Neoglacial Fluctuations of Terrestrial, Tidewater,

and Calving Lacustrine Glaciers,

Blackstone-Spencer Ice Complex,

Kenai Mountains, Alaska

by

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Abstract

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Chair of the Supervisory Committee: Professor Stephen Porter
Geological Sciences

The glaciers surrounding the Blackstone-Spencer Ice Complex display a variety of termini types: Tebenkov, Spencer, Bartlett, Skookum, Trail, Burns, Shakespeare, Marquette, Lawrence, and Ripon glaciers end in terrestrial margins; Blackstone and Beloit glaciers have tidewater termini; and Portage Glacier has a calving lacustrine margin. In addition, steep temperature and precipitation gradients exist across the ice complex from the maritime environment of Prince William Sound to the colder, drier interior.

The Neoglacial history of Tebenkov Glacier, as based on overrun trees near the terminus, shows advances ca. 250-430 AD (calibrated date), ca. 1215-1275 AD (calibrated date), and ca. 1320-1430 AD (tree ring evidence), all intervals of glacier advance around the Gulf of Alaska. However, two tidewater glaciers in Blackstone Bay retreated from their outermost moraines by 1350 AD, apparently asynchronously with respect to the regional climate signal.

The most extensive Kenai Mountain glacier expansions during Neoglaciation occurred in the late Little Ice Age. The outermost moraines are adjacent to mature forest stands and bog peats that yield dates as old as 5,600 BP. Prince William Sound glaciers advanced during two Little Ice Age cold periods, 1380-1680 and 1830-1900 AD.

The terrestrial glaciers around the Blackstone-Spencer Ice Complex all built moraines during the 19th century and began retreating between 1875 and 1900 AD. Portage and Burns glaciers began retreating between 1790 and 1810 AD, but their margins remained close to the outermost moraines during the 19th century.

Regional glacier fluctuations are broadly synchronous in the Gulf of Alaska region. With the exception of the two tidewater glaciers in Blackstone Bay, all glaciers in the Kenai Mountains, no matter their sizes, altitudes, orientations, or types of margins, retreated at the end of the Little Ice Age. The climate signal, especially temperature, appears to be the strongest control on glacier behavior during the last millennium.

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Climb the mountains and get their good tidings.
The winds will blow their own freshness into you
and the storms their energy,
while cares will drop away from you
like the leaves of autumn.

-- John Muir (1915)
Travels in Alaska

CHAPTER 1 - INTRODUCTION

Fluctuations in climate appear to be the norm during the Holocene Epoch. Following the last glacial maximum (LGM) and mid-Holocene warm period (Hypsithermal), widespread Neoglacial cooling episodes beginning ca. 4500 yr BP signaled the return to cooler late Holocene climates (Porter and Denton, 1967). The European chronology of the past 2000 years shows that climatic warming during the Roman period (ca. 100-300 AD) was followed by the cooler Dark Ages (ca. 300-900 AD), while the warm Medieval Optimum (ca. 900-1250 AD) preceded the colder Little Ice Age (ca. 1250-1900 AD)(Porter, 1986; Grove, 1988).

Matthes (1939) first described the Little Ice Age (LIA) as an "epoch of renewed but moderate glaciation which followed the warmest part of the Holocene". Porter and Denton (1967) defined the "climatic episode characterized by the rebirth and/or growth of glaciers following maximum shrinkage during the Hypsithermal" as Neoglaciation, but recognized that in some cases, these two intervals overlapped. The LIA is currently considered the most recent Neoglacial period of glacier expansion following the Medieval Optimum.

During the LIA, alpine glaciers advanced in virtually all mountainous regions of the world, including the Alps, Rocky Mountains, Sierra Nevada, Himalaya, East Africa, New Guinea, New Zealand, Andes, and Alaska (Grove, 1988). The early LIA began with a cold period ca. 1250-1500 AD, and was followed by a milder interval, the Little Climatic Optimum (Porter, 1986). The late LIA appears to be divided into two main cold stages each about a century long (17th and 19th centuries), with the coldest decades in the mid to late 1600's, early 1800's, and late 1800's (Crowley and North, 1991). Estimates of temperature decreases during the peak of the LIA are ca. 0.5-1.2^{o} C (Porter, 1986).

Several causal factors, including solar fluctuations (Seuss, 1968; Eddy, 1977; Kelley and Wigley, 1990; Wigley and Kelley, 1990; Crowley and Kim, 1996), and volcanic events (Bradley, 1978, 1987, 1988; Porter, 1981a, 1986; Oerlemans, 1988), the latter resulting in increased ice-core acidity (Hammer and others, 1980), have been postulated for the LIA. Nesje and Johannssenn (1992) suggest that volcanic eruptions coupled with reduced summer insolation allowed glacier expansion in the late Holocene. Comparisons of ice cores from polar ice caps and mid-latitude glaciers confirm that the LIA was global in extent (Houghton and others, 1990).

Glacier chronologies are studied worldwide as proxy climate data because glaciers may act as sensitive indicators of past climate (Porter, 1981b). The area surrounding the Gulf of Alaska has a good record of glacier-climate fluctuations because it is affected by storms emanating from the Aleutian low-pressure system (especially in the winter), because it has numerous coastal and inland glaciers, and because dense vegetation cover means that organic matter is often present in stratigraphic sections for dating glacial sediments.

Glaciers

A small local ice cap in the Kenai Mountains, a section of the arcuate Kenai-Chugach Mountains ringing the northern Gulf of Alaska (Fig. 1.1), was the focus of research. The icecap has no official name, but was informally called the Blackstone-Spencer Ice Complex by Field (1975). The ice cap extends from an altitude of 1525 m to sea level, has a variety of glacier types, and is marked by contrasting climate from its eastern to western margins. The highest peaks in the area rise to 1980 m. The firn limit ranges from 450 to 1000 m altitude, being highest on the southern and western sides of the ice cap (Field, 1975), the directions facing the prevailing source of moisture.

The northern section of this glacier complex, covering 312 km^2, is the source area for an array of glacier tongues (Fig. 1.1). The landward (western) side of the ice cap terminates in several valley glaciers. Trail, Bartlett, Skookum, and Spencer glaciers flow westward and terminate near the Alaska Railroad. Spencer Glacier supplies most of the discharge of the Placer River. Portage Glacier drains northward, ending as a calving terminus in Portage Lake. Burns and Shakespeare glaciers are part of the larger Portage Valley glacier system at the east end of Portage Lake.

Tidewater glaciers dominate the northeastern side of the icecap (Fig. 1.1). Beloit and Blackstone glaciers calve directly into Blackstone Bay. Marquette, Lawrence, and Ripon glaciers currently end close to tidewater, and undoubtedly had calving termini in the past. A large valley glacier, Tebenkov, drains the extreme northeastern portion of the ice complex.

Climate

The Blackstone-Spencer Ice Complex is located in a maritime climate, i.e., an area of fairly mild annual temperatures with moderate temperature fluctuations and high precipitation. Sea-level weather stations at Whittier and Portage have recorded a mean-annual temperature of 38^{o}F (3^{o}C)(Blanchet, 1983). This mean value is representative of sea-level conditions around the northern sector of the ice cap from Tebenkov Glacier to Portage Glacier. Inland, the mean-annual temperature at Moose Pass (133 m altitude) near the termini of Spencer and Bartlett glaciers is 34^{o}F (1^{o}C)(Blanchet, 1983).

The Kenai Mountains lie in the path of onshore-flowing winds associated with the Aleutian low that bring abundant precipitation to the coastal areas. A rain-shadow effect occurs on the landward (lee) side of the mountains (the western side of the ice cap). This results in a mean annual precipitation of 355 cm at the coast, contrasting with 150 cm on the west side of the mountains. Precipitation increases with altitude, reaching a maximum of 460 cm over the coastal peaks, but totaling only 305-350 cm over the inland peaks (Blanchet, 1983).

Snowpack varies over the ice complex as well. In the north and east, where the glaciers lie adjacent to Prince William Sound, the mean annual snow pack is 200 cm. Maximum snowpack is 358 cm as recorded at Whittier over a 13-year period (Blanchet, 1983). Inland, annual snowpack is 178 cm near the terminus of Bartlett Glacier (365 m elevation), and 127 cm at Spencer Glacier (61 m elevation). The top of the ice complex receives the thickest snowpack accumulation: 610 cm (Blanchet, 1983). Summer precipitation falls as rain at lower elevations, but as snow above 1,220-2,440 m. Winter precipitation is normally snow, but rain may fall at the lowest elevations (Blanchet, 1983).

Current equilibrium line altitudes (ELAs) for the Blackstone-Spencer Ice Complex are 400 m on Tebenkov Glacier, 450 m on Beloit Glacier, 500 m on Blackstone and Portage glaciers, and 950-1100 m on Trail Glacier (Field, 1975; Mayo and others, 1972; Viens, 1995). Thus, the ELA varies across this 17-km-wide ice cap from 400 m on eastern maritime side to 1100 m on the western landward side.

Geology

This section of the Kenai Mountains is composed of the Valdez Group (Cretaceous), a widely distributed flysch in southcentral Alaska (Kelly, 1985). The group is predominantly dark-gray mudstone, siltstone, argillite, and slate, with sandstone (mostly graywacke) interbeds (Zuffa and others, 1980). The rocks are locally calcareous and highly deformed. The Valdez Group is one of two rock units of the Chugach terrane which forms an arc around Prince William Sound from Kodiak Island to the Canadian border (Jones and others, 1981; Jones and others, 1984). The uniformity of lithologies, and the paucity of carbonates (Zuffa and others, 1980) provide a favorable substrate for colonization of lichens on moraines and below glacier-generated trimlines.

Holocene emergence-submergence history, deduced from shoreline morphology and radiocarbon dating, indicates that the Kenai Peninsula has subsided 91 m or more since Wisconsinan time (Plafker, 1969). Molinari and others (1994) describe the complex neotectonic history of Prince William Sound which includes isostatic loading and unloading from glacier fluctuations, elastic deformation associated with the Aleutian subduction zone, and local deformation from the 1964

Alaska earthquake. Isostatic rebound produced as much as 180 m of uplift between 16,000 and 5000 years BP. Elastic deformation added an additional 3-11 m of uplift. In contrast, long term subsidence measures 4-6 mm/yr for the past 5000 years, and post-1964 subsidence equals 9 mm/yr. Subsidence attributed to the 1964 Alaska earthquake, totaled 1.2 m for the Kenai-Chugach Mountains (Plafker, 1969), 2.0 m for the Portage area, and 1.5 m for Blackstone Bay (Blanchet, 1983; Post, 1967). This subsidence submerged moraines and drowned shoreline trees in Blackstone Bay, Tebenkov Bay, and Portage Valley.

Research Objectives

The intent of this research is to extend the knowledge of Holocene glacier fluctuations in the Gulf of Alaska region and determine if variations are based on characteristics of individual glaciers or larger climate patterns. The objectives are to: 1) identify and date the glacier fluctuations around the Spencer-Blackstone Ice Complex, 2) contrast glacier records from differing climate regimes on the landward and seaward flanks of the northern Kenai Mountains, 3) contrast the glacier records from terrestrial, tidewater, and calving lacustrine margins, and 4) compare the Spencer-Blackstone Ice Complex records to other Neoglacial chronologies in southern Alaska.

