), the bathymetric characteristic of Bridge Lake likely aids this rapid recession. Prior to 1975 the terminus of Bridge Glacier was grounded in the shallow waters of eastern Bridge Lake. At the present time, the snout of Bridge Glacier is floating in the deep waters of western Bridge Lake, which contributes to accelerated calving and ablation (Fig. 3.2).

An extensive terminal moraine complex is located 3.3 km downvalley of the 2003 terminus position, on the east side of Bridge Lake (Fig. 3.3). At least eight prominent nested recessional moraines are distinguishable, with several more ice-proximal moraine crests seen protruding above the lake surface. The outermost terminal moraine is colonized by mature whitebark pine and subalpine fir trees, with immature whitebark pine and subalpine fir seedlings and saplings populating the more proximal recessional moraines. Thickets of mountain alder (*Alnus incana* ssp. *tenuifolia*), trembling aspen



Fig. 3.2. Snout of calving icefront of Bridge Glacier in 2003 showing icebergs floating in Bridge Lake. Discharging into Bridge Lake in the foreground is South Creek, which drains across the floor of the former ice-dammed lake described by Ryder (1991). Photograph taken from summit of South Creek lateral moraines.

(*Populus tremuloides*), and willow (*Salix* sp.) shrubs have colonized the wetter inter-moraine troughs.

The prominent lateral moraine fringing the southern wall of the Bridge Glacier valley is a composite feature built from at least five nested moraines. The outermost moraine is colonized by mature whitebark pine, while the more proximal moraine crests host whitebark pine saplings and occasionally aspen, scrub birch (*Betula glandulosa*), and willow shrubs.

During the LIA the lateral moraine constructed by Bridge Glacier along its southern flank impounded a lake in the tributary valley of South Creek. This ice-marginal lake persisted for at least 530 years before four jökulhlaup events in the interval 1935 -

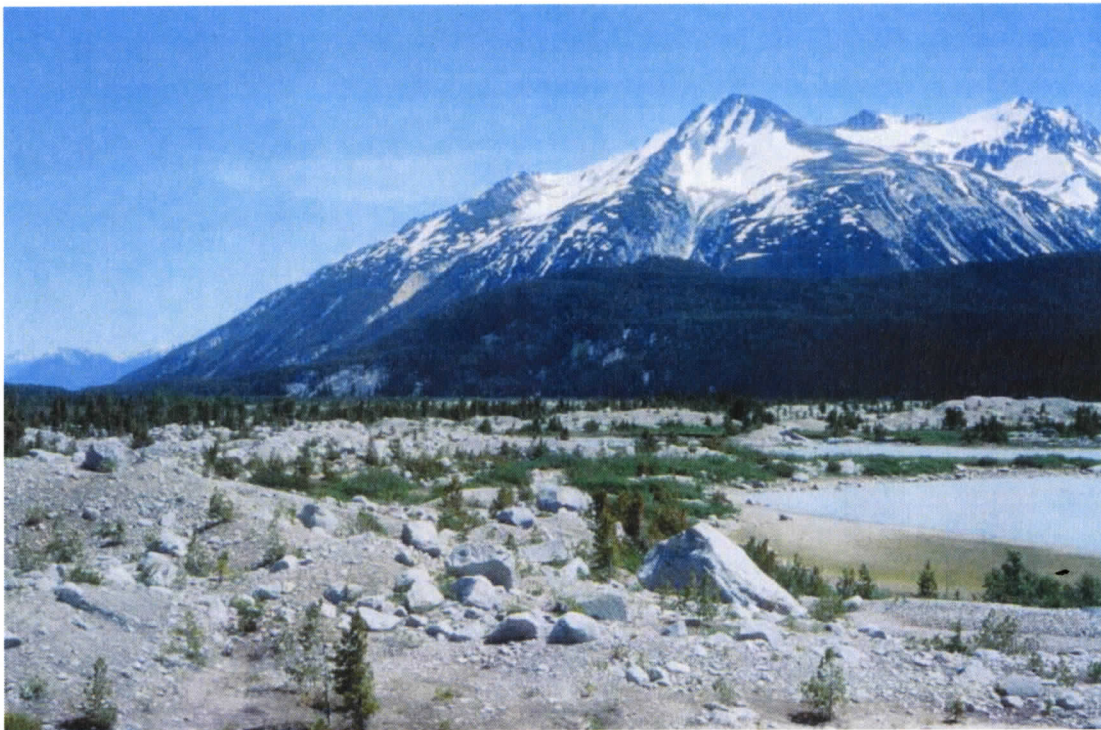


Fig. 3.3. Terminal moraines at Bridge Glacier, with Bridge Lake at right. Bridge River, seen at middle right, flows east and separates the terminal moraines into north (foreground) and south (background) sections. The glacier retreated to the west (right).

1970 drained the lake (Ryder 1991). Relic strandlines are evident around the perimeter of the former lake basin and a stream (informally called “South Creek”) currently runs through the incised moraine to drain into Bridge Lake.

Steep collapsing lateral moraines, colonized by mountain alder, delineate the northern flank of the valley walls above Bridge Glacier. The outermost northern lateral moraine hosts stands of mature whitebark pine. Water draining from the mountain slopes above has eroded the north lateral moraine complex in several places, forming a series of deeply incised gullies within the proximal moraine face. Water from these gullies flows onto the valley floor below and flows into Bridge Lake. Alder thickets colonize the banks of these streams.

A prominent arcuate-shaped mound with subdued topography (maximum height 5 m) is found on the northern shore of Bridge Lake 500 m downvalley from the 2003 terminus position. The surface of this mound is overlain by over 15 annual recessional moraines. The most ice-proximal of these moraines formed in 1980. An examination of historical aerial photographs reveals that Bridge Glacier receded westward across the mound between 1964 and 1975 at an average rate of 21 m/year (Tables 3.1 and 3.2).

Table 3.1. Air photographs used for photogrammetric analysis at Bridge Glacier.

Year	Air photo identifier	Area of interest
1964	BC4245-034 to 036, -038; BC4244-183	Snout, N. lateral gullies, Nunatak, South Lake
1975	BC7788-012, -149, -151, -153	Snout, South Lake, Gemini Pass, Nunatak
1979	30BC79100-070, -072, -074	Snout, Nunatak
1981	30BC81085-002, -004, -006, -038	Snout, Nunatak, South Lake
1987	30BCC676-022, -043	Snout, South Lake
1993	30BCC93086-032 to -038, -062 to -068	Snout, Nunatak, entire glacier
1997	30BCC97087-020	Snout
2005	30BCC05007-080	Snout

Table 3.2. Calculated rates of ice movement for Bridge Glacier.

Advance/Retreat	Rate (m/yr)	Calendar dates (A.D.) ^a	Area	Evidence
Advance	1.8	540-968	Snout, Area 1500 to Gravesite	¹⁴ C-dated <i>in situ</i> samples BG02-10 and BG03-90
Advance	1.3	968-1350	Snout, lateral, Gravesite to northern gullies	¹⁴ C-dated <i>in situ</i> samples BG03-90 and BG03-09
Retreat	6	1964-1975	Snout, Bridge Lake	Photogrammetry
Retreat	21	1964-1975	Lateral, Gravesite area recessional moraines	Photogrammetry
Retreat	0	1975-1979	Snout, Bridge Lake	Photogrammetry
Retreat	125	1979-1981	Snout, Bridge Lake	Photogrammetry
Retreat	29	1981-1987	Snout, Bridge Lake	Photogrammetry
Retreat	33	1987-1993	Snout, Bridge Lake	Photogrammetry
Retreat	38	1993-2005	Snout, Bridge Lake	Photogrammetry

^a Calendar dates for radiocarbon dated samples represented by median of 2 σ calibration dates.

A bedrock outcrop covering an area of 0.75 km² currently divides Bridge Glacier into two tongues (Fig. 3.1). Ice-covered at the maximum of the LIA, the outcrop emerged as a nunatak prior to 1964 (BC4245-038; Table 3.1) and continued to be surrounded by ice until at least 1993 (30BCC93086 no. 036; Table 3.1). The nunatak remains unvegetated with the exception of a few whitebark pine and subalpine fir seedlings newly established on its summit. The southern flank of the nunatak is mantled by a thick blanket of till incised by a series of deeply etched gully networks that expose a well-developed Orthic Humo-Ferric paleosol (Ryder and Thomson 1986).

3.4 Methods

Dendroglaciological and lichenometric dating methodologies were employed to establish the late Holocene history of Bridge Glacier. Dendroglaciology has the potential to provide an absolute calendar date for a specific glacial event (e.g., Luckman 1985), whereas lichenometric dating provides only a relative age of the termination of glacier activity (e.g., Andrews and Webber 1969).

3.4.1 Dendroglaciology

Dendroglaciology involves the use of dated tree rings to reconstruct the history of glacier fluctuations and glacial landforms. Dendroglaciological methods have been employed in PNA since the early 20th century (Lawrence 1950). Application of dendroglaciology in the B.C. Coast Mountains (e.g., Smith and Laroque 1996; Smith and Desloges 2000; Laroque and Smith 2003; Koch 2006; Reyes et al. 2006), the Canadian Rocky Mountains (e.g., Luckman 1988, 1995; Smith et al. 1995; Wood and Smith 2004) and Alaska (e.g., Barclay et al. 1999; Calkin et al. 2001; Wiles et al. 1999a, 1999b, 2002) has led to detailed descriptions of mid- to late-Holocene glacier activity.

The principles and limitations of dendroglaciology have been discussed elsewhere (*c.f.* Ryder and Thomson 1986; Luckman 1998*a*). Successful dendroglaciology is achieved through a combination of tree-ring analytical techniques that allow for the establishment of minimum ages for glacial landforms, and/or the dating of specific glacial events through determination of dates associated of tree death resulting from glacier advance. This latter application involves crossdating (i.e. matching variations in tree-ring width or other ring characteristics among several tree-ring series) subfossil tree ring samples with a local tree-ring chronology developed from living trees. Radiocarbon dating can provide a relative age for a subfossil sample in instances where the time since death of the subfossil sample exceeds the length of the living chronology, or where crossdating is prevented by significant inter-sample differences in the growth pattern of tree-rings (e.g., Wood and Smith 2004). In cases where a tree survives the physical stress of being pushed by a glacier, the date of the advance can be determined by establishing the age of the resulting scars or compression wood in the bole (e.g., Luckman 1988).

i. Wood sampling and treatment

Wood samples were collected from the Bridge Glacier study site in July 2002 and July 2003. Cross-sectional disks were cut from subfossil boles and detrital wood fragments found in the glacier forefield, within lateral and terminal moraines, and on the nunatak surface. In addition, saplings growing on recently deglaciated surfaces were sampled to establish the local ecesis interval (i.e. the time interval between surface exposure and vegetation establishment; McCarthy and Luckman 1993).

Living whitebark pine and subalpine fir trees close to Bridge Glacier were sampled to establish living tree-ring chronologies. A whitebark pine chronology was

constructed from cores ($n = 42$, from 21 trees) collected on a north-facing, mesic slope distal to the southeast lateral moraine complex ($50^{\circ} 50' 23''$ N, $123^{\circ} 28' 00''$ W; 1435 m asl). A subalpine fir chronology was constructed from cores ($n = 40$, from 20 trees) collected on a west-facing slope located 3 km south of the glacier terminus ($50^{\circ} 48' 36''$ N, $123^{\circ} 29' 11''$ W; 1544 m asl). Two cores were taken from each tree at breast height (1.3 m) (Luckman 1996) with a three-thread, 4.3-mm diameter Haglof increment borer. Each core was collected approximately 180° from its counterpart, along a plane that paralleled the contours of the terrain.

After air-drying, the disks and tree cores were prepared for analysis at the University of Victoria Tree-Ring Laboratory. The tree cores were mounted in slotted boards and sanded with progressively finer sand paper until a fine polish was achieved (Phipps 1985). The disks were similarly sanded to a fine polish, except for fragmented or rotten samples that were first stabilized by pouring melted paraffin wax onto their surface and allowing it to harden.

ii. Tree-ring analysis

Tree-ring widths of the increment core samples taken from living trees were measured to 0.001 mm accuracy using a WinDENDRO (v. 2003b) image analysis system (Guay et al. 1992). In order to ensure that each ring width series was complete (i.e. not missing any rings), the paired increment core samples from each living tree were crossdated. Once internally crossdated, species-specific tree-ring chronologies were developed with the aid of the visual crossdating program CDendro v.4.1.1 (Larsson 2003) and the quality-checking program COFECHA v 6.06 (Holmes 1999). The chronologies were standardized using the program ARSTAN (Cook and Holmes 1999) to remove

growth trends in the series related to tree-age and stand dynamics (Cook et al. 1990), producing indexed chronologies. The chronologies were standardised using a cubic smoothing spline (Cook and Peters 1981) of 50% response frequency of 67% of the series length; each ring width measurement was then divided by its respective spline (or “expected”) value to produce a dimensionless index.

The species of each subfossil wood sample was identified using a 40x microscope and a standard reference key (Hoadley 1993). The tree ring widths of each sample were measured along a minimum of two paths per disc and internally crossdated to ensure that no rings were missed. In cases where there was evidence of extreme ring compression and variability, additional paths were measured and crossdated. Once internally crossdated, attempts were initially made to crossdate these floating series with the living chronologies. For samples for which this was unsuccessful, attempts were then made to crossdate the floating series with other subfossil samples on the basis of species and location within the study site.

All series intercorrelation values (r) reported were calculated by the COFECHA program and are significant at the 99% confidence interval. Series transformations were completed using a cubic smoothing spline and the default parameters (i.e. 50% wavelength cutoff at 32 years, examined in 50-year segments lagged successively by 25 years). Where subfossil wood failed to crossdate with the living chronologies but successfully crossdated with a radiocarbon dated sample, relative dates were assigned to create floating chronologies. Radiocarbon ages of the subfossil samples are presented in the text as ^{14}C years BP. Standard radiometric analyses were conducted by Beta Analytic

Inc. on selected samples; samples submitted for analysis contained multiple perimeter annual growth rings.

3.4.2 Lichenometry

Lichenometry is a calibrated-age dating technique used to establish a minimum surface date of rocks using measurements of lichen thallus diameter. The technique is based on the assumption that the largest lichen growing on a substrate (e.g., a moraine) is the oldest individual, and that, if the growth rate for a given species is known, the maximum lichen size will provide a minimum stabilization date for the substrate (Grove 1988). Lichenometry requires the development of a lichen growth curve that expresses radial growth trends with time. Typically, lichen growth curves are established from age-controlled dating points, i.e. lichen thalli measurements from surfaces for which ages can be determined from historical accounts, photographs, or dendrochronology (Porter 1981).

In PNA lichenometry has been employed using several genera of lichen (e.g., Denton and Karlén 1973, 1977; McCarthy and Smith 1995; Wiles et al. 2002). Reconnaissance traverses of moraine surfaces at Bridge Glacier revealed that only *Rhizocarpon geographicum* lichen thalli were present in sufficient quantity for lichenometric dating. Local lichenometric control points were collected from dated headstones colonized by *R. geographicum* found in three cemeteries within 120 km of Bridge Glacier: the Mt. Garibaldi cemetery located within the city of Squamish (49° 44' 39" N, 123° 07' 58" W, 10 m asl), the Pemberton cemetery in Pemberton Meadows (50° 25' 46" N, 122° 54' 24" W, 225 m asl), and the William Haylmore Cemetery in the town of Goldbridge (50° 51' 21" N, 122° 50' 05" W, 685 m asl) (Appendix A). Although these sites are located at lower elevations than the Bridge Glacier site, the effects of elevation

on *Rhizocarpon* sp. growth rates has been negligible in other studies (*c.f.* Porter 1981; Smith and Desloges 2000; Larocque and Smith 2004; Lewis and Smith 2004).

Examination of historical vertical aerial photographs of the Bridge Glacier forefield provided additional dating control points, with lichen thalli measurements taken from recessional moraines surfaces where the exposure date was known.

Despite the numerous locations used for lichenometric control point collection, a shortage of reliable calibration data precluded the development of a lichen growth curve specific to the Bridge Glacier area. Several established *R. geographicum* growth curves were therefore reviewed to test their applicability to the Bridge Glacier site. The first growth curve examined was the 165-year record developed by Smith and Desloges (2000) for use at Tzeetsaytsul Glacier. Two radiocarbon-controlled lichen ages added by Larocque and Smith (2004) extend this curve to 680 years in the Mount Waddington area. The second growth curve included the 145-year long record presented by O'Neal and Schoenenberger (2003) from Washington and northern Oregon states that extends Porter (1981)'s 126-year long growth curve from Mt. Rainier. Both studies used historic data (aerial photographs, maps, and historical accounts) for lichenometric control points. Finally, the *R. geographicum* growth curve presented by Lewis and Smith (2004) for glacier sites on Vancouver Island was examined. Anchored by a 290-year old tree-ring dated control point, the curve was constructed from dated structures and headstones located within the surrounding region.

Correlation analyses of the relationship between the control points collected as part of this study (Appendix A) with those from previous regional studies demonstrate that the closest affinity ($r = 0.67$) is to the data presented by Larocque and Smith (2004).

As a consequence, the Larocque and Smith (2004) *R. geographicum* growth curve was used to assign relative ages to the Bridge Glacier deposits. Applying this particular lichen growth curve introduces an error margin associated with the 95% confidence interval. As reported by Larocque and Smith (2004: 410), this error envelope is greater for older samples. For instance, a 25-year old surface incorporates a +6 to -5-year error, whereas a 150-year old surface has an inherent dating error of +45 to -30 years.

Comparison of the observed ages of the lichenometric control points with the predicted ages calculated from the Larocque and Smith (2004) growth curve shows that this lichen growth curve appears to consistently underestimate the age of lichen thalli by 10 to 31 years. To account for this systematic error an increment of 17 years was added to all of the Bridge Glacier lichen age estimates derived from lichenometry.

Lichenometric sampling was completed on moraine crests found along transects crossing both terminal and lateral moraines. Moraines were identified according to their distal position along the transect, with the outermost moraine crest numbered as M1. These identifiers are unique to each transect and do not necessarily correlate to moraine crests with the same designation along other transects. At each crest intersection point a minimum of 30 lichen diameters (a- and b-axes) of the largest *R. geographicum* thalli located within 25 m of the transect were measured using handheld digital callipers (accuracy 0.01 mm). The maximum lichen diameter on each moraine was used to determine the substrate age (Smith and Desloges 2000; Larocque and Smith 2003).

3.5 Observations and results

3.5.1 Dendroglaciological investigations

Subfossil wood collected at Bridge Glacier was compared to the living whitebark pine and subalpine fir tree-ring chronologies. Spanning intervals of 322 and 196 years, respectively (Table 3.3 and Appendix B), both chronologies have similar growth trends and comparable inter-series correlation values, indicating a common radial growth response to climate (Larocque and Smith 2005). Both chronologies were initially used in an attempt to provide absolute dates for the wood samples collected at each of the locations described below. Radiocarbon dating of the subfossil wood samples revealed that their time since death exceeds the age of the oldest living tree chronology, which unfortunately prevented successful crossdating between the living chronologies and the subfossil wood at Bridge Glacier.

Table 3.3. Characteristics of chronologies developed from living trees at Bridge Glacier.

Chronology species	Location	Elevation (m asl)	N series	N trees	Series intercorrelation	Total length (years)	Mean length of series (yrs)	Span (Years A.D.)
Subalpine fir	50° 48' 36" N; 123° 29' 11" W	1544	40	20	0.617	322	212.4	1682-2003
Whitebark pine	50° 50' 23" N; 123° 28' 00" W	1435	41	21	0.462	196	157.2	1808-2003

i. Nunatak

Fourteen subfossil wood samples were collected from three locations on the nunatak. The first site is a gully below an incised till deposit plastered on the south flank of the nunatak at 1690 m asl. Historical aerial photographs indicate that this area became ice-free between 1975 and 1979. The gully is approximately 100 m above the

contemporary ice level, directly below a bedrock outcrop overlain by the Orthic Humo-Ferric paleosol described by Ryder and Thomson (1986).

Seven bole segments (4 subalpine fir, 3 whitebark pine) were sampled within the outcrop gully. Although the majority of these boles were found buried within the till and had been recently exposed by fluvial erosion, some of the wood found on the surface at this site may have been transported from locations higher on the nunatak where surface detritus was observed. The boles ranged in size from 20 to 50 cm in diameter, and 0.3 to 2.5 m in length, and had minimum ages of 96 to 318 years (Table 3.4).

Five of the subalpine fir samples (BG02-01, -03; BG03-57, -59, -60) from this gully crossdate ($r=0.673$) to create a 159-year long floating chronology (Bridge 1; Table 3.4 and Appendix C). The size and surface condition of the majority of the boles (no bark, but limited surface mastication) suggest they were not transported far from their growth position after being overridden by Bridge Glacier. Radiocarbon dating (BG03-57) shows that the chronology spans the interval from 3124 to 2966 ^{14}C years BP (revealed through crossdating) and shows that all five trees were killed simultaneously at ca. 2980 ^{14}C years BP (1390–1010 BC; Fig. 3.5 and Table 3.4).

A single piece of detrital whitebark pine was also found at this site (BG02-02). Unlike the other samples, it appeared to have been washed to this location after having been eroded from a prominent upslope gully. The fragment had a minimum of 40 annual tree rings, and an outer-ring radiocarbon age of 1190 ± 60 years BP (690-990 A.D.; Table 3.5). Due to its advanced deterioration the sample could not be used for crossdating, so it is unknown whether any of the other undated wood samples from this site have similar ages.



Fig. 3.4. Nunatak gully where sample BG03-62 was discovered below the Nunatak paleosol. This log crossdates into a 318-year long floating whitebark pine tree ring chronology. The whitebark pine sample (circled) embedded in till at the top of the gully (BG03-66) contained 281 rings, but failed to crossdate with any other samples.

The second sampling site on the nunatak was a neighbouring gully west of the outcrop gully (Fig. 3.4). At this location an exposed whitebark pine log (BG03-62) was found in the gully channel 10 m below the paleosol horizon. Approximately 35 cm in diameter, this tree was a minimum of 132 years old at time of death. It crossdated with an undated whitebark pine disk (BG03-58, 318 rings) from the first nunatak gully ($r = 0.533$), producing an undated 318-year long floating chronology (Bridge 6; Table 3.4).

Samples were also collected from exposed detrital wood found at sites positioned on the proximal slope of a distinct moraine on the nunatak crest at 1745-1790 m asl, approximately 100 m upslope from the gully sites. These boles were previously examined by Ryder (1991:38). Samples were collected from a large whitebark pine bole (BG03-64) and a large subalpine fir root (BG03-63). Both samples contained over 200 annual tree rings and crossdate ($r = 0.593$) to form a 312-year chronology (Bridge 5; Table 3.4 and Appendix C). Ryder and Thomson (1986) report that BG03-63 and -64 have radiocarbon ages of 540 ± 50 (1300 -1368 A.D. or 1381 - 1445 A.D., multiple intercepts) and 680 ± 50 years BP (1258 – 1400 A.D.), respectively (Table 3.5).

ii. South Creek

The southern lateral moraine complex at Bridge Glacier is incised by South Creek, which flows into Bridge Lake through the bottom sediments of the former moraine-dammed lake described by Ryder (1991). Examination of the west bank of the incised moraine revealed it to be a composite deposit consisting of a till overlain by a unit of lacustrine sands capped by a second till unit. The upper surface of the basal till is delineated by a >30 cm thick reddish brown paleosol containing remnant root fragments and an overlying 2-5 cm thick organic mat. A subalpine fir branch fragment (BG03-01) lying on the surface of the buried organic mat yielded an age of 430 ± 60 ^{14}C years BP (1410-1530 or 1545–1635 A.D., multiple intercepts; Table 3.5). The overlying sandy unit contained numerous bedded layers with small detrital pieces of wood.

Near where South Creek drains into Bridge Lake, the creek has incised through a mantle of ice-proximal sands and gravels, modified by postglacial alluvial sorting, into an underlying sandy till deposit. Exposed along 15 m of channel bank was a thin (<10 cm)

Table 3.4. Floating chronologies developed at Bridge Glacier. Years since death reported are assigned perimeter dates of the samples. Dates reported in bold indicate radiocarbon-dated samples. Perimeter dates for remaining samples are not absolute calendar dates, but instead are derived from their relative position within their associated radiocarbon controlled floating chronology.

