# LATE HOLOCENE FLUCTUATIONS OF LILLOOET GLACIER, SOUTHERN COAST MOUNTAINS, BRITISH COLUMBIA

by

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# ABSTRACT

Geomorphic reconstructions of glacier fluctuations are commonly hampered by poor preservation of landforms that predate the extensive Little Ice Age advances of the late Holocene. This thesis presents the results of a detailed study of lateral moraine stratigraphy at Lillooet Glacier in the southern Coast Mountains of British Columbia. Five tills, separated by laterally extensive organic horizons and lines of large woody debris, were found in three cross-sectional exposures through the lateral moraine and two shallow gullies incised into its steep proximal face. Eighteen new radiocarbon ages constrain the timing of five separate advances of Lillooet Glacier: (1) prior to 3000 $^{14}$C yr BP; (2) ~3000 $^{14}$C yr BP; (3) ~2500 $^{14}$C yr BP; (4) ~1700 to 1400 $^{14}$C yr BP; and (5) during the Little Ice Age, after 470 $^{14}$C yr BP. The Lillooet Glacier chronology is broadly synchronous with other glacier records from the Coast Mountains and adjacent areas. Data from Lillooet Glacier and other sites in the Coast Mountains provide clear evidence for a previously unrecognised period of glacier advance ~1700-1400 $^{14}$C BP, here termed the Bridge Advance. The glacier chronology emerging from the Coast Mountains suggests that regional late Holocene climate was more variable than is apparent from many proxy paleoenvironmental records. North Pacific ocean-atmosphere variability is the dominant control on the spatial and temporal variability of present glacier mass balance regimes in the coastal cordillera of North America. Thus, changes in the intensity of North Pacific ocean-atmosphere circulation patterns may provide a plausible forcing mechanism for late Holocene glacier advances in the region, particularly during coincident periods of summer drought.

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Most importantly, I thank my parents for providing an environment that encourages learning, respect and enjoyment of life. Particular thanks are due to my father, with whom I first discovered the mountains.

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# CHAPTER 1. Introduction and overview

Chronologies of past glacier fluctuations provide a proxy record of climate change that can be used to constrain and evaluate other paleoenvironmental proxy records. However, investigations of prehistoric glacier fluctuations are often limited to advances during the Little Ice Age (LIA) because this late Holocene period of widespread glacier expansion generally destroyed surficial evidence of prior glacier activity. Although LIA glacier chronologies are important, and can offer higher temporal resolution than is generally afforded by investigations limited to radiocarbon-based chronologies, they do not provide insight into the marked climatic variability that has recently been shown to characterize the remainder of the Holocene Epoch.

A variety of research approaches have been used to date pre-LIA glacier advances, including tephrochronology, analyses of downvalley lake sediment cores, lichenometry, cosmogenic exposure dating, dating of shorelines or sediments associated with glacier-dammed lakes, and stratigraphy of glacial sediments in forefields and lateral moraines.

This study employs the last of these approaches at Lillooet Glacier in the southern Coast Mountains of British Columbia. The regional chronology of Holocene glaciation in the Coast Mountains is limited to scattered, commonly poorly dated records. Recent high-resolution paleoenvironmental investigations in western North America confirm the sub-millennial variability of Holocene climate in the region, but the incomplete nature of the local and regional glacier chronologies frustrates efforts to compare the records and discern specific climate forcing mechanisms. Lateral moraine stratigraphy at Lillooet Glacier, which is characterized by stacked units of till separated by prominent organic horizons and accumulations of large woody debris, offers a unique opportunity to develop a detailed record of Holocene glaciation at one site.

This thesis is organized into two main chapters that are intended to be read as stand-alone papers, although some ideas and data are common to both.

Chapter 2 presents the scientific research that underpins the thesis. In it, the rationale for studying Holocene glacier fluctuations in the Coast Mountains is presented and lateral moraine stratigraphy at Lillooet Glacier is described. Radiocarbon ages are used in conjunction with stratigraphic relations to develop a detailed and coherent chronology of late Holocene activity of Lillooet Glacier. The chronology is compared to established and emerging glacier chronologies elsewhere in the Canadian and Alaskan Cordillera. I wrote the manuscript with editorial input from Dr. John Clague, my co-author. The manuscript has been submitted for publication to *Canadian Journal of Earth Sciences.*

Chapter 3 is an expanded discussion of a period of glacier advance at Lillooet Glacier about 1500 $^{14}C$ yr BP, which I term the Bridge Advance. This advance has not previously been described in the published literature on Coast Mountain Holocene glaciation. Drawing on new and previously published data from three other sites in the Coast Mountains, I argue that this period of glacier advance was likely regional in extent. In addition to presenting an abbreviated version of the relevant stratigraphy and radiocarbon ages from Lillooet Glacier, previously published data from the Mt. Waddington area and the Boundary Ranges in the northern Coast Mountains are reviewed. The chapter also draws on new supporting evidence from Bridge Glacier, although the discussion of this site is brief because the data will be presented in full in a forthcoming M.Sc. thesis[1]. The role of Pacific ocean-atmosphere variability in controlling climate along the Pacific coast of North America as a possible forcing mechanisms of Bridge Advance glacier activity is discussed. I wrote the manuscript with editorial input from the co-authors, Dr. Dan Smith, Dr. John Clague, Sandra Allen, and Dr. Sonya Larocque. Sandy Allen and Dr. Dan Smith (University of Victoria) provided the unpublished Bridge Glacier data. My co-authors and I will edit the manuscript prior to submission for publication in a peer-reviewed journal.

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[1] Sandra M. Allen, Department of Geography, University of Victoria, Victoria, British Columbia

## CHAPTER 2. Stratigraphic evidence for multiple Holocene advances of Lillooet Glacier, southern Coast Mountains, British Columbia

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### Abstract

Holocene lateral moraines in the Coast Mountains of British Columbia are commonly composed of multiple drift units related to several glacier advances. In this paper, we document lateral moraine stratigraphy at Lillooet Glacier in the southern Coast Mountains. Five tills, separated by laterally extensive organic horizons and lines of large woody debris, were found in three cross-sectional exposures through the lateral moraine and two shallow gullies incised into its steep proximal face. Eighteen new radiocarbon ages constrain the timing of five separate advances of Lillooet Glacier: (1) prior to 3000 $^{14}C$ yr BP; (2) ~3000 $^{14}C$ yr BP; (3) ~2500 $^{14}C$ yr BP; (4) ~1700 to 1400 $^{14}C$ yr BP; and (5) during the Little Ice Age, after 470 $^{14}C$ yr BP. The Lillooet Glacier chronology is broadly synchronous with other glacier records from the Coast Mountains. These records collectively demonstrate higher-frequency climate variability during the late Holocene than is apparent from many paleoecological reconstructions. Geomorphic reconstructions of glacier fluctuations are commonly hampered by poor preservation of landforms that predate the extensive Little Ice Age advances of the latest Holocene. Our results highlight the potential of lateral moraine stratigraphy for reconstructing these earlier events.

## Introduction

Climate variability during the Holocene Epoch was relatively limited, compared to the high-amplitude fluctuations evident in late Pleistocene proxy paleoclimatic records (e.g. Dansgaard et al. 1993; Petit et al. 1999). However, numerous ice core (e.g. O'Brien et al. 1995), marine (e.g. Bond et al. 2001), and terrestrial (e.g. Denton and Karlén 1973; Viau et al. 2002; Hallett et al. 2003; Hu et al. 2003) records suggest that Holocene climate was also characterized by marked, perhaps cyclical variability. One manifestation of this climatic variability is the resurgence of alpine glaciers during the middle and late Holocene, commonly termed Neoglaciation (Porter and Denton 1967). Alpine glacier termini fluctuate rapidly in response to changes in mass balance, and hence temperature and precipitation, so chronologies of past glacier activity have traditionally been used as climate proxies. These records can help constrain and evaluate other paleoenvironmental reconstructions.

Many studies of Holocene glacier fluctuations, though, are severely limited by sparse evidence for advances prior to the Little Ice Age (LIA), here defined as the period between ca. AD 1200 and 1900 (Grove 1988). During this time, many glaciers advanced to their most extended downvalley positions of the Holocene, and surficial evidence for earlier, less extensive advances was commonly destroyed, a process that has been termed "obliterative overlap" (Gibbons et al. 1984). In the Canadian Cordillera, fragmented moraines outside LIA limits have been indirectly dated using tephrochronology (Luckman and Osborn 1979), changes in lacustrine sedimentation rates and sediment organic content (Reasoner et al. 1994), and beds of clastic sediment in organic sequences (Fulton 1971). Lichenometry has successfully been used to date latest Holocene moraine sequences (Larocque and Smith 2003), but snow-kill, mass wasting, and uncertain calibration of the age-diameter relationship limit the use of the technique in investigations at longer time scales. Holocene moraines have been dated elsewhere using cosmogenic radionuclides (e.g. Finkel et al. 2003), but their utility for Holocene investigations in Coast Mountain settings has been questioned by Walker (2003).

Where pre-LIA moraines are missing or cannot be dated, records of Holocene glaciation have been derived from downvalley lake sediment cores (Souch 1994; Leonard and Reasoner 1999; Menounos 2002) and dated shorelines or sediments of former ice-dammed lakes (Clague and Rampton 1982; Clague and Mathews 1992). More direct dating of pre-LIA advances is possible where glacially overridden trees are exposed *in-situ* in glacigenic sediments (e.g. Luckman et al. 1993; Wiles et al. 1999; Wood 2002), or where glacially sheared tree stumps are present on nunataks (Mathews 1951).

Similarly, composite lateral moraines, consisting of drift related to multiple advances, can provide detailed information on the timing of glacier fluctuations. Separation and dating of tills in these moraines are facilitated by the presence of paleosols or weathered horizons, tephras, and prominent accumulations of woody debris derived from overridden trees and shrubs. Lateral moraine stratigraphy has been used extensively in Europe and New Zealand to develop chronologies of Holocene glacial activity (Röthlisberger et al. 1980; Gellatly et al. 1988; Holzhauser and Zumbühl 1996), but has only seen limited application in the North American Cordillera. Moraine stratigraphy and radiocarbon dating have provided constraints on Neoglacial advances in the Coast Mountains of British Columbia (Ryder and Thomson, 1986; Desloges and Ryder, 1990) and at Bugaboo Glacier (Osborn 1986; Osborn and Karlstrom 1988, 1989) and Stutfield Glacier (Osborn et al. 2001) in the Purcell and Rocky Mountains, respectively.

Ryder and Thomson (1986), expanding on earlier efforts by Mathews (1951) and Fulton (1971), established an important regional chronological framework for late Holocene glacier fluctuations prior to the LIA in the southern Coast Mountains. However, their work was limited to reconnaissance-scale investigations, and their regional glacier chronology was compiled from limited exposures at several sites.

Here, we present the results of a detailed study of lateral moraine stratigraphy at Lillooet Glacier in the southern Coast Mountains. Buried organic horizons and prominent layers of large woody debris were used to separate till units at four sites and to provide a detailed radiocarbon chronology of late Holocene glacier fluctuations. We compare our chronology to fragmented records of Holocene glacier activity elsewhere in the Coast Mountains and discuss the paleoclimatological implications of the regional glacier chronology.

**Setting**

The Coast Mountains are a belt of high-relief, northwest-trending mountain ranges that extend from southwest British Columbia to the St. Elias Mountains of southwest Yukon Territory and Alaska. Lillooet Glacier is located in the Pacific Ranges of the southern Coast Mountains. Bedrock in the study area is primarily late Cretaceous granodiorite and early Cretaceous volcanic and metasedimentary rocks of the Gambier Group (Monger and Journeay 1994).

Following decay of the late Pleistocene Cordilleran Ice Sheet, which was punctuated by several readvances of large valley glaciers (Friele and Clague 2002), glaciers in the southern Coast Mountains retreated and were probably no more extensive than today by ~9500 $^{14}C$ yr BP (Clague and James 2002). Presently, glaciers and icefields are common in the southern Coast Mountains, although all have thinned and retreated considerably since the end of the LIA (Mathews 1951; Ryder and Thomson 1986; Larocque and Smith 2003; Holm et al. in press).

Lillooet Glacier (Fig. 2.1; 50°45N, 123°46'W) is a large valley glacier that flows about 9.5 km southeast from icefields northeast and southwest of the valley axis, and terminates in the upper reach of Lillooet River valley at an elevation of about 1100 m. The highest peaks in the area are ~ 2950 m asl, and relief above the glacier locally exceeds 1300 m. Lillooet Glacier has retreated ~5 km from its maximum LIA terminus, which is marked by a terminal moraine that has been severely degraded by fluvial incision and mass wasting from steep, avalanche-prone valley slopes (Fig. 2.2). Farther upvalley, the glacier is fringed by large (height ~100 m), steep, sharp-crested LIA lateral moraines. The present surface of Lillooet Glacier is more than 200 m below these moraine crests, indicating a prolonged period of negative mass balance since the culmination of local LIA activity. A subdued, forested moraine is locally present directly outboard of the prominent lateral moraine ridge northeast of the glacier, and can be tracked as a fragmented ridge for ~3 km (Walker 2003). A large cliff of jointed basalt rises ~100 m from the recently deglaciated forefield and forms a bench underlying the northeast lateral moraine (Fig. 2.2). Though undated, the basalt probably erupted beneath or against glacier ice during the late Pleistocene, based on the presence of fan-shaped jointing patterns (e.g. Mathews 1958; Hickson 2000).

The study area is located in the Engelmann Spruce-Subalpine Fir and Mountain Hemlock biogeoclimatic zones, which are characterized by long, cool and wet winters, and short, cool summers. The growing season is short due to a typically thick, persistent winter snowpack. Arboreal vegetation at Lillooet Glacier is dominated by mixed stands of mountain hemlock (*Tsuga mertesiana*) and subalpine fir (*Abies lasiocarpa*). Disturbed sites near creeks and avalanche tracks support dense thickets of slide alder (*Alnus sinuata*).

The Coast Mountains are a substantial orographic barrier to moist, maritime air masses, and there is a strong west-east environmental gradient across the range. The Aleutian Low and Pacific High pressure systems are the dominant climatological controls on seasonality in the study area. During winter, the Aleutian Low intensifies and shifts southeastward to the Gulf of Alaska, driving moist air masses in a counter-clockwise direction toward the coast. In summer, weakening and northwest migration of the Aleutian Low are accompanied by intensification of the anticyclonic Pacific High, which results in drier conditions in the region as storm tracks are diverted around the zone of high pressure.

**Methods**

Field investigations in fall 2001 and summer 2002 focused on four sites along ~3 km of the northeast lateral moraine of Lillooet Glacier (Fig. 2.2). Three sites, informally named south gully (SG), middle gully (MG) and north gullies (NG), are continuous exposures through the lateral moraine that were eroded by streams emanating from the valley slope above. The fourth site, proximal gullies (PG), is a series of shallow gullies incised into the proximal face of the lateral moraine above the basalt bench. All of these sections are in steep, exposed positions on unstable morainal sediments, so investigations were necessarily brief to minimize exposure to rockfall, particularly at PG where the section could only be accessed by long rappels from the moraine crest and the rockfall hazard was acute.