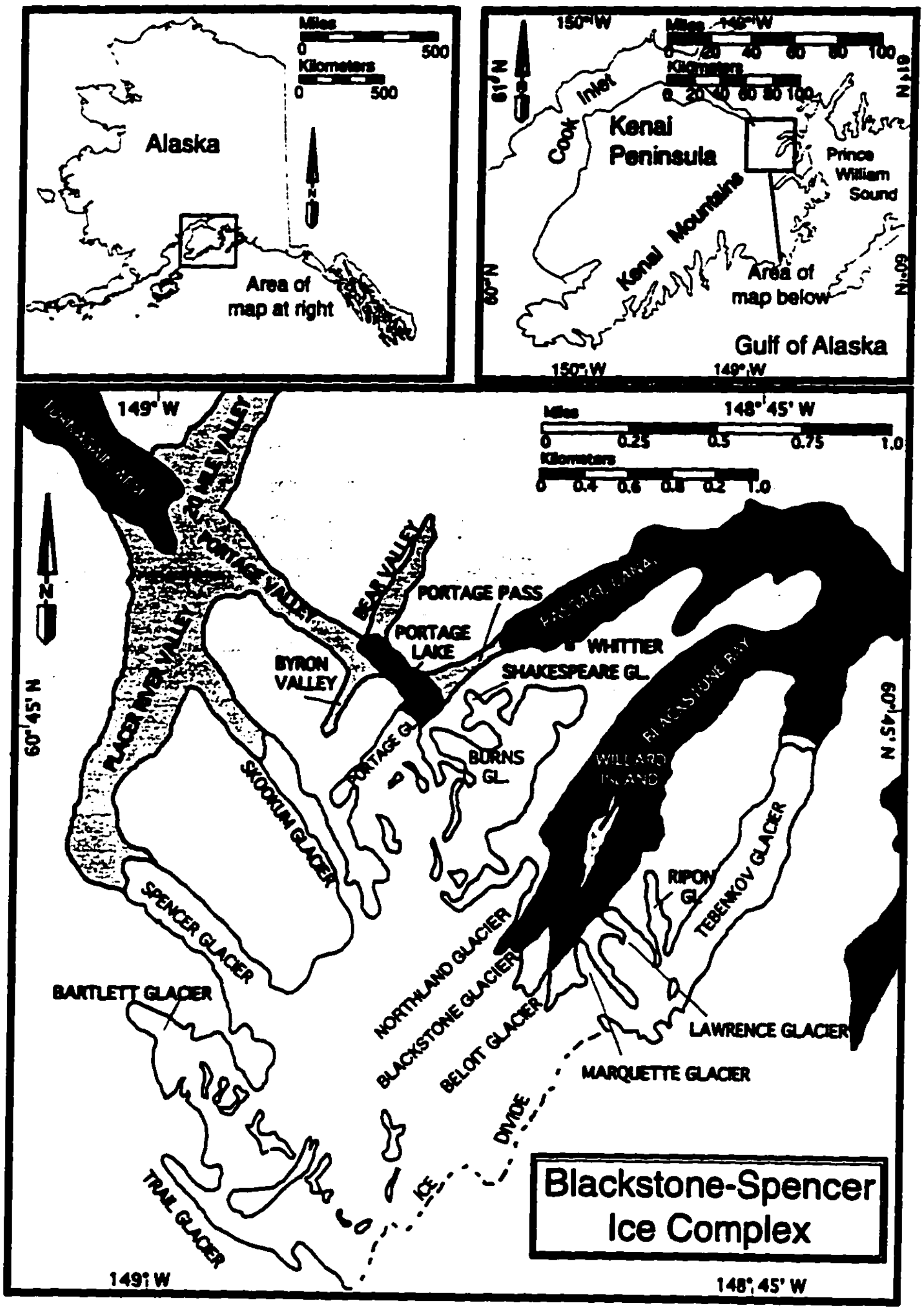

Fig. 1.1 - Spencer-Blackstone Ice Complex, northern Kenai Mountains, Alaska.

CHAPTER 2 - METHODOLOGY

Introduction

Holocene moraines around the Blackstone-Spencer Ice Complex are covered with various amounts of vegetation. Large and small trees on the moraines and intervening areas were dated by dendrochronology. Lichenometry was used to date moraines, trimlines, and areas of glacially abraded bedrock. Samples of buried trees and peats were collected for radiocarbon dating to place limiting ages on glacier advances or retreats.

Documentary and photographic evidence were used to determine the colonization time of spruce on the deglaciated areas, calibrate the lichen growth curve, and determine the past location of glacier margins.

Dendrochronology

Tree-ring ages were obtained from cores extracted using a Swedish increment borer and from cut sections. *Picea glauca* and *P. sitchensis* were the preferred species for analysis because they are slow growing, display distinctive growth rings, and are ubiquitous on the older moraines (Viereck and Little, 1972). Hybrids are common and often cannot be distinguished from either of the two non-hybrid species (D. Schmid, personal communication, 1985).

In the higher alpine areas, and in many areas outside the Little Ice Age (LIA) moraine limits, hemlock was also cored. Both Western and Mountain hemlock (*Tsuga heterophylla* and *T. mertensiana*, respectively) were used, but generally with less success than spruce due to the extreme density of the wood. In a few areas, cottonwoods (*Populus* spp.) were cored. These species, although easy to core, grow so fast and large that the center rings often could not be reached with the corer. However, on the LIA moraines of Spencer Glacier, no spruce or hemlock was found, necessitating the use of cottonwoods there. On the youngest moraines, cottonwood, alder (*Alnus* spp.), and willows (*Salix* spp.) <30 cm diameter were cut for trunk samples.

A total of 450 cores were collected and 75 trees were cut for this project (App. C). Trunks were cut as close to the base as possible, and cores were taken close to the base of the tree (usually within 30 cm of the ground), with the corer slanting downward into the center of the tree. A correction factor for cores taken above the base was determined using small spruce saplings of measured heights that were cut at their bases to determine their ages. Table 2.1 shows the heights and ages of these young trees, and indicates an average increase in height of ca. 3.5±0.4 cm/yr during the first 17 years of growth. A coring correction factor of 0.30 yr/cm was applied to the larger trees, based on the height they were cored above the ground surface.

Cut slabs were dried and sanded; cores were dried, mounted on mitered strips and sanded. Specimens were counted with a hand lens in the field and under a binocular microscope in the lab.

Tree Colonization

Average colonization times for broad regions have been determined by Viereck (1967): a 25-year ecesis period has been used previously by many authors (e.g., Sigafoos and Hendricks, 1969). Wiles and Calkin (1990) stated that a 14-year colonization period is appropriate for the area of the Harding Icefield, south of the Blackstone-Spencer Ice Complex.

In the southernmost Portage Pass, where 1974 aerial photographs show a small proglacial lake, the oldest alder collected from the drained area dated to 1975, implying a 1-yr lag for colonization. In areas deglaciated between 1974 and 1987 around the terminus of Bartlett Glacier, 1 of 10 alder seedlings was 12 years old in 1988, confirming that alder can colonize terrain as early as 1 year after deglaciation. In the Tebenkov Glacier foreland (Fig. 1.1), small push moraines were deposited along the ice margin between 1960 and 1980. Nineteen alder and hemlock seedlings were examined from an area in which no spruce seedlings were found. The oldest alder (22 years old) was 8 years older than the oldest hemlock (14 years old). These data show that in this area, both hemlock and alder colonize more rapidly than spruce, and that spruce colonization lagged deglaciation at least 22 years.

In front of Lawrence Glacier (Blackstone Bay) (Fig. 1.1), sparsely vegetated bedrock ridges support alder as old as 22 years and hemlock as old as 23 years. Spruce has not yet colonized the ridges, suggesting a colonization period of $\geq$ 23 years.

Based on these data, the colonization period used for spruce in this research is ca. 25 years. Alder and hemlock may colonize simultaneously and soon after deglaciation. M. Barker (personal communication, 1989) suggests that precipitation is the most important restricting variable in the colonization of many species, and in the maritime climate of the Gulf of Alaska and the Kenai Mountains this restriction appears to be minimal.

Lichenometry

Crustose lichens were used to estimate the time since deglaciation of moraine and bedrock substrates (Beschel, 1961) based on lichen growth rates. Sampling was restricted to *Rhizocarpon* sp. (Runemark, 1956; Benedict, 1988). Although four different species may be included in this genus in Alaska (Calkin and Ellis, 1980), because the growth rates are similar (Calkin and Ellis, 1980) the different species were used collectively without discrimination to construct a local lichen growth curve (Fig. 2.1).

The growth curve (Fig. 2.1) is based on lichens growing on substrates of known exposure age. In southcentral Alaska, such substrates are restricted to the last 100 years, the period since mining first occurred in this region. Shortly after mining began, the railroad was built and several bedrock quarries (metagraywacke) were opened. Both quarries and mine tailings within the field area provide suitable substrates. Historical and aerial photography place further limits on the age of moraines formed since 1950 AD. Dendrochronologic dating of moraines also provides some minimum age control for the last 200 years. Information antedating 1910 AD is based on written records and dendrochronology, and is considered reliable only for minimum age constraints (Webber and Andrew, 1973; Bickerton and Matthews, 1992).

A strong linear correlation ($r^2 = 0.955$) exists between substrate age and lichen size (Fig. 2.1). The relationship between these two variables is expressed by the following equation:

$$y = 1.943\ x - 3.818$$

where y represents substrate age (yrs) and x represents lichen size (mm). Due to the mild, moist climate in the field area, the lichen growth rate is among the highest yet reported, averaging 52 mm/100 yr (for comparisons see Webber and Andrews, 1973; Locke and others, 1979). I assumed that the period of rapid linear growth lasted at least 200 years, making this curve useful throughout the field area. This period is comparable to those from adjacent areas of Alaska and the Yukon (Denton and Karlen, 1973a; 1977; Rampton, 1970) and central Alaska (Reger and Pewe, 1969), as well as such areas as Steingletscher, Switzerland (Beschel, 1957) and Mt. Cook, New Zealand (Burrows and Lucas, 1967). The growth rate is somewhat higher than that derived by Wiles and Calkin (1990) for the southern Kenai Mountains (37 mm/100 yr), but their curve is based on dendrochronology alone.

Lichens were measured by clear plastic rulers to the nearest mm along the maximum axis, neglecting asymmetric and multiple thalli. Only large lichens were measured on each moraine; after preliminary measurements determined the largest size range, the smaller lichens were left unmeasured. This approach was based on the assumption that most small lichens colonized the moraine recently.

Lichen data (App. D) are displayed as absolute frequency distributions (Innes, 1983); additional data include location, sample size, maximum thallus diameter, and average of five largest thalli measurements (Appendix D). The maximum diameter recorded was used in estimating the age of each substrate, on the assumption that the maximum values each substrate, on the assumption that the maximum values indicate a minimum exposure time for the substrate.

The field area was exceptional in having a substrate lithology consisting entirely of Valdez Formation, a slightly metamorphosed graywacke turbidite, which promoted ubiquitous lichen coverage. All moraines sampled are stabilized, none apparently being ice-cored or actively slumping.

Ecological factors appear to affect the lichen sizes in some areas. In Portage Pass (Fig. 1.1), the largest lichens grow on the southeast sides of glacially transported boulders and ice-scoured strike ridges, likely the result of insolation differences within this narrow valley.

Competition among plant species also plays an important role on older moraines. On moraines in Portage Valley >130 years old, the *Rhizocarpon* are out-competed by mosses that overgrow and kill the thalli; trees that shade the surface promote moss and understory growth. Inside the glacier trimline in Portage Valley (between Portage Creek and Bear Valley), competition among different lichens is important. At lower elevations, individual thalli grow unimpeded, with margins distinct from adjacent lichens. At higher elevations, where the lichens should be older, a higher percent of rock surface is covered by lichens and individual thalli impinge on adjacent ones. The result is that the maximum diameters for *Rhizocarpon* are approximately the same in both locations implying that competition causes a decrease in growth rate for the older lichens (Innes, 1985).

The lichen data show that several assertions made by Innes (1984) are confirmed in this field area. First, the difference between the largest thallus diameter and the average of the five largest thalli is dependent on the sample size (Appendix D). The larger the sample size the more closely the average approaches the maximum. Second, there is a reduction in variation/standard deviation as the sample size increases. Third, on older substrates the importance of ecological factors must be considered when assessing growth rate.

Bickerton and Matthews (1992) suggest that error factors in lichen curves from the LIA period should be approximately 10%. Errors are likely to increase with age, approximating ±20 years on the older moraines and ±5 years on the younger ones.

Radiocarbon Dating

Samples collected for radiocarbon dating of Neoglacial ice advances include wood specimens found buried in or recently eroded from till exposures, as well as basal peats from bogs and buried organic matter located outside the LIA moraines (Appendix A).