Chronology name	Chronology length (years)	Temporal span (¹⁴ C years BP)	No. of samples	R	Samples included in chronology						
					Sample ID	Species ^a	Location	No. of rings	Yrs since death (¹⁴ C years BP)	2σ Calibration date ^b	2σ Calibration YBP ^b
Bridge 1	159	3124 - 2966	5	0.673	BG02-01	saf	Nunatak	96	2966		
					BG02-03	saf	Nunatak	126	2968		
					BG03-57	saf	Nunatak	144	2980 ± 60	BC 1390 - 1010	3340 - 2960
					BG03-59	saf	Nunatak	144	2973		
					BR03-60	saf	Nunatak	127	2987		
					BG02-10	wbp	Area 1500	86	1500 ± 50	A.D. 430 - 650	1520 - 1300
Bridge 2	216	1692 to 1477	4	0.597	BG03-13	wbp	Area 1500	166	1529		
					BG03-15	wbp	Area 1500	193	1480		
					BG03-33	wbp	Area 1500	99	1492		
					BG03-43	wbp	Area 1500	64	1502		
					BG03-82	wbp	Gravesite	388	1171		
Bridge 3	438	1577 to 1140	4	0.509	BG03-88	wbp	South Ck	358	1201		
					BG03-91	wbp	Gravesite	217	1360 ± 60	A.D. 600 - 780	1350-1170
					BG03-92	wbp	Gravesite	196	1140		
					BG02-05	wbp	N lat gullies	69	673 ^c		
Bridge 4	149	595 to 743	8	0.543	BG03-05	saf	N lat gullies	56	598		
					BG03-06	wbp	N lat gullies	57	680		
					BG03-07	wbp	N lat gullies	84	610		
					BG03-09	wbp	N lat gullies	97	620 ± 60	A.D. 1280 - 1420	670 - 530
					BG03-12	saf	N lat gullies	115	628		
					BG03-53	saf	N lat gullies	91	595		
					BG03-54	saf	N lat gullies	82	646		

Bridge 5	312	540 to 851 or 590 to 901	2	0.593	BG03-63 BG03-64	saf wbp	Nunatak Nunatak	240 221	540 ± 50 ^d or 590 A.D. 630 or 680 ± 50 ^d A.D.	1258 - 1400 1300 - 1368	692 - 550 650 - 582
Bridge 6	318	undated	2	0.533	BG03-58 BG03-62	wbp wbp	Nunatak Nunatak	318 132			
Bridge 7	253	undated	5	0.559	BG03-14 BG03-16 BG03-24 BG03-26 BG03-73	wbp wbp wbp saf wbp	Area 1500 Area 1500 Area 1500 Area 1500 Area 1500	45 62 119 105 248			
Bridge 8	324	undated	2	0.550	BG03-89 BG03-94	saf saf	South Ck Gravesite	139 305			
Bridge 9	252	undated	3	0.496	BG03-50 BG03-51 BG03-52	wbp saf saf	N lat gullies N lat gullies N lat gullies	90 93 241			

^a saf = subalpine fir; wbp = whitebark pine.

^b All radiocarbon dated samples' 2σ calibration dates provided by Beta Analytic using INTCAL98 (Stuiver and van der Plicht 1998), except where noted.

^c Radiocarbon dated to 690 ± 50 14C years BP.

^d Source Ryder and Thomson (1986); 2σ calibration dates calculated using program Calib 5.01 (Stuiver and Reimer 1993) applying INTCAL 2004 (Reimer et al. 2004).

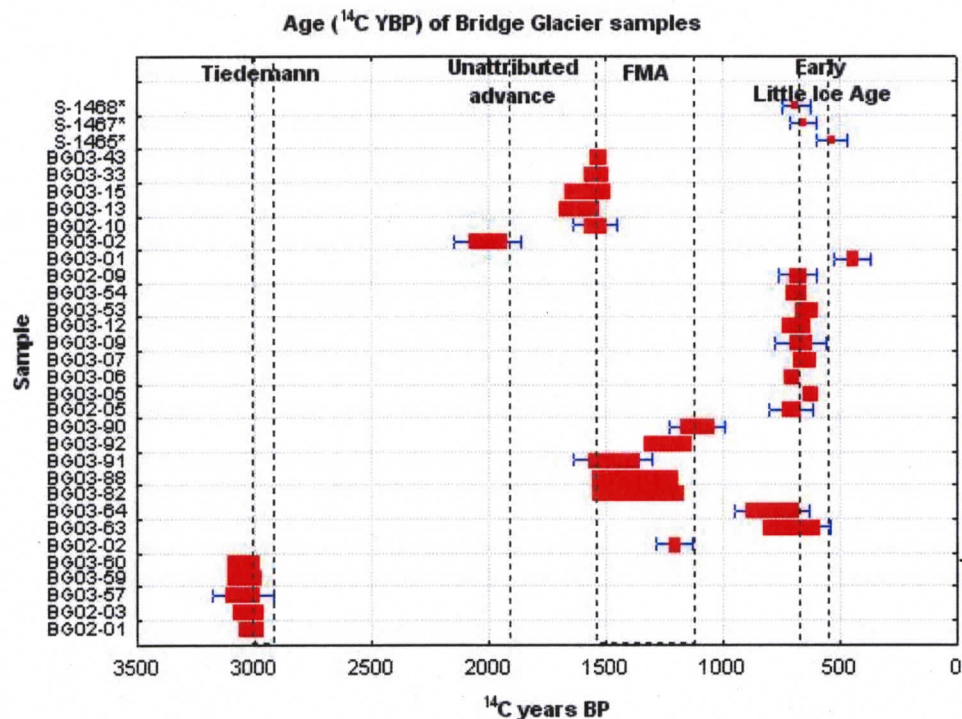


Fig. 3.5. Radiocarbon-dated samples and crossdated samples from Bridge Glacier. Rectangles represent sample age, and error bars represent one standard deviation margin associated with radiocarbon-dated samples. Asterisked samples' radiocarbon dates from Ryder and Thomson (1986).

reddish-brown paleosol on the till surface. This paleosol contained remnants of both sheared stumps and roots in growth position. Excavation of a 1.5-m long, glacially masticated subalpine fir bole (BG03-02, 149 rings) protruding from the bank showed that it was pressed into the underlying paleosol. Radiocarbon dating revealed that the tree was killed 1930 ± 70 ^{14}C years BP (60 BC-240 A.D.; Table 3.5).

Immediately downstream of the paleosol site, South Creek has built a small delta where it enters Bridge Lake. The surface of the delta and the adjoining stream channel contain numerous broken bole sections up to 2.5 m in length. Two of these detrital samples crossdate with samples collected at the Gravesite location. BG03-88 (whitebark pine, 358 rings) was determined to have been killed ca. 1201 ^{14}C years BP (Bridge 3;

Table 3.5. Summary of radiocarbon dated samples at Bridge Glacier.

Sample (Lab sample ID)	Species ^a	Location	¹⁴ C age YBP (no. rings dated)	2 σ Calibration date ^b	2 σ Calibration YBP ^c	Min age (yrs)	Latitude	Longitude	Altitude (m)	Description	Source
BG03-64 (S-1463) ^c	wbp	Nunatak	680 ± 50 (not reported)	A.D. 1258 - 1400	692 - 550	221	50°49' 20" N	123°34' 44" W	1788	Log on southern slope, a few metres below crest of lateral moraine	Ryder and Thomson, 1986
BG03-63 (S-1571) ^c	saf	Nunatak	540 ± 50 (not reported)	A.D. 1300 - 1368 A.D. 1381 - 1445	650 - 582 569 - 549	241	50°49' 21" N	123°34' 44" W	1788	Root fragment on southern slope, a few metres below crest of lateral moraine	Ryder and Thomson, 1986
BG02-01 (Beta-171546)	wbp	Nunatak	1190 ± 60 (45)	A.D. 690 - 990	1260 - 960	146	50°49' 19" N	123°34' 29" W	1960	Large log on southern slope	This study
BG03-57 (Beta-197976)	saf	Nunatak	2980 ± 60 (22)	BC 1390 - 1010	3340 - 2960	145	50°49' 19" N	123°34' 29" W	1689	Log in gully	This study
BG02-10 (Beta-171549)	wbp	Area 1500	1500 ± 50 (23)	A.D. 430 - 650	1520 - 1300	83	50°50' 23" N	123°30' 59" W	1397	<i>In situ</i> log in abandoned stream channel	This study
BG03-90 (Beta-181856)	saf	Gravesite	1040 ± 50 (35)	A.D. 895 - 1040	1055 - 910	144	50°50' 25" N	123°30' 16" W	1391	Log with bark, smeared in direction of ice flow	This study
BG03-91 (Beta-185807)	wbp	Gravesite	1360 ± 60 (unknown)	A.D. 600 - 780	1350 - 1170	218	50°50' 25" N	123°30' 16" W	1391	1 m-long log attached to exposed stump-like piece	This study
BG02-05 (Beta-171547)	wbp	North lateral gully	690 ± 50 (25)	A.D. 1260 - 1400	690 - 550	70	50°50' 44" N	123°30' 18" W	1430	1.5 m-long basal root stalk washed out from gully onto alluvial fan	This study
BG02-09 (Beta-171548)	saf	North lateral gully	650 ± 50 (16)	A.D. 1270 - 1410	680 - 540	67	50°50' 44" N	123°30' 18" W	1430	Stump, partially rooted	This study
BG03-09 (Beta-185806)	wbp	North lateral gully	620 ± 60 (unknown)	A.D. 1280 - 1420	670 - 530	98	50°50' 06" N	123°30' 31" W	1428	<i>In situ</i> , pushed by ice. Root system exposed during excavation	This study

BG03-01 (Beta-181854)	saf	West of South Ck	430 ± 60 (36)	A.D. 1410 - 1530 A.D. 1545 - 1635	540 - 420 405 - 315	36	50°49'32" 123°30'33"	1509	Bole fragment at contact between sandy lacustrine sediments and paleosol	This study
BG03-02 (Beta-181855)	saf	West of South Ck	1930 ± 70 (65)	BC 60 - A.D. 240	2010 - 1710	149	50°49'51" 123°30'02"	1400	Log in paleosol at contact between gravel and sand	This study
S-1468 ^c	saf	South Lk	685 ± 60 (not reported)	A.D. 1227 - 1234 A.D. 1237 - 1401	723 - 716 713 - 549	not reported	50°49'24" 123°29'24"	1482	Outer wood from trunk	Ryder and Thomson, 1986
S-1465 ^c	saf	South Ck	530 ± 65 (not reported)	A.D. 1291 - 1465	659 - 485	not reported	50°49'42" 123°29'30"	1465	Branch	Ryder and Thomson, 1986
S-1467 ^c	Hemlock sp.	South Ck	655 ± 60 (not reported)	A.D. 1264 - 1411	686 - 539	not reported	50°49'42" 123°29'42"	1465	Branch	Ryder and Thomson, 1986

^a saf = subalpine fir; wbp = whitebark pine.

^b All Bridge radiocarbon dates calculated by Beta Analytic using INTCAL98 (Stuiver and van der Plicht 1998); Ryder and Thomson (1986) 2σ calibration dates calculated using program Calib 5.01 (Stuiver and Reimer 1993) applying INTCAL 2004 (Reimer et al. 2004).

^c Sample identification reported by Ryder and Thomson (1986), calculated by the Saskatchewan Research Council.



Fig. 3.6. Paleosol developed on till of possible Tiedemann age on west bank of South Creek. Several root fragments were observed within the paleosol. A subalpine fir branch fragment (BG03-01) found sitting on the surface of the buried organic layer has an age of 430 ± 60 ^{14}C years BP.

Table 3.4 and Appendix C). BG03-89 (subalpine fir, 139 rings) is undated but crossdates with Gravesite sample BG03-94 to form a 324-year long floating chronology (Bridge 8; Table 3.4).

iii. Gravesite

Nested recessional moraines occur on the surface of the glacially-modified moraine mound on the north side of Bridge Lake. Two living tree saplings found growing on this surface were destructively sampled to establish a local ecesis interval. Based on an examination of vertical aerial photographs showing the length of time since Bridge Glacier retreated from these sampling sites, it is estimated that whitebark pine and

subalpine trees at this site have ecesis intervals of 20 and 14 years, respectively (Table 3.6).

Table 3.6. Seedlings sampled at the Gravesite to determine tree ecesis interval for Bridge Glacier forefield.

Sample no.	Species	Ring count (years)	Age of substrate (years)	Ecesis interval (years)
1	Whitebark pine	10	30	20
2	Subalpine fir	16	30	14

The morainal mound at the Gravesite location has been incised at several locations by meltwater. Examination of these now abandoned channels revealed that they had cut through a shallow surficial mantle of till to expose an older underlying till deposit. The surface of the underlying unit is mantled by a paleosol that in many places is overlain by buried downvalley-oriented tree boles killed during an advance of Bridge Glacier. Excavation of several of the boles (e.g. BG03-93, 307 rings) revealed that they were lying directly on the surface of a buried 10-cm thick organic layer underlain by a paleosol (Fig. 3.7).

Samples were collected from sixteen tree boles (12 whitebark pine, 4 subalpine fir), ranging in age from 388 to 146 years. Most of the samples were taken from a group of partially buried boles adjacent to one another and aligned in the direction of glacial flow (Fig. 3.8). Two of the longest boles examined (BG03-81, -82) were over 4 m long, retained small branches, and were traced to articulated *in situ* stumps, suggesting they were pushed over and buried as the glacier advanced downvalley. Some of the boles retained a partial covering of bark and a few showed evidence of bark beetle attack, indicating that the perimeter wood of the boles was largely intact.



Fig. 3.7. Gravesite sample BG03-93, lying on 10-cm thick organic horizon underlain by a paleosol. Located a few metres downvalley of the other Gravesite samples, this *in situ* subalpine fir bole contained 307 growth rings, indicating that a mature forest existed here prior to being run over by the late FMA advance of Bridge Glacier ca. 1040 ± 50 ^{14}C years BP.

Three whitebark pine samples from this site (BG03-82, -91, -92) and a detrital sample from South Creek (BG03-88) crossdate ($r = 0.509$) to produce a 438-year long floating chronology (Bridge 3; Table 3.4 and Appendix C). Anchored by radiocarbon-dated perimeter rings from BG03-91 (1360 ± 60 ^{14}C years BP [600–780 A.D.]; Table 3.5), the chronology has a temporal span of 1577 to 1140 ^{14}C years BP. Although a perimeter date of 1360 ± 60 ^{14}C years BP was assigned to BG03-91 at this location, the sample was not *in situ* and was assumed to have originated at an unknown location upvalley.

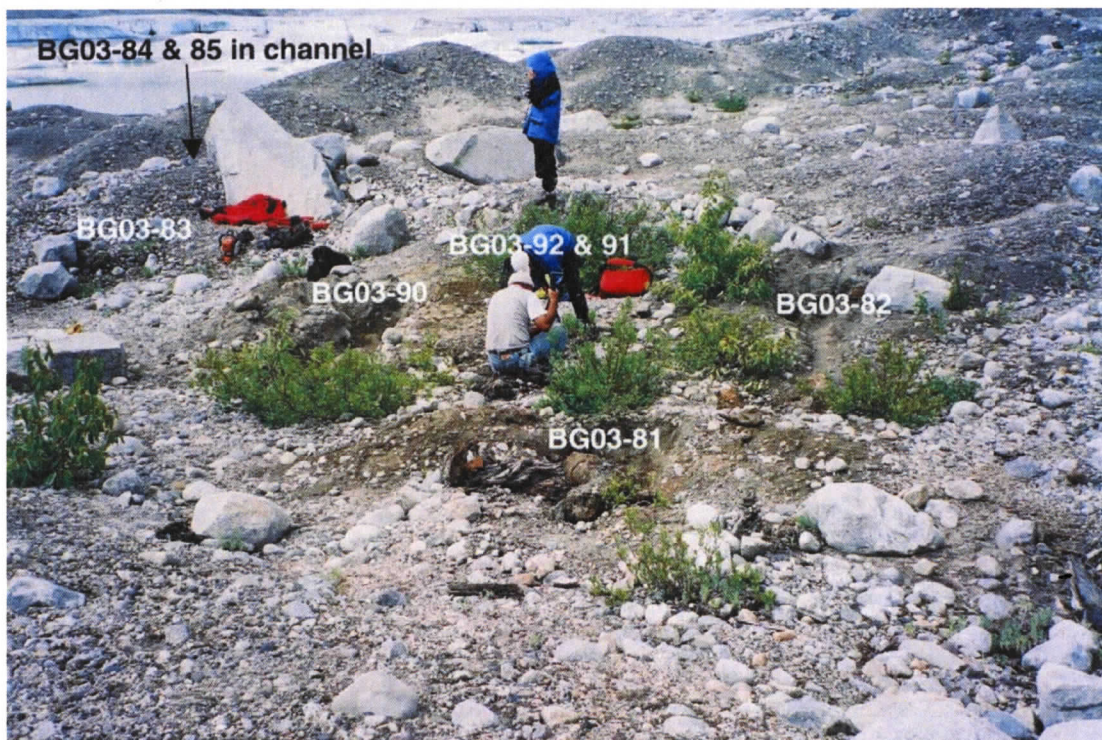


Fig. 3.8. The Gravesite study site showing locations of the subfossil boles buried by the advance of Bridge Glacier ca. 1360 ± 60 to 1040 ± 50 ^{14}C years BP. Shown in the background is a recessional moraine formed during the retreat of Bridge Glacier in 1977.

Two of the buried subalpine fir boles sampled at this site were similarly aligned in the direction of glacial flow and were killed by the same glacial advance. One of them (BG3-90, 142 rings) has a perimeter age of 1040 ± 50 ^{14}C years BP (895 – 1040 A.D.; Table 3.5). The second sample (BG03-94, 305 rings) had articulated *in situ* roots and crossdated ($r = 0.55$) with a detrital subalpine bole from South Creek (BG03-89, 139 rings) to create a 324-year long floating chronology (Bridge 8; Table 3.4).

iv. Area 1500

Five hundred metres upvalley from the Gravesite, and flanking the 2003 terminus position of Bridge Glacier, is a section of forefield characterized by abandoned meltwater

channels and a broad outwash fan. Numerous pieces of detrital wood of indeterminate origin litter the surface of this area.

Sample BG02-10 was collected from an *in situ* whitebark pine stump located within one of the most deeply incised abandoned stream channels (Fig. 3.9). The stump was rooted in till and the standing bole was pressed against the proximal face of a large protruding boulder. The sample was 22 cm in diameter, contained 86 tree rings, and was glacially sheared in a downvalley direction. Radiocarbon dating of a perimeter wood sample established that the tree was killed in 1500 ± 50 ^{14}C years BP (430-650 A.D.; Table 3.5).

Forty-one additional samples (26 whitebark pine, 15 subalpine fir) were collected from detrital wood at this site. Most samples were collected from the surface of the



Fig. 3.9. *In situ* sample BG02-10 located in an abandoned meltwater channel. Radiocarbon dated to 1500 ± 50 ^{14}C years BP, this whitebark pine tree was killed during the FMA advance of Bridge Glacier.

outwash fan downstream of the abandoned channel in which BG02-10 was located. As many of these samples contained relatively few annual rings, shorter segments were examined using COFECHA to assess the crossdating quality (50% wavelength cut-off at 32 years, examined in 25 year segments lagged successively by 12 years).

Dendrochronological analyses of the detrital samples led to the construction of two floating chronologies. Four whitebark pine samples located on the outwash plain downstream from BG02-10 crossdate ($r = 0.597$) with BG02-10 to form a 216-year long chronology spanning the interval from 1692 to 1477 ^{14}C years BP (Bridge 2; Table 3.4 and Appendix C). A second chronology was constructed from five samples collected at this location. Spanning a 253-year period, this undated floating chronology ($r = 0.559$) is made up of 1 subalpine fir and 4 whitebark pine samples containing from 66 to 194 tree rings (Bridge 7; Table 3.4).

v. Northern gullies

The steep exposed proximal slopes of the lateral moraine flanking the northern valley wall have a series of deeply incised gullies. Exposed at various locations within the gullies are *in situ* downvalley-oriented subfossil boles and detrital wood fragments. Additional pieces of subfossil wood detritus were found on the surface of the broad colluvial fans mantling the valley floor below the gullies.

At one site, an *in situ* 3 m-long whitebark pine bole (BG03-09) was found rooted within a buried paleosol with a distinct surface organic layer (Fig. 3.10). Partially buried beneath a mantle of till, this sample contained 98 annual rings, the outermost of which date to 620 ± 60 ^{14}C years BP (1280 – 1420 A.D.; Table 3.5).

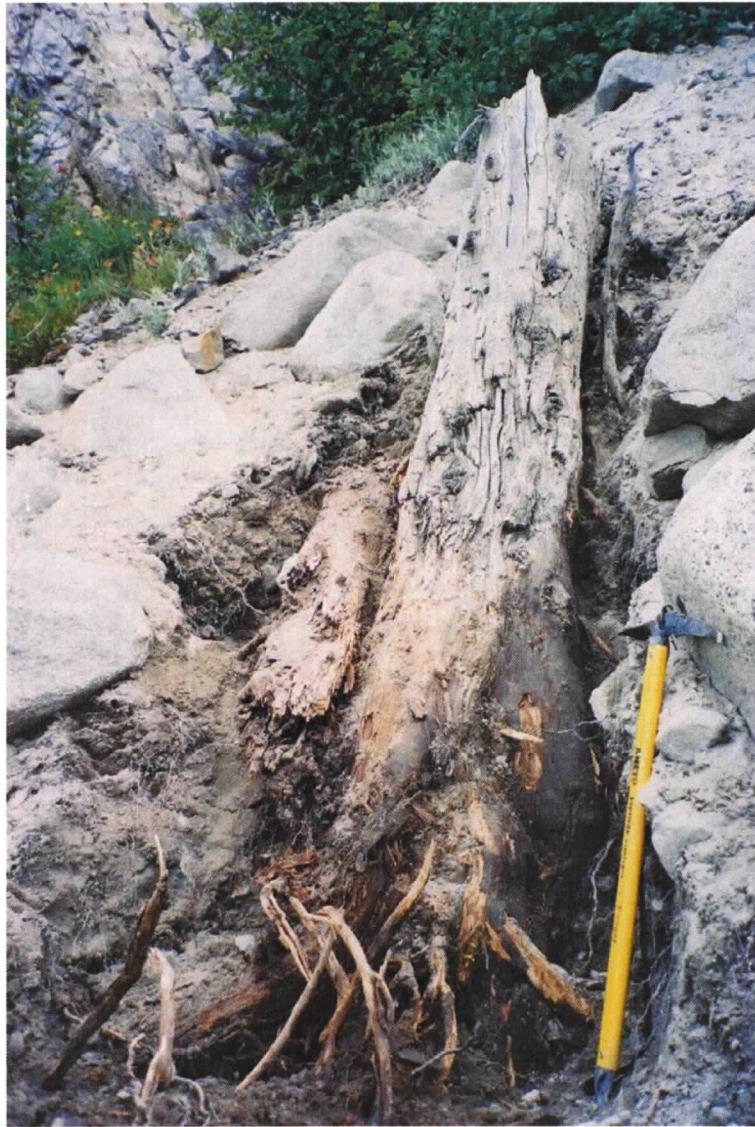


Fig. 3.10. *In situ* sample BG03-09 (620 ± 60 ^{14}C years BP) located on the proximal slope of the prominent northern lateral moraine at Bridge Glacier. Excavation of the site revealed the bole was rooted in a discontinuous paleosol. This whitebark pine bole was killed during an early LIA advance of Bridge Glacier, and contributes to the 149-year long tree ring chronology developed from samples at the northern gullies site.

In total, 22 samples (12 subalpine fir, 10 whitebark pine) were collected within or adjacent to the northern lateral moraine gullies. Two floating chronologies were constructed. The first chronology is a 149-year long whitebark pine chronology ($r = 0.543$) built from eight samples anchored to BG03-09 (Bridge 4; Table 3.4 and Appendix

C). Six samples were found in close proximity to BG03-09 and the remaining two were collected from boles located within a prominent gully approximately 500 m upvalley. Spanning the interval from 595 to 743 ^{14}C years BP, the chronology may record two distinct killing events, determined by the relative kill-dates of the crossdated samples: 670 ^{14}C years BP and 625-595 ^{14}C years BP (Fig. 3.5). A second undated 252-year long floating chronology ($r = 0.496$) was constructed from two subalpine fir samples (BG03-51, -52) and a whitebark pine sample (BG03-50) (Bridge 9; Table 3.4).

3.5.2 Lichenometric investigations

Lichenometric dates are reported in calendar years A.D., as calculated using the Larocque and Smith (2004) lichen growth curve, and are offered as minimum dates for the abandonment of moraine surfaces. The intent in assigning specific calendar years to these events is not to present absolute precise years of moraine stabilization (as these dates incorporate the systematic errors reported by Larocque and Smith (2004)), but to illustrate a sequence of moraine stabilization periods at Bridge Glacier that clearly date back to the early LIA.

i. Terminal moraines

The terminal moraine complex at Bridge Glacier comprises eight moraines (M1-M8, Fig. 3.11). Rising to heights of 2-3 m, the moraines are of variable width and are largely composed of bouldery sandy gravel. The moraines are bisected by the Bridge River, separating them into north (TMN) and south (TMS) segments. Examination of a vertical aerial photograph taken in 1993 (30 BC93086-033; Table 3.1) shows that the river had avulsed and shifted position northward by 2002. These adjustments have left behind a series of abandoned channels and fluvial deposits that partially fill some intermorainal

areas, notably within the TMN complex and between M4 and M7 within the TMS (Fig. 3.11).