At each site, moraine stratigraphy was logged using a barometric altimeter. We used the moraine crest as the stratigraphic datum at each measured section. Where organic horizons were present, we recorded their position and excavated laterally to determine their extent. These organic horizons are interpreted as remnants of soils developed on old moraine surfaces, and are hereafter termed paleosols. Bulk samples were collected from paleosol Ah and O horizons. The samples were wet sieved for macroscopic wood and charcoal. Large wood macrofossils, including tree stems up to 50 cm diameter, are commonly associated with the horizons of organic material. Cross-sections were cut from 35 buried tree stems for radiocarbon dating and chronostratigraphic correlation of till units based on tree-ring crossdating, which will be described elsewhere.

Interpretation of radiocarbon ages in moraine sediments requires consideration of the provenance and stratigraphic position of the dated material (Röthlisberger et al. 1980; Osborn 1986; Ryder and Thomson 1986). Radiocarbon analysis of bulk soil samples is problematic (e.g. Matthews 1980; Geyh et al. 1985), thus we only dated fragments of wood and charcoal from paleosols using the AMS radiocarbon method. The resulting ages are interpreted as maximum ages for deposition of overlying tills, i.e., the overlying till is no older than the dated sample. Conventional (proportional gas counter) radiocarbon ages of samples collected from laterally continuous accumulations of large tree stems lying directly on a paleosol are interpreted as direct dates for the glacier advance that deposited the overlying till. Ages from stems not in contact with paleosols are regarded as maximum ages for the surrounding till, although ages from well-preserved samples are probably close maxima or direct dates. Dated samples from buried tree stems were collected from the outermost 5 to 25 annual rings of the stems. Additional details concerning radiocarbon ages are presented in Table 2.1.

**Lateral moraine stratigraphy and radiocarbon ages**

*South gully*

Near the southeast edge of the LIA lateral moraine (Fig. 2.3), a large creek has cut through till and two woody horizons that record fluctuations of the glacier margin during LIA time (Fig. 2.4). About 5 m of light grey till with abundant large boulders overlie the upper horizon of woody debris, which contains *in-situ* roots. Another 5 m of similar till, of which the uppermost 1.5 m is locally oxidized and contains lenses of laminated sand, separates the upper and lower woody horizons. The lower woody horizon rests on 2 m of oxidized bouldery till, which in turn overlie, perhaps unconformably, over 20 m of dense, silty, dark gray, pervasively sheared diamicton draping a steeply sloping granitic bedrock surface. Clast content (~10%) and lithology (primarily basalt) distinguish the dark gray till from overlying, more bouldery tills that form the moraine. The bouldery tills contain 30-40% stones of dominantly granitic composition.

Samples recovered from the woody horizons yielded radiocarbon ages of 10±50, 170±60 and 290±60 $^{14}C$ yr BP (GSC-6600, GSC-6602 and TO-9753, respectively). Calibration of the radiocarbon ages yields overlapping ages ranging from AD 1690 to 1950, thus the south gully exposure provides little insight into the LIA chronology of Lillooet Glacier. However, it does suggest that the downvalley margin of Lillooet Glacier fluctuated enough during the LIA to allow colonization of the moraine by shrubs and perhaps small trees during periods of lesser ice extent. This inference is further supported by the multi-crested nature of the lateral moraine directly downvalley of the section. We associate the basal gray, silty diamict with late Wisconsin glaciation because it is lithologically and sedimentologically dissimilar to the Holocene morainal tills, and because a probable correlative till occurs on slopes high above the Holocene lateral moraines.

*Proximal gullies*

The LIA lateral moraine is steep and sharp-crested for about 1.5 km upvalley of south gully. Slopes on the proximal and distal sides of the moraine average 50° and 35°, respectively, and gullies incised into the proximal face have an average slope of ~45°. Near south gully, the proximal moraine face rises steeply from the edge of the basalt cliff, and is incised by gullies that extend from the moraine crest to the bedrock contact at the base of the moraine (Fig. 2.3). In several of the gullies, contacts between till units are highlighted by laterally continuous layers of woody debris that extend across gullies onto the open face of the moraine. The upper portion of the proximal face in this area is very steep (70-80°) and the basalt cliff at the base of the moraine prevents access from the glacier below, thus two adjacent gullies were investigated by rappelling down a rope anchored to trees at the base of the moraine's distal slope.

The two measured sections, termed PG-1 and PG-2 (Fig. 2.5), have nearly identical stratigraphy (Fig. 2.6). Till below the moraine crest is light gray and rich in granitic clasts (30-40%) up to 4 m in diameter. About 30 m below the moraine crest, this till (M1) overlies a prominent paleosol and woody horizon (Fig. 2.7a) that is traceable along both gullies and can be traced visually downvalley for several tens of metres. The paleosol, which includes a dark brown Ah or Oh horizon, is up to 20 cm thick and dips downvalley. Small, *in-situ* roots are common in the paleosol, and several large tree stems up to 50 cm in diameter and up to 173 years old lie directly on, and up to 1 m above, the paleosol. One of the stems, a ~15 cm diameter *Abies* stem with bark preserved on its lower side (Fig. 2.7b), yielded a radiocarbon age of 470±50 $^{14}C$ yr BP (GSC-6769). The stem was enclosed in till, 30 cm above the paleosol, but its excellent preservation, association with the prominent wood layer, and presence of bark suggest that it provides a direct date for deposition of the overlying till.

Unit M1 is underlain by about 6 m of till (M2), which is similar to M1 but contains lenses of oxidized, stratified medium to coarse sand. M2, in turn, sharply overlies another laterally continuous paleosol with abundant large woody debris (Fig. 2.7c). The light to dark brown paleosol is less continuous than the upper soil, but can nevertheless be traced across the two gullies. It has little apparent dip and was observed to intersect the more steeply dipping, upper woody horizon farther downvalley. The paleosol is mainly developed on till, although locally it overlies well-sorted medium to coarse sand up to 70 cm thick. A small twig recovered from the paleosol gave a radiocarbon age of 1527±41 $^{14}C$ yr BP (Wk-12310), which is a maximum age for the overlying till (M2). Deposition of M2 is further constrained by an age of 1390±50 $^{14}C$ yr BP (GSC-6760) from a 30-cm-diameter *Abies* stem lying directly on the paleosol. The tree from which this stem came was probably killed when it was buried by M2. Two large tree stems lying on the paleosol had 80 and 100 annual rings, respectively, suggesting that the glacier was sufficiently retracted prior to deposition of M2 that over 100 years of tree colonization could occur on the stabilized moraine surface.

The next unit in the moraine sequence consists of about 10 m of bouldery till (M3). This till overlies a discontinuous third paleosol and an associated layer of sparse woody debris. At PG-2, three large stems with up to 145 annual rings, one with preserved bark, were distributed in a sub-horizontal line across the steep gully wall (Fig. 2.7d). Limited excavations failed to locate a paleosol, but two of the three stems lay directly on 20-40-cm-thick bed of massive to faintly rippled, fine to medium sand. A third stem (*Tsuga*), entombed in till ~1.75 m above the sand, yielded a radiocarbon age of 2490±60 $^{14}C$ yr BP (GSC-6756). The woody horizon at PG-2 correlates with a paleosol/wood horizon at PG-1. The paleosol, a dark brown Ah horizon up to 7 cm thick, was excavated laterally over 3 m and traced across the broad gully as a sparse line of woody debris (Fig. 2.7e).

Investigations below M3 were aborted at PG-2 due to lack of rope. However, gentler slopes at PG-1 permitted access to a fourth paleosol/wood horizon over 65 m stratigraphically below the moraine crest (Fig. 2.7f). The overlying ~20 m of till (M4) are locally oxidized. The paleosol, which was traced across the entire downvalley side of the gully (>5 m horizontally), is a dark brown Om horizon up to 10 cm thick. Small rip-up clasts of peaty material are present in the overlying till, 2-3 cm above the contact. A radiocarbon age on a small wood fragment recovered from the paleosol indicates that the overlying M4 till was deposited after 3034±42 $^{14}C$ yr BP (Wk-12311). This age is statistically indistinguishable from an age of 2960±60 $^{14}C$ yr BP (GSC-6746), obtained from an *Abies* log with 47 annual rings lying on the paleosol, thus the ages are likely direct dates for deposition of M4. A fifth gray, granitic till (M5), less than 5 m thick, separates the fourth paleosol/wood horizon from what appears to be the dense, gray basal till at south gully. Thus, M5 was probably deposited during an earlier (pre ~3000 $^{14}C$ yr BP) phase of Holocene alpine glacier expansion.

*Middle gully*

A steep bedrock buttress (Fig. 2.2) truncates the lateral moraine about 1.7 km northwest of south gully. The moraine is less steep, and the distal slope distance from moraine crest to valley wall decreases, upvalley of the buttress. In contrast to the area near proximal gullies, where till accumulated on a bedrock bench, till upvalley of the buttress mantles steep granitic bedrock and is more easily mobilized by mass movement processes. Isolated detrital logs, which are rare downvalley of the bedrock buttress, are more common in the proximal moraine till.

About 500 m northwest of the bedrock buttress, the moraine is cut by a large stream. As at south gully, the lateral moraine sediments contain paleosols and wood that facilitate separation of till units (Fig. 2.8). The upper till (M1) is light gray and contains ~30% clasts up to 2 m in diameter. About 15 m below the moraine crest, M1 overlies a discontinuous oxidized horizon. The oxidized horizon dips distally toward a swale between the distal slope of the moraine and the valley wall to the east. Near the west end of the swale, the oxidized horizon thickens into a peat bed up to 40 cm thick containing interbeds and laminae of silt and sand in the lowest 20 cm. An age of 1090±50 $^{14}C$ yr BP (GSC-6606) from a branch near the top of the peat provides a maximum age for M1. However, the till was probably deposited about 440±60 $^{14}C$ yr BP (GSC-6604), the age of the large stem buried in M1 directly above the oxidized contact.

A distally thinning wedge of similar till (M2) separates the upper paleosol from a wood-bearing paleosol with sharp upper and lower contacts. The paleosol is a dark brown, compact, silty peat up to 15 cm thick that can be traced at least 10 m laterally. Where the peat is thickest, the upper 5 cm are rich in bryophytes and the lower 10 cm have common pebbles and granules and abundant silt and sand laminae. The advance that deposited M2 is younger than 1720±42 (Wk-12306) and 1600±70 $^{14}C$ yr BP (TO-9754), which are, respectively, the ages of charcoal and a branch recovered from the uppermost 3 cm of peat. Another till unit (M3), the uppermost 2 m of which are oxidized, underlies the peat and rests on steeply sloping bedrock. A wood fragment from the base of the peat overlying M3 has an age of 2443±40 $^{14}C$ yr BP (Wk-12307), indicating that M3 was deposited before that time.

*North gullies*

Several hundred metres northwest of middle gully, a series of adjacent stream-cut gullies provide excellent exposures through the moraine (Fig. 2.9). Here, the crest of the LIA lateral moraine stands only 3-5 m above its distal swale. Just east of the LIA moraine, a smaller, topographically subdued moraine ("outer moraine" of Walker 2003) impounds a small pond and bog. Basal peat in the bog dates to ~6200 $^{14}C$ yr BP (Walker 2003), thus the moraine is at least that old. A small stream draining the pond has incised though the main lateral moraine down to bedrock. This gully (NG-1) exposes ~15 m of till and stratified sand and gravel separated by two paleosols (Fig. 2.10). A wood-bearing light brown paleosol just below the moraine crest can be traced across gully interfluves into adjacent gullies. The paleosol has an apparent dip into the slope, suggesting that it developed on what was then the distal moraine surface. Near the proximal edge of the gullies, however, it defines a paleo-moraine crest. The paleosol is sharply overlain by ~30 cm of crudely stratified sand and gravel that pinch out to the east. Light gray till (M1) with clasts up to 50 cm in diameter, in turn, overlies these sediments and forms the crest of the LIA moraine.

A lower, wood-bearing paleosol is separated from the upper paleosol by up to ~10 m of gray till (M2) with 30-40% clasts up to 1 m in diameter. Oxidized lenses of cobble gravel are present in the lowest 2 m of M2. This till and M3, which underlies the paleosol, rest on steeply sloping granitic bedrock that crops out as high as 2 m below the moraine crest. The paleosol, a dark brown organic bed ~5 cm thick, can be traced upvalley into an adjacent gully. It contains abundant *in-situ* roots, and is associated with several large tree stems, with up to 104 annual rings, lying directly on its upper contact. A wood fragment from the paleosol yielded an age of 1549±45 $^{14}$C yr BP (Wk-12307), and the outer rings of an *Abies* log lying on the paleosol gave an age of 1700±80 $^{14}$C yr BP (GSC-6767). The calibrated 2$\sigma$ age ranges of the two ages overlap (Table 2.1), suggesting that the paleosol/forest horizon was buried by M2 about 1700-1550 $^{14}$C yr BP.

Sediments exposed in the gully directly downvalley of NG-1 (Figs. 2.10, 2.11; NG-2) shed additional light on the age of the upper till units at the north gullies (see also Walker 2003). The horseshoe-shaped gully cuts through a fragment of a lateral moraine ("middle moraine" of Walker 2003) that lies between the main and outer moraines. We excavated a paleosol beneath up to 125 cm of till (M1b) that form the middle moraine (Fig. 2.11, section NG-2b) and traced it downvalley to a point ~2 m below the LIA moraine crest. A thinner, poorly developed paleosol is developed on M1b. It defines a conspicuous paleo-moraine crest, drops distally to within ~40 cm of the lower paleosol, then rises to the surface near the proximal side of the middle moraine (Fig. 2.11, section NG-2a). At section NG-2a, the two paleosols are separated by sand and gravel that grade laterally into bouldery till (M1b) forming the paleo-moraine defined by the upper paleosol. A thin cap of bouldery till (M1a) forms the crest of the LIA moraine and, along with crudely stratified sand and gravel, fills the swale between the middle and LIA moraines, defined by the upper organic horizon.

Three radiocarbon ages from the lower paleosol constrain the age of the advance that deposited the middle moraine and hence unit M1b. A piece of charcoal from the lower paleosol (section NG-2b) yielded an age of 2086±49 $^{14}$C yr BP (Wk-12313). Two wood fragments from the same paleosol at section NG-2a gave ages of 890±40 and 1093±45 $^{14}$C yr BP (Beta-180885 and Wk-12308, respectively). The paleosol was buried by M1b at or after 890±40 $^{14}$C yr BP. We consider it unlikely that the paleosol persisted over ~1100 years, as implied by the radiocarbon age on charcoal at section NG-2b (i.e. between ~2100 and 900 $^{14}$C yr BP), because a thick till was deposited during that time in the adjacent gully (NG-1) and farther downvalley at middle and proximal gullies. The anomalously old age of Wk-12313 may be due to retransport of older charcoal (Hallet et al. 2003) or the inbuilt age of wood charcoal (Gavin 2001). Unit M1a could not be dated due to a lack of suitable material.