Air-dried samples were dated at the Quaternary Isotope

Laboratory of the University of Washington. Samples were pretreated to remove modern organic matter and processed using standard techniques. Dates are reported in both radiocarbon years and calibrated years (Stuiver and Reimer, 1993).

Historical and photographic records

A wealth of historical and photographic records exists in southern Alaska. Although not as thoroughly documented as at Glacier Bay, Columbia Glacier, and College Fiord, the record for this area spans more than 100 years.

The Alaska Railroad archival library in the Anchorage Historical and Fine Arts Museum provided numerous photographs taken before and during railroad construction (1910-1915). The archival section of the University of Alaska Fairbanks library provided manuscripts and oral histories from W.O. Field. The U.S.D.A. Forest Service Supervisor's Office, Anchorage, Alaska, and the Glacier Ranger District, Girdwood, Alaska, allowed me access to historical, aerial, and satellite photography from 1950-1990. R.A.M. Schmidt provided access to her photographic collection from the mid-1960's.

In addition, many archival resources from elsewhere were used to complete this research. The U.S. Geological Survey Photo Library (Denver, Colorado) provided unpublished photographs and field notes from S. Paige, 1906; U.S. Grant, 1908; L. Martin, 1911; B.L. Johnson, 1913; R.J. Wier, 1914; F.F. Moffitt, 1925; and F.F. Barnes, 1939-40. The World Data Center-Glaciology A archives at the University of Colorado (Boulder) provided unpublished photographs from the American Geographical Society, 1935; W.O. Field, 1935; Alaska Railroad, 1935; B. Washburn, 1937; U.S. Air Force,1950; M.T. Millett, 1957; U.S. Navy, 1957; L. Viereck, 1957; R.J. Goodwin, 1957; and E. Andrews. The U.S. Geological Survey-Glaciology Branch (Tacoma, Washington) retains numerous photos by A.S. Post, M. Meier, and others. All photographs used in this research are listed in Appendix B.

Maps were constructed from photographs in two ways: 1) vertical aerial photographs were superimposed using a zoom stereoscope, and 2) landmarks on oblique aerial and ground photography were located in relation to a glacier terminus, and the ice-margin position was transferred to maps made from vertical photography.

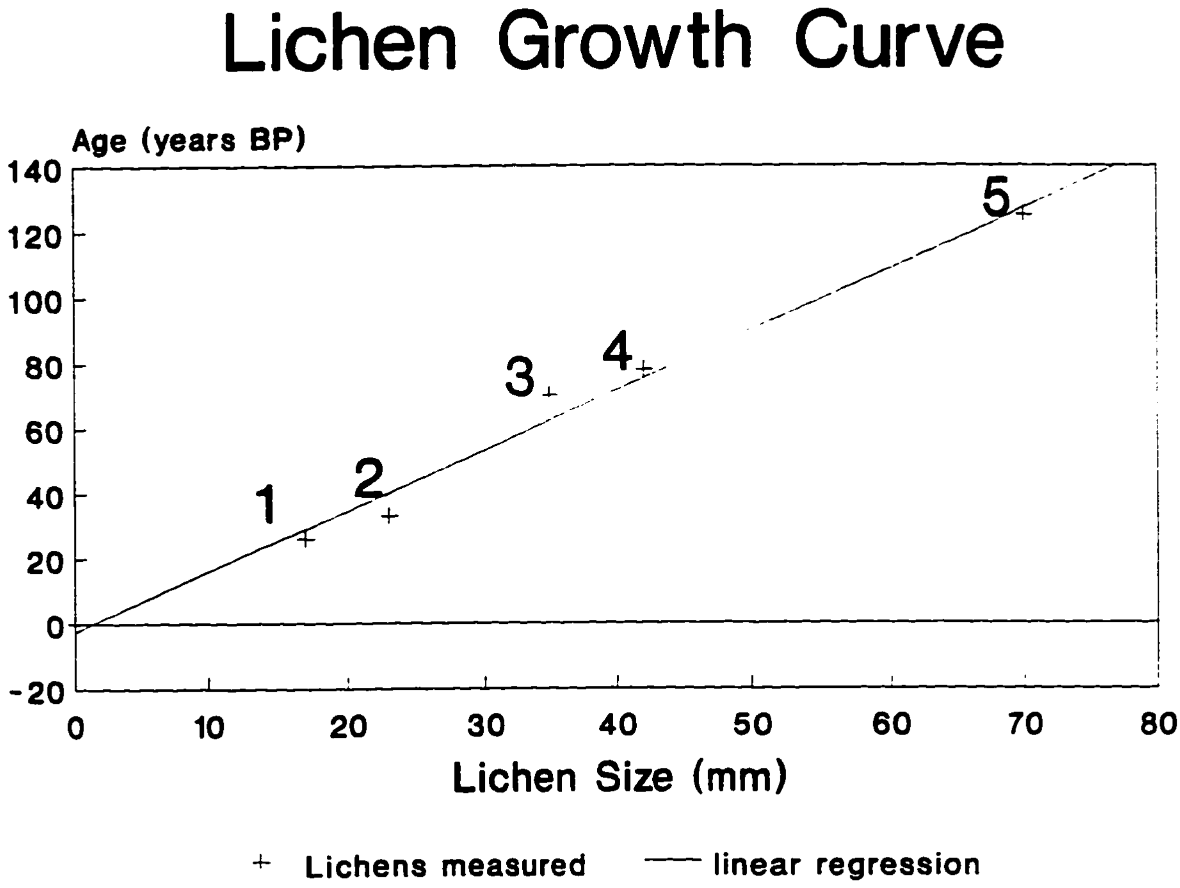

Fig. 2.1 - Relationship between substrate age (yrs) and lichen size. Numbers indicate the collection locations and methods used to determine age.

1 - Portage Pass youngest moraine - aerial photography
2 - Bartlett Glacier - Alaska Railroad quarry
3 - Bertha Creek, Turnagain Pass - mine tailings
4 - Spencer Glacier - Alaska Railroad quarry
5 - Portage Valley VC moraine - dendrochronology

TABLE 2.1 - Correction factor for coring height based on increasing height of spruce (*Piceas spp.*) seedlings over time

Location	No. of Nodes	Height (cm)	Age (yr)	Growth Rate (cm/yr)	Correction Factor (yr/cm)
Willard Is.	6	38	14	2.71	0.37
Portage Valley	8	35	8	4.38	0.23
	8	56	17	3.29	0.30
	11	45	15	3.00	0.33
	9	25	9	2.77	0.36
	8	40	14	2.85	0.35
	5	39	8	4.88	0.21
	4	27	8	3.38	0.30
	7	31	8	3.88	0.26

n=9
Average increase in height = 3.46 cm/yr
Average correction factor = 0.30 yr/cm height increase

CHAPTER 3 - PORTAGE VALLEY AND PORTAGE GLACIER

Introduction

Portage Glacier, 9 km long and covering 21 km^2, is the main outlet glacier on the northwestern side of the Blackstone-Spencer Ice Complex. The most recent advance of this glacier affected a large area, including Portage Valley, Portage Lake, Portage Pass, Bear Valley, Burns Glacier, and Shakespeare Glacier (Fig. 1.1). Portage Glacier and Burns Glacier are divided by a narrow bedrock slope, but previously Burns Glacier was a tributary to the larger ice tongue, and separated from it by a well-defined medial moraine. At that time, Portage Glacier terminated in Portage Valley, and Burns Glacier flowed uphill and terminated in Portage Pass. (Some earlier workers refer to Portage Glacier as a two-lobed ice tongue, but I will regard the northeastern lobe of the expanded Portage Glacier ending in Portage Pass as Burns Glacier.) Shakespeare Glacier, now located along the eastern wall of Portage Pass, was close to, or was a tributary of, Burns Glacier (Fig. 3.1). The histories of Burns and Shakespeare glaciers are presented in Chapter 4.

Moraines downvalley from Portage Lake

Portage Glacier previously terminated downvalley from the present western edge of the lake, and filled the lake basin. Its presence affected the drainage not only of the remaining portion of Portage Valley (between the modern lake and Turnagain Arm), but also Bear Valley, a northern tributary along the lake edge (Fig. 3.1). Glacier recession created Portage Lake (a process that continues at present).

A massive hummocky, multi-ridged end moraine blocks the western end of Portage Lake and extends across Bear Valley on the north and Byron Valley on the south (Fig. 3.2; labeled 1852). This moraine has been the location of the U.S. Forest Service Visitors Center since the 1960s, and I informally call this the Visitors Center (VC) moraine. Previous workers (Viereck, 1967; Field, 1975; Mayo and others, 1977) identified this moraine as marking the former maximum Holocene extent of Portage Glacier. This interpretation is incorrect, however, because two additional moraine segments lie downvalley from the VC moraine (Fig. 3.2; labeled 1810 and pre-1799). These narrow moraines have crests equal in altitude to the highest portions of the VC moraine, and they exhibit hummocky topography, kettles and small kettle lakes, and large boulders on their surfaces. Although these moraines have been truncated by a meltwater stream, their arcuate trend indicates the general location of the former ice margin.

No older moraines are obvious in Portage valley, and moraines predating the Little Ice Age were likely overrun by the latest advances, eroded by meltwater, or buried by outwash. The limited subsurface data suggest that outwash interfingers with intertidal sediments downvalley, revealing no buried bedrock or morainal ridges (Blanchet, 1984).

Age of the moraines downvalley from Portage Lake

No lichens >50 mm diameter were seen on the two moraines downvalley from the VC moraine (Fig. 3.2). This likely is due to growth of mosses on boulders and to shading by trees and shrubs that colonized the moraines. However, the trees provide a relatively good dendrochronological record (App. C). The younger of the two moraines has the better tree-ring record (Fig. 3.2; Table 3.1, App. C). Several trees on this moraine are 100-120 years old, and one tree was 149 years old in 1984 (Table 3.1, App. C). Assuming a 25-year ecesis time, the moraine dates to ca. 1810 AD. The older of the two moraines has numerous trees with rotten centers (Table 3.1, App. C). Because this moraine lies outside the ca. 1810 moraine, it predates it, but its exact age is unknown. The oldest tree predates 1799 AD, but the center was not recovered (Table 3.1, App. C). Although the oldest moraine predates the ca. 1810 moraine, it may be only several decades older, for Viereck (1967) recorded numerous trees 200-250 years old along the floor and walls of Portage Valley outside this moraine. No trees this old were found on the moraine.

Downvalley from these moraines, the forest floor is covered with thick peaty soil. The surfaces on which the

250-year-old trees grow are assumed to have been exposed during the late Pleistocene deglaciation of the valley. I therefore conclude that these two moraines represent the maximum advance of glaciers down Portage Valley during Neoglaciation, and that the greatest advance culminated in the middle to late 18th century, with ice recession from the oldest moraine beginning sometime before 1799 AD.

Aerial photography suggests that the two oldest moraines were eroded by meltwater when the glacier terminated at the VC moraine, because linear channels parallel to the valley and connected to the outwash plain appear along the northern margins of the eroded moraines, while the VC moraine displays no such erosion across its length. If segments of the older moraines existed downvalley, they likely were eroded or buried by outwash emanating from the glacier when it stood at the VC moraine position.

Age of the VC moraine

The VC moraine is a continuous arcuate ridge around the northwestern end of Portage Lake, including areas in Bear and Byron valleys (Figs. 3.1, 3.2). Trees were cored and lichens measured (Beschel, 1961; Locke and others, 1979) along the entire length of the moraine (App. C and D), and the results were consistent (Table 3.1, Fig. 3.2). The oldest trees measured were 100 years old in 1984, and the maximum *R.*

geographicum diameters were 70 mm at that time. Petroff (1884) stated that Portage Glacier stood at what is now the west end of the lake in 1880, and subsequent investigators have either chosen 1890 (Tarr and Martin, 1914; Schmidt, 1961) or 1880 (Mayo and others, 1977) as the date of the formation of the VC moraine. This research shows that the moraine stabilized as early as 1859 AD (based on 100-year-old trees and a 25-year colonization period) or 1852 AD (based on 70-mm lichens; Figs. 2.1, 3.3). The ice remained along the western shore of Portage Lake until 1900 AD (based on 59-year-old trees and 48-mm lichens found on ridges adjacent to the lakeshore) building the substantial moraine that dams Portage Lake (Fig. 3.2).