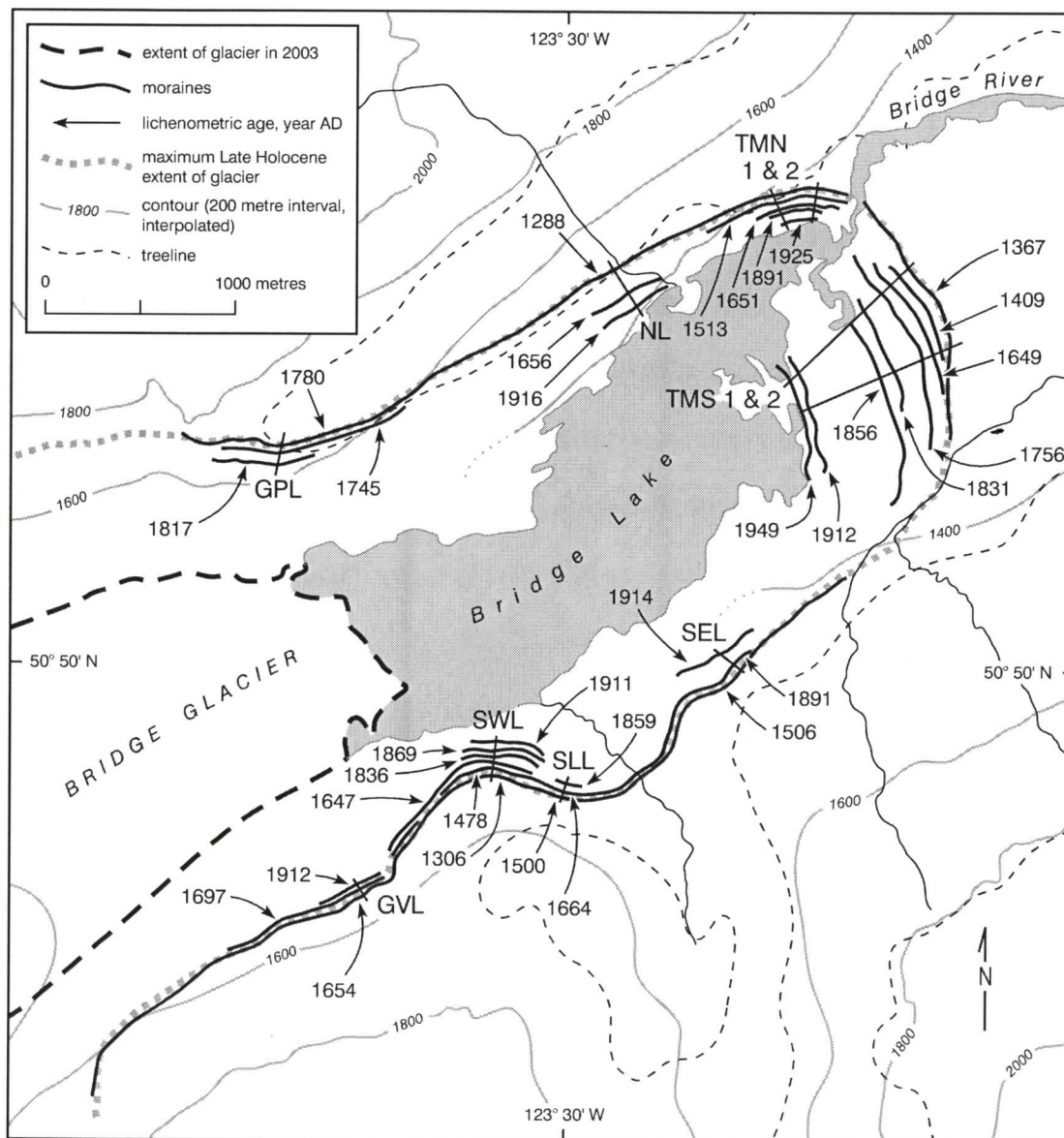


Fig. 3.11. Location of moraines and lichen transects (represented by straight lines through intact moraines series) at Bridge Glacier. Ten transects extended over eight distinct moraine sets (TMS and TMN transects were amalgamated for lichenometric analyses). TMN = terminal moraine north; TMS = terminal moraine south; GPL = Gemini Pass lateral; NL = north lateral; SEL = South Creek east lateral; SLL = South Lake lateral; SWL = South Creek west lateral; GVL = Goat Valley lateral.

Four lichenometric transects were established through the terminal moraine complex. Two transects were located on the north side of the river adjacent to the valley wall (TMN-1, -2), and two transects were located on the south side of the river in the centre part of the valley (TMS-1, -2). The most distal moraine was identified as M1, with subsequently more ice-proximal moraines identified as M2-M8. The outermost terminal moraine (M1) is vegetated by mature whitebark pine and subalpine fir trees, the oldest of which are over 100 years (Ryder 1991). Best preserved south of the river for only a few hundred metres between the present and 1993 river channel positions, all that remains of M1 within the TMN complex is a residual tree-covered mound of debris surrounded by outwash sediments. The largest lichen thalli (a-axis) found on the transects intersecting M1 range from 55.4 mm (TMN-M1) to 79.5 mm (TMS-M1) and indicate it was constructed following an advance of the glacier sometime before ca. 1367 A.D. (Table 3.7).

Segments of M2 were traced to locations on the north and south sides of the river (Fig. 3.11). On the north side of the river, M2 is the outermost continuous distal moraine and is composed of bouldery till surrounded by outwash sediment. Immediately south of the river M2 is initially distinct and laterally contiguous, but becomes less prominent towards the valley centre. The largest lichen thalli found on M2 measured 66.1 mm and 75.7 mm north and south of Bridge River, respectively, indicating an early 15th century (ca. 1409 A.D.) stabilization date (Table 3.7).

M3 and M4 were found in close proximity to one another north of the river, though the sizes of the largest lichen thalli found on each moraine differ significantly in size (M3: 53.6 mm; M4: 31.5 mm). South of the river M4 partially overlaps M3 creating

a broad flat-topped moraine crest with gently sloped sides. Lichen thalli on TMS-M3 were up to 53.8 mm in diameter, while those on TMN-M4 were up to 44.1 mm in size. Based on these measurements a mid-17th century age (ca. 1649 A.D.) was assigned to M3 and a mid-18th century age (ca. 1756 A.D.) to M4.

M5 and M6 are smaller than the more distal terminal moraines. Widely spaced with relic outwash channels and fluvial sediments separating them along the TMS transects (Fig. 3.11), the moraines appear to have stabilized within a short time of each other in the mid-19th century (TMS-M5: 38.1 mm [ca. 1831 A.D.]; TMS-M6: 35.8 mm [ca. 1856 A.D.]). On the north side of the river TMN-M5 is the most proximal terminal moraine found. It is a well-defined, steep-walled moraine 10 m from Bridge Lake. Lichen grow preferentially on its ice-proximal flank, but were a maximum of only 25.7 mm in diameter (i.e. stabilization ca. 1925 A.D., except for an anomalously large lichen measuring 49.5 mm).

TMS-M7 and -M8 are small (1-1.5 m high) recessional moraines protruding from the shallow water of Bridge Lake. Maximum lichen diameters on M7 and M8 measured 28.2 and 18.8 mm, respectively, describing moraines that stabilized in ca. 1912 and 1949 A.D.

ii. North lateral moraines

Lichenometric transects were completed at two locations on the north side of Bridge Valley. In both cases, the transects extended from the most ice-proximal moraine crest upward into the surrounding forest to M1.

One transect (NL; Fig. 3.11) was located north of Bridge Lake within a nested sequence of forested lateral moraine crests east of the Northern gullies site. The most

distal moraine crest (M1) is located at 1555 m asl some 150 vertical m above the valley floor. A thallus measuring 86.6 mm was found on the moraine crest, indicating abandonment during the late-13th century (ca. 1288 A.D.; Table 3.7). M2 was located approximately 30 m downslope at 1546 m asl. Lichen thalli ranging in maximum size from 48.6 to 53.1 mm place the stabilization of M2 at this location sometime in the late 17th to early 18th centuries (1656 to 1706 A.D.). M3 was found approximately 50 m downslope of M2 at 1526 m asl and was heavily colonized by alder bushes. The largest lichen thallus observed measured 27.2 mm and suggests that ice downwasted from the site early in the 20th century (ca. 1916 A.D.).

The second transect completed on the north side of the Bridge Valley is located below where a prominent tributary valley (Gemini Pass, GPL) joins the main trunk valley, approximately 2 km upvalley of the northern gullies site (Fig. 3.11). Partially forested, the blocky to bouldery moraine crests are located in close proximity to one another at 1663 m asl approximately 250 vertical m above the valley floor. The largest lichen thalli observed on M1 and M2 measured 41.9 and 45.1 mm respectively, suggesting they were both established in the 18th century (ca. 1780 to 1745 A.D.). The most proximal moraine sampled at this site is a thin, heavily vegetated ridge on which lichen colonization is relatively sparse. Of the few lichen sampled here, the largest measured 39.3 mm in diameter, placing stabilization of the moraine in the early 19th century (ca. 1817 A.D.).

iii. South lateral moraines

Four lichenometric transects were established on the lateral moraines lining the southern valley wall above Bridge Glacier (Fig. 3.11). Three of the transects were located

Table 3.7. Maximum lichen thallus measurements for surveyed moraines at Bridge Glacier. Lichenometric dates incorporate systematic errors reported by Larocque and Smith (2004).

Moraine ID	Latitude ^a (N)	Longitude ^a (W)	Elevation (m)	Year collected	Largest thallus (mm)	Calculated age (years) ^b	Moraine stabilization (year A.D.) ^c
TMS-M1	50° 51' 29.4"	123° 28' 35.0"	1380	2003	79.5	636.3	1367
TMS-M2	50° 51' 03.9"	123° 28' 08.5"	1389	2003	73.0	564.8	1438
TMS-M2	50° 50' 54.8"	123° 28' 13.4"	1399	2003	75.7	594.5	1409
TMS-M3				2003	47.6	285.4	1718
TMS-M3	50° 50' 53.7"	123° 28' 14.2"	1402	2003	53.8	353.6	1649
TMS-M4				2003	44.1	246.9	1756
TMS-M5				2003	38.1	171.8	1831
TMS-M6				2003	35.8	147.4	1856
TMS-M7	50° 50' 56.8"	123° 28' 54.8"	1391	2003	28.2	91.0	1912
TMS-M8	50° 50' 48.7"	123° 28' 48.6"	1392	2003	18.8	53.7	1949
TMN-M1	50° 51' 32.4"	123° 28' 44.7"	1391	2003	55.4	371.2	1632
TMN-M2			1398	2002	66.1	488.9	1513
TMN-M2				2003	63.1	455.9	1547
TMN-M3			1398	2002	53.6	351.4	1651
TMN-M3				2003	51.5	328.3	1675
TMN-M4				2003	31.5	111.6	1891
TMN-M5				2003	25.7	78.4	1925
TMN-M5				2002	49.5	306.3	1696
SLL-M1	50° 49' 29.3"	123° 29' 50.0"	1562	2003	67.4	503.2	1500
SLL-M2	50° 49' 29.3"	123° 29' 48.3"		2003	52.5	339.3	1664
SLL-M3	50° 49' 28.7"	123° 29' 45.4"	1555	2003	35.4	143.6	1859
SWL-M1			1586	2003	85.0	696.8	1306
SWL-M2	50° 49' 29.3"	123° 29' 50.0"	1573	2003	69.4	525.2	1478
SWL-M2	50° 49' 32.1"	123° 30' 02.3"		2003	68.9	502.7	1483
SWL-M3	50° 49' 29.3"	123° 30' 02.2"	1572	2003	54.0	355.8	1647
SWL-M4	50° 49' 32.8"	123° 30' 01.9"	1576	2003	37.7	167.3	1836
SWL-M5	50° 49' 33.3"	123° 30' 01.7"	1568	2003	34.4	134.5	1869
SWL-M6	50° 49' 33.5"	123° 30' 0.6"	1565	2003	28.4	92.1	1911
SEL-M1	50° 49' 26.1"	123° 29' 20.5"	1547	2003	66.8	496.6	1506
SEL-M2	50° 49' 47.0"	123° 29' 24.5"	1519	2003	31.6	95.3	1891
SEL-M3	50° 49' 57.3"	123° 29' 14.3"	1498	2003	27.9	89.4	1914
GVL-M1	50° 48' 59.7"	123° 31' 37.7"	1679	2003	53.4	349.2	1654
GVL-M2	50° 48' 59.9"	123° 31' 37.6"	1681	2003	49.5	306.3	1697
GVL-M3	50° 49' 60.0"	123° 31' 37.5"	1678	2003	28.2	91.0	1912
NL-M1			1555	2002	86.6	714.4	1288
NL-M2				2002	53.1	345.9	1656
NL-M2			1546	2002	48.6	296.4	1706
NL-M3	50° 50' 50.6"	123° 30' 31.9"	1526	2002	27.2	85.7	1916
GPL-M1	50° 50' 32.9"	123° 32' 17.1"	1663	2003	41.9	222.6	1780

GPL-M2	50° 50' 32.9"	123° 32' 17.1"	1663	2003	45.1	257.9	1745
GPL-M3	50° 50' 32.9"	123° 32' 17.1"	1663	2003	39.3	186.3	1817

^a Latitude and longitude are provided where recorded in the field at time of sampling; location was not recorded for every moraine.

^b Calculated by applying Smith and Larocque (2004) lichen growth curve.

^c A constant of 17 years was subtracted from year A.D. calculated by Smith and Larocque (2004) lichen growth curve to account for the slower lichen growth rate at Bridge Glacier site.

in the South Creek area: one transect to the east (SEL) and one to the west (SWL) of South Creek, and one above the western shore of the drained moraine-dammed lake (SLL). A fourth transect was established across a series of lateral moraines found further upvalley (GVL).

The transect at SEL crossed through 3 nested moraines located between two bedrock outcrops. The crest of M1 is 1547 m asl, approximately 150 m above the valley bottom. SEL-M1 had a maximum lichen thallus diameter size of 66.8 mm, suggesting an early 16th century moraine stabilization (ca. 1506 A.D.; Table 3.7). Lichen on M2 and M3 measured 31.6 and 27.0 mm respectively, placing their stabilization in the late-19th century (ca. 1891 A.D.) and early 20th century (ca. 1914 A.D.).

The transect at SLL-M1 crossed three lateral moraine crests and is located above the bevelled moraine described by Ryder (1991) at 1562 m asl. Lichen thalli on M1 were a maximum size of 67.4 mm, indicating the moraine stabilized early in the 16th century (ca. 1500 A.D.; Table 3.7). M2 and M3 had maximum lichen thalli a-axis lengths of 52.4 and 35.4 mm, respectively, suggesting they stabilized in the mid-17th (ca. 1664 A.D.) and mid-19th centuries (ca. 1859 A.D.).

The transect at SWL crossed five (M6-M2) lateral moraine crests and terminated at M1, 1586 m asl and approximately 150 m above the valley floor. Composed of angular

rubble and covered by a stand of mature whitebark pine and subalpine fir, a lichen thallus measuring 85.0 mm indicated the moraine stabilized in the early 14th century (ca. 1306 A.D.). The largest lichen thalli found on M2 ranged in size from 68.9 to 69.4 mm and indicate the moraine stabilized in the late 15th century (ca. 1478-1483 A.D.). The largest lichen thallus on SWL-M3 was 54.0 mm in diameter, indicating that the moraine stabilized in the mid 17th century (ca. 1647 A.D.). SWL-M4 and -M5 are located within 5 m of each other and have maximum lichen diameters of 37.7 and 34.4 mm, suggesting both were abandoned in the 19th century (ca. 1836 and 1869 A.D.). Extensive erosion of the ice-proximal face of M6 has left only a small segment of moraine. The largest lichen thallus found measured 28.4 mm across, suggesting that the moraine stabilized in the early 20th century (ca. 1911 A.D.).

The lichenometric transect below Goat Valley (GVL' Fig. 3.11) crossed three closely spaced moraine crests. M1 is located at 1679 m asl, approximately 200 m above Bridge Glacier, and had lichen thalli measuring up to 53.4 mm in diameter, placing its establishment in the mid-17th century (ca. 1654 A.D.; Table 3.7). Although M2 was located 5 m downslope of M1, it is a larger moraine with a crest height of 1681 m asl. A lichen thallus measuring 49.5 mm indicates M2 was constructed late in the 17th century (ca. 1697 A.D.). The largest lichen on M3 is 28.2 mm in diameter and indicates the moraine stabilized early in the 20th century (ca. 1912 A.D.).

3.6 Interpretation of results

Application of dendroglaciological and lichenometric techniques at Bridge Glacier allows for reconstruction of the late Holocene history of the glacier (Fig. 3.12). In the absence of a living tree-ring chronology of sufficient age to crossdate the subfossil

samples, these interpretations were derived from the radiocarbon dating of selected *in situ* and detrital subfossil wood samples, and the crossdating of these samples to other undated samples. Several Holocene glacial events were identified, minimum ages were assigned to moraines, and average rates of glacier movement were determined.

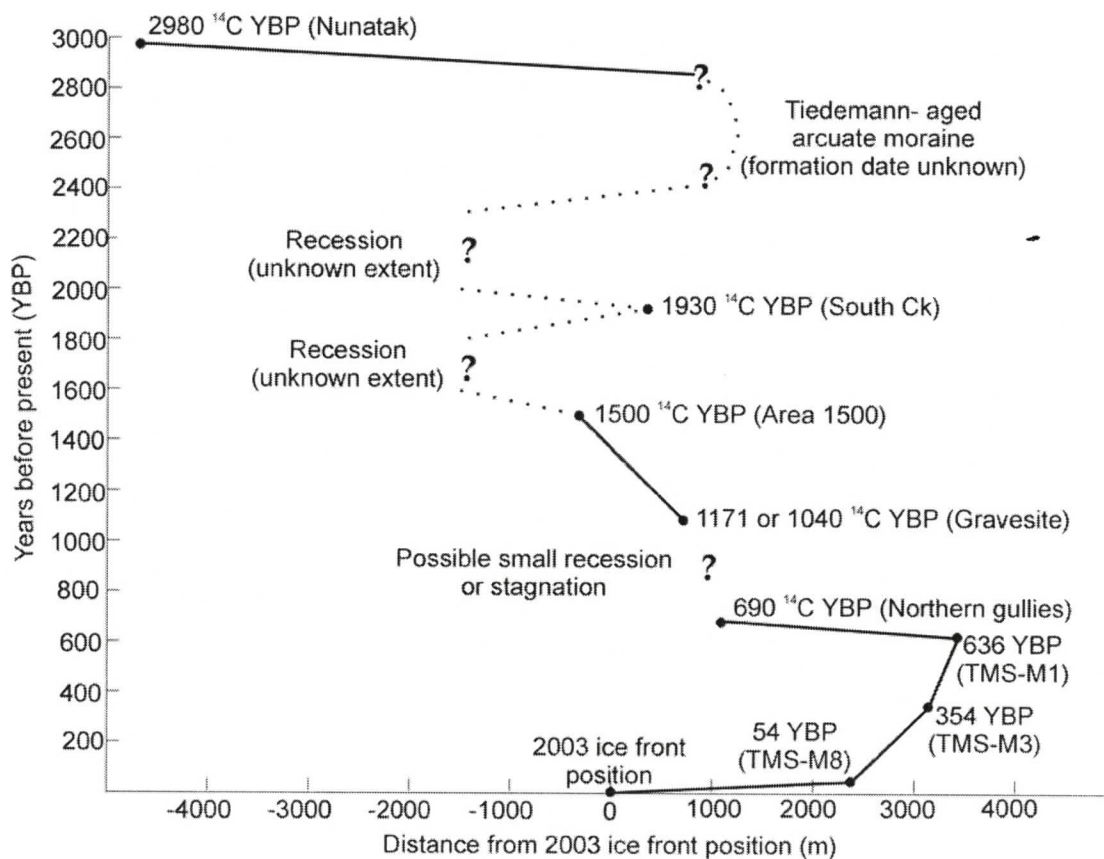


Fig. 3.12. Time-distance schematic of inferred Bridge Glacier ice front position during the late Holocene. Distance on the x-axis is measured in metres relative to the observed 2003 ice front position; negative values indicate upvalley (west) locations, and positive values indicate downvalley (east) locations.

3.6.1 Tiedemann Advance (ca. 3000 years BP)

Evidence for a Tiedemann-age advance of Bridge Glacier comes from the dendroglaciological analysis of five subalpine fir boles located within the first nunatak gully. Crossdating indicates that all of the trees were killed within a 20-year period ca.

2980 ± 60 ¹⁴C years BP, suggesting that Bridge Glacier overrode a mature forest growing at ca. 1700 m asl, 100 m above the present glacier surface. Based on their location, it is assumed that the trees were originally rooted in the paleosol found immediately above on the nunatak. Although undated, this deep and weathered buried Orthic Humo-Ferric paleosol has physical and chemical pedogenic attributes that would have required a lengthy time period for formation (Ryder and Thomson 1986).

3.6.2. Unattributed advance (ca. 1900 years BP)

Evidence for an ice-front oscillation of Bridge Glacier at 1930 ± 70 ¹⁴C years BP is derived from a buried bole fragment at South Creek (BG03-02; Table 3.5). This sample was found lying on the surface of a buried paleosol beneath a sandy till. Although Ryder and Thomson (1986) originally attributed glacial activity in this period to the Tiedemann Advance, there is dendroglaciological evidence from elsewhere in the Coast Mountains indicating that a distinct advance occurred from 2200 to 1900 ¹⁴C years BP (Laxton 2005; Koch 2006).

3.6.3. Bridge Advance (FMA, ca. 1500-1040 years BP)

There is unequivocal dendroglaciological evidence that Bridge Glacier advanced downvalley into a forest of mature whitebark pine and subalpine fir trees as early as 1500 ¹⁴C years BP. This evidence comprises the rooted and glacially sheared *in situ* stump (BG02-10; Fig. 3.9) in Area 1500 to which 4 detrital wood samples crossdate (Bridge 2; Table 3.4).

Further evidence for glacial activity during the Bridge Advance comes from dendroglaciological evidence collected at the Gravesite location 500 m downvalley of Area 1500 (Figs. 3.1 and 3.8). Consisting of a grouping of buried boles found lying on

the surface of a pre-Bridge Advance till, three of these samples and one South Creek sample crossdated to form the 438-year Gravesite whitebark pine chronology spanning the interval between 1577 to 1140 ^{14}C years BP (Bridge 3; Table 3.4 and Appendix C). Maximum ring counts indicate that this stand was established prior to 1500 ± 50 ^{14}C years BP, and was killed by the downvalley advance of Bridge Glacier 1360 ± 60 to 1140 ^{14}C years BP. This interpretation is supported by the radiocarbon age assigned to BG03-90, which has a perimeter age of 1040 ± 50 ^{14}C years BP (Fig. 3.5). Both BG03-90 and -92 are located within a few metres of each other and must have been killed at same time, with the 100 years difference in age attributable to radiocarbon dating errors and variations in the number of perimeter rings submitted for analysis.

Corresponding evidence for ice accumulation at this time comes from two other sites. BG02-02 at the nunatak site was killed 1190 ± 60 ^{14}C years BP and provides an indication of the minimum vertical extent of the ice surface during the Bridge Advance. Sample BG03-88 from the South Creek site crossdates into the radiocarbon dated Gravesite whitebark pine chronology (Bridge 3; Table 3.4), indicating it was killed at approximately the same time in 1201 ^{14}C years BP. Similarly, undated whitebark pine samples from the Gravesite (BG03-94) and the South Creek (BG03-89) crossdate and have perimeter dates within 19 years of each other (Bridge 8; Table 3.4). As BG03-94 was broken just above root level and lay oriented in the direction of glacial flow, this suggests that both trees were killed by the same advance of Bridge Glacier.

This interpretation of glacial activity during the Bridge Advance provides an opportunity to assess the contemporaneous glaciological behaviour of Bridge Glacier.

Assuming a uniform icefront, Bridge Glacier was advancing downvalley at an average rate of 1.8 m/year from ca. 1500 to 1040 ^{14}C years BP (BG02-10 to BG03-90; Table 3.2).

3.6.4 Early-LIA (ca. 700-500 years BP)

Lichenometric evidence at Bridge Glacier supports the initiation of LIA moraine-building activity during the late 13th to early 14th centuries. This interpretation is supported by dendroglaciological evidence collected in the north lateral moraine gullies and on the crest of the nunatak. Ryder (1991) hypothesized that BG03-64 (680 ± 50 ^{14}C years BP) was killed at least 100 years before Bridge Glacier advanced to its location. Crossdating shows that the kill date for BG03-64 precedes that of BG03-63 (540 ± 50 ^{14}C years BP) by at least 90 years; given the error associated with the radiocarbon dates these results substantiate an early 15th century LIA advance over the nunatak crest.

The dendroglaciological samples collected from the northern gullies provide a chronological perspective on the movement of Bridge Glacier during the LIA. The samples with the most recent kill-dates are located further east (downvalley). Sample BG03-09 (620 ± 60 ^{14}C years BP), located in the easternmost lateral gully, is the only north gully sample that was found *in situ*. Two subalpine fir samples (BG03-53 and -54) found 500 m west and 50 m upslope of BG03-09 crossdate with this sample (Bridge 4; Table 3.4 and Appendix C); the fact that they were killed at the same time as trees located further down-valley (BG02-05, 690 ± 90 ^{14}C years BP) is attributed to the gradual accumulation of glacier ice within the Bridge Glacier valley.

3.6.5. Late-LIA (ca. 500-100 YBP)

Lichenometric evidence from the terminal and lateral moraine complexes indicates that late LIA moraine stabilization events at Bridge Glacier occurred in the mid

15th, early 16th, mid-late 17th, early 18th, mid-late 19th, and early 20th centuries.

Corresponding evidence for late LIA glacial activity comes from detrital wood samples collected within morainal deposits at South Creek (Fig. 3.6). This evidence includes BG03-01 (430 ± 60 ¹⁴C years BP) and samples collected at the same site by Ryder and Thomson (1986), dating to the LIA (Table 3.5).

3.6.6. Twentieth century rates of glacial recession

The earliest available aerial photographs of the Bridge Glacier site were taken in 1964 (Table 3.1). Subsequent aerial photography took place on a bi-decadal basis. Using these photographs and the observed 2003 ice front position, the average rates of historical frontal recession were calculated (Table 3.2). Glacier retreat rates range from 6 m/year (1964-1975) to 125 m/year (1979-1981). During the period between 1975 and 1979, when Bridge Glacier's net mass balance was positive (1976-1977; Mokievsky-Zubok 1989), there was no retreat of the ice front.