## Synthesis and regional correlation

The stratigraphy and radiocarbon ages at the four study sites provide a detailed chronology of Lillooet Glacier advances during the late Holocene (Fig. 2.12). The earliest Holocene till (M5), exposed at the base of the proximal gullies, is at least ~3000 $^{14}C$ yr old. Two younger tills (M4 and M3) at the proximal gullies were deposited ~3000 $^{14}C$ yr BP and at or after ~2500 $^{14}C$ yr BP, respectively. A pre-LIA advance deposited M2 at all sites except south gully. At north and middle gullies, M2 buried paleosols and forest vegetation ~1700-1550 $^{14}C$ yr BP. Farther downvalley, ages from a buried paleosol and forest horizon indicate that M2 was deposited slightly later, ~1400 $^{14}C$ yr BP. Advances during the LIA are constrained by radiocarbon ages on woody material at the base of M1 at proximal and middle gullies. The earliest LIA advances were well underway by ~450 $^{14}C$ yr BP, and there is evidence for later fluctuations at south gully.

*Latest Pleistocene or early Holocene advance*

The time of the glacier advance that deposited M5, the oldest morainal till at Lillooet Glacier, is poorly constrained. It may be associated with the regional Garibaldi phase of glacier expansion, between about 6000 and 5000 $^{14}C$ yr BP, when several glaciers in the southern Coast Mountains are known to have advanced over forested valley floors and nunataks (Ryder and Thomson 1986; Koch et al. 2003; Smith 2003). Evidence for a correlative advance in the Canadian Rocky Mountains is lacking, but one or more advances of this age are recognized in interior and coastal Alaska (Calkin 1988). Alternatively, M5 may be associated with a glacier advance during the "8200-year cold event" (Alley et al. 1997). However, we consider this possibility unlikely because glacier termini during this period were probably severely retracted due to regional warm and dry conditions (e.g. Clague and Mathewes 1989; Hallett et al. 2003), and because the cold event is thought to have been very short-lived (Baldini et al. 2002). M5 could also be associated with a latest Pleistocene glacier advance, possibly during the Younger Dryas chronozone. A large valley glacier near Squamish, in the southern Coast Mountains, readvanced shortly after 10,650 $^{14}C$ yr BP (Friele and Clague 2002). Evidence for a possibly correlative advance in the Canadian Rockies, termed the Crowfoot advance, is preserved in terminal moraines overlain by Mazama ash (Luckman and Osborn 1979), composite lateral moraines (Osborn et al. 2001) and downvalley lake sediments (Reasoner et al. 1994). We prefer the latter, latest Pleistocene interpretation for the age of M5 for several reasons. The outermost moraine at the north gullies is older than ~6200 $^{14}C$ yr BP, based on basal radiocarbon ages from a pond impounded by the moraine (Walker 2003). It is probably much older than 6200 $^{14}C$ yr BP, because peat would not likely have accumulated at the site of the pond under the warm, dry conditions of the early Holocene. The glacier advance that built the outer moraine was of comparable extent to the climactic LIA advance, so it almost certainly would have deposited till in the vicinity of the proximal gullies, where M5 is exposed. Since M5 is the only bouldery till unit between the fourth paleosol and the compact, sheared, silty gray till that we associate with late Wisconsin glaciation, it probably correlates with the outer moraine. Thus, M5 is older than ~6200 $^{14}C$ yr BP and probably predates the early Holocene xerothermic period.

*Middle Neoglacial advances*

The advances that deposited M4 (~3000 $^{14}C$ yr BP) and M3 (at or after ~2500 $^{14}C$ yr BP) are correlative to previously recognized periods of Neoglacial glacier expansion in the Canadian Cordillera. Ryder and Thomson (1986) defined the Tiedemann Advance in the southern Coast Mountains and placed it at 3300-1900 $^{14}C$ yr BP on the basis of radiocarbon-dated moraine exposures at Tiedemann and Gilbert glaciers. The advance is thought to have culminated about 2300 $^{14}C$ yr BP. Similar ages have been obtained on glacially overridden wood near Whistler, north of Vancouver (Koch et al. 2003). Farther north in the Coast Mountains, alpine glaciers were advancing around this time near Bella Coola (Desloges and Ryder 1990) and in the Boundary Ranges near Stewart (Clague and Mathews 1992; Clague and Mathewes 1996; D.J. Smith, personal communication, 2003). A similar phase of glacier expansion, termed the Peyto Advance (Luckman et al. 1993), is well documented in the Canadian Rocky Mountains (Osborn et al. 2001; Wood 2002; Luckman in press). Non-surging glaciers were also advancing during this interval in the St. Elias Mountains (Denton and Karlén 1977).

These regional records are broadly synchronous, but individual glacier chronologies are less so. The diachronous nature of glacial deposits and the distribution of datable material at any one site complicate inter-glacier comparison, as does the variable quality of dating control. Our results indicate that the Tiedemann Advance at Lillooet Glacier encompassed at least two periods of ice advance, separated by a short interval when the glacier was sufficiently retracted that at least part of the moraine surface was colonized by vegetation. The vegetated moraine surface was overridden during the later Tiedemann advance. Ring-counts of large tree stems that were incorporated into M3 suggest that at least 145 years elapsed between deposition of M4 and subsequent burial by M3 *ca.* 2500 $^{14}C$ yr BP. This is a minimum estimate only, as it does not consider time for moraine stabilization and vegetation colonization.

*Post-Tiedemann, pre-Little Ice Age advance*

An advance subsequent to the Tiedemann Advance, but prior to the onset of LIA activity, is recorded by time-transgressive deposition of M2 at proximal, middle, and north gullies. Counts of annual rings in tree stems recovered from the base of M2 suggest that recession of Lillooet Glacier following its Tiedemann-age advances was sufficient to allow at least 100 years of tree colonization on a stabilized moraine surface. The advance was underway by ~1700-1550 $^{14}C$ yr BP and culminated at or after ~1400 $^{14}C$ yr BP. This event is well documented in Alaska, where an advance occurred about 1500 $^{14}C$ yr BP at Tebenkof Glacier (Wiles et al. 1999), in the southern Kenai Mountains (Wiles and Calkin 1994), and in coastal Alaska (Calkin 1988; Calkin et al. 2001). Some glaciers in the Canadian Rockies may have advanced at this time, based on radiocarbon-dated detrital and *in-situ* wood at one site (Luckman in press). Evidence for a glacier advance at this time is sparse in the Coast Mountains. Ryder and Thomson (1986), expanding on earlier work by Fulton (1971), suggested that recession of Tiedemann Glacier from its maximum Neoglacial position was slow and pulsatory until at least ~1300 $^{14}C$ yr BP, based on limited stratigraphic and morainal evidence. Lichenometric evidence suggests that moraines stabilized at the same site at about this time (Larocque and Smith 2003). Glaciers in the Duffey Lake watershed, on the leeward side of the southern Coast Mountains are thought to have expanded ~1500 $^{14}C$ yr BP, based on increases in mineral sediment flux recorded in lake sediment cores (Menounos 2002). In the northern Coast Mountains, near Stewart, Frank Mackie Glacier was advancing at 1600 $^{14}C$ yr BP (Clague and Mathews 1992). Similar results are emerging at Bridge Glacier, 20 km northeast of Lillooet Glacier, where a forest was overridden by the glacier ~1500 $^{14}C$ yr BP (Allen and Smith 2003).

*Little Ice Age advances*

Advances during the LIA deposited M1 at all four sites. The forest bed buried by M1 is particularly well-developed at proximal gullies, and counts of annual rings from large tree stems indicate that over 173 years of forest development occurred on the stabilized moraine deposited during the earlier, pre-LIA advance. Our investigations, however, provide only limited insight into LIA fluctuations of Lillooet Glacier due to substantial overlap in the calibrated age ranges of radiocarbon-dated wood samples, particularly at south gully (Table 1). Despite this limitation, our results can be broadly compared to other records of LIA glaciation in the region. Although we are unable to determine when Lillooet Glacier LIA advances began, the advance that deposited M1 at middle and proximal gullies was well underway by ~450 $^{14}C$ yr BP (AD 1330-1630). Ryder and Thomson (1986), Ryder (1987), Desloges and Ryder (1990), and Larocque and Smith (2003) report radiocarbon ages on *in-situ*, glacially overridden wood material at other sites in the Coast Mountains that suggest LIA expansion occurred at about the same time as, or up to several centuries earlier than, deposition of M1 at Lillooet Glacier. More precise records of regional LIA glacier activity have been obtained using lichenometry and tree-ring dating of overridden and detrital wood (Smith and Laroque 1996; Smith and Desloges 2000; Koch et al. 2003; Larocque and Smith 2003; Lewis and Smith in press). It is beyond the scope of this paper to review the literature on LIA advances in the Coast Mountains and elsewhere in the North American Cordillera. We direct the reader to comprehensive reviews by Luckman (2000), Calkin et al. (2001), and Luckman and Villalba (2001).

**Late Holocene climate variability**

The broad similarity of our Lillooet Glacier chronology to other records of glacier advance in the Coast Mountains suggests that a common climatic forcing mechanism may have led to prolonged positive glacier mass balance. However, deconvolution of the climatic controls on past glacier advances is complicated (Bradley 1999). Paleoecological investigations are sources of additional paleoclimatic insight that complement chronologies of glacier advance.

Pollen and plant macrofossil data indicate that warm, dry conditions during the early Holocene (~9500-7000 cal yr BP) were gradually replaced by a warm, moist climate regime, which, in turn, gave way to cool, moist conditions characteristic of modern climate about 4000-3000 cal yr BP (e.g. Hebda 1995; McLachlan and Brubaker 1995; Hansen and Engstrom 1996; Spooner et al 1997; Heinrichs et al. 2002). Many of these authors note the apparent link between this climatic deterioration and the regional onset of Neoglaciation, but they are unable to resolve specific periods of glacier advance from their paleoecological data. This is probably due in part to the large ecological range of the regional flora, which would dampen the response of vegetation to high-frequency climate change (Whitlock and Grigg 1999). Site selection is likely an additional factor. Results from two recent palynological investigations of ponds directly outside Neoglacial moraines in the Coast Mountains suggest that rapid increases in *Alnus* pollen are associated with independently dated Tiedemann-aged glacier advances nearby (Walker 2003; T.A. Arsenault, personal communication, 2003).

Fire frequency reconstructions based on high-resolution charcoal data potentially offer more detailed records of climate variability. Whereas many palynological reconstructions provide little or no evidence for climatic fluctuations since ~4000-3000 cal yr BP, fire frequency data from the southern Coast Mountains exhibit sub-millennial variability that is broadly synchronous with Tiedemann-age and LIA glacier advances (Fig. 12; Hallett et al. 2003). The glacier and fire frequency records, however, are in slight disagreement between 1900 and ~1300 cal yr BP. Well-dated glacier advances occurred at this time at Bridge and Lillooet glaciers, and possibly in the Boundary Ranges, but Hallett et al. (2003) suggest that this period was characterized by drought and dry fuel conditions, perhaps due to strengthening of the Pacific High. The apparent occurrence of glacier advances during a period that was probably dominated by conditions leading to negative glacier mass balance during the ablation season may reflect the importance of winter climate, in particular precipitation, for producing positive net mass balance and hence glacier advance.

## Conclusions

Paleosols and horizons of woody debris exposed in the northeast lateral moraine of Lillooet Glacier provide a detailed chronology of late Holocene advances of the glacier. Four periods of glacier advance are recognized in the moraine stratigraphy: (1) an advance prior to ~3000 $^{14}C$ yr BP, and probably before 6200 $^{14}C$ yr BP; (2) two advances at ~3000 and ~2500 $^{14}C$ yr BP, corresponding to the regional Tiedemann Advance; (3) an advance between ~1700 and 1400 $^{14}C$ yr BP that buried organic horizons at almost all of the study sites; and (5) several advances during the Little Ice Age, after 470 $^{14}C$ yr BP. The record presented here is in broad agreement with, and improves the resolution of, existing glacier chronologies in the Coast Mountains. In particular, we present the first direct evidence for a glacier advance in the Coast Mountains after the Tiedemann advance and before regional LIA glacier expansion. Detailed paleoclimatic interpretation of the glacier record is difficult, but comparison with paleoecological data, particularly fire frequency reconstructions, suggests that winter climate may have played an important role in controlling lengthy periods of positive net balance that would lead to glacier advance. Stratigraphic evidence for multiple glacier advances is exceptionally well preserved at Lillooet Glacier, and our results and earlier successes by other workers (e.g. Ryder and Thomson 1986; Osborn et al. 2001) highlight the importance of moraine stratigraphy in developing Holocene glacier chronologies. Future investigations of Holocene glaciation in the Coast Mountains, where composite moraines are relatively common, will likely benefit from detailed examination of moraine stratigraphy.