Age of deglaciation of the trimline area

An obvious trimline occurs along the bedrock cliff between Bear Valley and Portage Creek (adjacent to the Visitors Center) and extends into Bear Valley (Fig. 3.2). The trimline is recognized by an obvious vegetation contrast with dense forest above and ice-sculpted bedrock below. The less-vegetated bedrock area below the trimline is covered with lichens and scattered trees and is crossed by two bedrock channels parallel to the shoreline ca. 10 m above the modern lake level.

Lichens and trees growing below the trimline between Bear Valley and Portage Creek should give minimum ages for the deglaciation of the trimline. Although the trees are slightly younger than those on the 1852 moraine, the lichens are nearly the same size, suggesting that the trimline was deglaciated by ca. 1862 (Figs. 3.1, 3.2).

In a normal deglacial sequence, the ice would downwaste and retreat such that large lichens and old trees would be found at the trimline, with smaller and younger ones downslope. However, the evidence from this trimline area shows both the trees and lichens to be the same size and age at and below the trimline. This implies that deglaciation from the trimline occurred very rapidly and/or that water erosion at the ice-bedrock interface rapidly produced an ice-free channelway. The presence of abondoned meltwater channels cutting across this outcrop are consistent with the latter interpretation.

Bear Valley chronology

If Portage Glacier filled the lake basin to the VC moraine between ca. 1799 and 1900, then one consequence may have been an ice-dammed lake in Bear Valley, along the north side of Portage Valley (Fig. 3.1). Mayo and others (1977) suggested that strandlines in Bear Valley are evidence of this lake, as are kettles in the Bear Valley outwash plain. I found no evidence of strandlines in Bear Valley either on aerial photographs or in the field. Neither wave-cut benches nor lines of driftwood appear around the periphery of the valley. The 1852 moraine certainly protrudes into Bear Valley (and contains kettles), but the outwash along the northern edge of the moraine (and north of the Alaska Railroad tracks) is not pitted.

An ice-dammed lake in Bear Valley is certainly a logical consequence of Portage ice along the western lake shore, but lacking field evidence, the lake hypothesis seems less valid. Because the trimline was deglaciated by 1862 and provided an outlet, it is unlikely that an ice-dammed lake existed in Bear Valley in 1880 (as suggested by Mayo and others, 1977). Although an ice-dammed lake may have existed between 1799 and 1862, no evidence for it was found.

Ground photography (Fig. 3.3) shows that by 1915 a narrow channel, called "The Canyon", extended along the trimline area between Bear Valley and Portage Creek, and was walled by bedrock along the northeast side and by ice cliffs (>30 m high) on the southwest. Therefore, by the early 20th century at least, the drainage out of Bear valley was flowing along the valley wall (and glacier margin), and no ice-dammed lake existed in that valley.

Mayo and others (1977) and Blanchet (1984) also suggested that such a lake produced outburst floods that scoured Portage Valley. The only field evidence for such an occurrence might lie in the erosion of the previously mentioned older moraines found downvalley of the VC moraine. Alternatively, these moraines may have been eroded by glacial meltwater unrelated to flooding.

Recession from the VC moraine

A map surveyed by Brooks for the U.S. Geological Survey during 1911-1912 (Fig. 3.4) shows Portage Glacier filling the lake basin and extending to the site of the modern Visitors' Center. A stream exiting Bear Valley flowed adjacent to the ice along the bedrock trimline area, and then into Portage River. Small proglacial lakes existed in the Bear Valley and Visitors Center's areas at this time.

Ice remained at the northwest end of the lake from 1852 to 1914 (Fig. 3.3). Although the northern margin of the glacier remained close to the bedrock cliff below the trimline until 1914, the southern ice margin retreated northward, leaving a series of lateral moraines along the shoreline (Fig. 3.2). Erosion by Byron Stream entering the lake basin along the southern ice margin may have promoted this retreat. During the early 20th century, Portage Valley may have seen the type of terminal evolution and proglacial lake development described by Kirkbride (1993) for New Zealand lacustrine glaciers.

Observers noted that Portage Lake opened by 1914 (Schmidt, 1961), and continuing ice retreat is documented with oblique ground photographs (Barnes, 1943) in 1939 (Figs.

3.1, 3.2). At that time the ice had receded 840 m from its 1914 western terminus and had deposited a small, narrow, zig-zag-shaped moraine along the southern shore of the lake. This moraine dates prior to 1937 based on dendrochronological evidence (Table 3.1) and to 1922 based on lichenometric evidence (Fig. 3.2, App. D).

Changes in glacial regime 1799-1994

The time-distance diagram (Fig. 3.5) and Table 3.2 indicates how the marginal retreat rates are tied to the different glacier regimes (terrestrial vs. calving). From the end of the 18th century until the beginning of the 20th century, Portage Glacier had a terrestrial terminus. The mass balance was dominantly negative and the retreat rate averaged 3.8 m/yr.

Once the glacier terminus retreated to the lake basin (Fig. 3.6), the rate of retreat increased by an order of magnitude as the glacier developed a calving margin (although melting still occurred). Retreat rates averaged 33.6 m/yr from 1914 to 1939. From 1939-50 the ice retreated across a portion of the lake basin up to 196 m deep (Fig. 3.7) and, consistent with a theoretical calving law (Meier and Post, 1987; Post, 1975), this exceptional water depth caused the calving rate to increase. The highest average retreat rate recorded for Portage Glacier, 145.5 m/yr, occurred during the period 1939-50. By 1950, the ice margin stood at a mid-basin shoal where the water is 120-140 m deep, and ice retreat slowed to 44.4 m/yr. This value is also in close agreement with a prediction based on the theoretical calving law (Mayo and others, 1977). From 1959 through 1994, the calving rate ranged between 33.3 and 60.0 m/yr (Table 3.2). Mayo and others (1977) report a slight 1-year advance in 1974, and this may have influenced the lower retreat rate calculated for the 1965-1974 period. From 1914-1994, the retreat rate averaged 56.3 m/yr, more than an order of magnitude greater than the rate prior to lake formation (Table 3.2). Drawdown of the ice also accompanied catastrophic retreat (Fig. 3.6). Mayo and others (1977) calculated 200 m of total thinning between 1880 and 1972, with 100-120 m of terminal thinning between 1950 and 1972.

Recent retreat 1984-1996

The last 12 years have seen immense changes in Portage Glacier (Fig. 3.1). In 1984, the glacier terminus trended north-south across Portage Lake, but by 1994, the terminus trended east-west across the head of the lake basin. In the interim, ice-margin lakes, supraglacial lakes, and subglacial channels had developed and disappeared. (Appendix B contains references to photographic evidence.)

In August 1984, a small ice-dammed lake existed at the base of Portage Pass in an embayment behind the calving terminus. The terminus ended in Portage Lake adjacent to the westernmost bedrock knoll at the base of Portage Pass. The water level in this lake appeared to be the same as the level of Portage Lake some 110 m away, suggesting an englacial or subglacial conduit connecting the two. By 1987 this small lake became connected to Portage Lake along a narrow subaerial channel ca. 10-15 m wide and ca. 100 m long. A large ice-cored moraine along the westernmost wall at the base of Portage Pass was separated from the main ice mass by this channel. During 1988 both the channel and lake enlarged, so that an arm of Portage Lake now lapped at the base of Portage Pass. Like a door swinging open, calving concentrated the recession along the northeast edge of the lake.

During 1989, two significant features developed in the Portage Glacier/Portage Lake/Portage Pass region. The meltwater stream from Burns Glacier joined with a subglacial channel exiting Portage Glacier below the Burns Glacier terminus. This subaerial stream flowed along the outer eastern edge of Portage Glacier before becoming englacial at the base of Portage Pass. Huge semi-circular arches formed as the stream flowed into the ice, and a calving front developed.

The englacial or subglacial stream exited approximately midway along the calving margin in Portage Lake, and was apparent by the water turbulence and velocity (Capt. Robt. Patronsky, personal communication, 1989). Above the subglacial tunnel, large ice bergs calved off by upward tilting and rotation. Multiple waterlines developed along the ice cliff prior to calving, indicating that upward rotation occurred. The subglacial channel was located in a 250-m-deep depression (Capt. Robt. Patronsky, personal communication, 1989), the deepest area in Portage Lake.

The second unusual feature seen during 1989 was a supraglacial lake (Figs. 3.1, 3.8, 3.9) located in a closed depression between the two medial moraines ca. 100 m behind the calving margin.

By the 1990 ablation season, the entire hydrologic system of Portage Glacier was rearranged. The 1989 englacial stream abandoned its route for a subaerial course along the terminus at the base of Burns valley and Portage Pass, ending in the enlarged arm of Portage Lake. No subglacial stream was evident along the lacustrine ice cliffs, and 10-m-deep stream (Capt. Robt. Patronsky, personal communication, 1990) meandered through ice-cored moraines, lateral moraines, and stream deposits at the base of Portage Pass. The supraglacial lake connected to Portage Lake, initially as a narrow channel, later widened by calving (Fig 3.9).

During 1991, the meltwater steam remained subaerial outside the ice margin and no apparent subglacial stream exited the lacustrine ice cliffs. The supraglacial lake was connected to Portage Lake by a channel that opened and closed seasonally.

By 1992, the supraglacial lake became connected to Portage Lake, first as a 28-m-deep supraglacial embayment, and later as a 200-m-deep embayment (Capt. Robt. Patronsky, personal communication, 1992). Portage Lake enlarged by calving, and the northeastern ice margin (adjacent to Burns Glacier) increased its stagnant zone as shown by areas of extensive ice-cored lateral moraine. Similar processes continued in 1993.

Two important changes were recorded in 1994. First, depth soundings showed the glacier terminating in water ca. 100 m deep, suggesting that the margin was approaching the head of the lake basin. Second, the northwestern margin retreated around the corner of a bedrock peninsula, and calving in September 1994 revealed the bedrock cliff under the center of Portage Glacier. The bedrock outcrop lies between the two medial moraines, and in approximately the same location as the prior supraglacial lake. In 1977, Mayo and others suggested that Portage Glacier would reach the end of the lake basin by 2020 AD, but this position was reached 26 years earlier than predicted.

Calving and melting of the extensive stagnant ice zone during early summer 1995 exposed the eastern shoreline adjacent to Burns Glacier, and calving enlarged the bedrock exposure. In June 1996, the exposure measured 35 m high and 100 m long, with 100 m deep water at its base and ice cliffs towering to 75 m above it.

Capt. Robt. Patronsky (personal communication, 1994) reports that the glacier advanced 15-40 m each winter between 1988 and 1994 due to the decrease in winter calving. The amount of ice that advances during the winter is commonly calved away by July, and an additional 40 m or more calves from the ice front by the end of the ablation season. From 1988 to 1994, the ice front retreated approximately 500 m. In 1972, Mayo and others (1977) calculated forward motion of the ice to be 165 m/yr, a terminal retreat of 30 m/yr, and a total loss of 195 m/yr from the terminus. Assuming the same forward motion in 1988-94 as in 1972, this would result in a total loss of 205 m/yr or more from the terminus.

Predictions

Three conditions are possible for Portage Glacier in the future: 1) continued retreat, 2) stabilization at a shoreline position, and 3) advance. In the first case, the glacier would continue retreating up valley, beyond the calving environment, and leave Portage Lake behind. Although other lacustrine calving glaciers have behaved in this way, Mayo and others (1977) suggest that the mass balance of Portage

Glacier will prevent this from happening. They suggest that the second option is the most likely because once the glacier retreats from the lake basin and the calving rate decreases, the flow velocity and mass balance are sufficient to maintain the glacier terminus near the shoreline where small ice masses would break off and tumble into shallow water.