3.7 Discussion

Investigations at Bridge Glacier provide evidence for four episodes of late-Holocene glacier activity (Fig. 3.12): a Tiedemann-age advance dating to 2980 ± 60 ¹⁴C years BP, an unattributed advance at 1930 ± 70 ¹⁴C years BP, a First Millennium A.D. advance at 1500 ± 50 ¹⁴C years BP, and the early LIA advances 690 to 600 ± 50 ¹⁴C years BP. These findings complement and add to our understanding of late-Holocene glacier activity in the B.C. Coast Mountains.

There is widespread evidence for a major expansion of glaciers in the B.C. Coast Mountains at ca. 3000 years BP (e.g., Ryder and Thomson 1986; Jackson and Smith 2005). The kill date established for the radiocarbon-dated tree ring chronology ($2980 \pm$

60 ^{14}C years BP) at the nunatak site confirms that Bridge Glacier was also advancing at this time (Bridge 1; Table 3.4). Of particular interest is the similarity between this date and that established for the Tiedemann Advance by Reyes and Clague (2004; 2960 ± 60 ^{14}C years BP) at nearby Lillooet Glacier, which suggests a local synchrony in Tiedemann-age glacier events.

The Tiedemann Advance was originally proposed to describe glacial activity during the interval between 3300 to 1900 ^{14}C years BP (Ryder and Thomson 1986). Although Koch and Clague (2005) have suggested that this interval be expanded to cover the interval from 3900 to 2200 ^{14}C years, no evidence supporting continued glacial expansion over this interval was recovered at Bridge Glacier. The discovery of a distinct event at ca. 1930 ^{14}C years BP at the South Creek site that buried an immature soil horizon suggests that Bridge Glacier retreated some distance upvalley following the Tiedemann Advance. Supporting evidence for this interpretation comes from nearby Gilbert Glacier, where Ryder and Thomson (1986) report that a glacial advance initiated at ca. 2200 ^{14}C years BP had ended by ca. 1900 ^{14}C years BP.

Since the initial reporting of dendroglaciological evidence for an expansion of Bridge Glacier at ca. 1500 ^{14}C years BP (Allen and Smith 2003), there has been a growing body of research confirming the regional nature of the FMA. Recognized at other sites within the southern B.C. Coast Mountains (Larocque and Smith 2003; Koch et al. 2004a, 2004b; Reyes and Clague 2004), correlative events have recently been reported at sites throughout the coastal cordillera of northwestern North America (Reyes et al. 2006). The findings at Bridge Glacier are distinct, however, as they are interpreted to indicate that Bridge Glacier was still advancing downvalley at the Gravesite some 400

years later at 1040 ± 50 ^{14}C years BP (Table 3.5). Supporting evidence for continued ice expansion of Bridge Glacier at this time is provided by Ryder and Thomson (1986) who report that an ice-impounded lake formed in an adjacent tributary valley at 1115 ± 40 ^{14}C years BP. Given that Bridge Glacier was advancing downvalley from this point within 300 years, it may be that the glacial expansion during the Bridge Advance was an influential factor in the significant downvalley extension of glacier ice during the subsequent LIA.

Evidence for the initiation of the LIA glacial activity in the B.C. Coast Mountains ranges from as early as the 10th century (Koch and Clague 2005), but is more commonly reported as occurring in the 11th and 12th centuries (Ryder and Thomson 1986; Desloges and Ryder 1990; Larocque and Smith 2004). The findings of this study confirm those of Ryder and Thomson (1986) and Ryder (1991) who report that the maximum downvalley extension of Bridge Glacier occurred during the early LIA from 685 ± 60 to 530 ± 65 ^{14}C years BP. This interpretation is substantiated by the lichenometric ages assigned to the two outermost terminal moraines (1367 and 1409 A.D.; Table 3.7) at Bridge Glacier, and by the discovery of *in situ* dendroglaciological samples from within the north lateral moraine gullies dating from 595 to 680 ± 60 ^{14}C years BP (Bridge 4; Table 3.4). The lichenometric evidence shows that the earliest terminal moraine stabilization at Bridge Glacier occurred prior to 1367 A.D., indicating that the LIA at Bridge Glacier was initiated prior to this date. Following this event a series of increasingly proximal end moraines stabilized in each subsequent century (Table 3.7), a discovery consistent with that of Larocque and Smith (2003). The most commonly identified periods of lateral moraine stabilization were those in the 17th, 19th, and 20th centuries (Table 3.7). These

findings are similar to those reported from elsewhere in the B.C. Coast Mountains (Lewis and Smith 2004; Larocque and Smith 2003) and in the cordillera of southern Alaska (Wiles and Calkin 1994; Wiles et al. 2002).

The LIA ice front oscillations of Bridge Glacier obliterated or buried much of the geomorphologic evidence for pre-LIA glacial activity. Although evidence for glacial advances dating to 2980 ± 60 , 1930 ± 70 , and 1500 ± 50 ^{14}C years BP was discovered, only the arcuate morainal mound found abutting the northern shore of Bridge Lake provides an indication of the downvalley extent of these earlier advances (Fig. 3.12). Believed to be the remains of a terminal moraine associated with the Tiedemann Advance of Bridge Glacier, the discovery 500 m upvalley of an 86 year-old glacially-sheared stump dated to 1500 ± 50 ^{14}C years BP suggests that following this advance there was a period of substantial glacial retreat. The discovery of 400-year old boles dating to 1040 ± 50 ^{14}C years BP rooted within a buried paleosol suggests that Bridge Glacier did not extend this far downvalley for almost another 2000 years after the Tiedemann Advance.

Whether the nunatak area remained ice-free during the earlier Holocene advances is unknown. If the Bridge Advance persisted until ca. 1040 ^{14}C years BP, as suggested by dendroglaciological evidence at the Gravesite, then there was little time for ice on the summit of the nunatak to recede the 100 vertical m required to allow the germination of sample BG03-64 ca. 900 ^{14}C years BP. This may indicate that the summit of the nunatak remained ice-free during the Bridge Advance. Historic air photos of Bridge Glacier and the valley trimline elevation indicate that the nunatak was ice-covered during the LIA.

3.8 Conclusions

This paper demonstrates how complementary geobotanical dating techniques were used to reconstruct the late Holocene history of Bridge Glacier. Four distinct glacial advances of Bridge Glacier were successfully identified through the application of dendroglaciology and lichenometry. These findings highlight glacial expansion at 3000 ^{14}C years BP during the Tiedemann Advance and substantiate the existence of a period of glacier expansion at ca. 1900 ^{14}C years BP. Whereas the discovery of a glacially-sheared *in situ* stump dating to 1500 ± 50 ^{14}C years BP confirmed that Bridge Glacier was advancing downvalley during the now widely recognized FMA, the discovery of a second site dating to 1040 ± 50 ^{14}C years BP indicates that the glacier continued to advance downvalley for an additional four centuries. Finally, the lichenometric investigations at Bridge Glacier successfully identified eight distinct icefront oscillation and moraine stabilization events dating from the early-LIA to the late-LIA.

These interpretations of the late Holocene history of Bridge Glacier provide a long-term perspective on the extent and character of a glacier within the southern B.C. Coast Mountains. Although the maximum downvalley extent of the Bridge Glacier during the first three late Holocene advances is difficult to reconstruct due to the more extensive LIA advances, the dendroglaciological discoveries suggest that the maximum extent of Bridge Glacier during the Tiedemann Advance is recorded by the arcuate morainal mound situated along the northwestern shore of Bridge Lake. The buried forest found rooted on this Tiedemann-aged surface show that Bridge Glacier advanced at least to this location during the succeeding Bridge Advance.

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Chapter Four: Summary

4.1 Summary

Dendroglaciological and lichenometric analyses have enabled the reconstruction of Bridge Glacier's late Holocene history. During field investigations in July 2002 and 2003 geobotanical evidence was discovered that led to the identification of four distinct episodes of late Holocene glacial expansion.

The earliest episode is a Tiedemann-aged advance recorded by five glacially-overridden trees that died ca. 2980 ^{14}C years BP. This advance was followed by an interval of ice front retreat, forest invasion, and pedogenesis prior to a subsequent, but less extensive, readvance ca. 1930 ^{14}C years BP. Following this unattributed period of glacier expansion, Bridge Glacier once again retreated up valley allowing for the establishment of a mixed forest of subalpine fir and whitebark pine trees. By ca. 1500 ^{14}C years BP Bridge Glacier was readvancing downvalley during the Bridge Advance. This interval of expansion persisted until at least 1040 ± 50 ^{14}C years BP and may have been influential in the significant downvalley extension of glacier ice during the subsequent LIA.

Complimentary dendroglaciological and lichenometric data allowed for the establishment of a detailed chronology of LIA glacial activity at Bridge Glacier. Radiocarbon evidence presented by Ryder (1991) and additional discoveries during this investigation confirm that the LIA expansion of Bridge Glacier was well underway by ca. 650 ^{14}C years BP. As with most glaciers in the Coast Mountains, Bridge Glacier experienced its Holocene glacial maximum during the LIA. Lichenometric analyses

identified moraine stabilization in the late 13th to early 14th, mid 15th, early 16th, mid-late 17th to early 18th, mid-late 19th, and early 20th centuries.

These interpretations of the late-Holocene glacial history of Bridge Glacier provide a long-term perspective of the glacier's extent and character. Based on the discovery of glacially-killed trees at numerous locations at the site, it is suggested that during the Tiedemann Advance the glacier front extended downvalley to build the large arcuate terminal moraine found abutting the northwestern shore of present-day Bridge Lake sometime after 2980 ± 60 ¹⁴C years BP. Following this advance Bridge Glacier retreated and maintained ice front positions upvalley from this point until 1040 ± 50 ¹⁴C years BP, when during the waning stages of the Bridge Advance the moraine was overrun by ice advancing at an average rate of 1.8 m/year. After advancing over the Tiedemann-aged moraine, Bridge Glacier appears to have maintained a terminal position somewhere downvalley of the moraine before advancing downvalley at 1.3 m/year to construct its outermost terminal moraine prior to ca. 1367 A.D.

For most of the LIA, the Bridge Glacier ice front appears to have periodically advanced and then retreated, with each succeeding retreat building an additional end moraine proximal to the most recently deposited moraine. Historical aerial photographs from the past 50 years show that Bridge Glacier has retreated upvalley from its LIA terminal moraine complex at average rates ranging from 0 to 125 m/year. Within the last decade (1993-2005) the rate of frontal retreat has averaged 38 m/year.

This thesis demonstrates how complimentary geobotanical dating techniques can be used to reconstruct the late Holocene history of glaciers found within the Coast Mountains of British Columbia. The findings of this investigation build upon previous

research (e.g., Mathews 1951; Ryder and Thomson 1986; Desloges and Ryder 1990; Clague and Mathewes 1996; Smith and Desloges 2000) and complement the recent discoveries reported by Larocque (2003), Reyes (2003), Laxton (2005), and Koch (2006). A significant contribution provided by this thesis is the discovery in 2002 of the glacially-sheared whitebark pine stump dating to 1500 ± 50 ^{14}C years BP. Providing irrevocable proof for an expansion of Bridge Glacier during a time when nearby Lillooet Glacier was also advancing (Reyes and Clague 2004), this finding helped to catalyze discussions leading to the proposed FMA advance (Reyes et al. 2006).

4.2 Limitations and future research opportunities

1. Considering the wealth of glacially-killed trees discovered during site investigations at Bridge Glacier in 2002 and 2003, it was hoped that a tree-ring chronology could be developed that extended beyond the temporal span of the living chronologies through crossdating with subfossil wood samples. Ultimately, nine floating tree-ring chronologies of varying length (149 to 438 years) were compiled, five of which were assigned relative dates by radiocarbon dating a representative sample. Two additional tree-ring chronologies were developed from living trees: a whitebark pine chronology spanning the period 1808 to 2003 A.D., and a subalpine fir chronology spanning the interval between 1682 to 2003 A.D. (Appendix B).

Despite numerous attempts and iterations, it proved impossible to crossdate any of the floating chronologies with either of the living chronologies. It is believed that most of the subfossil wood collected from the Bridge Glacier site predates the living chronologies, or, if there is overlap, the number of coincident

years is too few to produce reliable crossdating results. Many samples suspected to be of contemporary origin contained too few rings to achieve any significant series intercorrelation results. Future dendrochronological research at this site should focus on collecting tree ring samples from mature and/or historically killed trees. This approach would hopefully bridge the time span between the existent living chronologies and the floating chronologies collected as part of this study.

2. Numerous authors have used tree-ring chronologies to construct proxy climate records (e.g., Sziecz 1995 et al.; Barclay et al. 1999; Watson and Luckman 2004b; Larocque and Smith 2005a) and to reconstruct glacier mass balance histories (e.g., Villalba 1990 et al.; Lewis and Smith 2004; Watson and Luckman 2004a; Larocque and Smith 2005b). Such reconstructions require that significant correlations exist between the independent variable (i.e. climate or mass balance) and tree-ring indices for either ring density or width.

Efforts were made to correlate climate and mass balance data for Bridge Glacier with the ring-width indices of the living tree-ring chronologies. Several interesting findings emerged in my analyses that may be of use to future investigators pursuing the construction of proxy records from tree-ring data in this region. These preliminary analyses are detailed in Appendices D and E.

Glacier mass balance surveys were conducted at Bridge Glacier from 1976 to 1984 (Mokievsky-Zubok 1985). Although a strong correlation to the longer mass balance data set from Place Glacier permitted extension (through regression)

of the Bridge Glacier mass balance data to 36 years, efforts to correlate these data with the two living tree ring chronologies were unsuccessful.

Future dendroclimatological investigations at Bridge Glacier should focus on submitting wood samples for densitometric analysis. Numerous researchers report success in using indices of maximum tree-ring density, early wood width or density, or late wood width or density to correlate with climate or mass balance data (e.g., Briffa et al. 1988; Splechna et al. 2000; Watson and Luckman 2004*b*).

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Appendix A. Lichenometric data used to date moraines at Bridge Glacier

Table A.1. Lichenometric control points collected for Bridge Glacier area. These data were used to assess three existing *Rhizocarpon* spp. growth curves for their applicability to lichen at the Bridge Glacier study site.

Location	Substrate description	Year (A.D.) of surface stabilization	Age of substrate (years)	Largest a-axis diameter (mm)
Goldbridge cemetery	Headstone	1939	64	18.6
Goldbridge cemetery	Headstone	1940	63	17.4
Goldbridge cemetery	Headstone	1942	61	17.1
Goldbridge cemetery	Headstone	1947	56	13.6
Goldbridge cemetery	Headstone	1952	51	16.9
Goldbridge cemetery	Headstone	1953	50	17.5
Goldbridge cemetery	Headstone	1957	46	14.4
Garibaldi cemetery	Headstone	1959	44	17.8
Bridge Glacier	Rock on recessional moraine	1977	26	10.1
Pemberton cemetery	Headstone	1980	23	13.4
Garibaldi cemetery	Headstone	1983	20	9.8

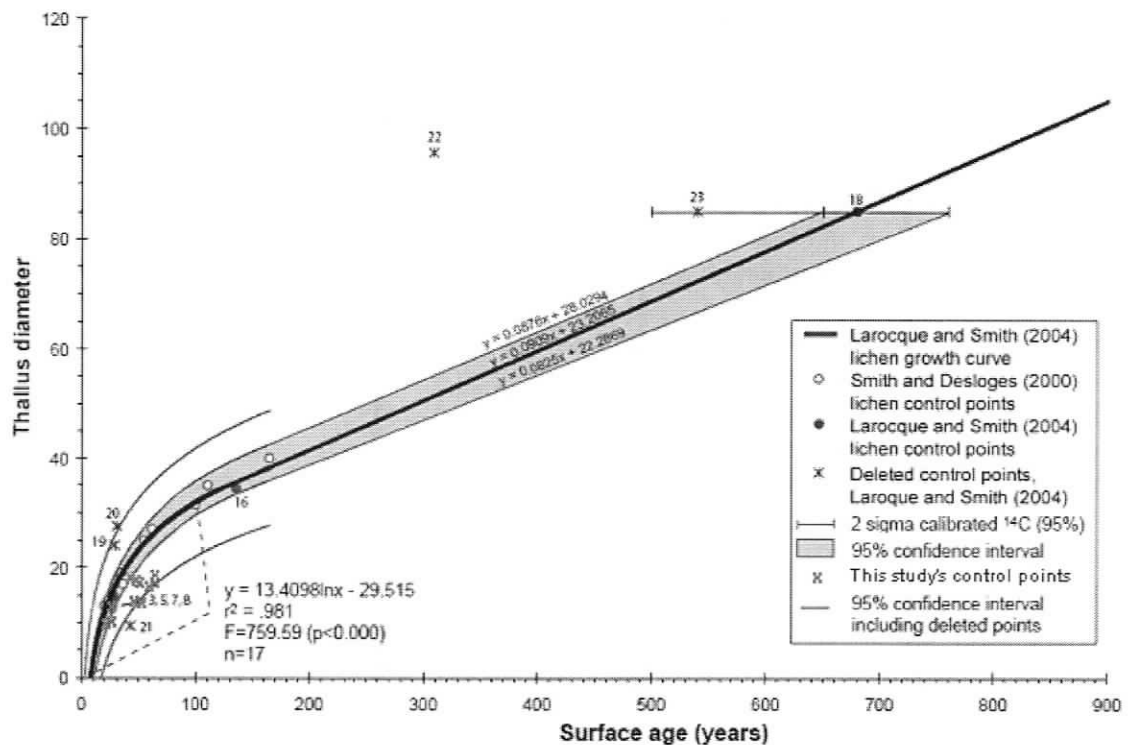


Fig. A.1. Larocque and Smith (2004) calibrated *Rhizocarpon* spp. growth curve showing control points used in this study.

Appendix B. Bridge Glacier living tree ring chronologies

B.1 Tree ring chronology preparation

Two living tree ring chronologies were produced from forest stands at the Bridge Glacier site. A subalpine fir stand was sampled on an east-facing slope in Nasty Valley (*unofficial name*) at 1544 m asl elevation, approximately 3 km south of the 2003 snout of Bridge Glacier ($50^{\circ} 48' 36''$ N, $123^{\circ} 29' 11''$ W, m asl; Fig. B.1). Whitebark pine samples were collected on the north-facing flank of the glacier valley at 1435 m elevation, approximately 2 km east of the 2003 terminus of the glacier ($50^{\circ} 50' 23''$ N, $123^{\circ} 28' 00''$ W; Fig. B.1).

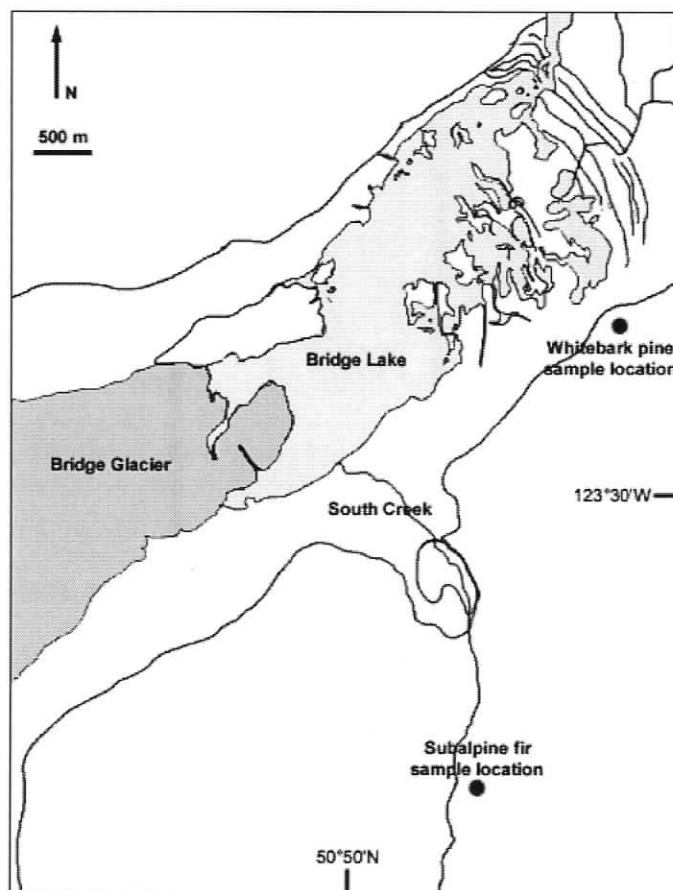


Fig. B.1. Location of stands sampled to construct the Bridge Glacier living tree-ring chronologies.

After air-drying, the increment samples were mounting in slotted boards and sanded to a fine polish (Stokes and Smiley 1968). The WinDendro v.2003b image process software and a flatbed scanner were used to measure annual ring-widths to 0.01 mm (Guay et al. 1992). Tree ring series were visually crossdated with the help of the programs ITRDB Viewer (Varem-Sanders 1996) and CDendro (Larsson 2003). Quality checking of the crossdated series was done using program COFECHA v 6.06 (Holmes 1999). A cubic smoothing spline with a 50% wavelength cutoff of 32 years was used to remove low-frequency variance for quality checking, and the series were examined in 50-year segments lagged successively by 25 years.

Living ring-width series were standardized using the program ARSTAN (Cook and Holmes 1999) to remove growth trends in the series related to tree-age and stand dynamics (Cook et al. 1990), producing indexed chronologies for whitebark pine and subalpine fir. Because individual trees may require different detrending techniques to extract a common climate signal, the interactive function of ARSTAN for Windows was used to determine the most appropriate standardization method (i.e., the method that removes distinct low frequency components while retaining the variation in growth attributable to climate) for each series (Cook and Krusic 2005). Most series in the subalpine fir and whitebark pine chronologies were standardised using a cubic smoothing spline (Cook and Peters 1981) of 50% response frequency of 67% of the series length. Series for which this method was inappropriate were detrended by applying a straight line through the mean ring-width value of the series. Each ring-width measurement was then divided by its respective spline (or “expected”) value to produce a dimensionless index.

In contrast to the double-detrending method commonly applied to standardise *Pinus* species chronologies (*c.f.* Gedalof and Smith 2001), a single detrending method proved sufficiently robust to remove age and stand dynamics related trends from the Bridge Glacier series. The whitebark pine samples were extracted from an open canopy stand, and the samples did not display abrupt differences in the juvenile to mature portions of the tree ring series that are commonly observed in *Pinus* species.

The indexed series were then submitted to autoregressive modelling to remove autocorrelation from the series. Site chronologies were calculated using a biweight robust mean to lessen the relative importance of outliers. Four variations were produced for each species' chronology: a) a raw ring width chronology, calculated by simply averaging the ring widths of all rings for a single year in all series; b) a standardized chronology, calculated from the mean of the detrended series; c) a residual chronology, calculated from the mean of the detrended series after applying autoregressive modelling; and, d) an ARSTAN chronology, for which the autocorrelation common to all series (pooled autoregression) was added to the chronology (Figs. B.2 and B.3).

B.2 Results

The characteristics of the final chronologies are summarized in Table B.1. The subalpine fir chronology is 322 years long, spanning the interval between 1682-2003 A.D. The whitebark pine chronology is 196 years long, spanning the interval between 1808-2003 A.D. The mean series intercorrelations reported for the subalpine fir ($R = 0.617$) and whitebark pine (0.462) chronologies are similar to those developed at other tree-ring sites in the B.C. Coast Mountains (Johnson 2003; Larocque and Smith 2005).

Table B.1. Characteristics of chronologies developed from living trees at Bridge Glacier.

Chronology species ^a	N series	N trees	Series intercorrelation	Total length (years)	Mean length of series (yrs)	Span (years A.D.)	Mean sensitivity
saf	40	20	0.617	322	212.4	1682-2003	0.203
wbp	41	21	0.462	196	157.2	1808-2003	0.170

^asaf = subalpine fir; wbp = whitebark pine

Mean sensitivity is a measure of the relative difference between ring widths in consecutive years (Fritts 1976). Values can range from 0 to 2, with higher values indicating less ring-width complacency in a series, and therefore greater potential for dendroclimatologic utility. Although neither chronology demonstrated high mean sensitivity values, the values are again comparable to those from other chronologies from the B.C. Coast Mountains (Johnson 2003; Larocque and Smith 2005).