**Table 2.1.** Radiocarbon ages, Lillooet Glacier.

| $^{14}C$ age (yr BP)[a] | Calibrated age range (cal yr before AD 1950)[b] | Laboratory number[c] | Sample code[d] | Dated material[e] | Significance[f] |
|---|---|---|---|---|---|
| | | | *South gully* | | |
| 10±50 | 260 – 0 | GSC-6600 | CIA-01-18-1 | Root from paleosol | |
| 170±60 | 300 - 0 | GSC-6602 | CIA-01-18-4 | Root from paleosol | |
| 290±60 | 500 - 0 | TO-9753 | CIA-01-18-3 | Root or branch from paleosol | |
| | | | *Proximal gullies* | | |
| 470±50 | 620 - 330 | GSC-6769 | AVR-02-43-3 | Log (*Abies* sp.) in till (outer 25 rings) | M1-direct |
| 1390±50 | 1410 - 1180 | GSC-6760 | AVR-02-43-8 | Log (*Abies* sp.) on paleosol (outer 15 rings) | M2-direct |
| 1527±41 | 1520 - 1330 | Wk-12310 | AVR-02-43-6 | Twig from paleosol | M2-max |
| 2490±60 | 2740 - 2360 | GSC-6756 | AVR-02-51-1 | Log (*Tsuga* sp.) in till (outer 10 rings) | M3-max |
| 2960±60 | 3320 - 2950 | GSC-6746 | AVR-02-52-2 | Log (*Abies* sp.) on paleosol (outer 10-15 rings) | M4-max |
| 3034±42 | 3360 - 3080 | Wk-12311 | AVR-02-52-1 | Wood fragment from paleosol (upper 2 cm) | M5-min |
| | | | *Middle gully* | | |
| 440±60 | 620 - 320 | GSC-6604 | CIA-01-21-2 | Log on paleosol (outer 5 rings) | M1-direct |
| 1090±50 | 1170 - 930 | GSC-6606 | CIA-01-21-3 | Branch on peat (outer 10 rings) | M1-max; M2-min |
| 1600±70 | 1690 - 1330 | TO-9754 | CIA-01-21-1 | Branch from top of paleosol (outer 5 rings) | M2-direct |
| 1720±42 | 1710 - 1530 | Wk-12306 | CIA-01-21-7 | Charcoal from paleosol (upper 3 cm) | M2-max |
| 2443±40 | 2710 - 2350 | Wk-12307 | CIA-01-21-8 | Wood fragment from paleosol (lower 3cm) | M3-min |

**Table 2.1. (cont.)** Radiocarbon ages, Lillooet Glacier

| | | | | | |
|---|---|---|---|---|---|
| | | | *North gullies* | | |
| 890±40 | 920 - 710 | Beta-180885 | LG-24-2 | Wood fragment from paleosol (upper 3 cm) | M1b-max |
| 1093±45 | 1170 - 930 | Wk-12308 | AVR-02-41-3 | Wood fragment from paleosol (upper 3 cm) | M1b-max |
| 1549±45 | 1530 - 1330 | Wk-12309 | AVR-02-42-2 | Wood fragment from paleosol (upper 1 cm) | M2-max |
| 1700±80 | 1820 - 1420 | GSC-6767 | AVR-02-42-6 | Log (*Abies* sp.) on paleosol (outer 20 rings) | M2-direct |
| 2086±49 | 2300 - 1900 | Wk-12313 | LG-24-1 | Charcoal from paleosol (upper 3 cm) | reworked |

[a] Error terms are $1\sigma$ for Beta, TO and Wk ages, and $2\sigma$ for GSC ages. GSC, TO and WK ages are normalized to $\delta^{13}C$ = -25.0‰ PDB. Beta, TO and Wk ages were derived using the AMS technique. GSC ages are derived from a proportional gas counter.

[b] Determined from the decadal data of Stuiver et al. (1998) using the probability distribution method within the program CALIB 4.4 (Stuiver and Reimer, 1993). Age ranges are $\pm 2\sigma$ and were calculated with an error multiplier of 1.0.

[c] Beta - Beta Analytic; GSC - Geological Survey of Canada; TO - IsoTrace Laboratory, University of Toronto; Wk - University of Waikato, New Zealand.

[d] Collector: AVR - A.V. Reyes; CIA - J.J. Clague; LG – L.A. Walker.

[e] Wood identification by R.J. Mott.

[f] M1, M2, etc. are lithostratigraphic units from text. Max, min and direct refer to maximum, minimum and direct ages for the corresponding lithostratigraphic unit.

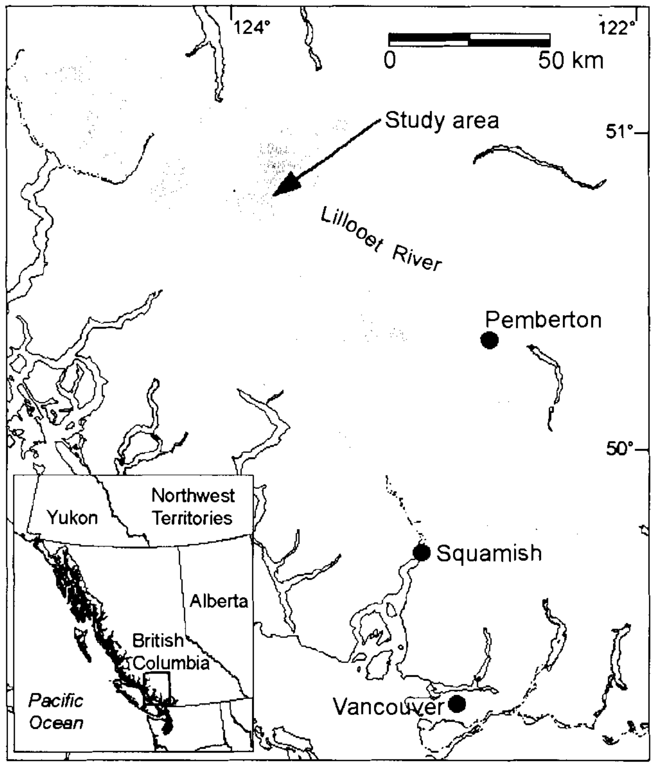


**Figure 2.1.** Map showing location of Lillooet Glacier. Shaded gray areas are glaciers and icefields.

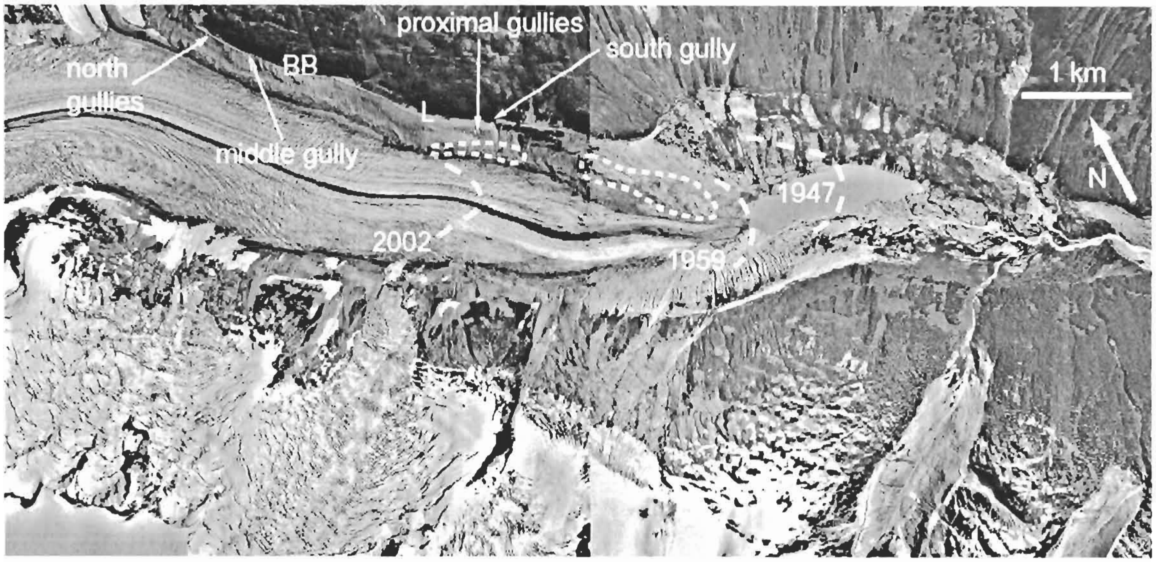


**Figure 2.2.** Airphoto mosaic, showing Lillooet Glacier terminus and forefield in 1965, and localities mentioned in the text. Approximate ice limits in 1947, 1959 and 2002 are delineated by long dashed lines. The 1947 and 1959 ice limits are from British Columbia airphoto BC478-17 and Hutchison (1961), respectively. Short dashed lines enclose extensive outcrops of jointed basalt. L is the lichen sampling site and BB is the prominent bedrock buttress. British Columbia airphotos BC5149-052, 053; reproduced with permission of British Columbia Ministry of Sustainable Resource Management, Base Mapping and Geomatic Services Branch.

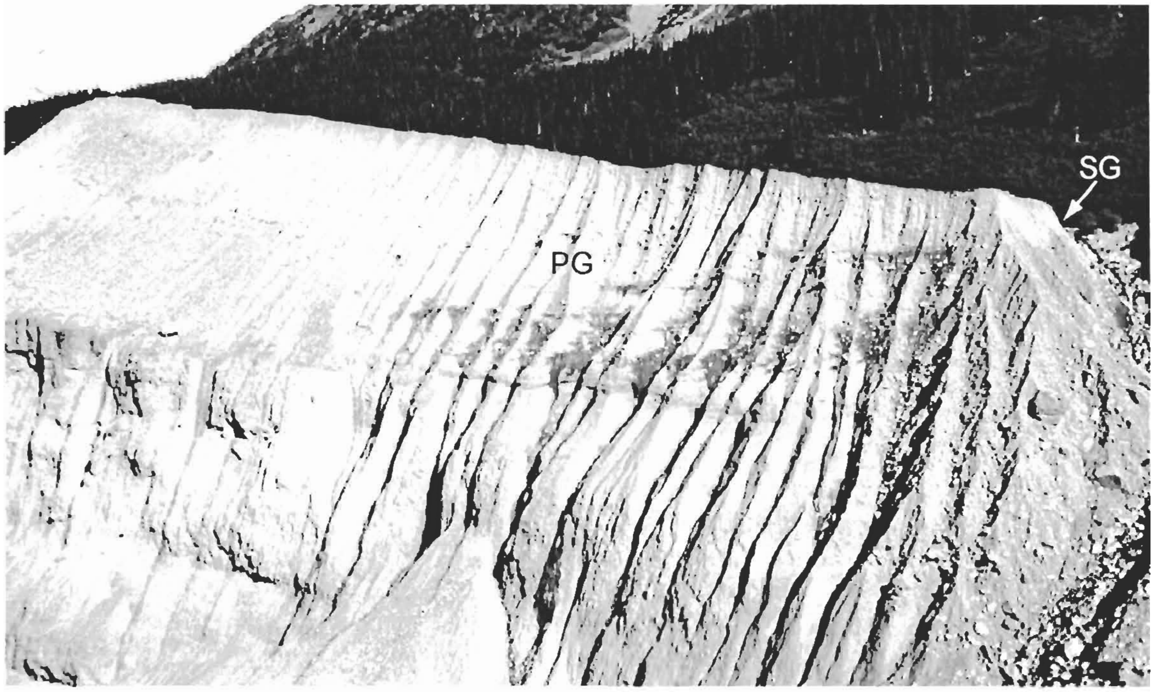


**Figure 2.3.** Proximal face of the northeast lateral moraine in the vicinity of south gully (SG) and proximal gullies (PG).

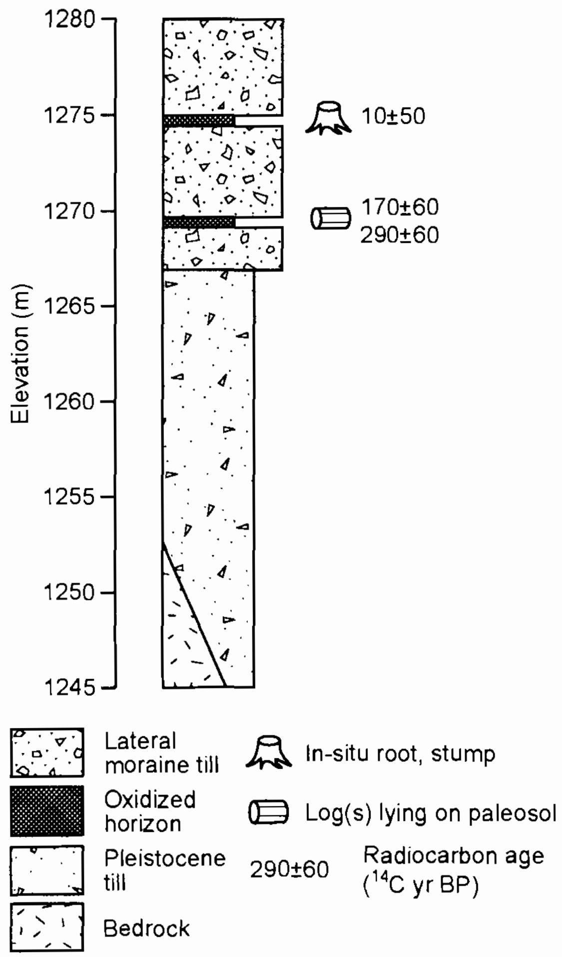


**Figure 2.4.** Stratigraphy of sediments exposed at south gully. Thickness of oxidized horizons is not to scale.

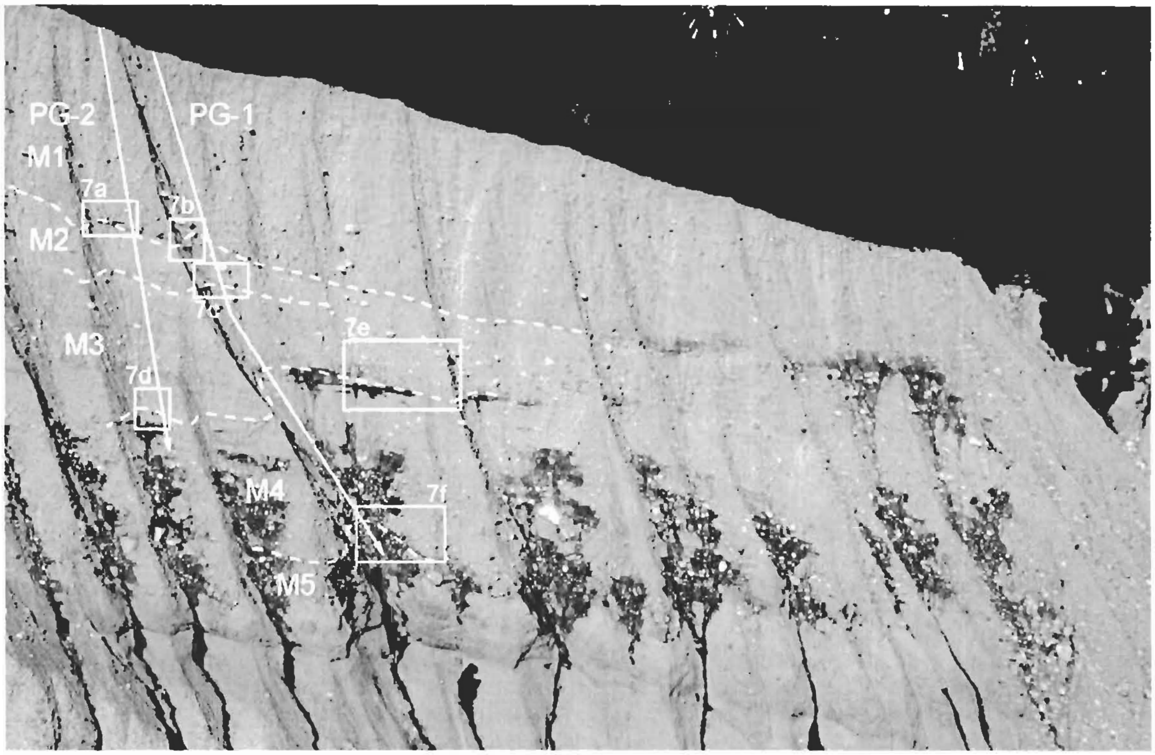


**Figure 2.5.** Locations of measured sections at proximal gullies. M1, M2, etc. are lithostratigraphic units from Figure 2.6. Dashed lines are paleosol/wood horizons shown in Figure 2.6. Boxes are locations of Figures 2.7a-f.