The third scenario (ice advance) is not predicted. Based on tidewater glacier theory (Meier and Post, 1987), an advance across the basin can happen if the terminus remains in shallow water during the advance. This might be possible if Portage Lake fills with sediment and the ice advances across a shallow, sediment-filled basin. Studies in Glacier Bay (Powell, 1980) have recorded sedimentation rates as high as 2 m/day, but this rate occurs only during the melt season in areas directly adjacent to the active ice front. No sedimentation studies have been done in Portage Lake, but if sediment flux is low, it could easily take hundreds of years to fill this lake with sediment.

A more likely possibility, suggested by tidewater glacier theory, is one in which Portage Glacier builds a moraine/ subaqueous fan at its terminus and progressively erodes and redeposits the sediment as it advances downbasin. Advancing tidewater glaciers, including Taku (Motyka and others, 1992), Hubbard (Trabant and others, 1991), Harriman and Meares (Viens and Post, 1995) appear to be moving in this manner.

Tidewater glacier advance likely is much less rapid than retreat (Porter, 1989), implying that it would have taken Portage Glacier 100 years or more prior to 1799 to advance to its LIA maximum. If Portage Glacier advanced at a rate of 12-32 m/yr, similar to rates of other recorded calving glaciers (Veins, 1995), its advance down the Portage Lake basin could have begun as early as 1382 AD or as late as 1643 AD, assuming a negligible resting period on its 1799 AD terminal moraine.

Prior studies (Bartsch-Winkler and Schmoll, 1992; Schmoll and others, 1984; Combellick, 1994; Combellick and Reger, 1994) indicate subsidence at the head of Turnagain Arm (7 km downvalley from Portage Lake) where a submerged peat layer, dated to 1788 AD, is attributed to an earthquake. Alternatively, if this subsidence instead was related to glacial advance and concurrent isostatic depression, then it may correlate with the LIA advance of Portage Glacier.

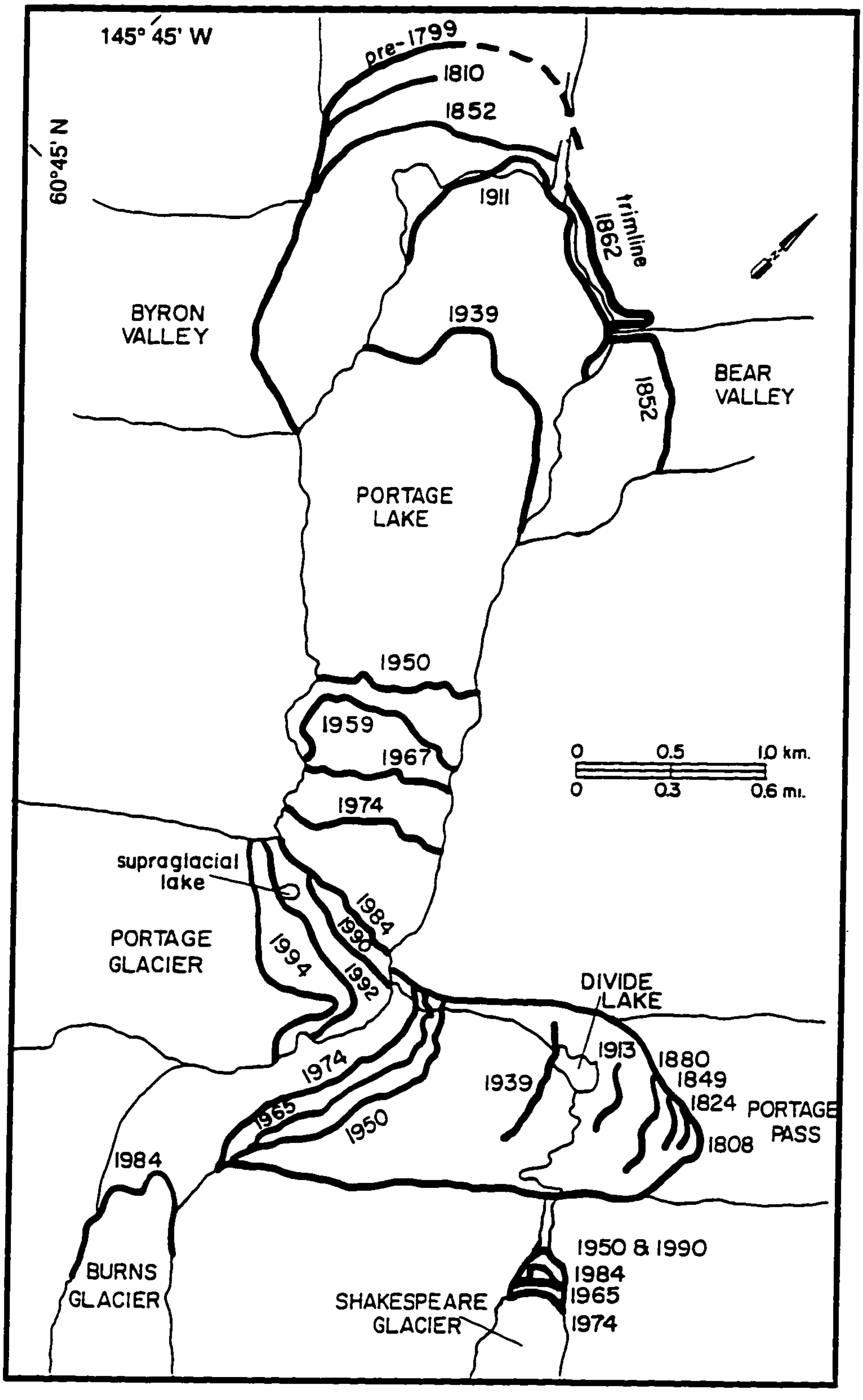

Fig. 3.1 - Map of Portage Valley showing place names and historical positions of glacier termini.

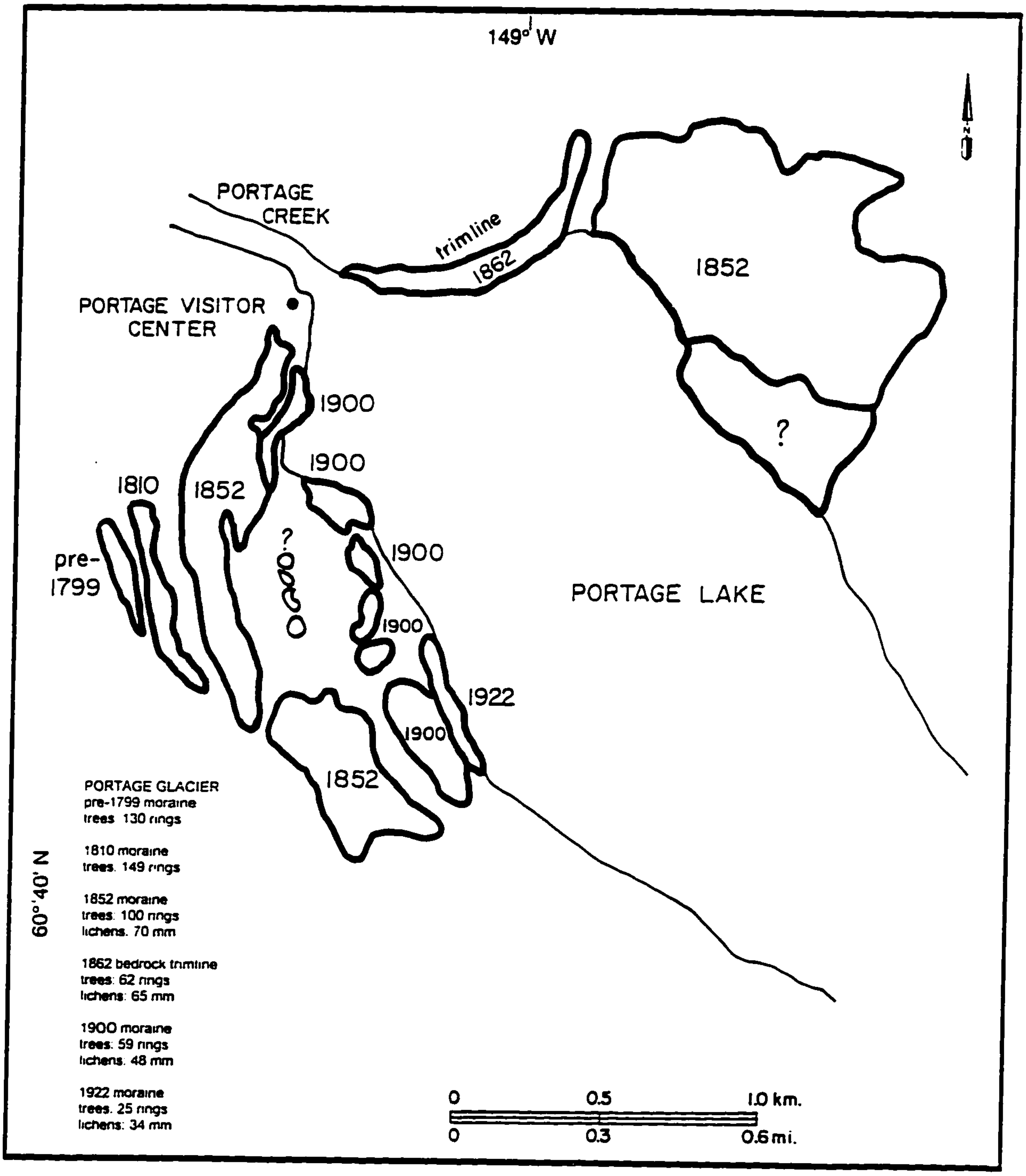

Fig. 3.2 - Map showing moraines along western end of Portage Lake. Ages assigned moraines are based on dendrochronology and lichenometry. Ages are approximate; tree-ring ages assume a 25-year colonization period.

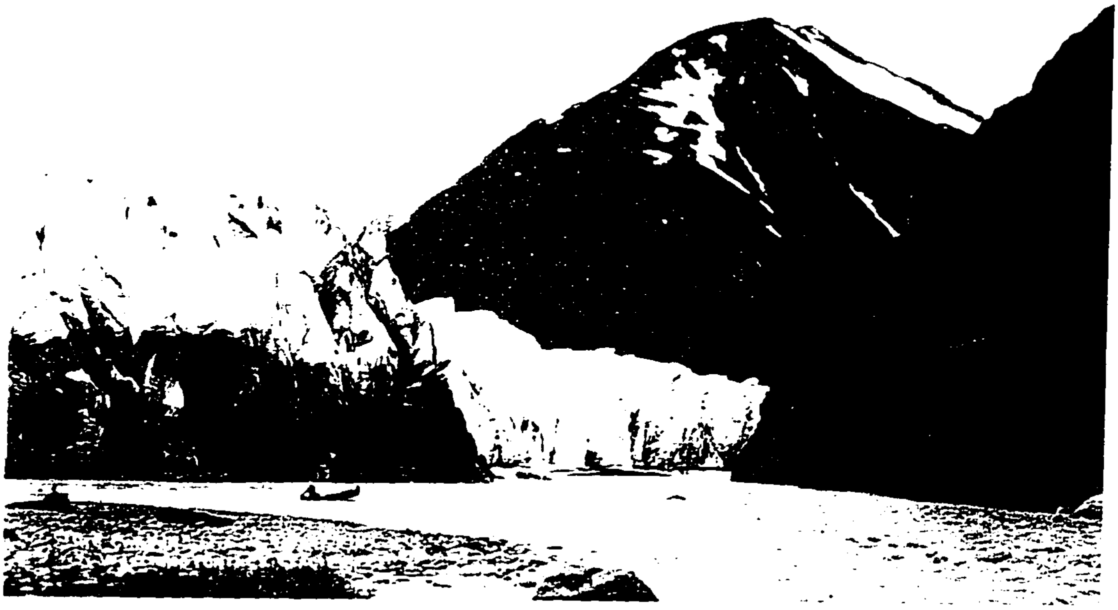

Fig. 3.3 - "The Canyon", 1910-15, west end of Portage Glacier. P.S. Hunt, photographer. B82.99.8 Anchorage Museum of History and Art.