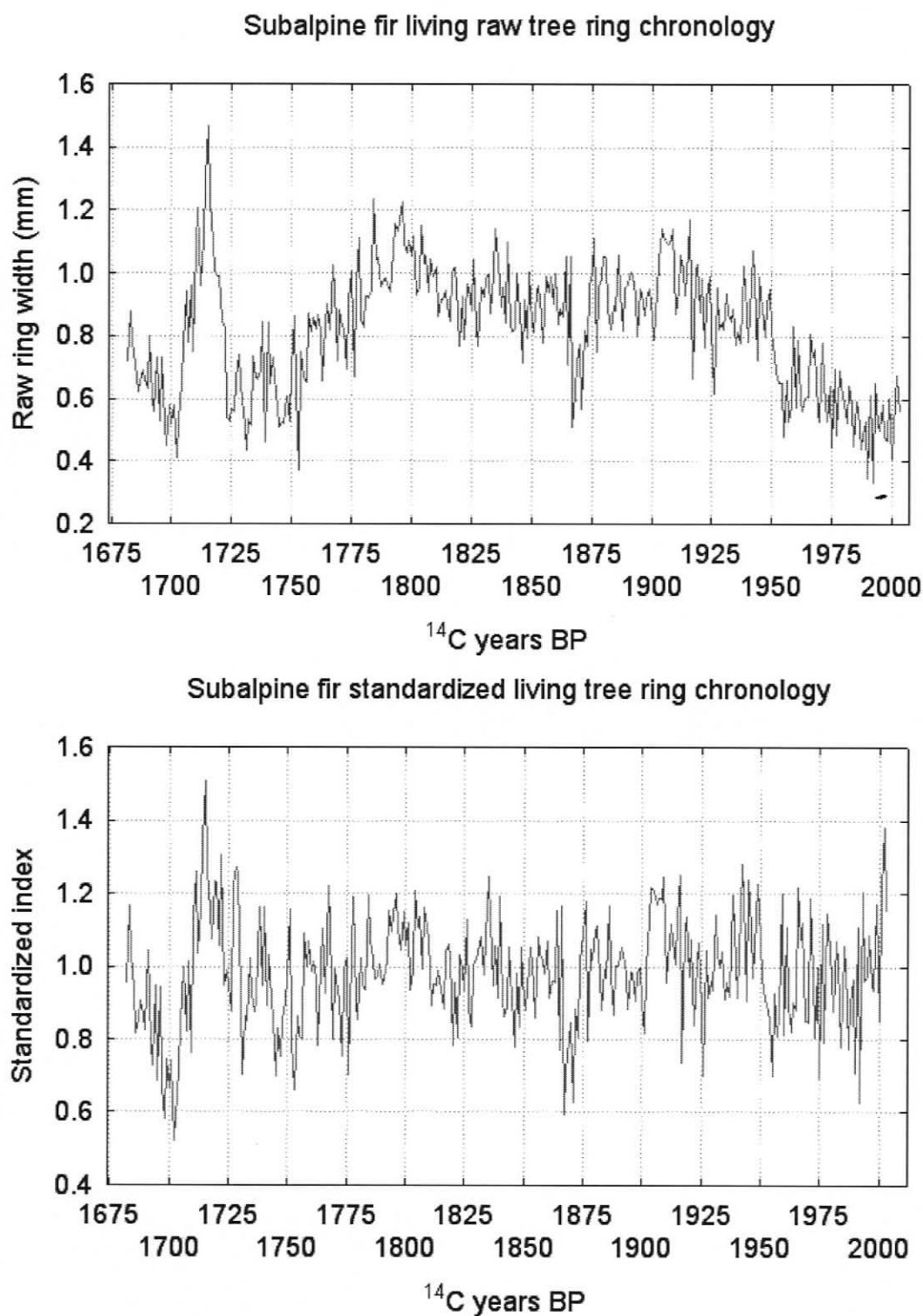


Fig. B.2. Bridge Glacier subalpine fir living tree-ring chronology (raw ring width and standardized chronologies).

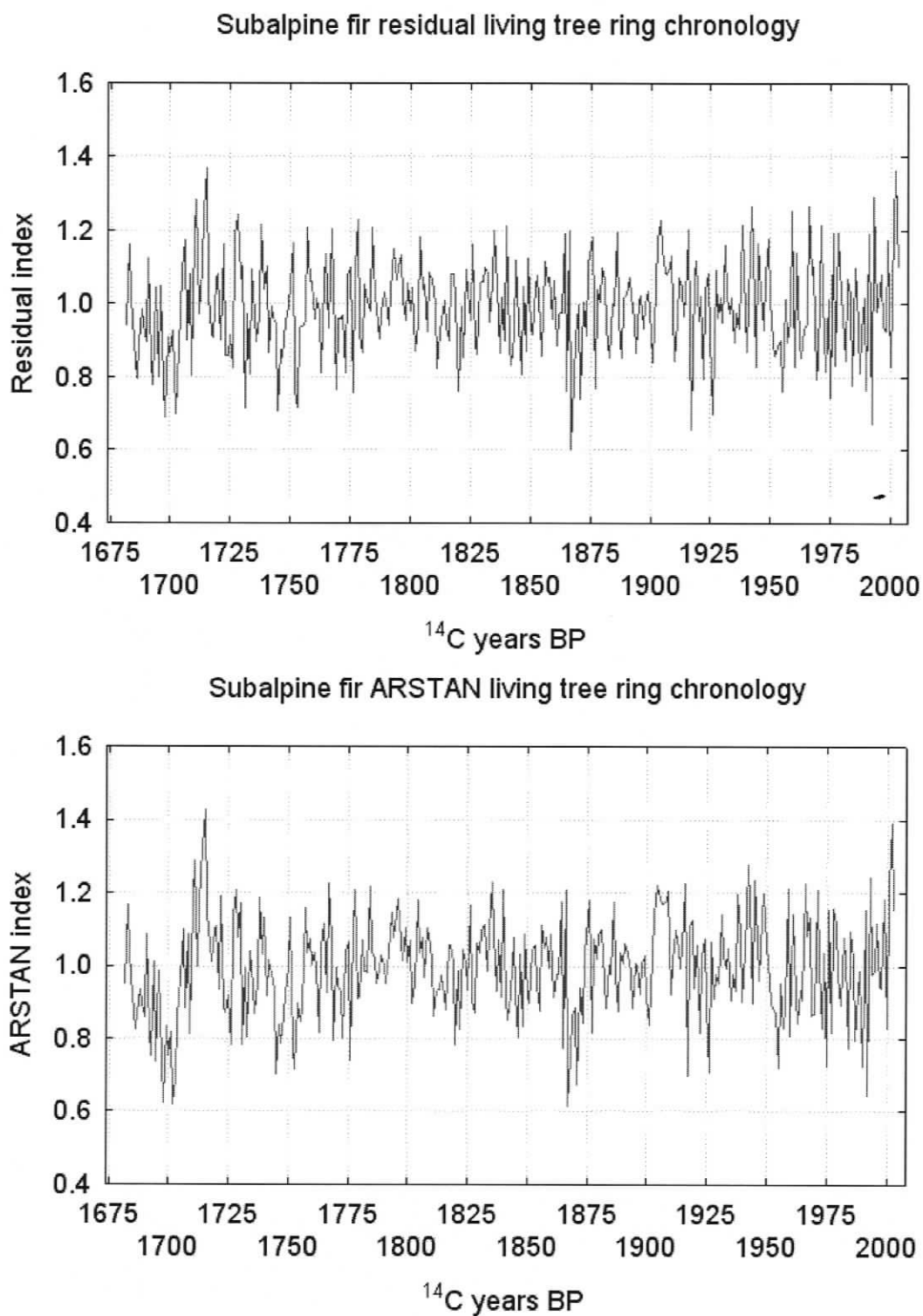


Fig. B.2. continued. Bridge Glacier subalpine fir living tree-ring chronology (residual and ARSTAN chronologies).

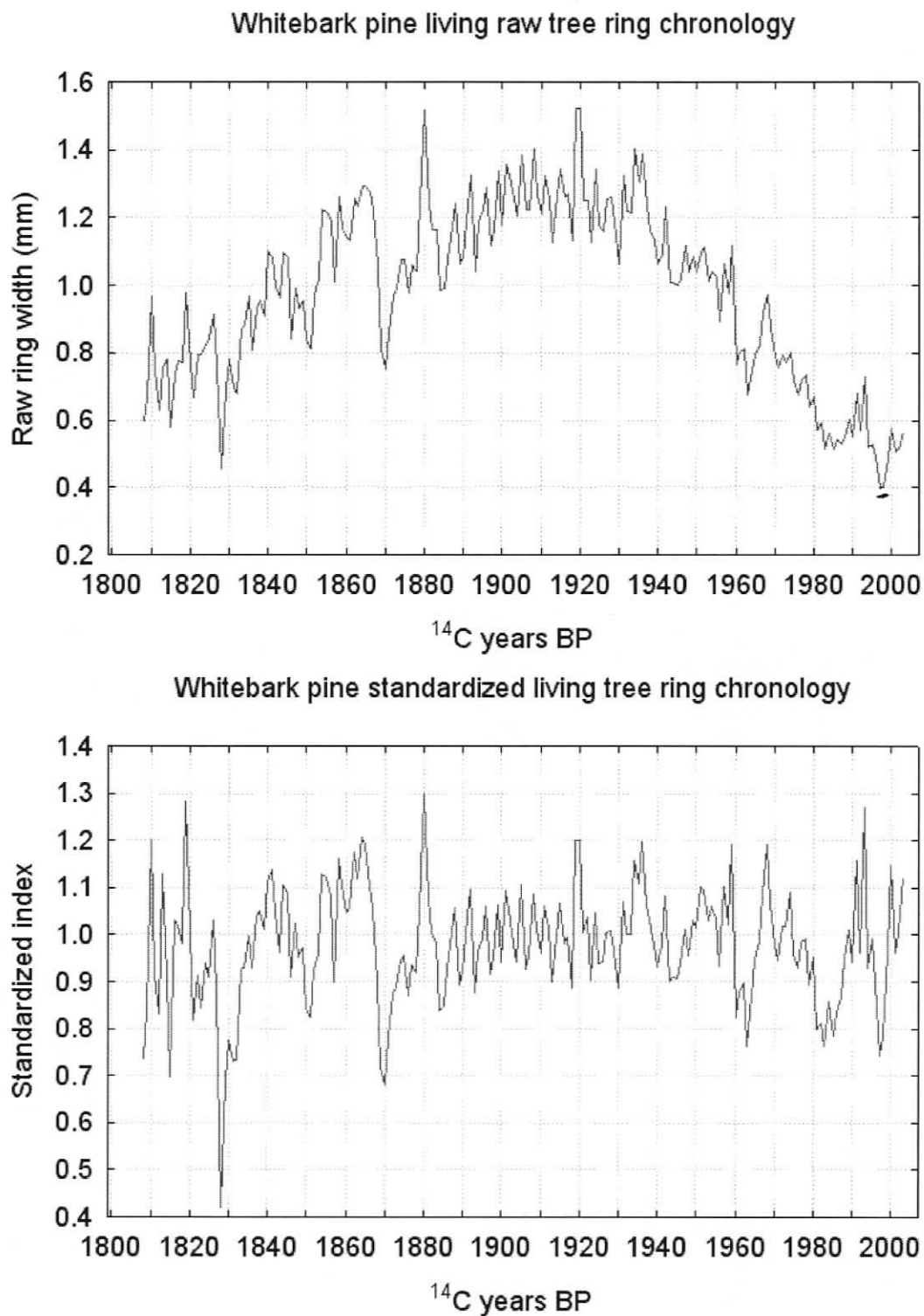


Fig. B.3. Bridge Glacier whitebark pine living tree-ring chronology (raw ring width and standardized chronologies).

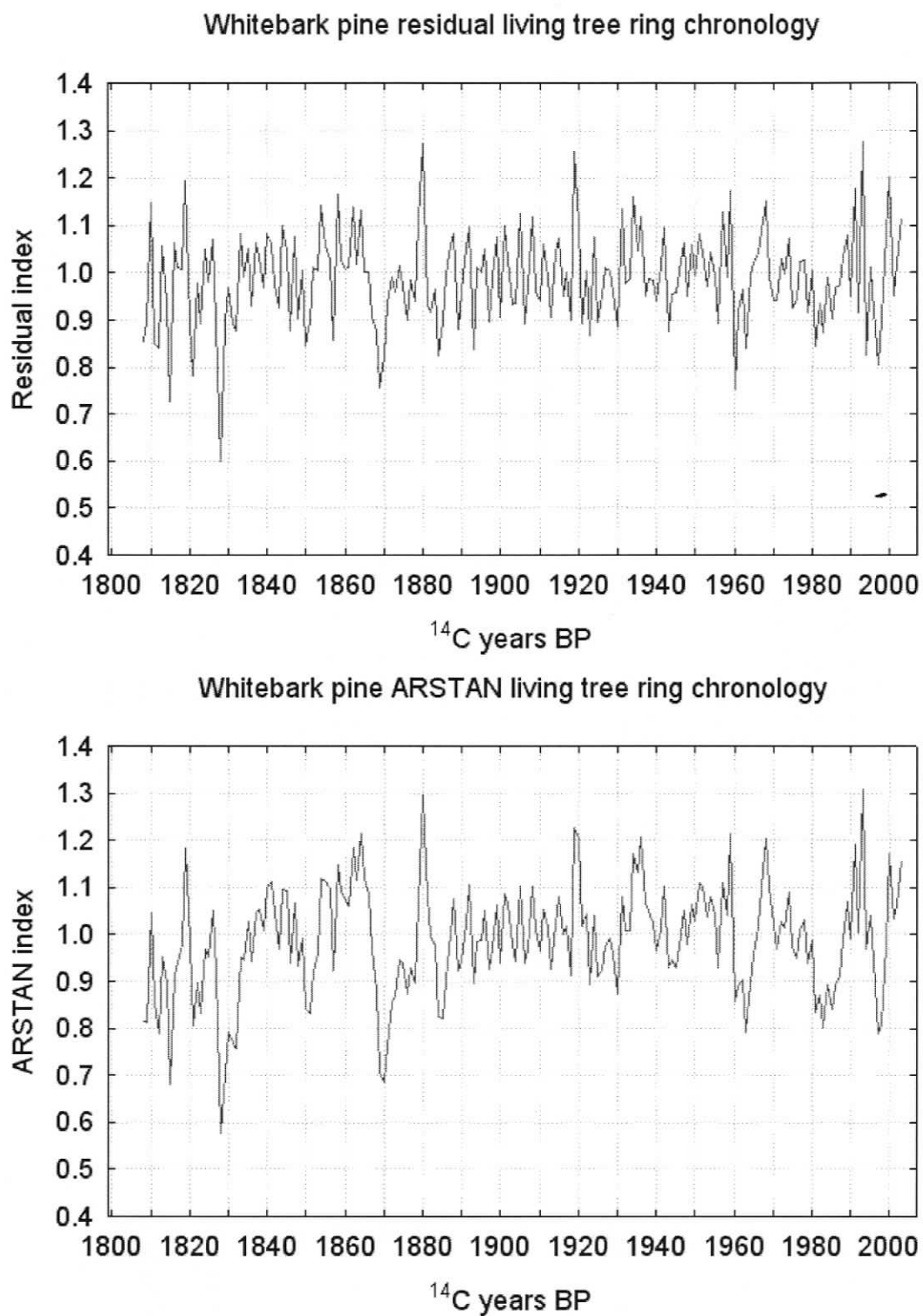


Fig. B.3. continued. Bridge Glacier whitebark pine living tree-ring chronology (residual and ARSTAN chronologies).

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Appendix C. Bridge Glacier subfossil chronologies

Nine tree-ring chronologies (Bridge 1 to 9; Table 3.4) were developed from subfossil wood samples collected at the Bridge Glacier site (Fig. C.1). Sample preparation and crossdating methods are outlined in Chapter 3. Standardization techniques are identical to those described in Appendix A.

Tree ring chronologies Bridge 1, 2, 3, 4, and 5 (Figs. C.2 – C.6) include at least one radiocarbon-dated sample, which allows relative dates to be associated with the chronologies. Chronologies Bridge 6, 7, 8, and 9 are undated floating chronologies.

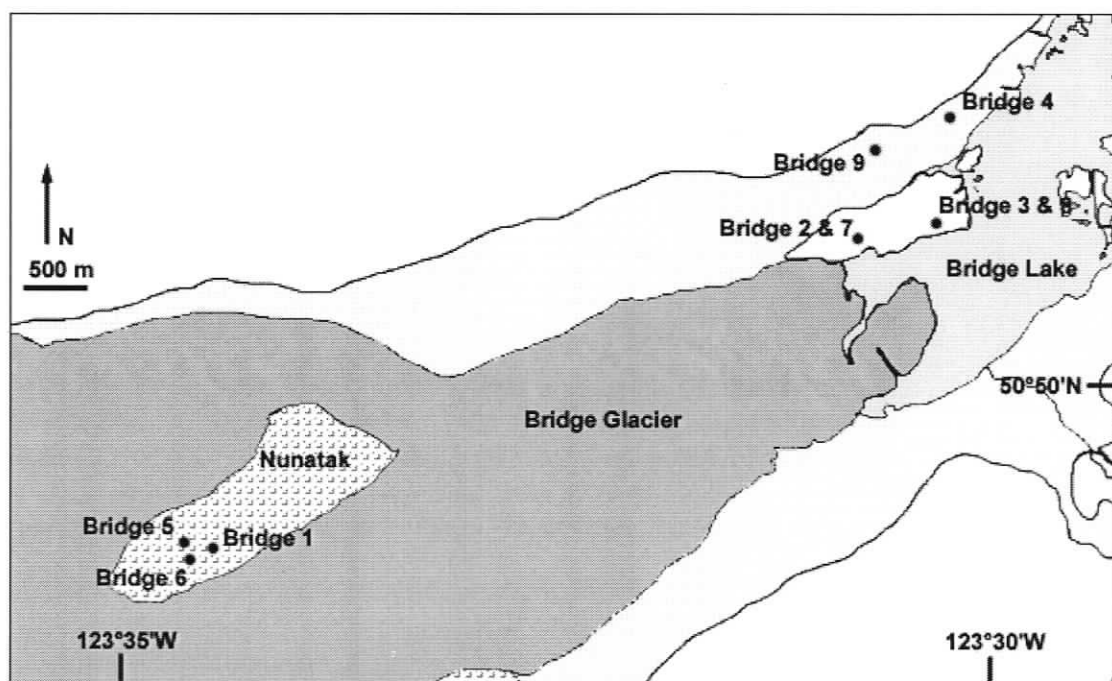


Fig. C.1. Locations of sites for Bridge Glacier subfossil chronologies, Bridge 1 to 9.

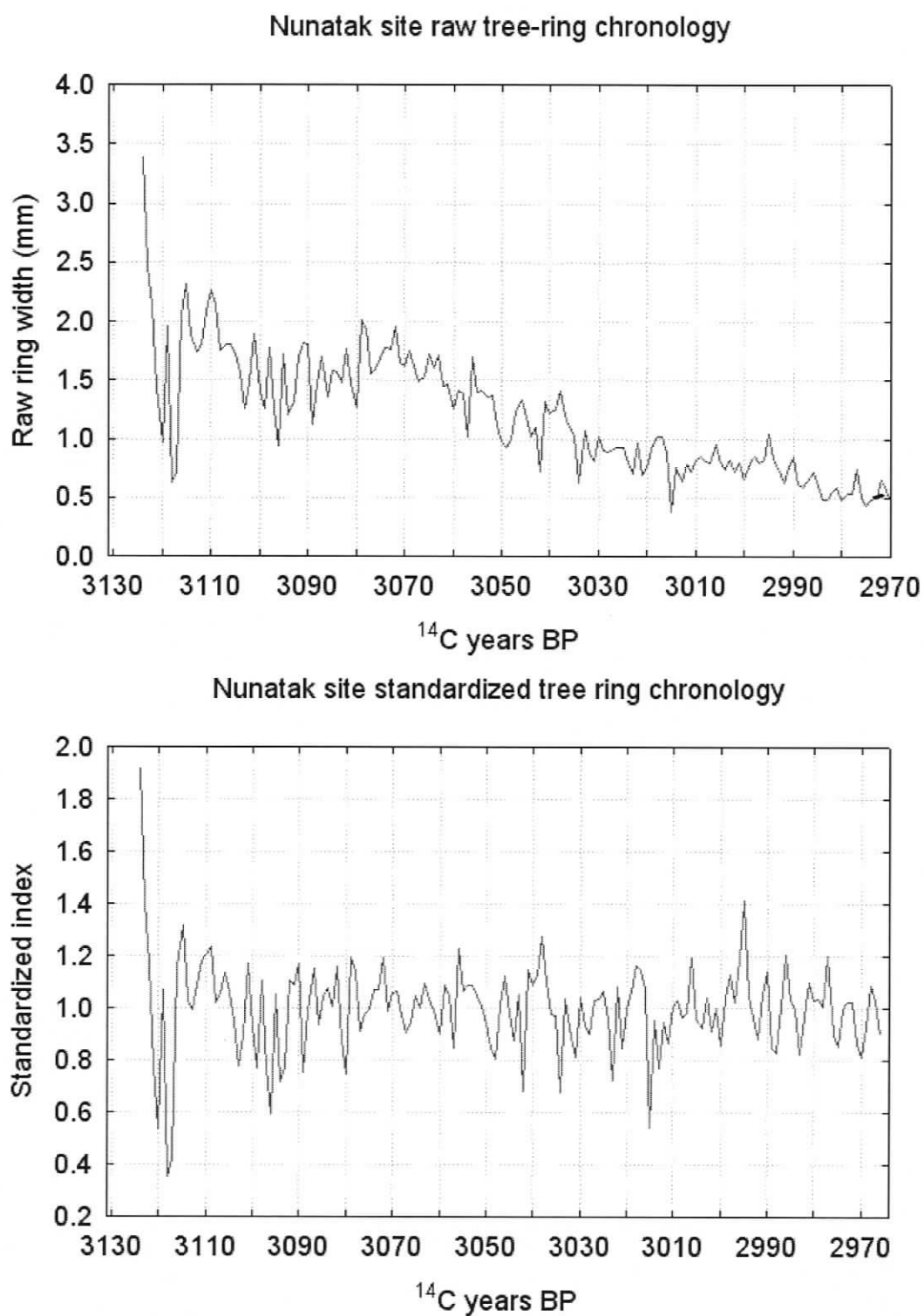


Fig. C.2. Bridge 1 (Nunatak site raw and standardized tree ring chronologies).

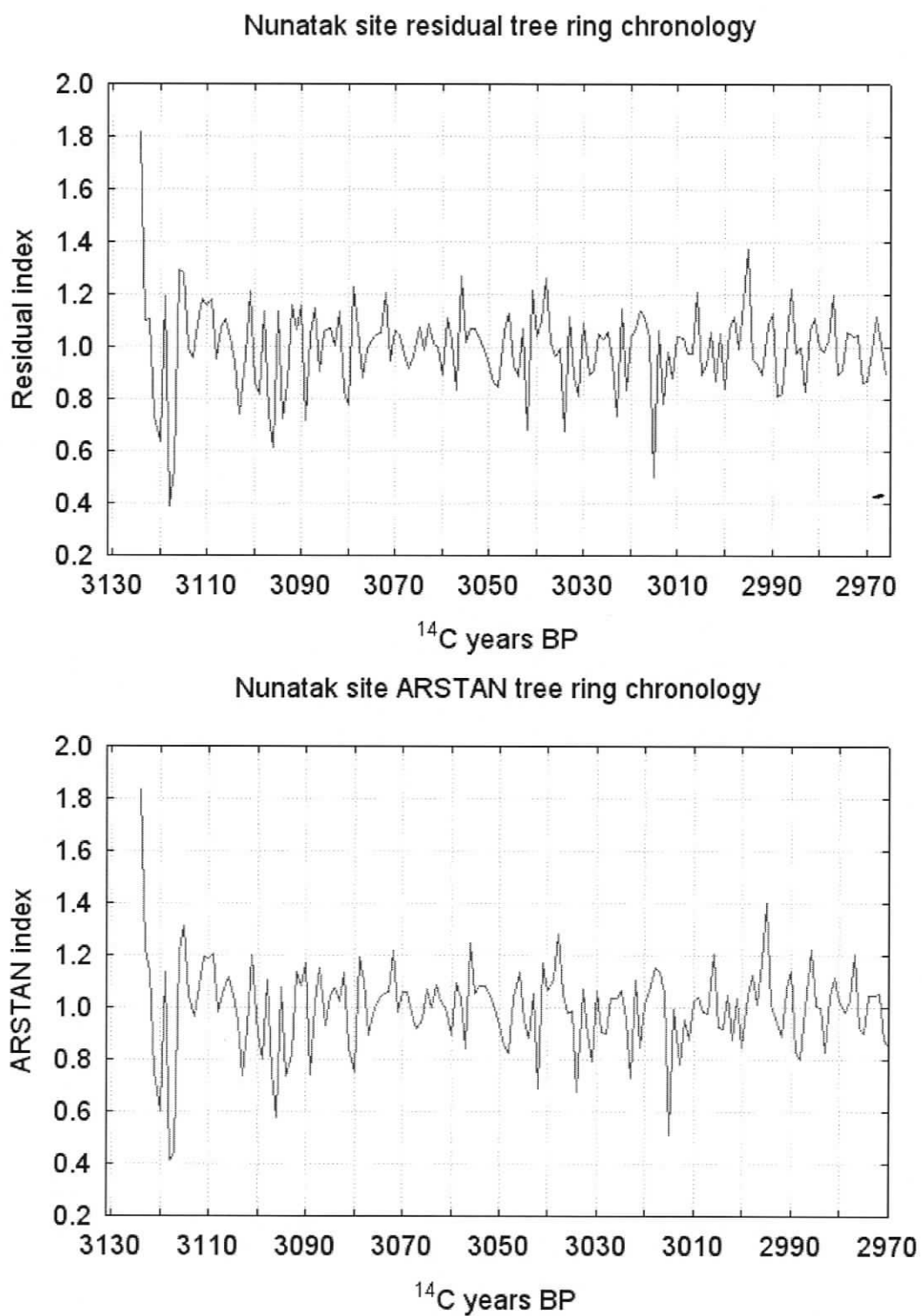


Fig. C.2. continued. Bridge 1 (Nunatak site residual and ARSTAN tree ring chronologies).