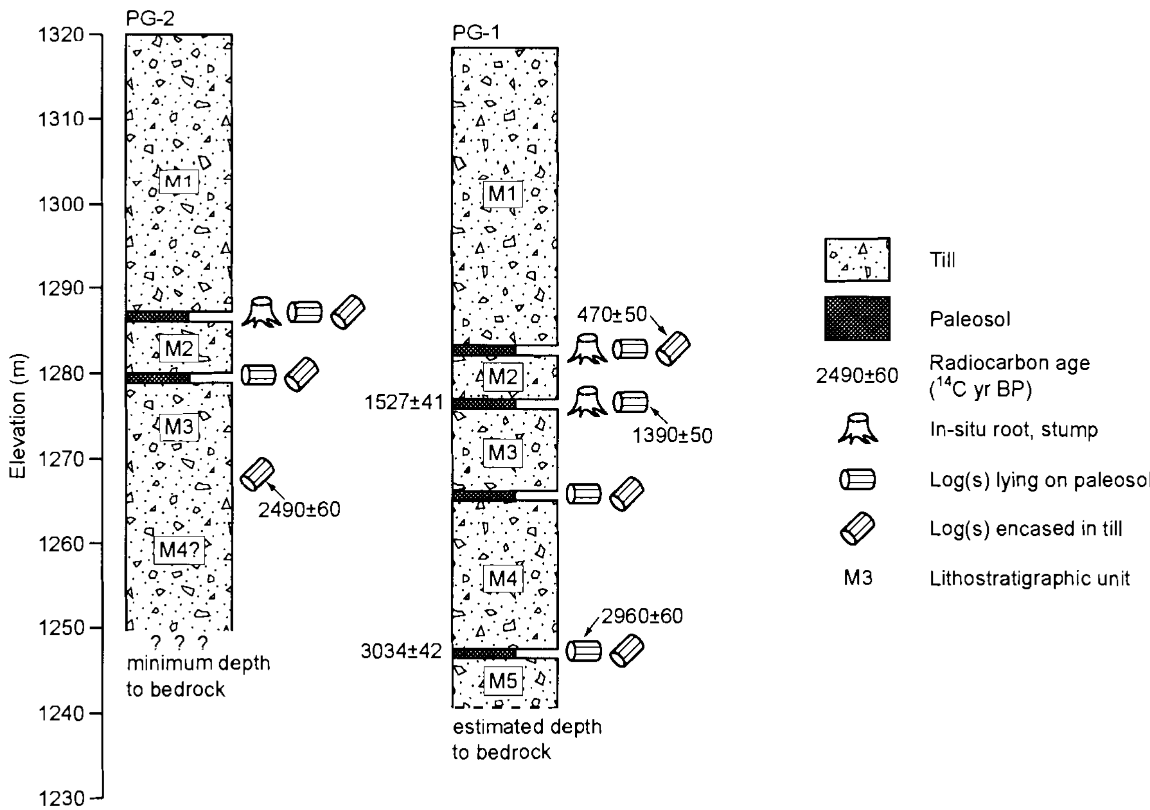


**Figure 2.6.** Stratigraphy of sediments exposed at proximal gullies. Paleosol thickness is not to scale. Radiocarbon ages to the left of PG-1 are on macrofossils recovered from paleosols.

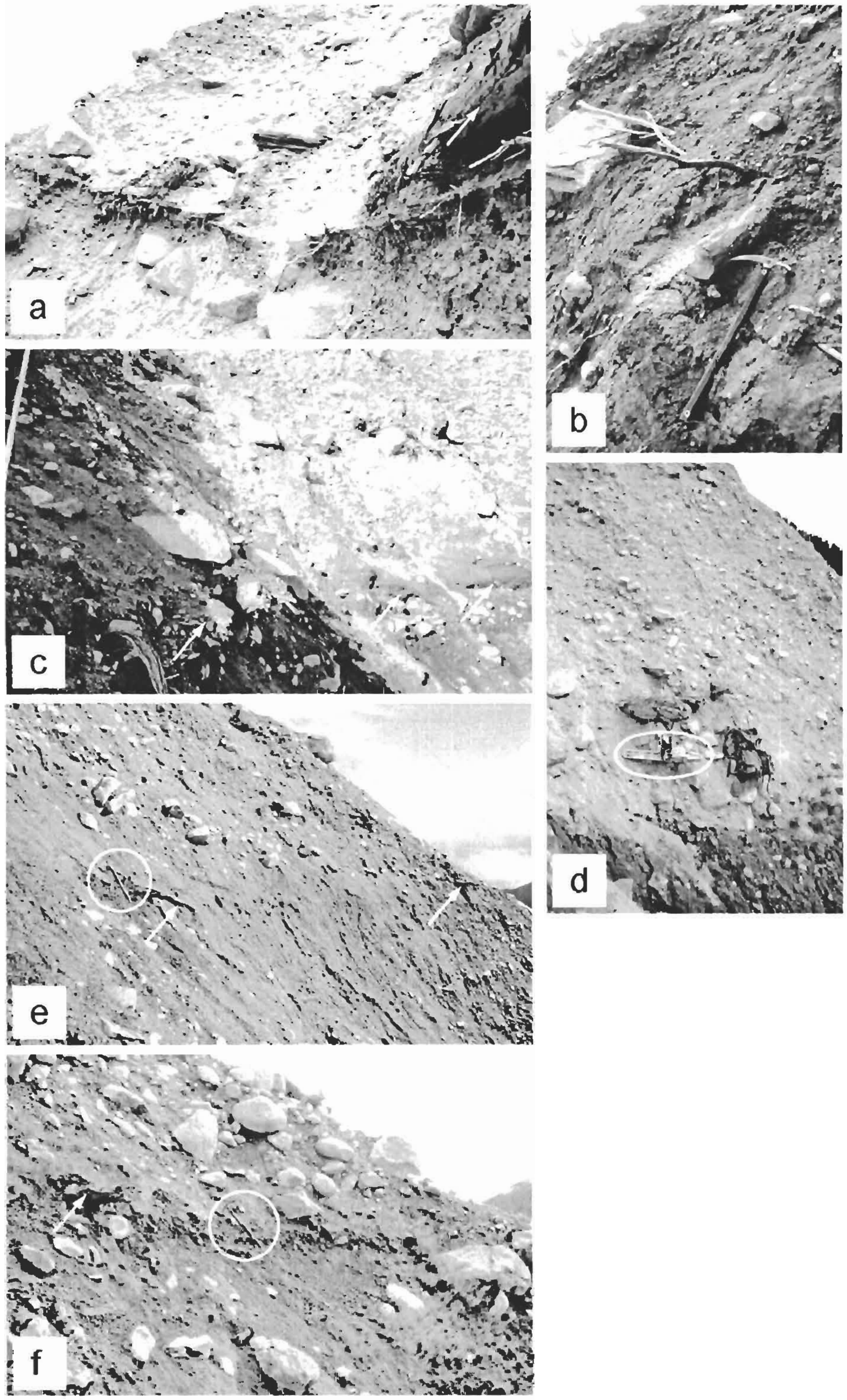

**Figure 2.7.** Paleosols and wood horizons exposed in northeast lateral moraine at proximal gullies. (*a*) Uppermost paleosol at PG-2. Tree stem at far right (arrow) is ~50

cm in diameter. (*b*) Uppermost paleosol at PG-1. Cut stem yielded a radiocarbon age of 470±50 $^{14}$C yr BP (Table 1). (*c*) Second paleosol at PG-1 (arrows). (*d*) Lowest wood horizon at PG-2. Chainsaw (circled) is ~80 cm long. (*e*) Third paleosol at PG-1. Wood lying on paleosol is marked by arrows. (*f*) Fourth (lowest) paleosol at PG-1. Stem at left (arrow) gave a radiocarbon age of 2960±60 $^{14}$C yr BP (Table 1). Circled ice axe in (*b*), (*e*) and (*f*) is 70 cm long.

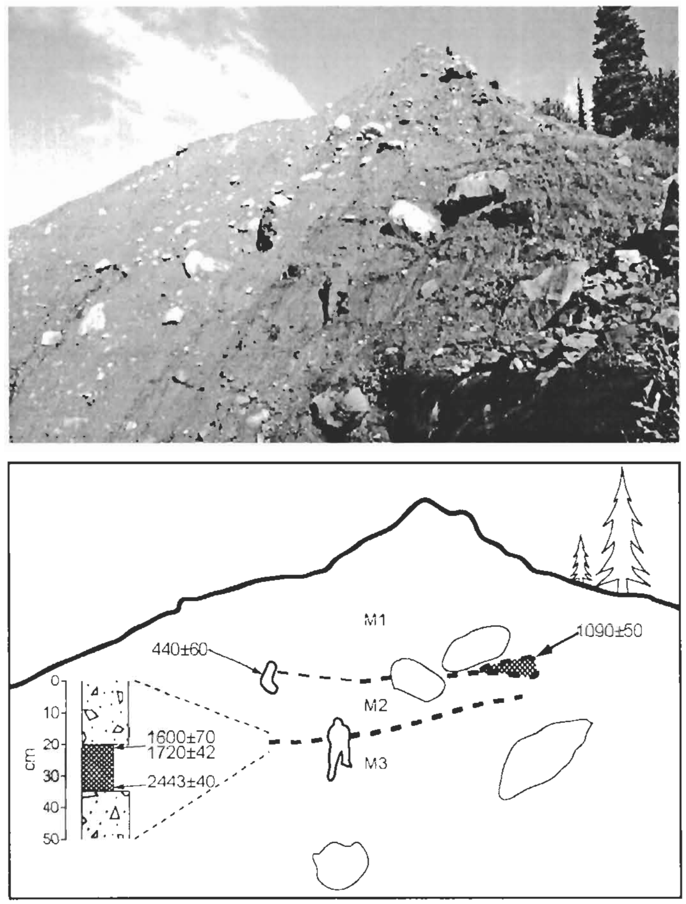


**Figure 2.8.** Sediments exposed at middle gully; view upvalley. Thick dashed lines mark paleosols. The thin dashed line indicates an oxidized horizon. Stratigraphic symbols and units as in Figure 2.6.

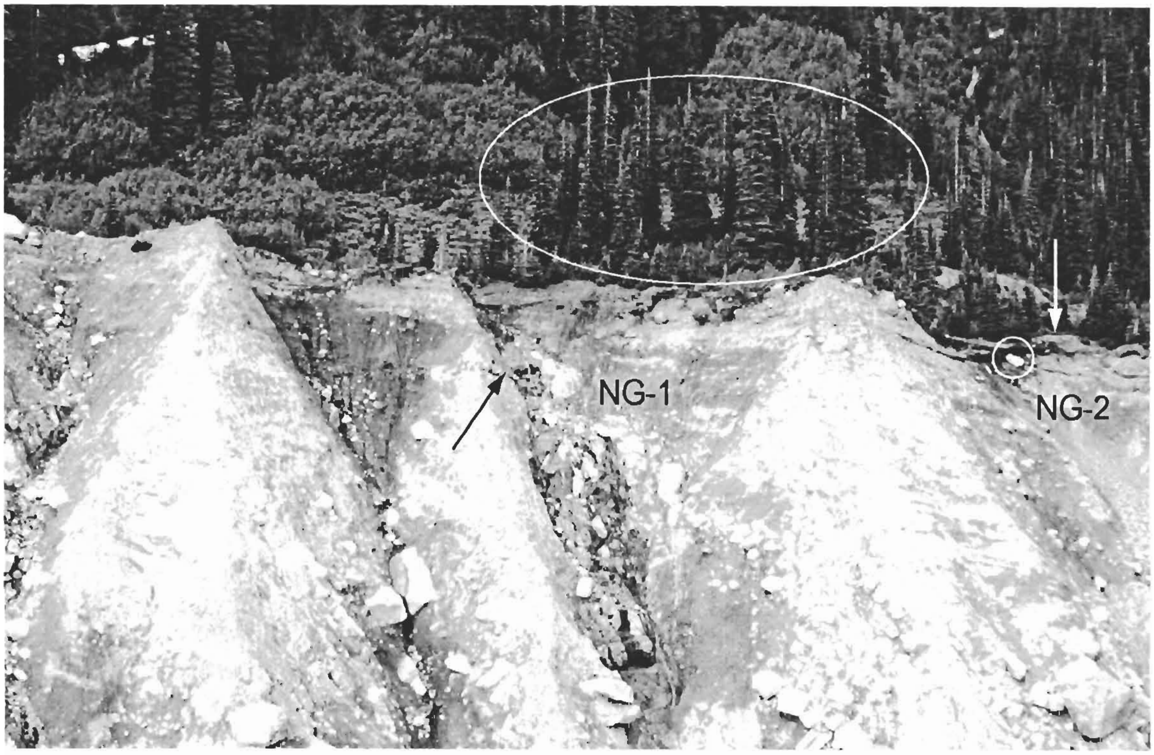


**Figure 2.9.** Proximal face of lateral moraine at north gullies. The locations of the outer and middle moraines are marked by a prominent stand of subalpine fir (circled) and a white arrow, respectively. The white boulder at far right (circled) is also shown in Figure 2.11. The location of dated samples at NG-1 is indicated by a black arrow.