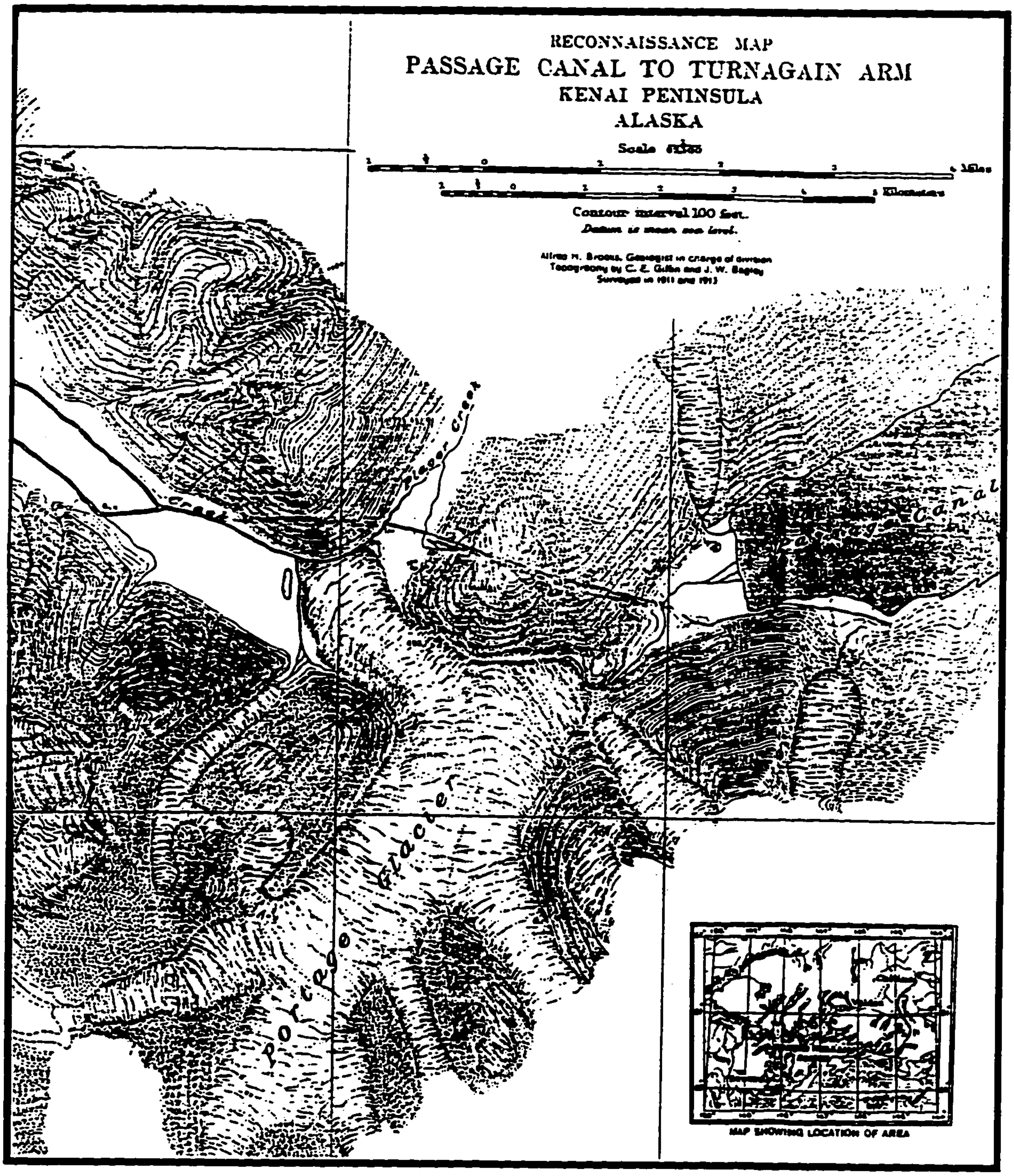

Fig. 3.4 - Map of Portage Glacier in 1911 (from Grant and Higgins, 1913).

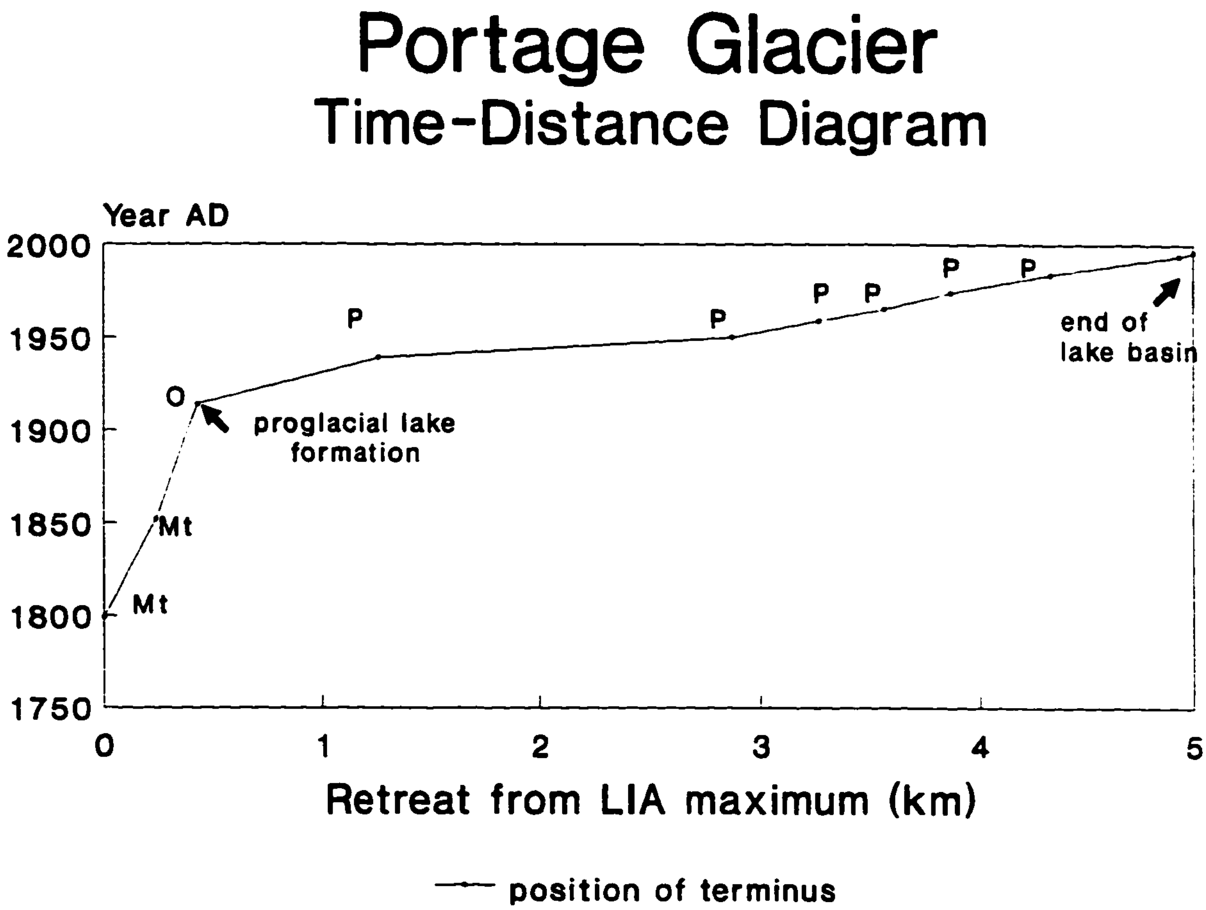

Fig. 3.5 - Time-distance diagram showing the LIA retreat of Portage Glacier based on evidence from moraines dated by tree rings (Mt), observations (O), and photographs (P).

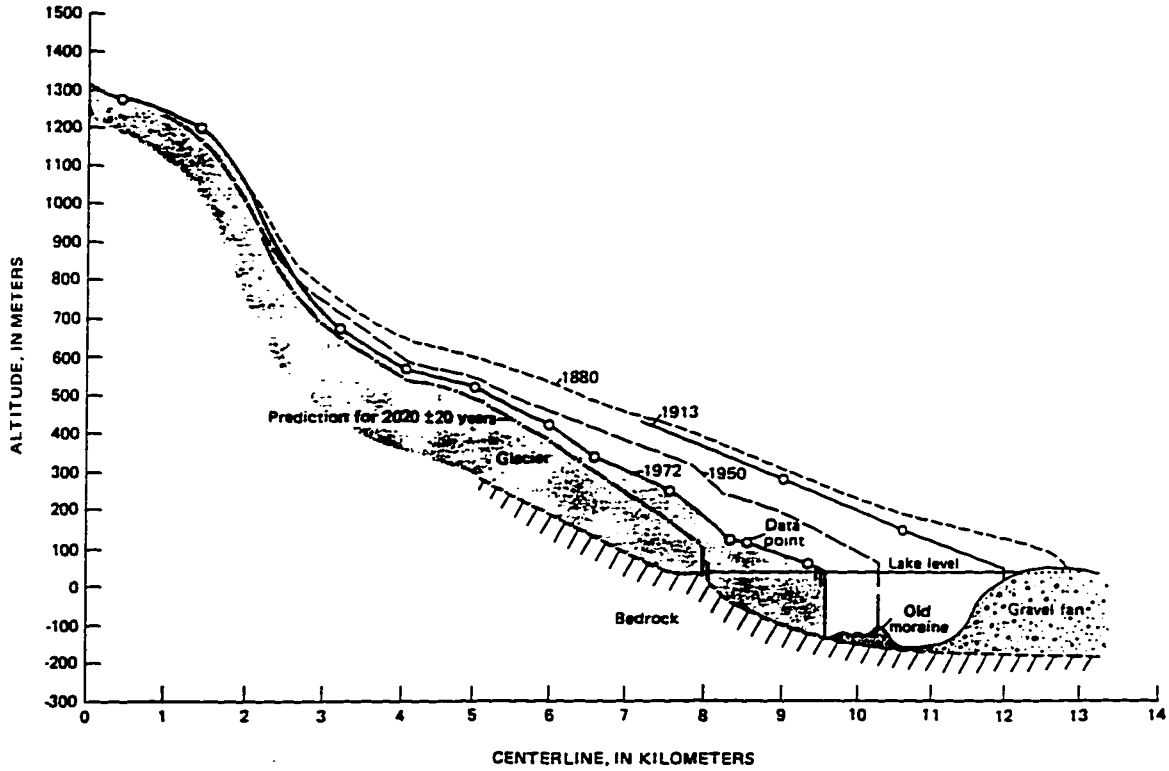

Fig. 3.6 - Portage Glacier retreat across Portage Lake (from Mayo and others, 1977).

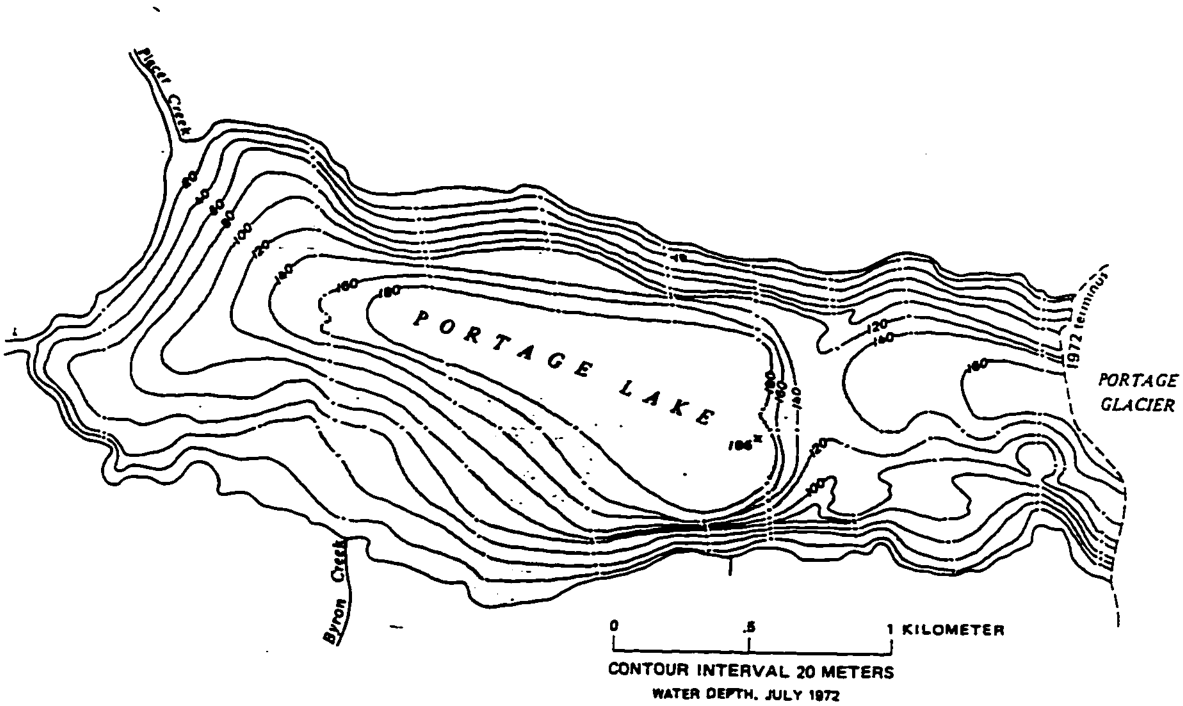

Fig. 3.7 - Bathymetry of Portage Lake (from Mayo and others, 1977).