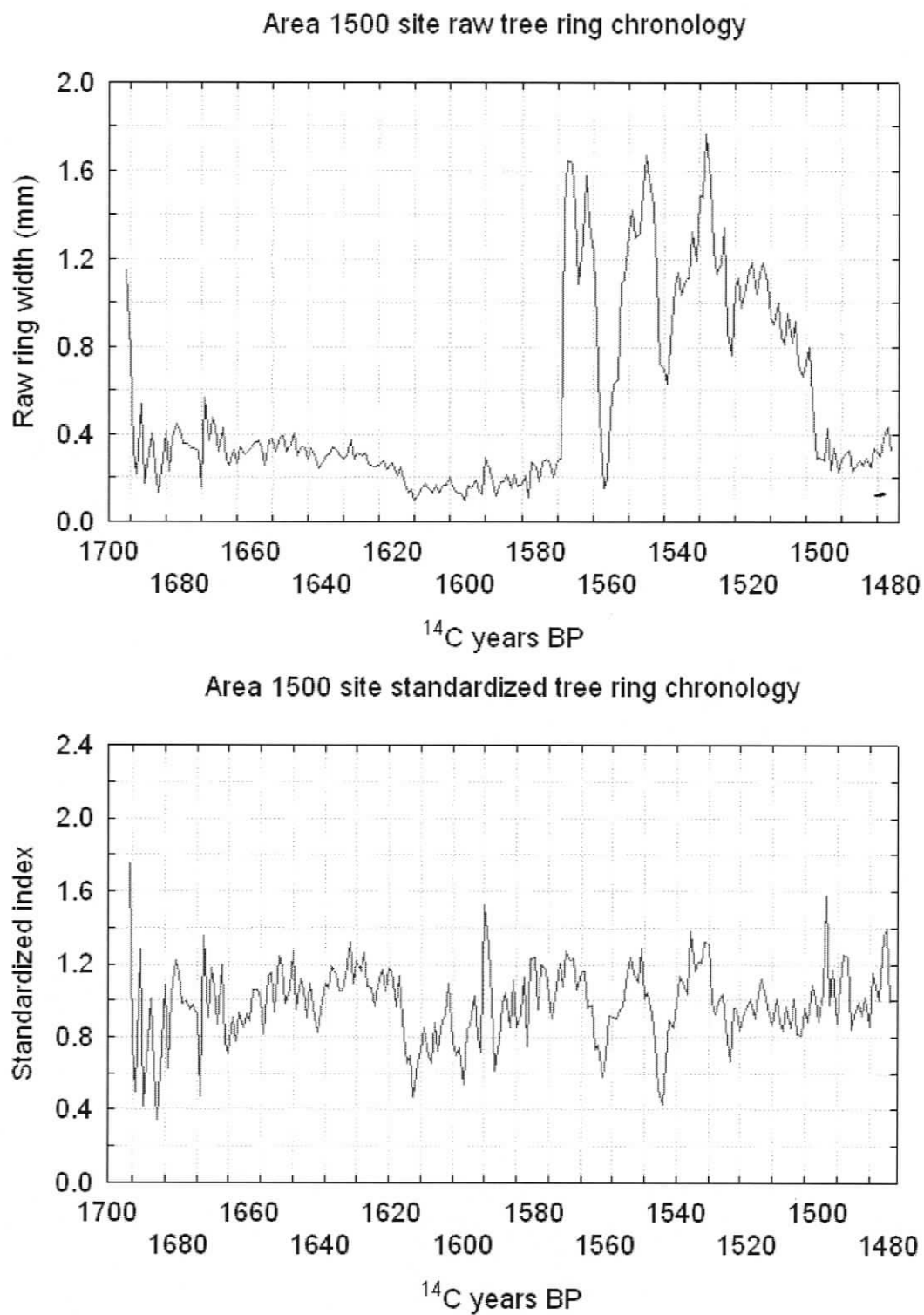


Fig. C.3. Bridge 2 (Area 1500 site raw and standardized tree ring chronologies).

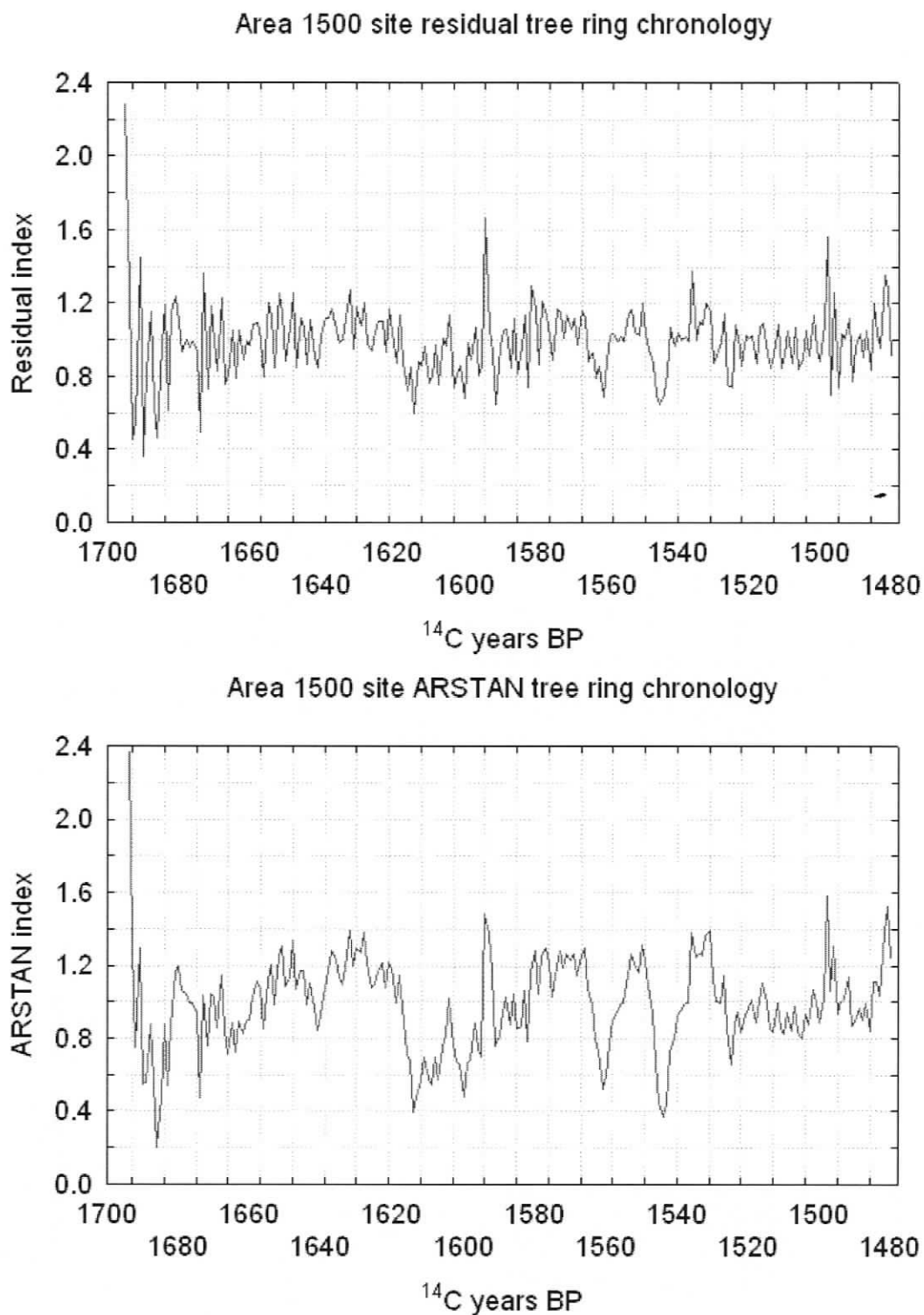


Fig. C.3. continued. Bridge 2 (Area 1500 site residual and ARSTAN tree ring chronologies).

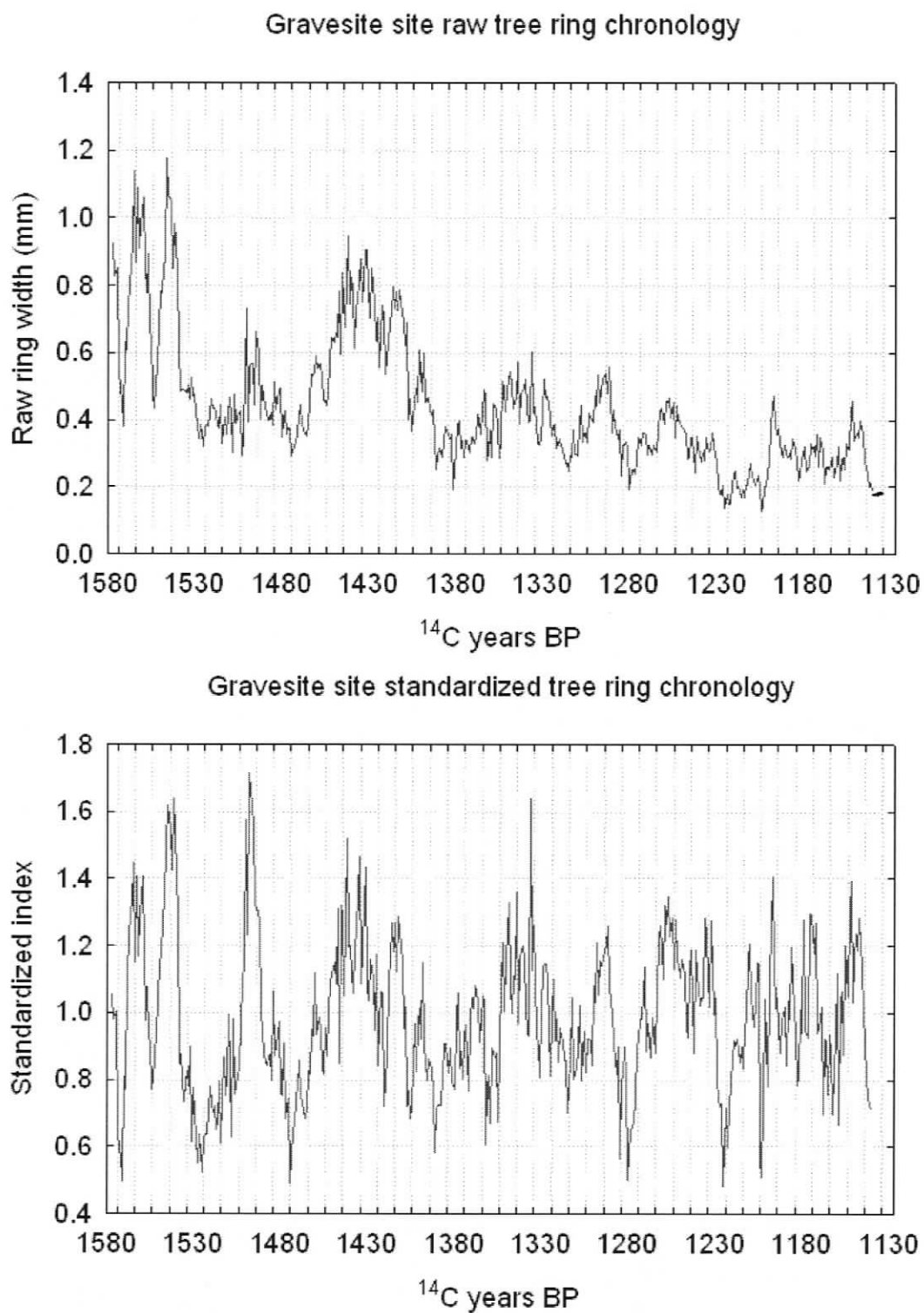


Fig. C.4. Bridge 3 (Gravesite site raw and standardized tree ring chronologies).

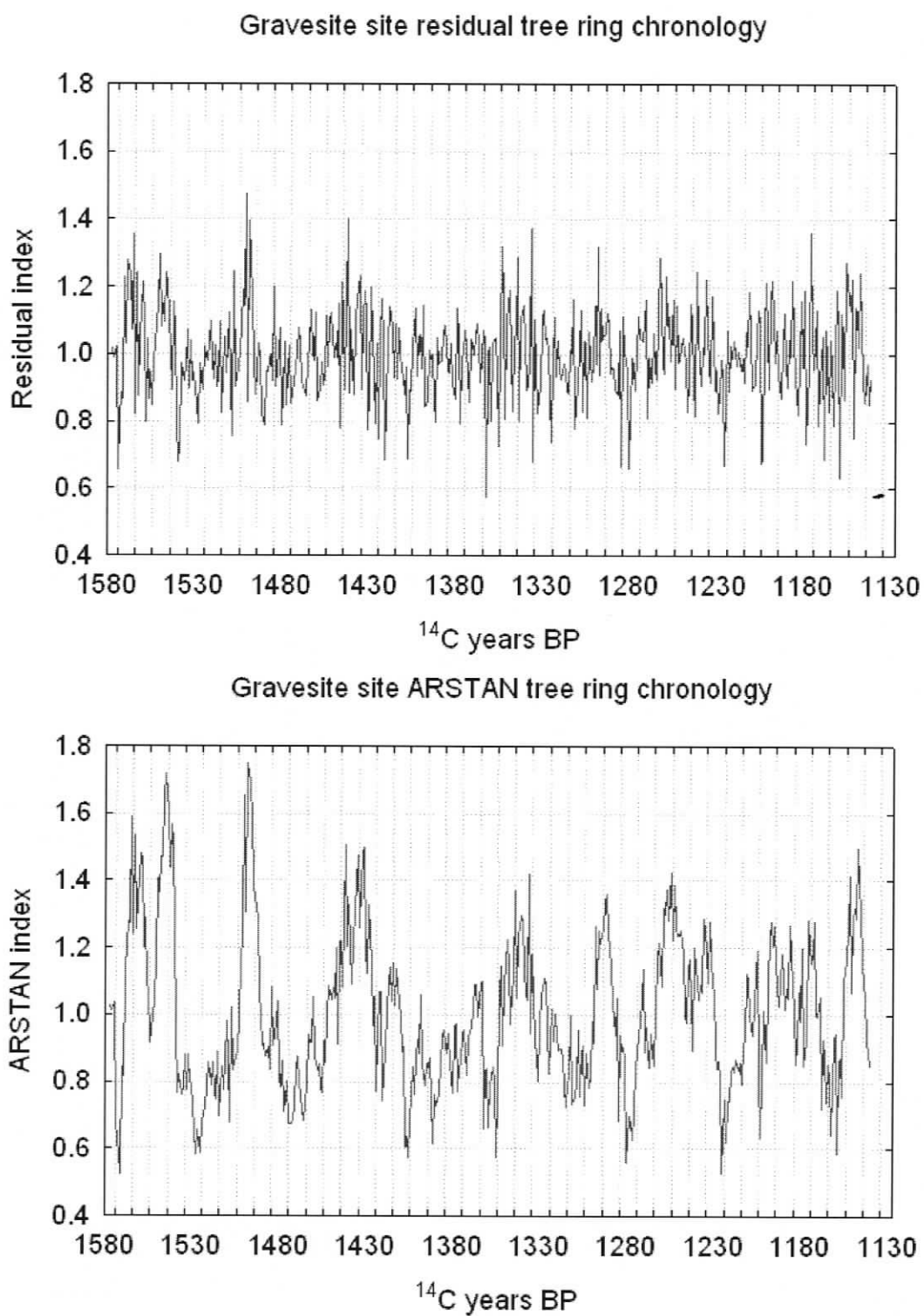


Fig C.4. continued. Bridge 3 (Gravesite site residual and ARSTAN tree ring chronologies).

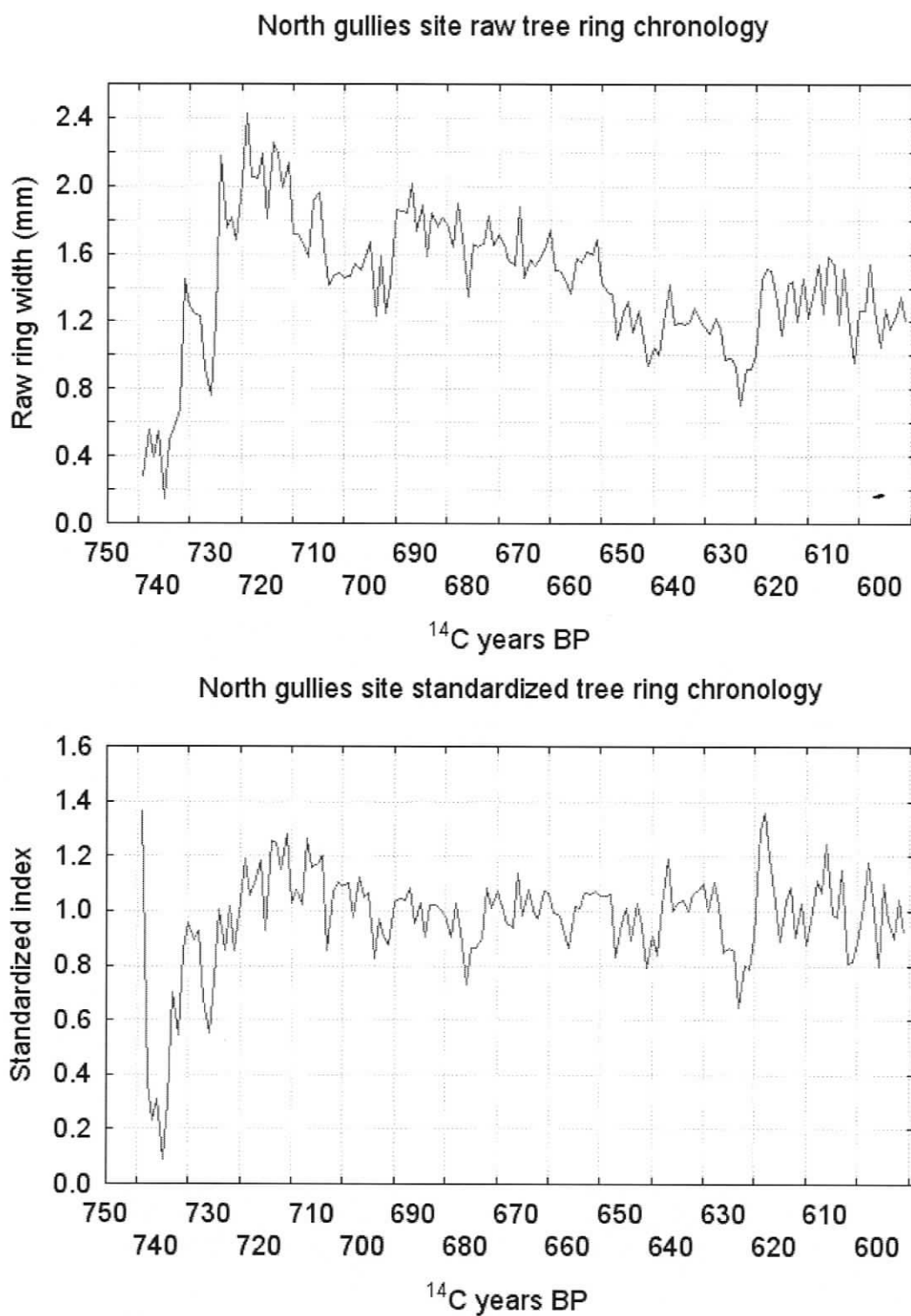


Fig C.5. Bridge 4 (North lateral gullies site raw and standardized tree ring chronologies).

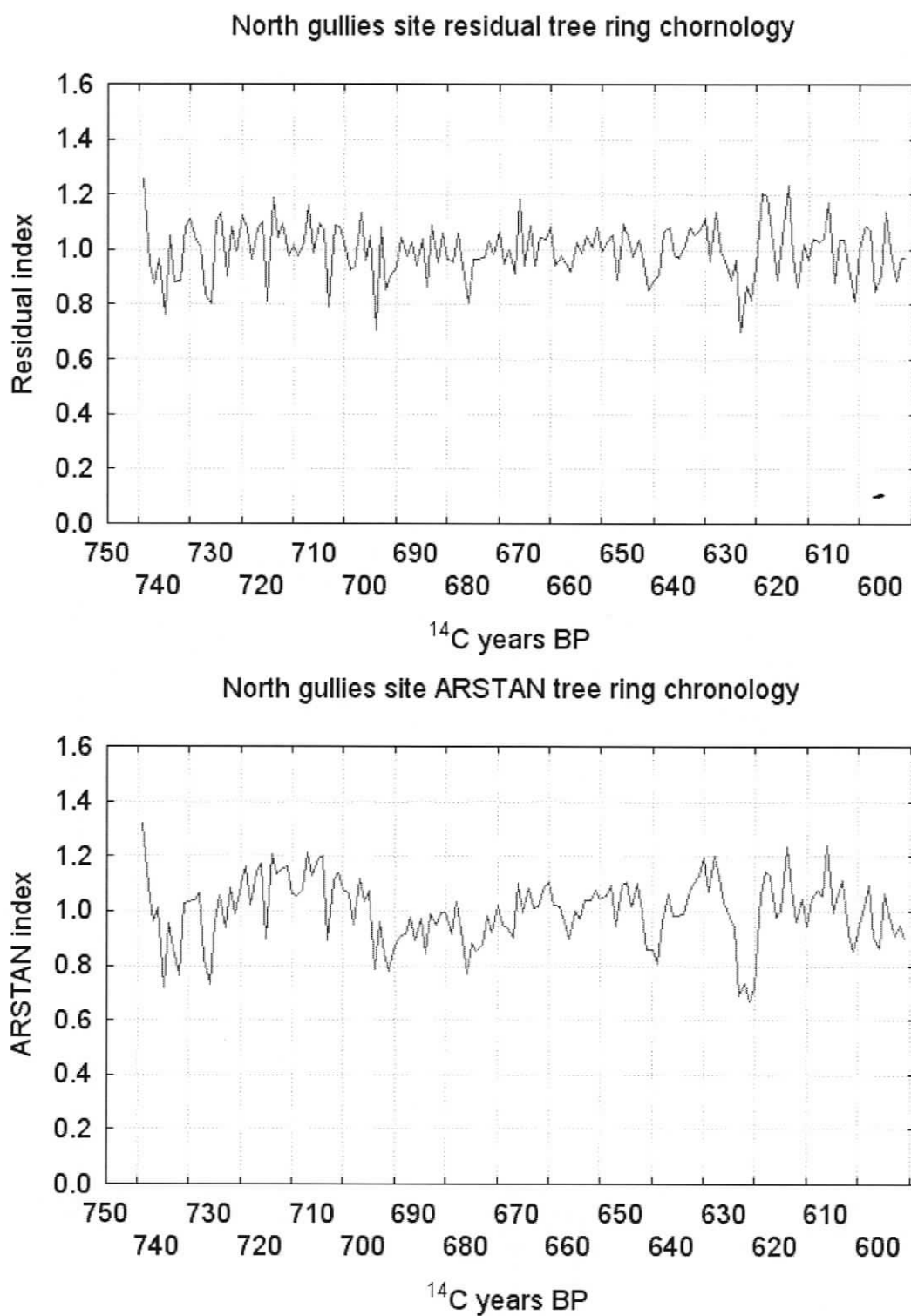


Fig C.5. continued. Bridge 4 (North lateral gullies site residual and ARSTAN tree ring chronologies).

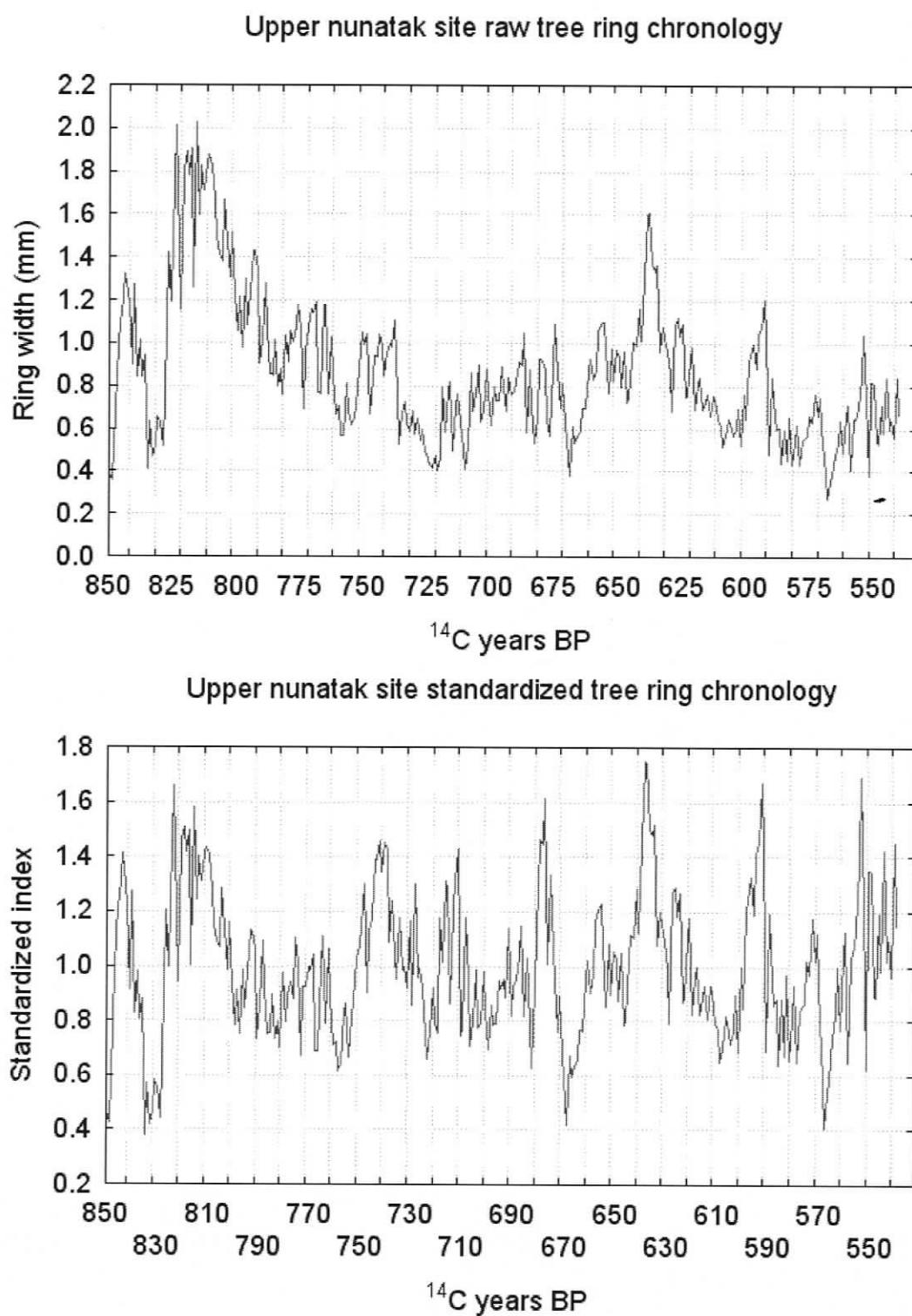


Fig. C.6. Bridge 5 (upper nunatak site raw and standardized tree ring chronologies).