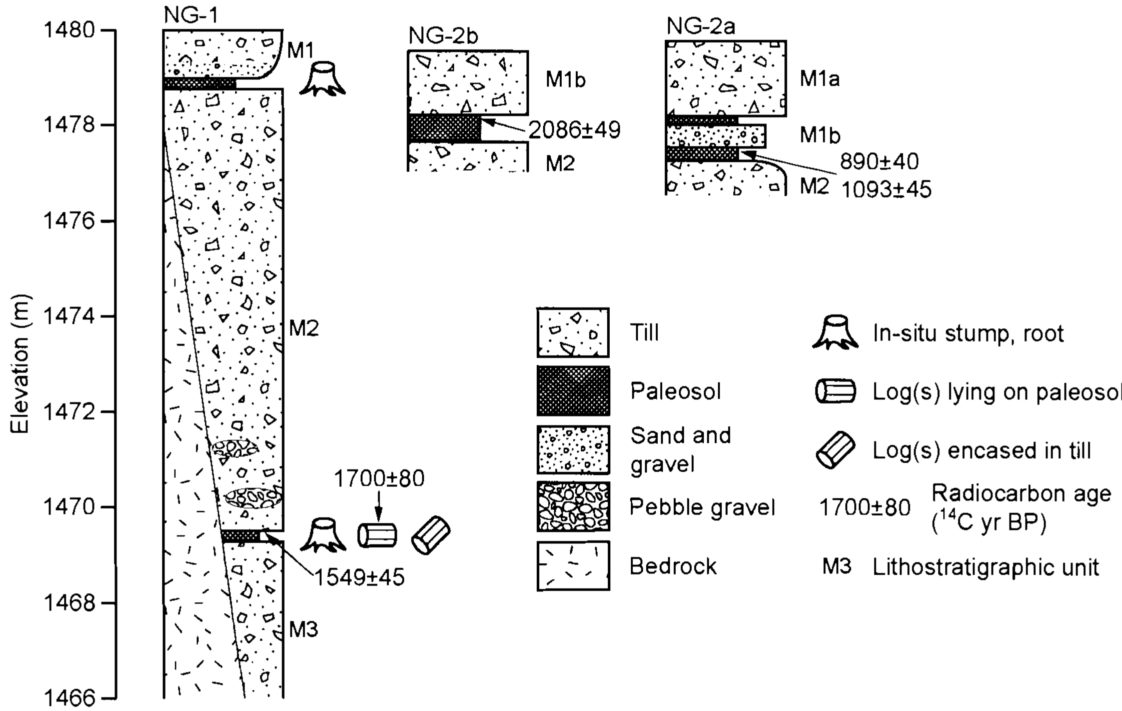


**Figure 2.10.** Stratigraphy of sediments exposed at north gullies. Paleosol thickness not to scale.

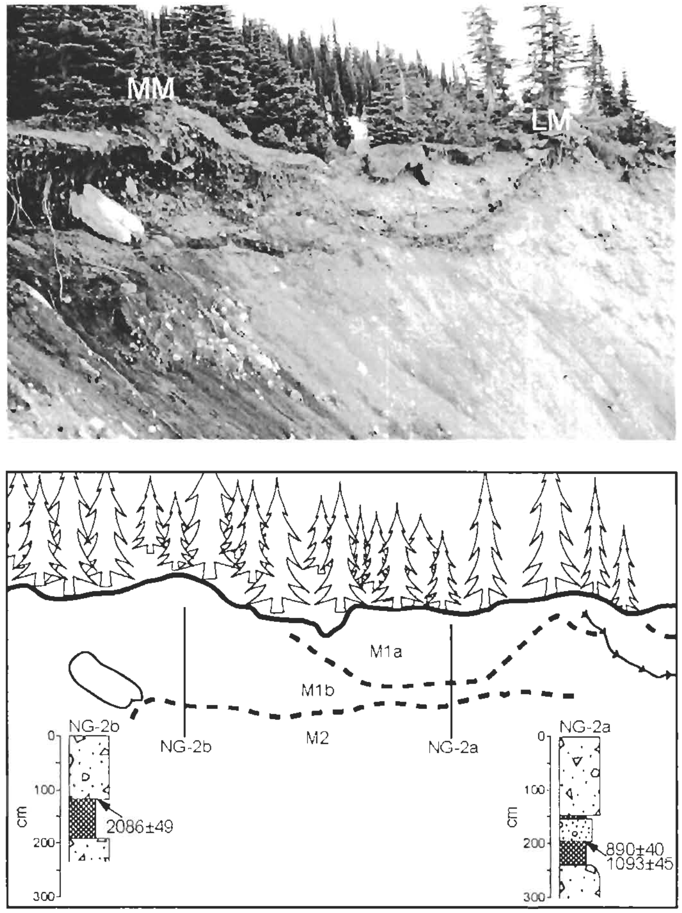


**Figure 2.11.** Sediments exposed at NG-2. MM is the middle moraine and LM is the LIA moraine. Dashed lines mark locations of excavated paleosols, and stratigraphic symbols are as Figure 2.6. Solid arrowed line at far right marks the ridge crest separating NG-2 from the adjacent downvalley gully. The white boulder at far left is also shown in Figure 2.9. Stratigraphic symbols and units as in Figure 2.10.

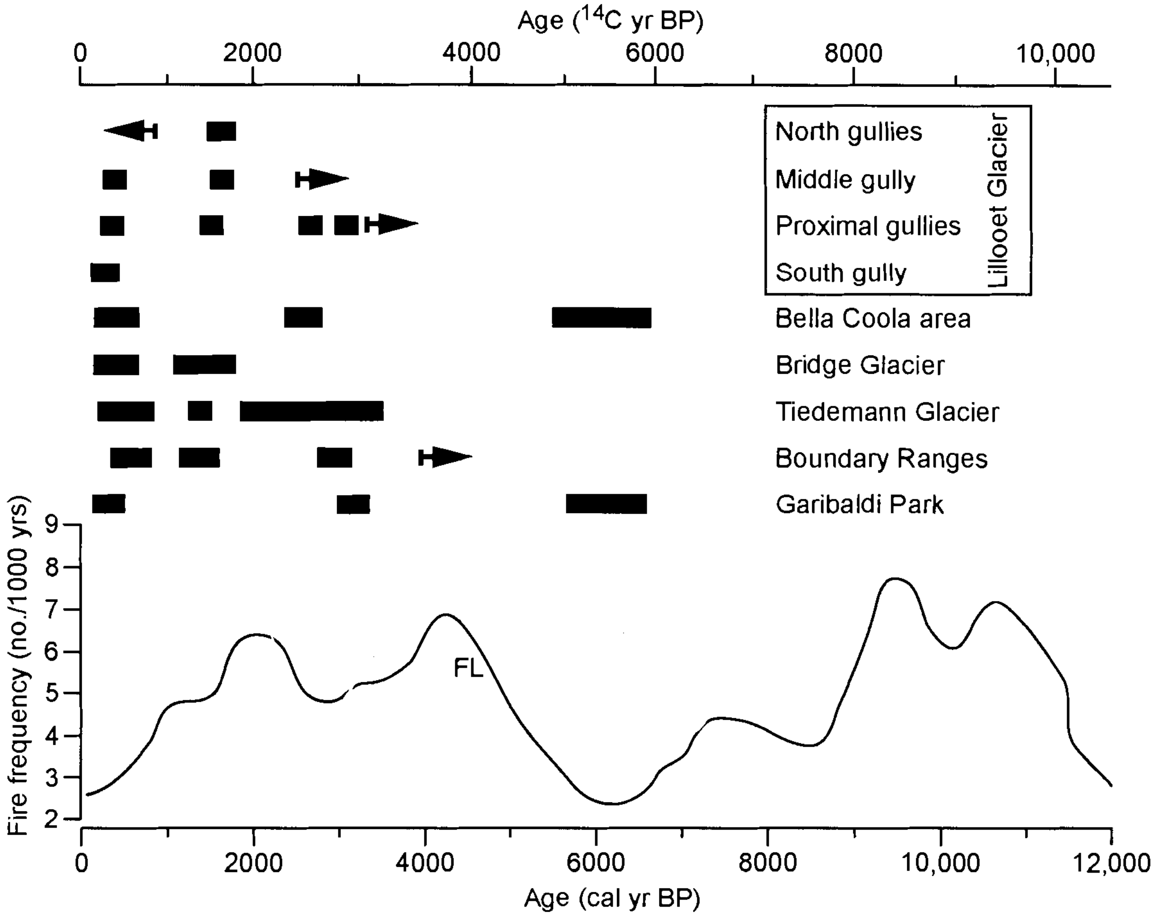


**Figure 2.12.** Summary chronology of dated Holocene advances at selected sites in the Coast Mountains (top) and southwestern British Columbia Holocene fire frequency (bottom; Hallett et al. 2003). FL is Frozen Lake (Coast Mountains) and MBC is Mount Barr cirque (Cascade Mountains). Horizontal bars represent dated periods of glacier advance. Left and right arrows represent, respectively, maximum and minimum ages for glacier advances. Vertical shaded bars mark generalized periods of glacier advance in the Coast Mountains. Sources of glacier data: Lillooet Glacier (this study); Bella Coola area (Desloges and Ryder 1990; Smith and Desloges 2000; Smith 2003; D.J. Smith, personal communication, 2003); Bridge Glacier (Ryder 1987; Allen and Smith 2003; D.J. Smith, personal communication, 2003); Tiedemann Glacier (Ryder and Thomson 1986; Larocque and Smith 2003); Boundary Ranges (Clague and Mathews 1992; Clague and Mathewes 1996; D.J. Smith, personal communication, 2003); Garibaldi Park (Ryder and Thomson 1986; Koch et al. 2003).

# CHAPTER 3. The Bridge Advance: a pre-Little Ice Age advance of alpine glaciers in the Coast Mountains of British Columbia

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## Abstract

New and previously published data from four sites in the Coast Mountains of British Columbia provide evidence for a previously unrecognized period of glacier advance several hundred years prior to the beginning of the Little Ice Age. We name this event the Bridge Advance. Radiocarbon ages on buried, *in-situ* tree stems indicate that Bridge Glacier in the southern Coast Mountains was advancing 1500 $^{14}$C yr BP. A paleosol and buried forest bed exposed in a composite lateral moraine at nearby Lillooet Glacier suggest that soil development and colonization of morainal surfaces following the mid-Neoglacial Tiedemann Advance were interrupted by a glacier advance that began as early as ~1700 $^{14}$C yr BP and culminated after 1400 $^{14}$C yr BP. Farther north, lichen-dated lateral moraines at Tiedemann Glacier stabilized about 1330 cal yr BP. An advance of Frank Mackie Glacier, in the northern Coast Mountains, about 1600 $^{14}$C yr BP, blocked Bowser valley and impounded Tide Lake. The Bridge Advance is coincident with a period of increased summer drought inferred from fire frequency.  North Pacific ocean-atmosphere circulation patterns control the present mass balance regimes of glaciers in the region, and changes in the intensity of these circulation patters may provide a plausible forcing mechanism for late Holocene glacier advances in the region, particularly during coincident periods of summer drought.

## Introduction

Advances of alpine glaciers during the Little Ice Age (LIA), here defined as the period between *ca.* AD 1200 and 1900 (Grove 1988), commonly destroyed geomorphic evidence of earlier Holocene glacier activity. Thus, many regional chronologies of Holocene glaciation prior to the LIA are limited to sparse, poorly dated records. Glaciers are good indicators of climate because changes in their mass balance, and thus extent, are an integrated response to climate forcing through temperature and precipitation. For this reason, reconstructions of historic and prehistoric glacier activity have been the focus of many paleoclimatic investigations. However, the equivocal nature of many glacier records, particularly for the period prior to the LIA, can frustrate efforts to infer past climate and to compare the glacier record to other proxy paleoenvironmental indicators.

This limitation has become particularly acute in light of recent efforts to evaluate climatic variability throughout the Holocene. Proxy indicators in ice cores, marine sediments, lake cores, and tree-rings, for example, demonstrate that Holocene climate was variable at millennial and sub-millennial time scales (e.g. O'Brien et al. 1995; Bond et al. 2001; Grudd et al. 2002; Viau et al. 2003).

Recent high-resolution paleoenvironmental investigations in western Canada and Alaska have shown that late Holocene climate in this region was also characterized by sub-millennial climate variability (e.g. Pienitz et al. 2000; Hu et al. 2001, 2003; Finney et al. 2002; Menounos 2002; Hallett et al. 2003). Several decades of research in Alaska have led to the development of detailed chronologies of glacier activity (see reviews in Calkin 1988; Calkin et al. 2001) that can be compared to these high-resolution paleoenvironmental proxy records. In comparison, the Holocene glacier record in the

British Columbia Coast Mountains is poorly resolved. Glacier advances during the middle Neoglacial period, locally known as the Tiedemann Advance (Ryder and Thomson 1986), are documented at several sites in the Coast Mountains (Ryder and Thomson 1986; Desloges and Ryder 1990; Clague and Mathews 1992; Clague and Mathewes 1996; Koch et al. 2003). However, the full complexity of the Tiedemann Advance is only now being appreciated, and evidence for later, pre-LIA advances has proven elusive.

In this paper, we present evidence from four sites in the Coast Mountains for a period of glacier expansion about 1700-1400 $^{14}C$ years ago, which we term the Bridge Advance. This phase of glacier activity followed the Tiedemann Advance, but preceded the LIA advances of the last millennium. Our synthesis combines previously published investigations at Frank Mackie Glacier (Clague and Mathews 1992) and Tiedemann Glacier (Fulton 1971; Ryder and Thomson 1986; Larocque and Smith 2003) with new work at Lillooet Glacier (Reyes and Clague submitted) and nearby Bridge Glacier. We discuss the paleoclimatic implications of a glacier advance at this time and speculate on possible climate forcing mechanisms in the context of the relation between regional glacier mass balance and North Pacific ocean-atmosphere dynamics.

**Evidence for the Bridge Advance**

Radiocarbon ages relevant to the Bridge Advance are presented in Table 3.1. Radiocarbon ages are presented in the text as both uncalibrated $^{14}C$ years ($^{14}C$ yr BP) and calibrated years before AD 1950 (cal yr BP). Non-radiocarbon chronological data, i.e.

lichen ages, are presented as calibrated years before AD 1950 (cal yr BP) to facilitate comparison with radiocarbon-based chronologies.

*Bridge Glacier*

Bridge Glacier (50°49'N, 123°29'W) is a large valley glacier draining an icefield in the Pacific Ranges of the southern Coast Mountains, about 175 km north-northwest of Vancouver, British Columbia (Fig. 3.1). Little Ice Age advances of Bridge Glacier impounded an ice-marginal lake in a tributary valley. The lake probably drained catastrophically due to breaching of moraine and ice dams during recession from the LIA ice limit (Ryder 1991). Relatively little is known of the Holocene history of Bridge Glacier (Ryder and Thomson 1986). Relevant results of ongoing investigations by researchers at the University of Victoria Tree-Ring Laboratory (Allen and Smith 2003) are presented here and will be described in greater detail in a forthcoming thesis[2].

Recession of Bridge Glacier from its maximum LIA position has exposed scattered detrital wood and *in-situ* mats of buried tree stems in the glacier forefield. A sheared whitebark pine (*Pinus albicaulis*) snag in growth position, exposed in an abandoned proglacial stream channel about 200 m north of the 2002 glacier snout, yielded a radiocarbon age of 1500±50 $^{14}C$ yr BP (1520-1310 cal yr BP; Beta-171549). Several large boulders are lodged against the snag (Fig. 3.2a), thus it was not possible to determine the depth to its rooting horizon. In any case, the tree was killed about 1500 $^{14}C$ yr BP when Bridge Glacier advanced over the site and sheared off the top of the tree stem.

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Recent fluvial incision about 500 m downvalley of the standing snag has exposed a cluster of sheared subfossil tree stems (Fig. 3.2b). The stems have been pushed downvalley into a paleosol developed on diamicton and are oriented downvalley. They are buried in till associated with the glacier advance that knocked the trees down, but subsequent fluvial action has eroded the till and covered it with a thin veneer of poorly sorted gravel. One of the buried stems gave a radiocarbon age of 1040±50 $^{14}$C yr BP (1060-790 cal yr BP; Beta-181856). This age is either a maximum age for the culmination of the Bridge Advance or a minimum age for initiation of LIA expansion of Bridge Glacier.