Fig. 3.8 - Portage Glacier, July, 1989, looking south. Note supraglacial lake behind southern margin (right side) of the terminus. F. Gerhard, photographer.

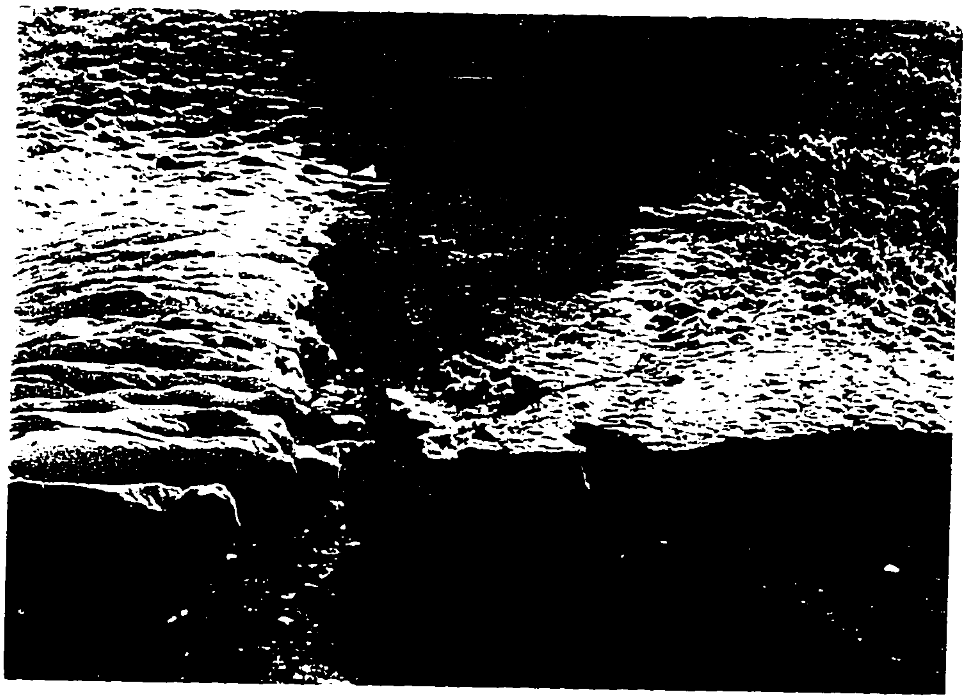

Fig. 3.9 - Supraglacial lake connected to Portage Lake, October, 1990, view looking southeast. Ice cliff bordering lake is ca. 15 m high. F. Gerhard, photographer.

TABLE 3.1 - PORTAGE VALLEY
CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Year of sampling	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Portage Valley - oldest moraine - west of VC moraine								
27	cored	*Picea*	1984	130	none	100	25+30	<1799
101				117	wide	30	25+9	<1833
140				115	good	20	25+6	1838
164				120	wide	15	25+5	<1834
Portage Valley - middle moraine - between VC and oldest moraine								
41	cored	*Picea*	1984	149	good	U	25	1810
Portage Valley - VC moraine - west shore								
179	cored	*Picea*	1985	57	good	15	25+5	1898
Portage Valley - VC moraine - south side - near Byron (1980?)								
17	cored	*Picea*	1984	66	good	U	25	1893
Portage Valley - VC moraine - central portion - near VC								
T-18	cut	*Picea*	1984	100	good	0	25	1859
158	cored	*Picea*	1984	79	good	10	25+3	1877
Portage Valley - VC moraine - Bear Valley								
122	cored	*Picea*	1984	83	good	10	25+3	1873
Portage Valley - trimline area								
192	cored	*Picea*	1985	62	good	0	25	1898
Portage Valley - lakeshore - between Byron channels								
203	cored	*Picea*	1985	48	good	10	25+3	1909
Portage Valley - lakeshore - Byron delta - near boulder with X								
306	cored	*Picea*	1985	35	good	10	25+3	1922
Portage Valley - lakeshore - zigzag moraine								
T-29	cut	*Salix*	1985	30	good	0		1955
174	cored	*Picea*	1985	23	good	U	25	1937

* Correction factors
25 - years for spruce colonization
0 - years for alder and hemlock colonization
n - years for correction of height cored (spruce only)
(see Table 2.1 for correction calculation)
Data condensed from Appendix C.

TABLE 3.2

MORAINAL AND PHOTOGRAPHIC EVIDENCE OF

THE RETREAT OF PORTAGE GLACIER

DATE	DISTANCE of RETREAT(m)	RETREAT RATE (m/yr)
<u>Terrestrial valley glacier</u>*		
1799-1859	240	4.0
1859-1914	190	3.5
Average		3.8
<u>Calving lacustrine glacier</u>**		
1914-39	840	33.6
1939-50	1600	145.5
1950-59	400	44.4
1959-65	300	50.0
1965-74	300	33.3
1974-84	460	46.0
1984-94	600	60.0
Average		56.3
1799-1994 TOTAL RETREAT	4930	

* Extrapolated margins required for 1799-1810 positions (due to subsequent erosion of moraines)
** Averaged margins required for 1914-1994 (due to irregularities across glacier terminus)

CHAPTER 4 - PORTAGE PASS AND BURNS GLACIER

Introduction

Portage Pass has been a route for travelers since the 18th century (Carberry, 1979). The topography rises steeply from the eastern shore of Passage Canal and levels to a flat crest at 245 m altitude before dropping southward to meet Portage Lake, Portage Glacier, and Burns Glacier.

Four distinct geomorphic/ecologic zones are found when crossing modern Portage Pass. The first zone, at the extreme northern end adjacent to Passage Canal, is covered by mature maritime spruce and hemlock forest. Thick peaty soils cover the bedrock together with large trees as much as 328 years old (Table 4.1). Younger trees, to 205 years old, occur halfway up the north side of the pass. In the second zone, at the north side of the crest, tundra vegetation covers glacially scoured bedrock, and trees exhibit stunted krummholz growth. Sedimentary rocks of the Valdez Formation are inclined at high angles, producing strike ridges and valleys that contain small rock-basin lakes and wet swampy areas.

A dramatic change in topography and vegetation occurs at the crest of the pass where strike valleys end against embankments of glacial debris, and narrow arcuate moraine ridges (Fig. 4.1) bury the strike ridges. This third zone marks a former ice limit. The arcuate moraines are covered with boulders and thickets of alder, willow, spruce, and hemlock containing krummholtz trees. The ridges, clearly defined in aerial photography, but difficult to follow on the ground, cover the area from the crest to the southern end of Portage Pass. The ridges are clear evidence of past glacier expansion into the pass, and are interpreted as having formed along the Burns Glacier terminus. When Burns Glacier acted as a tributary to Portage Glacier, a medial moraine separated the two ice tongues, and Burns Glacier advanced up Portage Pass (Fig. 3.4).

The fourth distinct area lies at the southern base of Portage Pass, adjacent to Portage and Burns glaciers and Portage Lake. Here a rocky open area with sparse newly colonized plants is found amidst meltwater streams, ice-cored debris, and stagnating ice blocks.

Area outside the LIA Limit

Although striated and ice-scoured bedrock are common on the north side of Portage Pass, many features attest to a considerable time since deglaciation: a mature forest (with trees >300 years old; App. C) and the presence of lichens up to 210 mm in diameter (App. D). Two basal peat samples from rock basins in Portage Pass outside the glacial limit have ages of 4900 ± 25 yr BP (calibrated age of 5724-5697 and 5656-5596 yr BP with 2 intersects; App. A) and 2770 ± 25 yr

BP (calibrated age of 2934-2905 and 2892-2846 yr BP with 2 intersects; App. A), indicating minimum dates for deglaciation.

Little Ice Age Landforms

At the Holocene glacial limit, the oldest moraine (Fig. 4.1, Table 4.1) is covered by trees up to 119 years old and lichens up to 105 mm in diameter. The 119-year tree-ring age is a minimum for deglaciation because the krummholz growth of the oldest trees forces their branches and trunks to spread laterally, making determination of the oldest limb difficult. Every lichen thallis >100 mm in diameter is growing on a single boulder, which may have been colonized as supraglacial debris. The maximum lichen size (excluding this boulder) is 92 mm diameter, suggesting deglaciation of the outermost moraine by 1808 AD. This date is consistent with the ages of the oldest moraines in Portage Valley. (Figs. 3.1, 4.1).

The second-oldest moraine (Figs. 4.1) is covered with trees as much as 75 years old, but all exhibit krummholz growth. Lichens measuring 85 mm maximum diameter imply that the moraine was deglaciated by ca. 1824 AD (Fig. 4.1). The third moraine south of the crest of Portage Pass (Figs. 4.1) is so small that few large trees grow on it, but the maximum lichen diameter is 72 mm, suggesting a date of 1849 AD. The similar chronological sequences of the three oldest moraines in both Portage Pass and Portage Valley imply that Burns and Portage glaciers were retreating at approximately the same time (Figs. 3.1, 4.1).

A series of narrow, undulating ridges covered with alder, willow, and spruce occupy the area adjacent to the north side of Divide Lake (Fig. 4.1). Deglaciation of the oldest of these moraines by 1880 AD can be inferred from lichens measuring 59 mm in diameter and from 51-year-old trees (Fig. 4.1).

The moraine directly adjacent to the north side of Divide Lake is covered with trees as much as 47 years old. This indicates deglaciation of this ridge by 1913, in excellent agreement with the photographic evidence. The 1914 photograph by Weir (Fig. 4.2) shows a steep ice front terminating north of present Divide Lake, with a narrow proglacial lake draining eastward along the ice front.

The 1939 Barnes photograph (Fig. 4.3) shows that the terminus had retreated beyond the south shore of Divide Lake. Divide Lake appears to be at its maximum extent, with a morainal isthmus separating the southern shore of Divide Lake from an extensive proglacial lake. This places the terminus at approximately the first moraine south of Divide Lake. Only the crest is exposed, with most of the moraine being submerged.

Post-1950 AD. ice limits are mapped from vertical aerial photography (App. A). From 1950 to 1974 Burns Glacier flowed as a tributary to the larger Portage Glacier; a medial moraine separated the two glaciers. Between 1950 and 1965, Burns Glacier retreated 300 m southward (average 20 m/yr). Between 1965 and 1974, the terminus retreated 140 m, with the retreat rate slowing to 15.5 m/yr (Fig. 4.4). However, during the 1970's this glacier experienced considerable thinning, and it stagnated at the base of Portage Pass (Fig. 4.4). Terminal retreat of 920 m (averaging 92 m/yr) occurred between 1974 and 1984. By 1984, Burns Glacier had completely abandoned Portage Pass, leaving only debris-covered ice slowly melting outside the former medial moraine. By 1984, the Burns terminus was located ca. 580 m from Portage Glacier, and separated from it by a glacially scoured bedrock surface (Fig. 4.1).