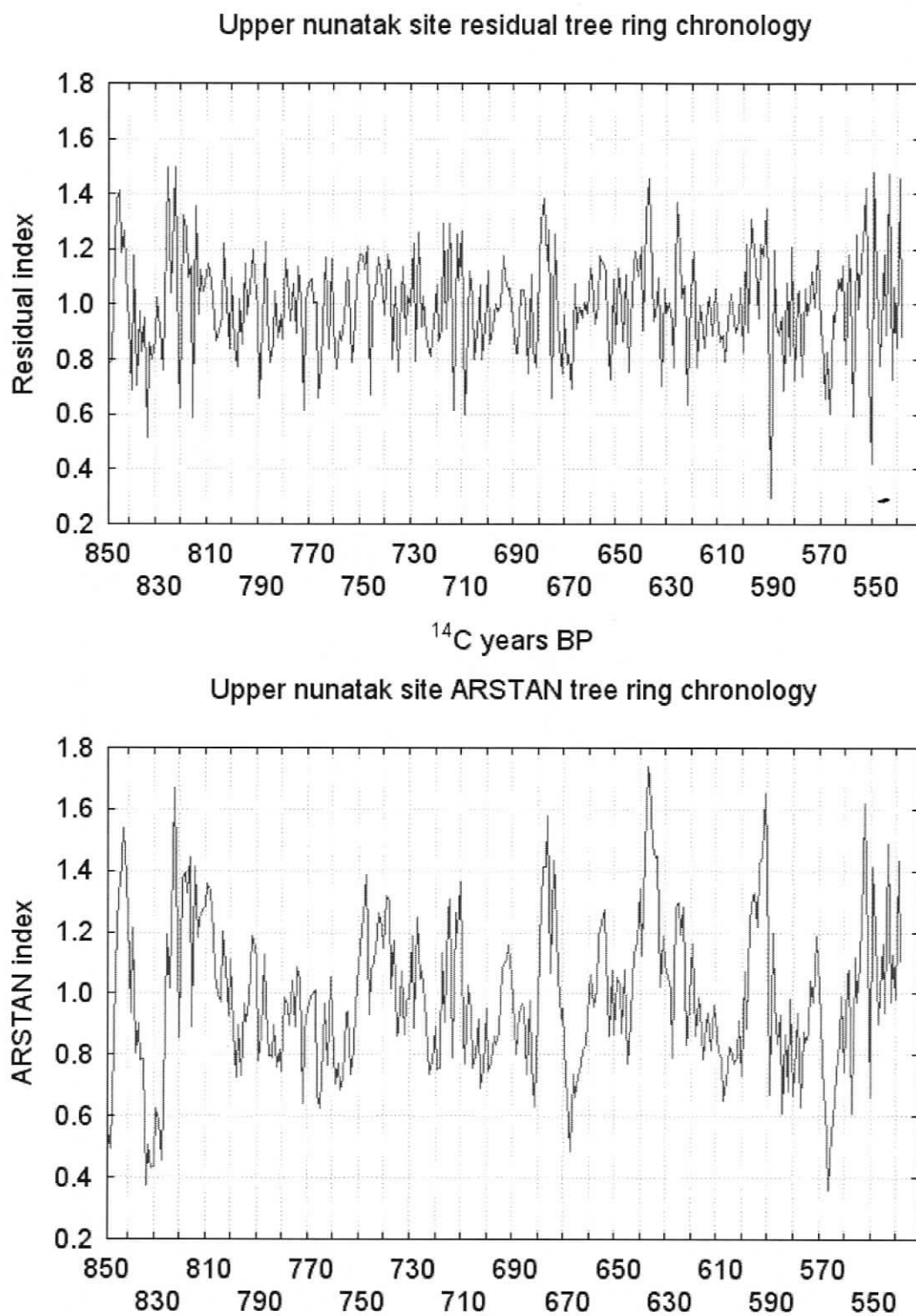


Fig. C.6. continued. Bridge 5 (upper nunatak site residual and ARSTAN tree ring chronologies).

Appendix D. Dendroclimatology at Bridge Glacier

D.1 Introduction

The shortage of long-term instrumental records in western North America has led to a reliance on proxy paleoclimatic reconstructions to understand climatic variability prior to the 20th century. Tree rings have long been recognized as annually resolved indicators of climatic conditions, and their analysis has provided long-term indices of meteorological variables on every continent with the exception of Antarctica. Trees growing at sites on the fringes of their ecological requirements are especially sensitive to variability in climate conditions, and are of great utility in dendroclimatological studies.

In the Coast Mountains of B.C., tree-ring based paleoclimatic studies have been conducted on Vancouver Island (Laroque and Smith 2005), in the Mount Waddington area (Larocque and Smith 2005), the Bella Coola region (Desloges 1987), and the Stewart area (Penrose and Smith 2006). Larger regional dendroclimatological reconstructions have included the Coast Mountain region (Schwiengruber 1988; Briffa et al. 1992; Esper et al. 2002). This appendix outlines my efforts to explore the relationship between climate and the radial growth of subalpine fir and whitebark pine in the vicinity of Bridge Glacier.

D.1.1 Subalpine fir in dendroclimatological studies

Subalpine fir thrives at both cold, high elevation sites (up to 2250 m asl) and at moist lower elevation sites (minimum 600 m asl). It prefers locations where the summers are cool, the winters cold, and the snowpacks deep (Parish and Thomson 1994). Its range extends from mountainous areas of the southern Yukon and southeastern Alaska, to western Alberta and interior British Columbia, to southern Colorado (Henderson 1981).

Subalpine fir trees have been the subject of numerous dendroclimatological studies (e.g., Smith and Desloges 2000; Splechtna et al. 2000; Larocque and Smith 2005). Most researchers report that its climate-growth response is elevation dependent (Ettl and Peterson 1995*a*; Splechtna et al. 2000; Peterson et al. 2002). The ring-widths of subalpine fir trees growing at lower elevation sites show weaker relationships to climate factors than high elevation trees, and are negatively related to May and June air temperatures of the current growing season. In the high elevation trees, ring-widths are positively related to July air temperatures of the current growing season and autumn air temperature of the previous year. High elevation tree ring-widths are negatively related to air temperatures in May of the growth year (likely due to an association with low precipitation) and to the previous year's August air temperature and autumn precipitation.

In high elevation, moist sites in the Olympic Mountains, Washington state, the ring-widths of subalpine fir trees were shown to be negatively related to winter precipitation (Ettl and Peterson 1995*b*) and spring snowpack depth (Peterson et al. 2002). At all elevations the tree ring-widths were positively related to the air temperature of the current growing season. In dry, low and medium elevation sites, tree ring-widths proved to be negatively related to the previous year's summer air temperature, and positively related to summer precipitation in the year of growth. Similar to the findings of Splechtna et al. (2000), Ettl and Peterson (1995*b*) found a negative relationship between tree ring-width growth and August air temperatures in the previous year. Individual trees growing at the same site showed a tremendous variation in their response to climatic factors, and many samples failed to correlate significantly to either air temperature or precipitation (Ettl and Peterson 1995*a*). At a medium elevation, moist site in the Canadian Rockies,

Wood (2002) identified a positive relationship between the annual radial growth of subalpine fir trees and July air temperatures in the year of growth.

D.1.2 Whitebark pine in dendroclimatological studies

Whitebark pine is a hardy, long-lived, high elevation tree species that thrives in cold, moist climates. Despite these characteristics, whitebark pine trees have seldom been used in the western Canadian cordillera for dendroclimatic reconstruction due an inability to discern the climatic variables responsible for its annual radial growth (Luckman and Youngblut 1999; Youngblut 1999; Larocque and Smith 2005). Luckman and Youngblut (1999) report that whitebark pine samples from the Canadian Rockies are long-lived and cross-date with great success, but their efforts to identify the climatic variables controlling the annual radial growth were unsuccessful.

By contrast, Peterson et al. (1990) report that whitebark pine trees growing in the Sierra Nevada forests of California demonstrate a strong positive relationship with the current year's spring precipitation and air temperature, and negative relationships with summer precipitation. Other significant relationships vary depending on the age of the sample, and none of the relationships explain more than 40% of the variance (i.e. the relationships are relatively weak). Similar radial-growth relationships exist in the Sawtooth Salmon River region of Idaho where Perkins (1995) determined that the annual radial growth of whitebark pine was positively correlated with winter and spring precipitation, and negatively correlated with the current year's April temperature. Further investigations by Perkins and Swetnam (1996) revealed that whitebark pine trees in this area also have an inverse relationship with May temperature. In their study of whitebark pine in Idaho, Biondi et al. (1999) discovered a strong positive relationship to July air

temperatures, which contributed to the development of an 858-year long proxy regional temperature record.

D.2 Methods

To reveal any inherent relationships between local climate and radial growth, correlation analyses of several meteorological data sets (see below) and raw, standardized, residual, and ARSTAN tree ring chronologies were run. The climate variables used in these analyses were monthly mean temperature and total monthly precipitation. The data were analysed by hydrologic year (i.e. October of the previous year to September of the current year) to reflect an entire growing year (*c.f.* Ettl and Peterson 1995a). Because tree ring growth is commonly related to climatic aspects of the previous year's growing season (Fritts 1976), analyses were also executed with a one-year lag for the dependent variables.

The International Dendrochronology Program Library program PRECON v5.17b (Fritts et al. 1991) was used to perform response function analyses. Response functions are used to identify significant relationships between climatic variables (i.e. mean monthly temperature and total monthly precipitation and annual radial growth). A 14-month window was used, beginning with the September of the previous year and ending with the October of the current year. Analyses were run for year t and $t-1$.

D.2.1 Meteorological data for the Bridge Glacier site

Data from several meteorological stations were used to compile a long-term local data set for temperature and precipitation to be used in a correlation analyses with tree ring growth (Table D.1). In addition to this reconstruction of local temperature (BRrt)

and precipitation (BRrp) for Bridge Glacier, data from the Canada Gridded Climate Data 1961-1990 (CGCD; Hopkinson 2000) were also included in the analyses.

Table D.1. Meteorological station location and duration of climate data used for dendroclimatic analyses

Meteorological station name	Latitude (N)	Longitude (W)	Elevation (m)	Variable	Years used in analysis
Agassiz	49° 15'	121° 46'	15	Temperature	1889-2002
				Precipitation	1889-2002
Pemberton Meadows	50° 27'	122° 56'	223	Temperature	1912-1967
				Precipitation	1912-1967
Pemberton BC Forest Service	50° 19'	122° 49'	218	Temperature	1969-1984
				Precipitation	1969-1984
Pemberton Airport	50° 18'	122° 44'	204	Temperature	1984-2000
				Precipitation	1984-2000
Water Survey Canada gauging station no. 08ME023	50° 51'	123° 27'	1377	Temperature	1981-2005
				Precipitation	1984-2005
Canada Gridded Climate Data cell 1246	50° 41'	123° 14'	n/a	Temperature	1961-1990
				Precipitation	1961-1990

D.3 Results

Tables D.2 to D.9 show the results of the correlation analyses. The following significant relationships were consistently observed with temperature:

BRrt data for year t:

- Whitebark pine standardized and residual chronologies positively related to May temperature of current growing season

BRrt data for year t-1:

- Subalpine fir raw, standardized, residual, and ARSTAN chronologies negatively related to August temperature of previous growing season

- Whitebark pine raw, standardized, residual, and ARSTAN chronologies positively related to June temperature of previous growing season

CGCD for year t:

- Whitebark pine raw, standardized, and ARSTAN chronologies negatively related to January temperature

CGCD for year t-1:

- Subalpine fir standardized, residual, and ARSTAN chronologies positively related to July temperature of previous growing season
- Whitebark pine standardized, residual, and ARSTAN chronologies positively related to September temperature of previous growing season

The following significant relationships were consistently observed with precipitation:

BRrp for year t:

- No consistent significant relationships observed

BRrp for year t-1:

- No consistent significant relationships observed

CGCD for year t:

- Subalpine fir raw, standardized, residual, and ARSTAN chronologies negatively related to November precipitation
- Whitebark pine standardized, residual, and ARSTAN chronologies negatively related to February precipitation
- Whitebark pine raw, standardized, and ARSTAN chronologies negatively related to July precipitation

CGCD for year t-1:

- Subalpine fir raw, standardized, residual, and ARSTAN chronologies negatively related to April precipitation of previous growing season
- Subalpine fir standardized, residual, and ARSTAN chronologies positively related to October precipitation of previous growing season

D.4 Summary

Several significant findings reported in the literature are consistent with the relationships observed in these analyses. Like Splechna et al. (2000), the Bridge Glacier subalpine fir trees are negatively related to the August air temperature of the previous growing season. The strong negative relationship between November precipitation in the current growth year and radial growth of subalpine fir at Bridge Glacier is comparable to those observed within interior B.C. (Splechna et al. 2000).

The radial growth of whitebark pine trees at the Bridge Glacier site is positively related to April air temperature. This finding agrees with those of Peterson et al. (1990) who report positive relationships with spring temperatures, but contrasts with Perkins (1995)'s observations of negative relationships with April air temperature. Like the trees in the Sierra Nevada region, the Bridge Glacier whitebark pine trees show a positive relationship with July precipitation.

Table D.2. Correlation matrix for Bridge River reconstructed monthly temperature data for growth year t (BRrt) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	BRt: 1	BRt: 2	BRt: 3	BRt: 4	BRt: 5	BRt: 6	BRt: 7	BRt: 8	BRt: 9	BRt: 10	BRt: 11	BRt: 12
SAFraw	-0.21	-0.15	-0.52	-0.32	-0.29	-0.41	-0.30	-0.08	-0.06	0.04	-0.08	0.24
SAFstd	-0.18	-0.06	-0.42	-0.19	-0.26	-0.35	-0.20	-0.08	-0.07	0.10	0.10	0.36
SAFres	-0.24	-0.09	-0.42	-0.13	-0.14	-0.28	-0.13	-0.01	0.02	0.17	0.14	0.39
SAFars	-0.20	-0.07	-0.42	-0.16	-0.22	-0.33	-0.16	-0.05	-0.04	0.10	0.09	0.37
WBPraw	-0.13	0.01	-0.06	0.18	0.15	0.03	-0.11	0.00	0.14	-0.02	-0.05	0.10
WBPstd	-0.09	0.21	0.18	0.47	0.22	0.11	0.11	0.02	0.18	0.16	0.34	0.36
WBPres	-0.11	0.26	0.22	0.40	0.20	0.14	0.04	0.11	0.20	0.30	0.35	0.42
WBPars	-0.11	0.18	0.14	0.44	0.19	0.10	0.09	-0.01	0.16	0.16	0.35	0.36

^aSAF = subalpine fire; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

Table D.3. Correlation matrix for Bridge River reconstructed monthly temperature data for growth year $t-1$ (t-1 BRrt) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	t-1BRt: 1	t-1BRt: 2	t-1BRt: 3	t-1BRt: 4	t-1BRt: 5	t-1BRt: 6	t-1BRt: 7	t-1BRt: 8	t-1BRt: 9	t-1BRt: 10	t-1BRt: 11	t-1BRt: 12
SAFraw	0.11	0.22	0.13	-0.26	-0.31	-0.17	-0.25	-0.55	-0.48	-0.40	-0.30	-0.40
SAFstd	0.11	0.26	0.18	-0.13	-0.29	-0.17	-0.17	-0.52	-0.42	-0.29	-0.07	-0.23
SAFres	0.14	0.26	0.24	-0.06	-0.22	-0.09	-0.12	-0.48	-0.38	-0.26	-0.09	-0.25
SAFars	0.12	0.27	0.20	-0.11	-0.26	-0.13	-0.14	-0.50	-0.41	-0.28	-0.06	-0.24
WBPraw	0.38	0.11	0.16	0.28	0.37	0.48	0.27	0.24	0.14	0.01	0.03	-0.04
WBPstd	0.34	0.25	0.31	0.48	0.38	0.46	0.39	0.22	0.23	0.27	0.46	0.23
WBPres	0.51	0.22	0.31	0.30	0.28	0.47	0.37	0.25	0.12	0.23	0.29	0.08
WBPars	0.35	0.25	0.33	0.50	0.38	0.45	0.39	0.21	0.20	0.25	0.44	0.22

^aSAF = subalpine fire; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

Table D.4. Correlation matrix for CGCD monthly temperature data for growth year t (Gridt) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	Gridt: 1	Gridt: 2	Gridt: 3	Gridt: 4	Gridt: 5	Gridt: 6	Gridt: 7	Gridt: 8	Gridt: 9	Gridt: 10	Gridt: 11	Gridt: 12
SAFraw	-0.17	0.05	-0.19	-0.14	-0.04	-0.20	-0.12	-0.22	-0.20	-0.01	0.03	0.06
SAFstd	-0.04	-0.01	-0.18	-0.06	0.03	-0.15	-0.01	-0.04	-0.11	0.11	0.01	0.09
SAFres	0.03	-0.01	-0.26	0.02	0.04	-0.17	0.01	-0.04	-0.07	0.15	0.00	0.09
SAFars	-0.03	-0.01	-0.20	-0.05	0.02	-0.16	-0.01	-0.06	-0.11	0.12	-0.01	0.10
WBPraw	-0.48	-0.05	-0.09	-0.13	-0.22	-0.01	-0.09	-0.40	-0.13	-0.10	0.12	0.08
WBPstd	-0.39	-0.25	-0.04	0.04	-0.12	0.12	0.11	-0.17	0.16	0.05	0.26	0.19
WBPres	-0.14	-0.11	-0.10	0.05	-0.26	0.15	0.04	-0.10	0.19	0.24	0.24	0.13
WBPars	-0.38	-0.28	-0.02	0.05	-0.09	0.21	0.16	-0.11	0.20	0.09	0.28	0.19

^aSAF = subalpine fire; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

Table D.5. Correlation matrix for CGCD monthly temperature data for growth year t-1 (t-1 Grid) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	t-1Grid: 1	t-1Grid: 2	t-1Grid: 3	t-1Grid: 4	t-1Grid: 5	t-1Grid: 6	t-1Grid: 7	t-1Grid: 8	t-1Grid: 9	t-1Grid: 10	t-1Grid: 11	t-1Grid: 12
SAFraw	-0.10	0.17	-0.09	-0.22	-0.27	-0.06	0.23	-0.01	-0.19	-0.17	-0.28	-0.26
SAFstd	0.04	0.13	-0.01	-0.15	-0.22	0.07	0.43	0.17	-0.11	-0.15	-0.33	-0.28
SAFres	0.10	0.10	-0.11	-0.11	-0.24	0.07	0.41	0.16	-0.09	-0.13	-0.34	-0.27
SAFars	0.04	0.12	0.00	-0.15	-0.24	0.07	0.42	0.16	-0.10	-0.16	-0.34	-0.28
WBPraw	-0.34	0.07	-0.05	-0.07	-0.05	-0.14	-0.14	-0.18	0.01	-0.06	0.12	0.00
WBPstd	-0.13	-0.10	0.02	0.12	0.15	0.01	0.06	0.20	0.36	0.11	0.23	0.03
WBPres	0.15	0.09	0.06	0.14	0.23	-0.07	0.01	0.40	0.38	0.15	0.12	-0.08
WBPars	-0.10	-0.10	0.05	0.13	0.16	0.05	0.11	0.26	0.40	0.15	0.25	0.05

^aSAF = subalpine fire; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

Table D.6. Correlation matrix for Bridge River reconstructed monthly precipitation data for growth year t (BRrp) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	BRrp: 1	BRrp: 2	BRrp: 3	BRrp: 4	BRrp: 5	BRrp: 6	BRrp: 7	BRrp: 8	BRrp: 9	BRrp: 10	BRrp: 11	BRrp: 12
SAFraw	-0.11	0.19	0.20	0.18	0.15	-0.37	-0.19	0.32	-0.09	-0.14	-0.27	0.22
SAFstd	-0.13	0.15	0.28	0.20	0.10	-0.44	-0.15	0.27	-0.04	-0.06	-0.13	0.29
SAFres	-0.19	0.17	0.20	0.15	0.12	-0.36	-0.14	0.35	-0.12	-0.12	-0.14	0.23
SAFars	-0.15	0.17	0.24	0.20	0.10	-0.40	-0.17	0.30	-0.08	-0.08	-0.15	0.27
WBPraw	-0.43	-0.08	-0.14	0.04	0.22	0.34	0.00	0.47	-0.42	-0.07	-0.30	-0.37
WBPstd	-0.44	-0.02	0.06	0.11	0.10	0.16	0.16	0.45	-0.28	0.04	-0.12	-0.17
WBPres	-0.33	-0.10	0.05	-0.04	0.31	0.03	0.42	0.36	-0.27	-0.03	-0.15	-0.38
WBPars	-0.45	-0.06	0.07	0.14	0.11	0.16	0.11	0.45	-0.29	0.03	-0.13	-0.16

^aSAF = subalpine fire; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

Table D.7. Correlation matrix for Bridge River reconstructed monthly precipitation data for growth year t-1 (p-1 BRrp) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	p-1BRrp: 1	p-1BRrp: 2	p-1BRrp: 3	p-1BRrp: 4	p-1BRrp: 5	p-1BRrp: 6	p-1BRrp: 7	p-1BRrp: 8	p-1BRrp: 9	p-1BRrp: 10	p-1BRrp: 11	p-1BRrp: 12
SAFraw	-0.21	-0.34	-0.28	-0.16	-0.19	-0.07	-0.02	-0.52	-0.11	0.09	0.19	-0.08
SAFstd	-0.17	-0.29	-0.18	-0.06	-0.25	-0.15	0.01	-0.45	-0.01	0.05	0.21	0.07
SAFres	-0.21	-0.33	-0.27	-0.09	-0.25	-0.05	-0.02	-0.43	-0.08	0.09	0.21	0.01
SAFars	-0.18	-0.31	-0.20	-0.08	-0.24	-0.11	0.01	-0.45	-0.04	0.06	0.22	0.03
WBPraw	-0.02	0.08	-0.44	-0.25	-0.14	0.47	-0.09	0.07	-0.54	-0.21	0.18	-0.11
WBPstd	0.10	0.22	-0.31	-0.22	-0.29	0.32	0.06	0.03	-0.41	-0.34	0.34	0.18
WBPres	0.31	0.20	-0.27	-0.29	-0.28	0.22	0.03	-0.21	-0.30	-0.26	0.39	0.22
WBPars	0.05	0.20	-0.32	-0.20	-0.29	0.32	0.03	0.04	-0.42	-0.32	0.30	0.17

^aSAF = subalpine fire; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

Table D.8. Correlation matrix for CGCD monthly precipitation data for growth year t (Gridp) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	Gridp: 1	Gridp: 2	Gridp: 3	Gridp: 4	Gridp: 5	Gridp: 6	Gridp: 7	Gridp: 8	Gridp: 9	Gridp: 10	Gridp: 11	Gridp: 12
SAFraw	0.32	0.02	0.12	-0.21	-0.33	-0.08	-0.06	0.08	0.14	-0.21	-0.54	0.07
SAFstd	0.22	0.03	0.21	-0.15	-0.14	0.02	0.02	0.10	0.20	-0.25	-0.58	0.01
SAFres	0.24	0.06	0.25	-0.22	-0.13	-0.03	0.05	0.13	0.09	-0.28	-0.55	-0.03
SAFars	0.22	0.03	0.21	-0.16	-0.15	-0.02	0.01	0.11	0.17	-0.26	-0.58	-0.02
WBPraw	0.23	-0.30	-0.04	-0.18	-0.57	-0.27	-0.39	0.05	-0.01	-0.19	-0.25	0.21
WBPstd	0.25	-0.38	0.28	-0.08	-0.35	-0.17	-0.42	0.03	-0.11	-0.33	-0.18	0.16
WBPres	0.37	-0.41	0.43	0.02	-0.17	-0.03	-0.24	0.07	-0.06	-0.34	-0.24	0.07
WBPars	0.22	-0.42	0.30	-0.03	-0.29	-0.14	-0.40	0.04	-0.07	-0.35	-0.21	0.16

^aSAF = subalpine pine; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

Table D.9. Correlation matrix for CGCD monthly precipitation data for growth year t-1 (p-1 Grid) with Bridge Glacier tree ring chronologies. Highlighted correlations are significant at $p < 0.05$. Numbers represent months (i.e., 1=January...12=December), by growth year (i.e., October of previous calendar year to September of current calendar year).