*Lillooet Glacier*

Lillooet Glacier (Fig. 3.1; 50°45'N, 123°46'W) flows southeast from icefields and terminates in the headwaters of Lillooet River, about 20 km southwest of the Bridge Glacier terminus. Buried paleosols and prominent horizons of large woody debris exposed in the northeast lateral moraine indicate that the moraine is a composite feature composed of drift deposited by up to five Holocene advances. The moraine stratigraphy and Holocene chronology of Lillooet Glacier are described in detail by Reyes and Clague (submitted). Here, we limit our discussion to stratigraphy and radiocarbon ages pertaining to the Bridge Advance.

The Bridge Advance deposited till that buried a paleosol and forest vegetation at three sites along the northeast lateral moraine. At the farthest upvalley site, termed north gullies (NG; Fig. 3.3), till was deposited on top of a paleosol after 1549±45 $^{14}$C yr BP (1530-1330 cal yr BP; Wk-12309), the age of a wood fragment from the uppermost part of the paleosol. The outer rings of a log encased in till but lying on the palesol yielded an age of 1700±80 $^{14}$C yr BP (1820-1420 cal yr BP; GSC-6767). The calibrated ages of these two samples are statistically indistinguishable at the 2σ level (Table 3.1), and they are interpreted as direct dates for the glacier advance that deposited till at north gullies.

Till and paleosols are also exposed in a stream-cut gully through the lateral moraine ~350 m southeast of north gullies. At this exposure, termed middle gully (MG; Fig. 3.3), the upper 3 cm of a laterally extensive paleosol yielded charcoal that gave an age of 1720±42 $^{14}$C yr BP (1710-1530 cal yr BP; Wk-12306). A branch in the paleosol yielded an age of 1600±70 $^{14}$C yr BP (1690-1330 cal yr BP; TO-9754), suggesting that Bridge Advance till buried the paleosol at this site at or shortly after that time. A higher paleosol grades laterally into a peat sequence up to 40 cm thick. Both the paleosol and the peat are developed on Bridge Advance till and are, in turn, overlain by LIA till, which forms the present moraine crest (Reyes and Clague submitted).

About 2 km downvalley, shallow gullies incised into the proximal face of the lateral moraine (proximal gullies, PG, Fig. 3.3) expose five Holocene tills separated by laterally extensive paleosols and associated lines of woody debris (Reyes and Clague submitted). One of the paleosols, about 42 m below the moraine crest, is covered by till associated with the Bridge Advance. A twig from the paleosol and a log lying directly on the paleosol yielded ages of 1527±41 $^{14}$C yr BP (1520-1330 cal yr BP; Wk-12310) and 1390±50 $^{14}$C yr BP (1410-1180 cal yr BP; GSC-6760), respectively. These ages suggest that Bridge Advance till buried the paleosol and wood horizon at proximal gullies ~1400 $^{14}$C yr BP. As at middle gully, the Bridge Advance till at proximal gullies is overlain by a paleosol and abundant woody debris that were subsequently buried by a LIA advance of Lillooet Glacier (Reyes and Clague submitted).

*Tiedemann Glacier*

Tiedemann Glacier (Fig. 3.1; 51°21'N, 124°56'W) emanates from an icefield surrounded by the highest peaks in the Coast Mountains (4000 m asl) and flows 24 km east before terminating in an extensive zone of stagnant ice and ice-cored supraglacial debris. It is one of the few sites in the Coast Mountains where there is clear geomorphic evidence for a pre-LIA, Holocene glacier advance (Tiedemann Advance of Ryder and Thomson 1986). Ryder and Thomson (1986) suggested that recession of Tiedemann Glacier from its maximum Neoglacial position ~2300 $^{14}C$ yr BP was slow and pulsatory until at least 1330±65 $^{14}C$ yr BP (1350-1080 cal yr BP; S-1473). This is based on the age of a transported log in ice-marginal fluvial sediments in a lateral moraine exposure (Site 1, unit F2 of Ryder and Thomson 1986). Several fragmented moraines ("middle moraines" of Ryder and Thomson 1986), located between the LIA and Tiedemann-age moraines, were deposited during this period, based on an age of 1270±140 $^{14}C$ yr BP (1510-920 cal yr BP; GSC-977) on basal peat from a bog at the distal side of one of the middle moraines ("inner moraine bog" of Fulton 1971; Ryder and Thomson 1986). Larocque and Smith (2003) suggested that this moraine stabilized about 1330 cal yr BP, using a linear extrapolation of a Coast Mountain lichen growth curve modified from Smith and Desloges (2000). The modified growth curve includes a local calibration point dated at 540±60 $^{14}C$ yr BP (Larocque 2003).

*Frank Mackie Glacier*

Indirect evidence for the Bridge Advance is also present in the Boundary Ranges, about 500 km north of Tiedemann Glacier. The Neoglacial history of Frank Mackie Glacier (Fig. 3.1; 56°20'N, 130°05'W) has been reconstructed from sediments of Tide

Lake (Clague and Mathews 1992). The lake, which at its maximum extent had a volume of ~1 $km^3$, formed when Frank Mackie Glacier advanced far enough to block drainage of Bowser River. Tide Lake last drained around AD 1930, and Bower River has since incised thick deposits related to several prehistoric phases of Tide Lake (Fig. 3.4a). Major phases of Tide Lake are associated with advances of Frank Mackie Glacier about 2700-2600 $^{14}$C yr BP (Tiedemann Advance) and during the LIA. There is also stratigraphic evidence for a less extensive advance of Frank Mackie Glacier at the time of the Bridge Advance (Clague and Mathews 1992). Three sections at the north end of the Tide Lake basin, less than 1 km from Frank Mackie Glacier's LIA moraine, expose deltaic sand and gravel overlain by rhythmically bedded glaciolacustrine sediments (Fig. 3.4b). *Abies* needles in the deltaic sediments gave an age of 1600±40 $^{14}$C yr BP (1590-1390 cal yr BP; TO-2898), indicating that Frank Mackie Glacier had advanced enough to impound a shallow lake at that time. Wood fragments in glaciolacustrine mud overlying the deltaic sediments yielded ages of 1520±50 $^{14}$C yr BP (1520-1310 cal yr BP; GSC-5386) and 1440±40 $^{14}$C yr BP (1410-1290 cal yr BP; TO-2897). These ages suggest the Bridge Advance phase of Tide Lake probably persisted for several hundred years, but overlap of the calibrated age ranges and the possibility that the wood fragments in the glaciolacustrine mud were redeposited preclude a more precise chronology of Tide Lake and thus advances of Frank Mackie Glacier.

*Summary*

Stratigraphic and chronological data from the four study sites provide evidence for a period of glacier advance in the Coast Mountains that is distinct from both the earlier Tiedemann Advance and later LIA advances of the last millennium. Bridge

Glacier was advancing over forest near its present terminus ~1500 $^{14}C$ yr BP and may have continued advancing until at least ~1040 $^{14}C$ yr BP. Ongoing work may help resolve whether the latter age is related to the Bridge Advance or to early LIA activity (Allen and Smith 2003). Nearby Lillooet Glacier was advancing over paleosols and forest vegetation at three sites between 1700 and 1400 $^{14}C$ yr BP. A paleosol and buried forest horizon developed on Bridge Advance till were subsequently buried during a LIA advance (Reyes and Clague submitted), suggesting that the Bridge Advance was distinct from earlier and later glacier activity. The moraine stratigraphic record at Tiedemann Glacier is poorly constrained, but limiting ages on lateral moraines between the prominent moraine ridges associated with Tiedemann-age and LIA advances indicate that Tiedemann Glacier constructed moraines during the Bridge Advance. Lacustrine sediments of Tide Lake indicate that Frank Mackie Glacier advanced about 1600 $^{14}C$ yr BP following a period when it was less extensive. A subsequent period of recession, when the Tide Lake basin was empty, preceded more extensive advances of the LIA.

## Discussion

### *Regional correlation*

Corroborating evidence for the Bridge Advance in the Coast Mountains comes from several lakes in the southern Coast Mountains (Menounos 2002). Sedimentation increased at lakes in two glaciated basins near Bridge and Lillooet glaciers between 2000 and 1500 cal BP, though detailed interpretation of this record is complicated by a high signal to noise ratio in sedimentation indices and poor dating control in the relevant portions of the cores.

Evidence for a Bridge-age advance in the Canadian Rockies has been found at Peyto Glacier (Fig. 1), where trees were overridden ~1500 $^{14}C$ yr BP (Luckman 1994, in press). At Cavell Glacier, ~150 km north of Peyto Glacier, detrital wood recovered from the glacier forefield yielded radiocarbon ages between ~1900 and 1600 $^{14}C$ yr BP (B.H. Luckman, personal communication, 2003). The lack of stratigraphic context, however, limits interpretation of these ages.

Glacier activity broadly coeval with the Bridge Advance has been, comparatively, well documented at several land-terminating, non-surging glaciers in Alaska (Fig. 3.1). Overridden organic material and *in-situ* stumps at Nizina and Kuskulana glaciers, in the Wrangell Mountains, date to 1810 and ~1750 $^{14}C$ yr BP, respectively (Wiles et al. 2002). Tebenkof Glacier in Prince William Sound overrode a forest about 1450 $^{14}C$ yr BP (Wiles et al. 1999a), as did Grewingk and Dinglestadt glaciers in the southern Kenai Mountains (Wiles and Calkin 1994). Advances occurred at about this time at Bering, Beare, and Sheridan glaciers (Wiles et al. 1999b; Calkin et al. 2001) and at several calving glaciers along the Alaska coast (Calkin 1988; Calkin et al. 2001).

*Paleoclimatic implications*

Bridge Advance glacial activity is broadly synchronous in the Coast Mountains and coastal Alaska, within the limits of the dating techniques that were used to develop the glacier chronologies presented in this paper. Synchroneity is also evident in dated LIA moraine sequences from the same region (Larocque and Smith 2003). The similar timing of advances along the Pacific coast of North America suggests that glaciers, in some cases separated by up to 2000 km, have responded to the same climate forcing mechanisms. This forcing likely originated in the Pacific Ocean and led to periods of positive mass balance that were long enough to result in appreciable glacier expansion. Direct characterization of the climatic conditions that caused the Bridge Advance is not possible, but some insight can be gleaned from paleoecological investigations and from modern relationships between climate, glacier mass balance, and North Pacific ocean/atmosphere dynamics.

Holocene paleoecological records from Pacific North America reveal a transition to cool, moist conditions about 4000-3000 cal yr BP, but do not suggest environmental change associated with specific periods of Neoglacial glacier advance (Hedba 1995). Fire frequency reconstructions based on charcoal accumulation rates are perhaps better sources of high-frequency paleoclimatic information, both because of the excellent chronological control achieved through extensive AMS radiocarbon dating, and because periods of increased fire frequency can be directly linked to prolonged periods of hot, dry summers, particularly in the moist, high-elevation forests of Pacific North America (Agee 1993). Mid-Neoglacial and LIA glacier advances in the Coast Mountains are broadly synchronous with intervals of low fire frequency at a site in the southern Coast Mountains (Hallett et al. 2003), but an intervening peak in fire frequency seems inconsistent with the time of initiation of the Bridge Advance. Hallet et al. (2003) suggested that enhancement of the Pacific High during this period may have led to dominance of warm and dry summer air masses over western North America. Such conditions would have produced negative summer glacier mass balance anomalies, thus winter climate likely had to be more favourable for glaciers to expand.

Present-day winter climate in Pacific North America is dominated by the Aleutian Low pressure system, which directs storm tracks into western North America (Latif and

Barnett 1996). Changes in the intensity of the Aleutian Low are linked to interdecadal variability in North Pacific sea surface temperature, commonly termed the Pacific Decadal Oscillation (PDO), whereby warm (cold) PDO regimes are associated with an enhanced (diminished) Aleutian Low (Mantua et al. 1997; Mantua and Hare 2002). Intensification of the Aleutian Low generally leads to increased storminess in coastal Alaska and warm, drier conditions in the Pacific Northwest. Conversely, coastal Alaska is anomalously dry and the Pacific Northwest is subject to more frequent storms during periods of a diminished Aleutian Low (Leathers et al. 1991; McCabe et al. 2000).

These spatial patterns of surface climate variability have been detected in several environmental datasets in Pacific North America, including spring snowpack (Moore and McKendry 1996) and streamflow (Moore and Demuth 2001). Glacier mass balance records from the southern Coast Mountains and Alaska are particularly helpful in confirming the spatial pattern of winter climate variability in the North Pacific. In spite of the limited number and distribution of monitored glaciers and the short length of the mass balance records, the data indicate that positive (negative) winter mass balance anomalies in the southern Coast Mountains are associated with a diminished (enhanced) Aleutian Low and cold (warm) PDO phases; the opposite is true for glaciers in maritime Alaska (Hodge et al. 1998; Bitz and Battisti 1999; McCabe et al. 2000; Moore and Demuth 2001). Thus, variations in sea surface temperature and atmospheric circulation can be linked to variability in glacier mass balance, particularly winter balance.

It is not known if modern North Pacific ocean-atmosphere dynamics are representative of conditions throughout the Holocene. Proxy records suggest that the PDO was active during the LIA and that its spatial and temporal coherence differed through time (Gedalof et al. 2002). It is, however, unclear how atmospheric circulation affected regional climate earlier in the Holocene. Spooner et al. (2003) suggested that intensification of the Aleutian Low may have been responsible for regional climatic change in northwestern North America between 3000 and 2000 $^{14}C$ yr BP. The broad coherence of Bridge Advance activity in Alaska and the British Columbia Coast Mountains likewise suggests that the North Pacific plays a role in regulating winter climate conditions that are conducive to regional glacier advance. The mechanisms that drive North Pacific ocean-atmosphere variability are not understood, but may be related to solar forcing (e.g. Hu et al. 2003) and global teleconnections of cyclic North Atlantic climate oscillations (Hendy and Kennett 1999; Mix et al. 1999).

## Conclusions

Evidence from four sites in the British Columbia Coast Mountains supports a regional advance of alpine glaciers between ~1700 and 1400 $^{14}C$ yr BP. The evidence includes overridden forests in glacier forefields and lateral moraine sediments, lichen- and radiocarbon-dated moraines, and ice-dammed lake sediments. The disparate nature of these records emphasizes the problem of reconstructing glacier advances prior to the Little Ice Age. The success, however, of long-term detailed research in Alaska suggests that similarly detailed investigations in the Coast Mountains will yield additional evidence for this event. Limited high-resolution paleoenvironmental data from the southern Coast Mountains suggest that summer climate during the Bridge Advance was unfavourable for long periods of positive net glacier mass balance. Modern North Pacific ocean-atmosphere dynamics, particularly in winter, have an appreciable impact on glacier mass balance, and may have provided the climatic conditions conducive to glacier expansion during the Bridge Advance.