Chronology ^a	p-1Grid: 1	p-1Grid: 2	p-1Grid: 3	p-1Grid: 4	p-1Grid: 5	p-1Grid: 6	p-1Grid: 7	p-1Grid: 8	p-1Grid: 9	p-1Grid: 10	p-1Grid: 11	p-1Grid: 12
SAFraw	-0.11	-0.28	-0.60	-0.21	-0.49	-0.21	-0.33	-0.12	-0.07	0.30	0.01	0.30
SAFstd	-0.21	-0.30	-0.60	-0.11	-0.29	-0.15	-0.29	-0.08	-0.05	0.38	0.09	0.25
SAFres	-0.22	-0.29	-0.57	-0.14	-0.26	-0.20	-0.29	-0.06	-0.13	0.36	0.16	0.16
SAFars	-0.20	-0.29	-0.60	-0.12	-0.28	-0.16	-0.31	-0.08	-0.08	0.37	0.12	0.25
WBPraw	0.25	-0.18	-0.07	-0.32	-0.53	-0.23	-0.26	-0.19	-0.12	-0.08	-0.16	0.23
WBPstd	0.30	-0.15	0.22	-0.28	-0.24	-0.17	-0.14	-0.36	-0.26	-0.16	-0.03	0.19
WBPres	0.24	0.02	0.13	-0.25	-0.02	-0.08	0.12	-0.46	-0.26	0.05	0.09	0.11
WBPars	0.26	-0.19	0.22	-0.21	-0.17	-0.14	-0.12	-0.34	-0.24	-0.12	-0.02	0.17

^aSAF = subalpine pine; WBP = whitebark pine; raw = raw ring width; std = standardized; ars = ARSTAN.

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Appendix E. Mass balance reconstruction at Bridge Glacier

E.1 Introduction

The mass balance of a glacier is the difference between winter gains (i.e. accumulation through snowfall) and summer losses (i.e. ablation through sublimation or melting) of glacial ice mass measured over a specific time, usually one year (Benn and Evans 1998). Glacial mass balance is an effective proxy measure of climate variables, as the determining processes (i.e. accumulation and ablation) are controlled by the atmospheric environment of the glacier (Hodge et al. 1998). Specifically, a glacier's inputs and outputs are determined by the snowfall and air temperature experienced at a site.

Several recent studies in the western Canada have used tree rings as proxy determinants of glacier mass balance (Lewis and Smith 2004; Watson and Luckman 2004; Larocque and Smith 2005). Although this was not a primary objective of this research program, exploring the mass balance dynamics of Bridge Glacier is relevant to reconstructing its recent advance and retreat history.

Bridge Glacier's mass balance was monitored by B.C. Hydro from the 1976-1977 to the 1983-1984 mass balance year (September to August) (Mokievsky-Zubok 1985; Table E.1). Data for the 1983-1984 year were published by the Institute of Arctic and Alpine Research (Dyurgerov 2002). Zavisha and Sykora Glaciers are within the same drainage as Bridge Glacier, and were monitored over the same interval as Bridge Glacier (Table E.1).

Longer mass balance records are available from Place (1965–2001) and Helm (1982–2001) glaciers located 80 and 100 km southeast of Bridge Glacier (Table E.1).

Both sets of mass balance data show good correlation with the data from Bridge Glacier (Helm Glacier: $r = 0.854$; Place Glacier $r = 0.893$), indicating a regional synchrony in recent glacier behaviour. Given the strength of these relationships, the glacier mass balance data from the longer Place Glacier record was used in a regression relationship to extend the Bridge Glacier record. Fig. E.1 illustrates the correspondence between the reconstructed record at Bridge Glacier and the record from Place Glacier from 1965 to 2001 (Table E.1).

E.2 Bridge Glacier mass balance 1965-2001

Bridge Glacier has experienced consistent negative net mass balance over the 35 years of data reconstruction, with the exception of the following years: 1968-1969, 1975-1976, 1996, and 1999-2000 (Fig. E.1 and Table E.1). The period of positive net mass balance observed in 1975-1976 is substantiated by aerial photographs, which show no retreat of the Bridge Glacier terminus for the years 1975-1979, and by the mass balance records of other glaciers in the region (Dyurgerov 2002). The subsequent period of consistent and large losses in glacial net mass balance has been observed regionally (Luckman and Watson 2006), and is attributed to an established shift from the Pacific Decadal Oscillation's cold phase to its warm phase in 1976-1977 (Hodge et al. 1998; Moore and Demuth 2002).

E.3 Proxy glacier mass balance reconstruction

Previous research in western Canada has shown that the radial growth trends of climatically-sensitive trees have the potential to serve as glacier mass balance proxies (Lewis and Smith 2004; Watson and Luckman 2004; Laroque and Smith 2005). In this

Table E.1. Net mass balance data for selected glaciers in the southern B.C. Coast Mountains. Data are in mm water equivalent per year. Year represents the later calendar years in the mass balance year (e.g., 1965 = 1964 to 1965 mass balance year). Bridge Glacier reconstruction was developed through regression analysis with Place Glacier data ($R = 0.893$, significant at $p < 0.05$).

Year	Place Glacier ^a	Helm Glacier ^a	Zavisha Glacier ^b	Sykora Glacier ^b	Bridge Glacier ^b	Bridge Glacier reconstruction
1965	-514					-217
1966	115					322
1967	-1212					-814
1968	-131					111
1969	-216					39
1970	-1510					-1069
1971	-340					-68
1972	-344					-71
1973	-300					-33
1974	-564					-259
1975	-240					18
1976	877	568	1100	830		974
1977	-1227	-1464	-260	-510	-770	-827
1978	-433	-783	390	-160	-260	-147
1979	-2212	-2301	-690	-1080	-1360	-1670
1980	-923	-944	-670	-510	-770	-567
1981	-1097		-204	128	-404	-716
1982	-754	-347	-524	-115	-509	-422
1983	-443	-209	366	647	206	-156
1984	-341	-328	314	750	-228	-68
1985	-1887	-1732	-1215	-1340	-1869	-1392
1986	-1317	-1333				-904
1987	-867	-789				-519
1988	-961	-559				-599
1989	-1010	-1623				-641
1990	-938	-1790				-580
1991	-990	-2239				-624
1992	-790	-2798				-453
1993	-2280	-2342				-1728
1994	-2010	-1885				-1497
1995	-2486	-1465				-1905
1996	-211	211				43
1997	-888	-1073				-537
1998	-2450	-2850				-1874
1999	620	1500				754
2000	130	110				335
2001	-760	-600				-427

^aSource: Dyurgerov 2002.

^bSource: 1977-1984 data Mokievsky-Zubok 1885; 1984-1985 data from Dyurgerov 2002.

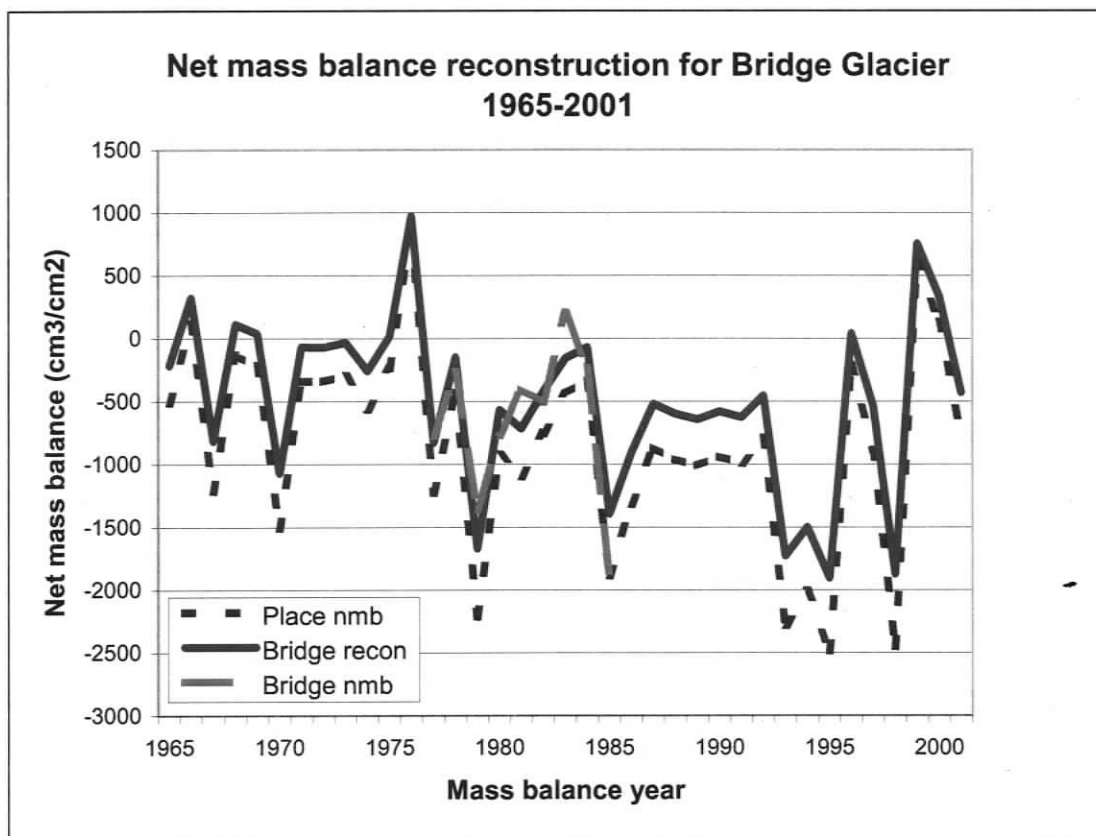


Fig. E.1. Net mass balance (nmb) reconstruction for Bridge Glacier 1965-2001 (solid line). Also shown are the actual nmb measurements for Place Glacier (1965-2001; short-dashed line) and Bridge Glacier (1977-1985; long-dashed line).

instance, the two ring-width chronologies collected at the Bridge Glacier site were correlated against the net mass balance reconstruction. The results were disappointing (Table E.2), as no significant relationships between the mass balance record and annual ring growth records were observed. There are several possible reasons for this:

- Although the reconstructed net mass balance record for Bridge Glacier extends for 35 years, there may not have been sufficient number of data points to robustly identify any relationship between mass balance and tree rings.

Table E.2 Correlation matrix of the Bridge Glacier net mass balance reconstruction (BGnmb) with the subalpine fir (SAF) and whitebark pine (WBP) raw ring width (raw), standardized (std), residual (res), and ARSTAN (ars) chronologies. None of the correlations are significant at $p < 0.05$.

Chronology	BGnmb
SAFraw	0.21
SAFstd	0.04
SAFres	0.03
SAFars	0.04
WBPrw	0.26
WBPstd	0.12
WBPres	0.14
WBPars	0.06

- Only four years of positive net mass balance conditions exist within the reconstructed record for Bridge Glacier (Table E.1). This may be an insufficient number of years to establish the corollary response of tree-rings to the changed climate conditions.
- A simple linear correlation model may not be appropriate to account for the potential relationship between mass balance and tree ring-width. Persistence is likely to exist in both data sets, and the response times to the forcing mechanisms (i.e. climate variability) may differ for trees ring formation and glacier mass balance.
- The differential radial growth response of trees at this site to July air temperatures (subalpine fir) and February precipitation totals (whitebark pine) (Appendix D) suggests that annual net mass balance data may not correspond to the factors limiting their growth. Further analyses of the data sets might reveal better correspondence to summer and/or winter mass balance records (e.g., Watson and Luckman 2004).

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Appendix F. Subfossil wood samples collected at Bridge Glacier

Table F.1. Summary of subfossil wood samples collected at Bridge Glacier study site.

Sample ID	Species ^a	Location	N 50 deg	W 123 deg	Elevation (m)	Min age (years)	No. rings analysed ^b	Measurable radius (cm)	Chronology	Field notes
BG02										
1	saf	South slope of nunatak	49°18.78"	34°28.98"	1690	146	96	10.2	Bridge 1	3m-long mast, no roots
2	wbp	South slope of nunatak	49°18.78"	34°28.98"	1690	>40	0	0.0	1190 ± 60	Perimeter sample from very large log
3	saf	South slope of nunatak	49°18.78"	34°28.98"	1690	127	126	19.8	Bridge 1	1/4 section from log on upper bench
4	saf	North lateral gully	50°44.16"	30°18.24"	1430	46	41	9.1		Bole with root stalk left, either slid down slope or pushed up onto slope by glacier as it moved down.
5	wbp	North lateral gully	50°44.16"	30°18.24"	1430	70	69	17.0	690 ± 50 Bridge 4	1.5 m-long basal root stalk washed out below gully onto fan apex
6	wbp	North lateral gully	50°44.16"	30°18.24"	1430	87	73	14.2		1.2 m-long basal root stalk washed out onto gully
7	wbp	North lateral gully	50°44.16"	30°18.24"	1430	96	93	10.0		Root stalk in soil below organic material and overlain by till
8	saf	North lateral gully	50°44.16"	30°18.24"	1430	61	53	12.8		Small section of fragmented log, rotted, found on outwash immediately below gully.
9	saf	North lateral gully	50°44.16"	30°18.24"	1430	67	62	9.9	650 ± 50	
10	wbp	Area 1500	50°23.16"	30°59.22"	1397	86	86	10.9	1500 ± 50 Bridge 2	<i>In situ</i> stump in abandoned stream channel
11	wbp	Area 1500	50°23.16"	30°59.22"	1397	98	87	13.8		Broken fragment of large tree, upstream of BG02-10.
12	wbp	Area 1500	50°23.16"	30°59.22"	1397	87	87	8.5		Bole fragment with stump attached, downstream on alluvial fan

BG03

1	saf	West of South Creek	49° 31.5"	23° 33.1"	1509	36	35	4.6	430 ± 60	Branch from lateral by ice dam lodged in sandy lacustrine sediments where they met with brown soil
2	saf	West of South Creek	49° 50.5"	30° 02.3"	1400	149	149	11.1	1930 ± 70	Bole fragment from terrace at contact between soil layers, glacially scrapped.
3	saf	West of South Creek	49° 50.0"	29° 59.7"	1407	131	130	6.7		70 cm-long, on same horizon as BG03-02
4	saf	West of South Creek	49° 50.3"	29° 58.2"	1406	>20	0	0.0		Partial rim of fragmented bole next to chunk of root mass (possibly <i>in situ</i>), 25 m upstream of BG03-03
5	saf	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	69	56	6.7	Bridge 4	Masticated at both ends
6	wbp	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	58	57	7.1	Bridge 4	
7	wbp	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	85	84	13.8	Bridge 4	Contains charcoal
8	saf	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	96	90	14.7		Large log under rock outcrop, oriented down slope, 2-3 m long
9	wbp	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	98	97	13.7	620 ± 60	3m-long <i>in situ</i> , pushed by ice. Root system intact. Weak soil horizons with some organic material
10	wbp	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	~300	287	7.0		Upper east corner of gullies
11	wbp	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	77	75	7.5		Upper east corner of gullies
12	saf	Northern ("Beefcake") gully	50° 40.1"	30° 31.4"	1428	126	115	14.5	Bridge 4	Old stump, not <i>in situ</i> , upper east corner of gullies

13	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	185	166	4.4	Bridge 2	Outwash fan
14	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	45	45	4.5	Bridge 7	Outwash fan
15	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	194	193	5.0	Bridge 2	Outwash fan
16	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	62	62	9.5	Bridge 7	Outwash fan
17	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	66	66	10.5		Outwash fan
18	saf	Area 1500	50' 20.9"	30' 50.8"	1394	124	124	15.6		Outwash fan
19	saf	Area 1500	50' 20.9"	30' 50.8"	1394	50	50	9.0		Outwash fan
20	saf	Area 1500	50' 20.9"	30' 50.8"	1394	304	304	11.5		Outwash fan
21	saf	Area 1500	50' 20.9"	30' 50.8"	1394	123	122	7.4		Outwash fan
22	saf	Area 1500	50' 20.9"	30' 50.8"	1394	78	78	6.6		Outwash fan
23	saf	Area 1500	50' 20.9"	30' 50.8"	1394	247	247	19.5		Outwash fan
24	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	119	119	9.6	Bridge 7	Outwash fan
25	saf	Area 1500	50' 20.9"	30' 50.8"	1394	40	37	4.8		Outwash fan
26	saf	Area 1500	50' 20.9"	30' 50.8"	1394	105	105	7.1	Bridge 7	Outwash fan
27	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	119	119	10.4		Outwash fan
28	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	98	98	16.5		Outwash fan

29	saf	Area 1500	50' 20.9"	30' 50.8"	1394	87	87	8.0	Outwash fan
30	saf	Area 1500	50' 20.9"	30' 50.8"	1394	170	168	8.7	Outwash fan
31	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	178	178	13.9	Outwash fan
32	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	101	101	9.6	Outwash fan
33	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	100	99	12.9	Bridge 2 Outwash fan
34	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	84	84	15.9	Outwash fan
35	wbp	Area 1500	50' 20.9"	30' 50.8"	1394	82	80	5.1	Outwash fan
36	saf	Area 1500	50' 20.9"	30' 50.8"	1394	53	53	6.0	Outwash fan
37	saf	Area 1500	50' 20.9"	30' 50.8"	1394	127	127	5.7	Outwash fan
38	saf	Area 1500	50' 20.9"	30' 50.8"	1394	79	78	7.4	Outwash fan
39	unk	Area 1500	50' 23.2"	30' 59.3"	1403	>70	0		Close to BG02-10 site in abandoned channel.
40	unk	Area 1500	50' 19.3"	31' 08.8"	1400	>60	0		Below bedrock, avalanche sample
41	saf	Area 1500	50' 19.9"	31' 07.6"	1404	106	105	13.1	
42	wbp	Area 1500	50' 19.9"	31' 07.6"	1404	90	90	7.6	
43	wbp	Area 1500	50' 19.3"	31' 08.8"	1400	66	64	21.9	Bridge 2 Below bedrock
44	saf	Area 1500				60	59	8.1	On dead ice below bedrock

45	saf	Area 1500				68	68	11.2		On dead ice below bedrock
46	wbp	Area 1500			84	85	85	9.5		On dead ice below bedrock
47	wbp	Area 1500			111	112	112	17.5		On dead ice below bedrock
48	wbp	Area 1500	50' 23.2"	30' 59.3"	1403	39	39	9.0		Close to BG02-10 site in abandoned channel.
49	wbp	Area 1500	50' 23.2"	30' 59.3"	1403	74	74	20.8		Close to BG02-10 site in abandoned channel.
50	wbp	Northern gully (with till slab)	50' 32.8"	31' 00.8"	1485	90	90	5.0		Bridge 9
51	saf	Northern gully (with till slab)	50' 32.8"	31' 00.8"	1485	93	93	9.2		Bridge 9
52	saf	Northern gully (with till slab)	50' 32.8"	31' 00.8"	1485	241	241	10.8		Bridge 9
53	saf	Northern gully (with till slab)	50' 32.8"	31' 00.8"	1485	92	91	13.3		Bridge 4
54	saf	Northern gully (with till slab)	50' 32.8"	31' 00.8"	1485	83	82	12.0		Bridge 4 Wet, stuck under boulders in creek
55	wbp	Northern gully farthest west	50' 30.3"	31' 06.1"	1510	82	74	12.4		Farthest upslope in gully
56	wbp	Nunatak	49' 18.9"	34' 28.7"	1689	214	214	24.4		poss. x-d w dated sample. 66
57	saf	Nunatak	49' 18.9"	34' 28.7"	1689	145	144	17.9	2980 ± 60	Bridge 1 In gully beneath bedrock where we found 2002 ¹⁴ C- dated sample.
58	wbp	Nunatak	49' 18.9"	34' 28.7"	1689	318	318	14.8		Bridge 6 In gully beneath bedrock where we found 2002 ¹⁴ C- dated sample.
59	saf	Nunatak	49' 18.9"	34' 28.7"	1689	146	144	15.1		Bridge 1 In gully beneath bedrock where we found 2002 ¹⁴ C- dated sample.
60	saf	Nunatak	49' 18.9"	34' 28.7"	1689	127	127	16.4		Bridge 1 In gully beneath bedrock where we found 2002 ¹⁴ C- dated sample.

61	saf	Nunatak	49' 18.9"	34' 28.7"	1689	177	177	11.1	In gully beneath bedrock where we found 2002 ¹⁴ C-dated sample.
62	wbp	Nunatak	49' 13.4"	34' 45.8"	1701	132	132	11.2	Gully farther west of previous 6 samples, wet sample
63	saf	Nunatak	49' 20.7"	34' 43.7"	1788	241	240	21.0	Sample from log sampled by Ryder, who reports <i>in situ</i> . Below highest lateral crest
64	wbp	Nunatak	49' 20.7"	34' 43.7"	1788	221	221	13.6	Smaller log next to BG03-63, Ryder reports <i>in situ</i>
65	saf	Nunatak	49' 18.7"	34' 38.1"	1746	213	212	10.8	Down slope of BG03-63 & 64, on lower lateral crest
66	wbp	Nunatak	49' 20.4"	34' 32.2"	1745	281	275	17.7	poss. x-d w 56 Hanging over gully where BG03-62 was collected.
67	wbp	Alluvial fan ("Mordor")				98	97	15.7	
68	saf	Alluvial fan ("Mordor")				111	110	13.5	
69	wbp	Northern ("Chucky") gully	50' 52.2"	30' 07.3"	1443	250	246	25.8	Large log in upper west corner
70	saf	Northern ("Chucky") gully	50' 52.2"	30' 07.3"	1443	76	70	14.7	<i>In situ</i> , in direction of ice flow. Soil surrounding it, large branch still attached.
71	wbp	Alluvial fan				93	93	7.6	
72	saf	Alluvial fan				82	76	11.8	
73	wbp	Alluvial fan				248	248	11.7	Bridge 7
74	wbp	North terminals	51' 32.1"	28' 39.3"	1379	202	202	13.8	Old log in riverbank.
75	wbp	Gravesite ("Diagon Alley")	50' 24.1"	30' 12.9"	1391	229	211	19.3	Large stump with bark, redeposited fluvially.
76	wbp	Gravesite ("Diagon Alley")	50' 24.3"	30' 12.8"	1390	140	139	7.2	1.5 m-long bole in a bar in old stream channel. 2 piths. Sheared and masticated.

77	wbp	Gravesite ("Diagon Alley")	50' 24.3" 30' 12.8"	1390	194	194	194	9.0	Long bole protruding from centre of same bar as BG03-76
78	wbp	Gravesite ("Diagon Alley")	50' 24.3" 30' 12.8"	1390	167	161	161	13.8	Large bole from same place as BG03-76 & 77.
79	wbp	Gravesite ("Diagon Alley")	50' 24.3" 30' 12.8"	1390	226	226	226	17.0	In stream bed south of previous 3 samples, 2.5 m long log, has bark
80	wbp	Gravesite ("Diagon Alley")	50' 24.3" 30' 12.8"	1390	146	146	146	7.4	Just behind BG03-79, smaller fragment on upper surface of moraine.
81	wbp	Gravesite	50' 25.0" 30' 16.0"	1391	183	181	181	11.0	Has large branch
82	wbp	Gravesite	50' 25.0" 30' 16.0"	1391	388	388	388	19.5	Big, <i>in situ</i> , some perimeter wood loss
83	wbp	Gravesite ("Diagon Alley")	50' 24.3" 30' 12.8"	1390	150	150	150	9.8	On moraine crest upslope of BG03-84, small branches in tact.
84	wbp	Gravesite ("Diagon Alley")	50' 24.3" 30' 12.8"	1390	227	227	227	11.6	On same bench as BG03-85
85	saf	Gravesite ("Diagon Alley")	50' 24.3" 30' 12.8"	1390	191	189	189	7.3	On bench next to major channel, 2.5 m long bole, in lee of rock, weathered.
86	saf	South Creek	49' 28.7" 29' 45.4"	1562	374	373	373	21.4	Large tree lying on top of 1974 moraine, has fire scar.
87	wbp	South Creek	49' 28.7" 29' 45.4"	1562	274	273	273	14.8	Wedge from snag in midst of large boulders near BG03-86
88	wbp	South Creek	49' 51.1" 30' 04.4"	1390	359	358	358	12.6	Likely from South Lake
89	saf	South Creek	49' 51.1" 30' 04.4"	1390	139	139	139	14.3	Snag 3m-long, likely from South Lake
90	saf	Gravesite	50' 25.0" 30' 16.0"	1391	144	142	142	12.3	Bark flaking off, beetle galleries, aligned in glacier flow direction
91	wbp	Gravesite	50' 25.0" 30' 16.0"	1391	218	217	217	11.2	1m-long log attached to exposed stump-like piece.
92	wbp	Gravesite	50' 25.0" 30' 16.0"	1391	198	196	196	8.2	Long, thick bole, aligned with glacier flow, gold-brown soil present

93	saf	Gravesite ("Diagon Alley")	50' 24.1"	30' 12.9"	1391	307	306	10.8	<i>In situ</i> , in side of moraine, organic layer with soil beneath, bark and pith present
94	saf	Gravesite ("Diagon Alley")	50' 24.1"	30' 12.9"	1391	305	305	17.8	Big, wet, next to BG03-93, pointed in direction of glacier flow, broken off at root level.

^a saf = subalpine fir; wbp = whitebark pine

^b In cases where tree rings are exceptionally compressed or damaged the number of rings analysed is fewer than the sample's minimum age.

^c Radiocarbon age from Ryder and Thomson (1986).