**Table 3.1.** Radiocarbon ages pertaining to the Bridge advance in the Coast Mountains.

| $^{14}C$ age (yr BP)[a] | Calibrated age range (cal yr before AD 1950) | Laboratory number[c] | Dated material[d] | Collector (Reference) |
|---|---|---|---|---|
| | | | *Lillooet Glacier* | |
| 1390±50 | 1410 – 1180 | GSC-6760 | Log (*Abies*) on paleosol (15 rings) | A.V.Reyes (Reyes and Clague submitted) |
| 1527±41 | 1520 – 1330 | Wk-12310 | Twig from paleosol | A.V.Reyes (Reyes and Clague submitted) |
| 1549±45 | 1530 – 1330 | Wk-12309 | Wood fragment from paleosol | A.V.Reyes (Reyes and Clague submitted) |
| 1600±70 | 1690 – 1330 | TO-9754 | Branch from top of paleosol (5 rings) | J.J. Clague (Reyes and Clague submitted) |
| 1700±80 | 1820 – 1420 | GSC-6767 | Log (*Abies*) on paleosol (20 rings) | A.V.Reyes (Reyes and Clague submitted) |
| 1720±42 | 1710 – 1530 | Wk-12306 | Charcoal from paleosol | J.J. Clague (Reyes and Clague submitted) |
| | | | *Bridge Glacier* | |
| 1040±50 | 1060 – 790 | Beta-181856 | Rooted bole in paleosol (35 rings) | S. Allen and D.J. Smith (unpublished data) |
| 1500±50 | 1520 – 1310 | Beta-171549 | Standing snag (94 rings) | S. Allen and D.J. Smith (unpublished data) |
| | | | *Tiedemann Glacier* | |
| 1270±140 | 1510 – 920 | GSC-977 | Basal peat in a moraine-dammed pond | R.J. Fulton (Fulton 1971) |
| 1330±65 | 1350 – 1080 | S-1473 | Reworked log in fluvial sediments | B. Thomson, N.F. Alley (Ryder and Thomson 1986) |

**Table 3.1. (cont.).** Radiocarbon ages pertaining to the Bridge advance in the Coast Mountains.

| | | | | |
|---|---|---|---|---|
| | | | *Frank Mackie Glacier* | |
| 1440±40 | 1410 – 1290 | TO-2897 | Wood fragment in glaciolacustrine mud | J.J. Clague (Clague and Mathews 1992) |
| 1520±50 | 1520 – 1310 | GSC-5386 | *Abies* fragment in glaciolacustrine mud | J.J. Clague (Clague and Mathews 1992) |
| 1600±40 | 1590 – 1390 | TO-2898 | *Abies* needles in deltaic sands | J.J. Clague (Clague and Mathews 1992) |

[a] Error terms are $1\sigma$ for Beta, TO and Wk ages, and $2\sigma$ for GSC ages.

[b] Determined from the decadal data of Stuiver et al. (1998) using the probability distribution method within the program CALIB 4.4 (Stuiver and Reimer 1993). Age ranges are $\pm 2\sigma$ and were calculated with an error multiplier of 1.0.

[c] Beta - Beta Analytic; GSC - Geological Survey of Canada; S - Saskatchewan Research Council; TO - IsoTrace Laboratory, University of Toronto; Wk - University of Waikato, New Zealand.

[d] Logs were sampled at their perimeter and the number of annual rings sampled is noted in parentheses.

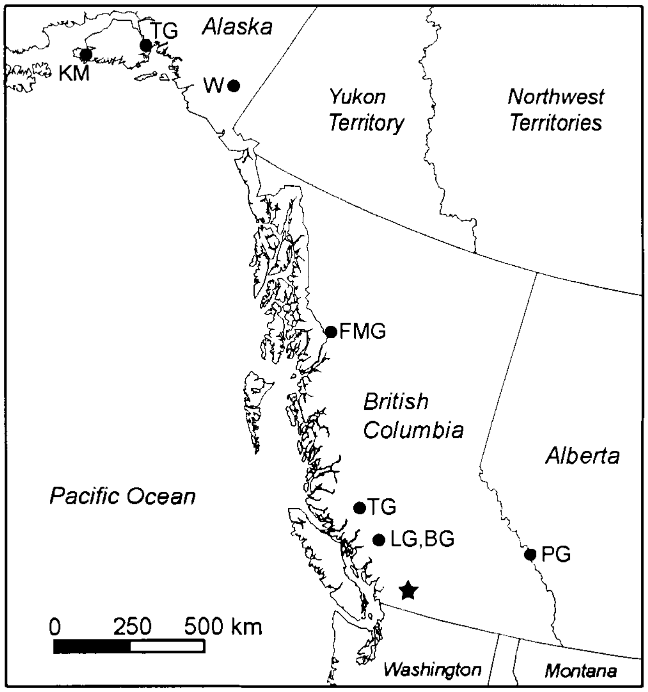


**Figure 3.1.** Map showing locations of sites mentioned in the text. TeG, Tebenkof Glacier; KM, Kenai Mountains; W, Wrangell Mountains; FMG, Frank Mackie Glacier; TiG, Tiedemann Glacier; LG, Lillooet Glacier; BG, Bridge Glacier; PG, Peyto Glacier. Star marks location of fire frequency reconstruction of Hallett et al. (2003).

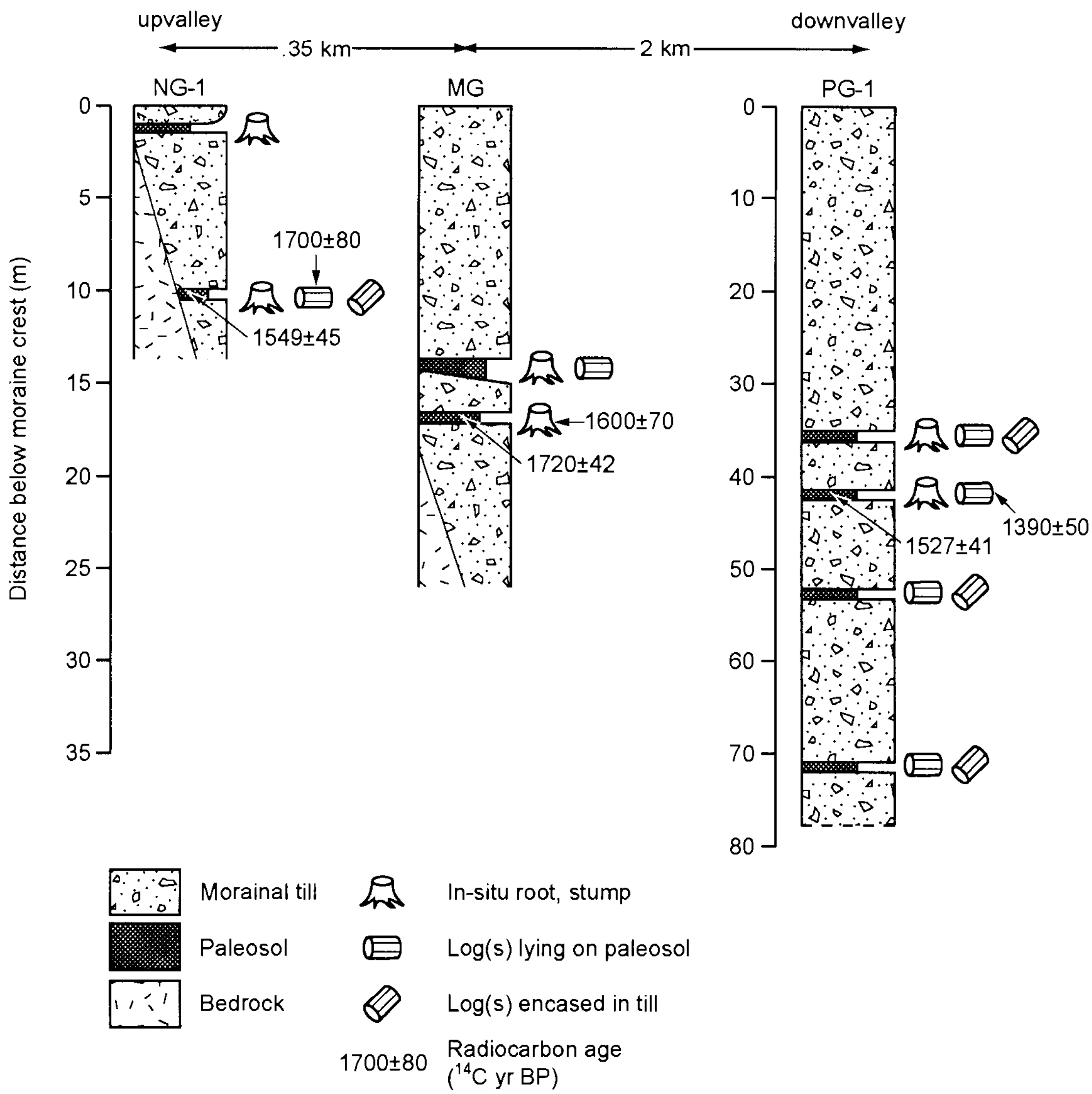


**Figure 3.3.** Lateral moraine stratigraphy at Lillooet Glacier pertaining to the Bridge Advance (adapted from Reyes and Clague submitted). NG, north gully; MG, middle gully; PG, proximal gully. Note different depth scale for section PG. Refer to Table 3.1 for radiocarbon ages. Paleosol thickness is not to scale.

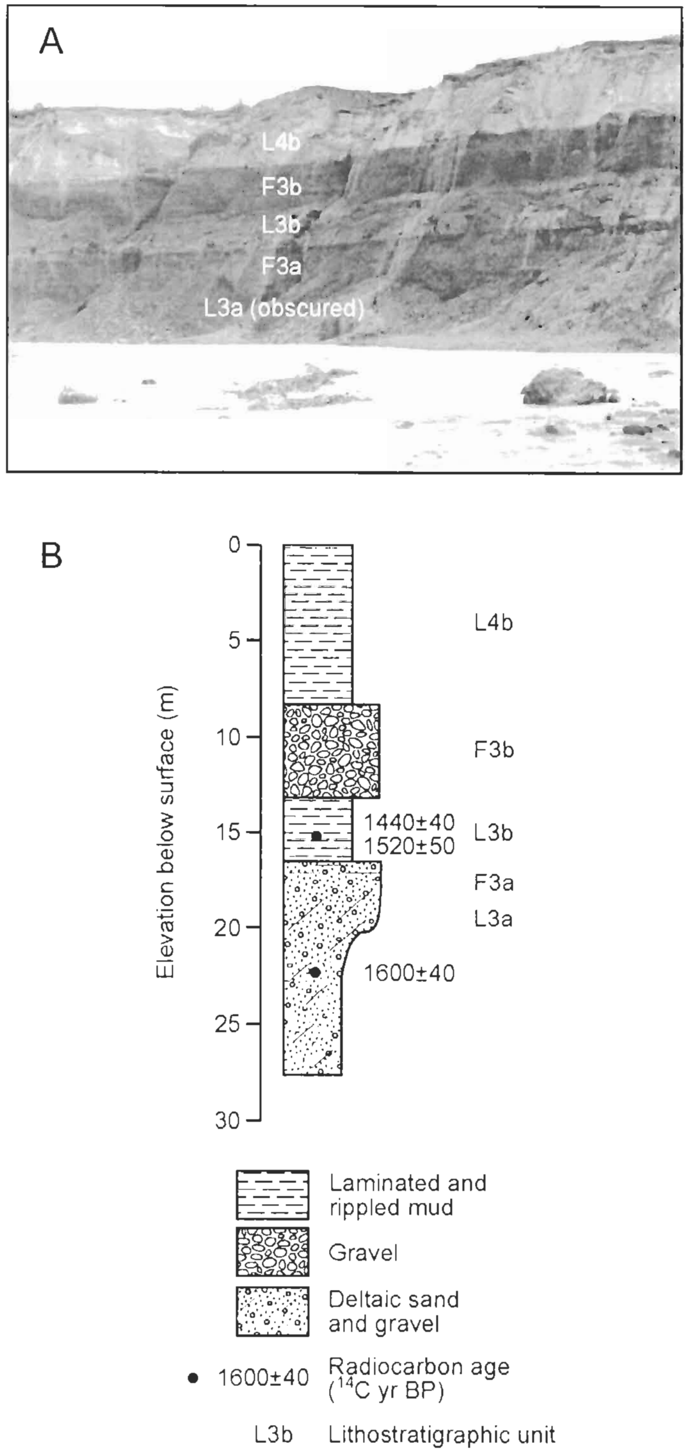


**Figure 3.4.** Photograph and composite stratigraphy of sediments at the north end of the Tide Lake basin. (*a*) Photograph of alternating fluvial and glaciolacustrine sediments (site 8 of Clague and Mathews 1992). Lithostratigraphic units are from Clague and

Mathews (1992). Photograph by J.J. Clague; reproduced with permission (*b*) Composite stratigraphy of three sections at the north end of the Tide Lake basin (sites 8, 9 and 10 of Clague and Mathews 1992). Little Ice Age moraine sediments are omitted. Refer to Table 3.1 for radiocarbon ages.

# CHAPTER 4. Conclusions

Paleosols and horizons of woody debris exposed in the northeast lateral moraine of Lillooet Glacier provide a detailed chronology of late Holocene advances of the glacier. Four periods of glacier advance are recognized in the moraine stratigraphy: (1) an advance prior to ~3000 $^{14}C$ BP, and probably before 6200 $^{14}C$ BP; (2) two advances at ~3000 and ~2500 $^{14}C$ BP, corresponding to the regional Tiedemann Advance; (3) an advance between ~1700 and 1400 $^{14}C$ BP that buried organic horizons at almost all of the study sites; and (5) several advances during the Little Ice Age, which culminated after 470 $^{14}C$ BP. The record presented here is in broad agreement with, and improves the resolution of, existing glacier chronologies in the Coast Mountains. In particular, the Lillooet Glacier record provides the first direct evidence for a glacier advance in the Coast Mountains between ~1700 and 1400 $^{14}C$ BP. Evidence compiled from Lillooet Glacier and three other sites in the British Columbia Coast Mountains suggests that this was a regional advance of alpine glaciers.

Detailed paleoclimatic interpretation of the glacier record is difficult, but comparison with paleoecological data, particularly fire frequency reconstructions, highlights the importance of winter climate. The similar timing of advances along the Pacific coast of North America suggests that glaciers, in some cases separated by up to 2000 km, have responded to a common set of climate forcing mechanisms that led to periods of positive mass balance long enough to produce appreciable glacier expansion. The association between present-day glacier mass balance and ocean-atmosphere dynamics in the North Pacific suggests that the Aleutian Low pressure system may play an important role in controlling regional climatic conditions that have a direct impact on glacier mass balance.

The results presented in this thesis highlight the importance of detailed site examination for investigations of Holocene glacier fluctuations. The reconnaissance scale of investigations limited earlier efforts in the Coast Mountains, and it is likely that repeat visits to classic sites will lead to new, important insights. The incomplete surficial record of pre-Little Ice Age glacier records requires that a variety of research approaches be applied in Holocene glacier investigations. In particular, in the Coast Mountains, where composite moraines are relatively common, such investigations will likely benefit from detailed examination of moraine stratigraphy.

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