Burns Glacier retreated downhill from 1808 to 1974 (Crossen, 1995) to the base of Portage Pass. Streams flowing from high snowfields and outflow from Divide Lake ponded at the ice margin forming small ice-dammed lakes. These lakes are present in the 1950, 1951, and 1959 photographs, but it is not known whether they existed continuously from 1950-1959. No lakes appear in the 1974 and 1984 aerial photographs, but field reconnaissance in 1984 and 1985 showed that a small lake existed at the base of Portage Pass.

Although Burns and Portage glaciers differ in their dynamics and rates of retreat, their retreat histories are generally similar. Both advanced prior to the early 1800's, and retreated slowly during the 19th century (Figs. 3.5, 4.1, 4.4; Table 3.2). Retreat then accelerated in the 20th century. From 1890 to 1974 Burns Glacier retreated downslope across the southern side of Portage Pass (Fig. 4.4). Portage Glacier also retreated more rapidly during this period, but its retreat rate was considerably higher once calving began (Fig. 3.5). Burns Glacier experienced an unusually high retreat rate (Figs. 4.1, 4.4) from 1974-1984 when the stagnant ice mass at the base of Portage Pass wasted away.

Stream Piracy

Stream piracy has played an important role in the history of Portage Pass. Prior to 1939, Burns Glacier terminated in Divide Lake; subsequently, a series of small moraines was deposited between 1939 and 1977 from Divide Lake to the base of Portage Pass. In 1950, Divide Lake exhibited the same drainage pattern it had since its formation: a kidney-shaped lake draining east through a channel toward Shakespeare Glacier, then north through a scoured bedrock canyon and, after combining with discharge from Shakespeare Glacier, ending in Passage Canal. On the southwest side of Divide Lake, a hanging stream discharged ca. 25 m south of the lake and flowed downhill toward Burns Glacier and the base of Portage Pass. In 1950, a small tributary stream from Divide

Lake entered this stream (Fig. 4.5).

After 1950, this stream enlarged the canyon along the southwestern flank of Portage Pass (Table 4.2). In 1959 the canyon head reached the third moraine below Divide lake, but it is unknown what percent of Divide Lake discharged southward at that time. By 1965, the canyon trending south toward the ice front continued to deepen and widen, and a large landslide (possibly a result of the Alaska 1964 earthquake) blocked the earlier drainage channel from Divide Lake to Passage Canal. These two events led to the reversal of Divide Lake drainage and reduced the size of the lake. Subsequent aerial photographs show the increasing size of the dry lake bed surrounding the shrinking lake. After 1965 the added discharge from Divide Lake increased erosion of the southern canyon. Retreat and downwasting of Burns Glacier further promoted canyon deepening. From 1984-1995, headward erosion had drained the southern lobe of Divide Lake and increased the exposure of the lake basin (Fig. 4.5). Table 4.2 shows measured rates of headward erosion and canyon widening. The stream achieved a stable profile by 1987 when an arm of Portage Lake reached the base of Portage Pass.

Shakespeare Glacier

This small glacier descends 2.5 km along the east side of Portage Pass. The 1911 map of Portage Pass (Fig. 3.4) and a photograph from 1914 (App. B) show the terminus of Shakespeare Glacier very close to Burns Glacier. Earlier, Shakespeare Glacier may have been a tributary of Burns Glacier, because the 1914 Weir photography (App. B) shows a cleared area and boulder line in the foreland below the ice margin. However, by 1911 Shakespeare Glacier had separated from Burns Glacier. Although the older photography does not allow mapping of the terminus, excellent vertical aerial coverage after 1950 does.

A clearly defined set of bedrock strike ridges between the glacier terminus and the valley bottom is visible on all aerial photographs and was used to plot marginal positions (Fig. 4.1). The terminus was in retreat from 1950 (when the ice margin was located behind the 3rd strike ridge) to after 1974. From 1950 to 1965, the margin retreated 120 m (8 m/yr) and lay between the 4th and 5th strike ridges. By 1974, another 20 m of retreat (2.2 m/yr) brought the glacier behind the 5th strike ridge. By 1984, an advance (4 m/yr) brought the terminus to the 4th strike ridge, 40 m beyond the 1974 position. By 1990, following another 90 m of advance (6 m/yr) the ice margin lay at the 3rd strike ridge, the same position it occupied in 1950 (Fig. 4.6).

This advance likely is linked to measured climate changes about 1976, when a step increase in annual Alaskan temperatures occurred (S.A. Bowling, unpublished data). The

Past Global Changes (PAGES) program has determined that the period 1977-1994 was 0.25^{o}C warmer and 10 mm wetter in southcentral Alaska (Anderson, 1995). Subsequently, glaciers in this area experienced positive mass balances (Mayo and Trabant, 1984; Mayo and March, 1990). Although the accumulation areas of large glaciers have increased in thickness, their termini have not yet advanced. In comparison, small, steep glaciers like Shakespeare Glacier, began to advance.

Little Ice Age Advance of Portage Glacier

Portage Glacier advanced to its outer Holocene limit in Portage Valley prior to 1799 AD (see Chapter 3). Burns Glacier (as a tributary to the larger Portage Glacier) advanced prior to 1808 AD. A considerable difference of opinion exists as to when this ice advance occurred. The earliest written record of Portage Glacier and Portage Pass comes from Lt. Joseph Whidbey, a member of George Vancouver's crew, who on June 7, 1794 observed Portage Pass from Passage Canal and noted a valley of "tolerably even surface" which was "nearly destitute of any vegetable productions" and "equally passable in all directions" and free of snow (Vancouver, 1798). At this time, Russians and Indians, with canoes, were able to cross "the isthmus" between Passage Canal and Turnagain Arm (Carberry, 1979).

Teben'kov (1852) also reported the use of this portage, and in 1880 Ivan Petroff recorded that a "glacial formation forms the portage route between Chugatch Bay (now Passage Canal) and Cook Inlet" (Tarr and Martin, 1914). By the 1898 gold rush, prospectors and others were using Portage Pass as a short route to the goldfields, and observers from 1880 to 1909 reported that the glacier in Portage Pass terminated approximately 1 km from Passage Canal (Tarr and Martin, 1914), with its surface at ca. 300 m elevation (Mendenhall, 1900); it was then connected to Portage Glacier which flowed west to a gravelly outwash plain leading west to Turnagain Arm. Portage Glacier derives its name from this easy traverse from tidewater to tidewater.

These historical references have led some workers (Tarr and Martin, 1914; Field, 1975) to conclude that Portage Pass and Portage Valley were free of ice in 1794, but ice-covered by 1880. Likewise, Mayo and others (1977) concurred that "in 1794, the then unnamed Portage Lake was probably larger than it is today" and that Portage and Burns glaciers "advanced approximately 5 km between 1794 and 1880-1900".

This study suggests an alternative interpretation. The moraines in Portage Valley (dating as early as 1799 AD) and in Portage Pass (minimum date of 1808 AD) suggest an advance that culminated close to 1794 when Whidbey observed Portage Pass. Because he apparently viewed Portage Pass from the boat or the shore, no clear evidence would have been seen of a glacier advancing up the opposite side of the pass, or of a glacier terminating just below the crest opposite his viewing point.

As calculated in Chapter 3, Portage Glacier may have advanced sometime between 1382 AD and 1643 AD. Because Portage Glacier reached its maximum extent downvalley from the western lake shore at approximately the same time that Burns Glacier was reaching its maximum extent in Portage Pass, these two glaciers would have been an easy avenue for travel between Passage Canal and Turnagain Arm for at least 100 years.

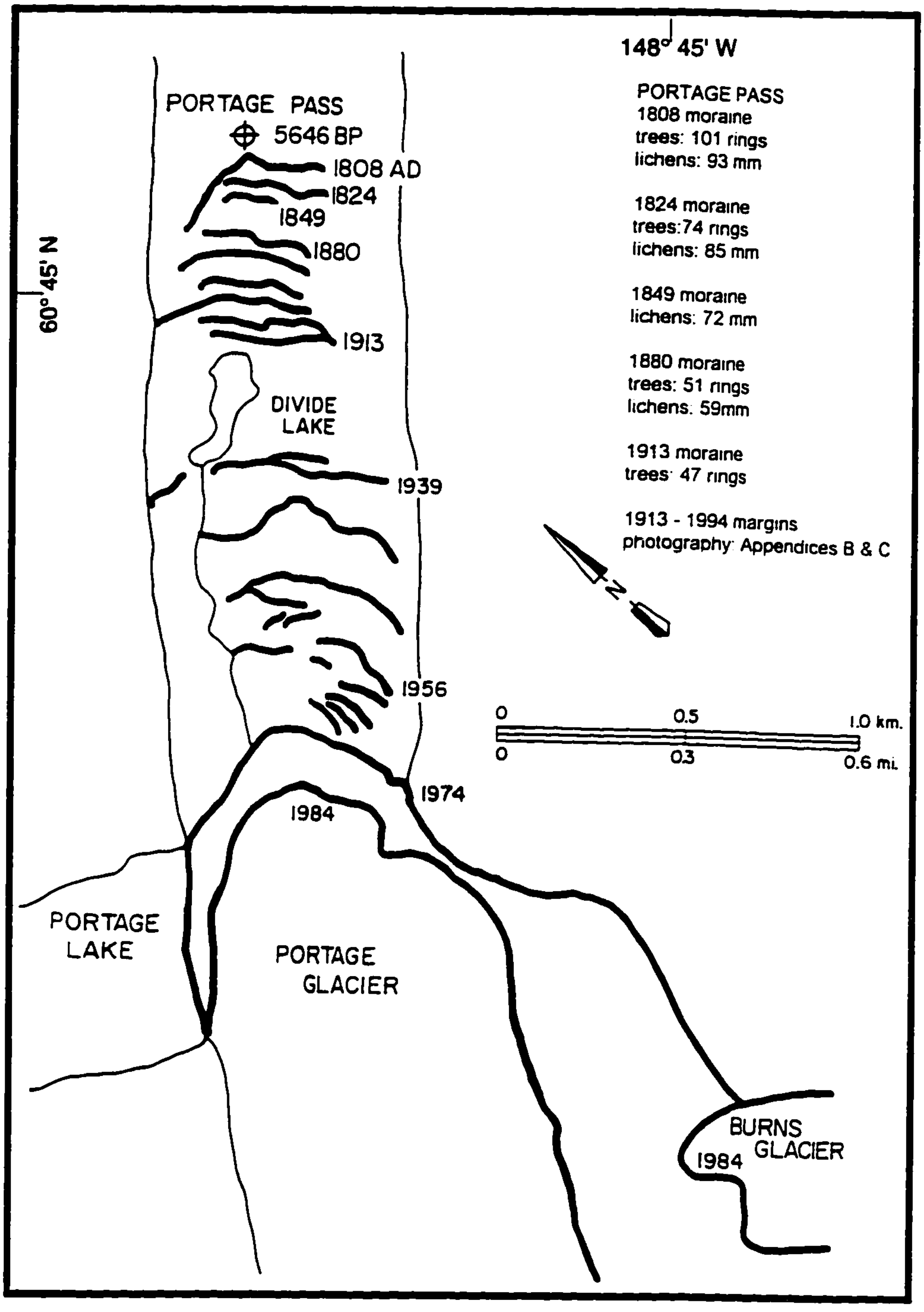

Fig. 4.1 - Map of moraines from the crest of Portage Pass to the terminus of Portage Glacier. Moraines dated by dendrochronology and lichenometry.

Fig. 4.2 - Photograph (1914) from Portage Pass, looking south, showing the margin of Burns Glacier north of Divide Lake, and the 1808 AD moraine along the right side of the photo.
(From F.F. Barnes, No. 261, U.S. Geological Survey Archives, Golden, Colorado.)

Fig. 4.3 - Photograph (1939) from Portage Pass, looking south, showing Burns Glacier terminus with proglacial Divide Lake at its maximum extent and draining eastward.
(From F.F. Barnes, No. 122, U.S. Geological Survey Archives, Golden, Colorado.)

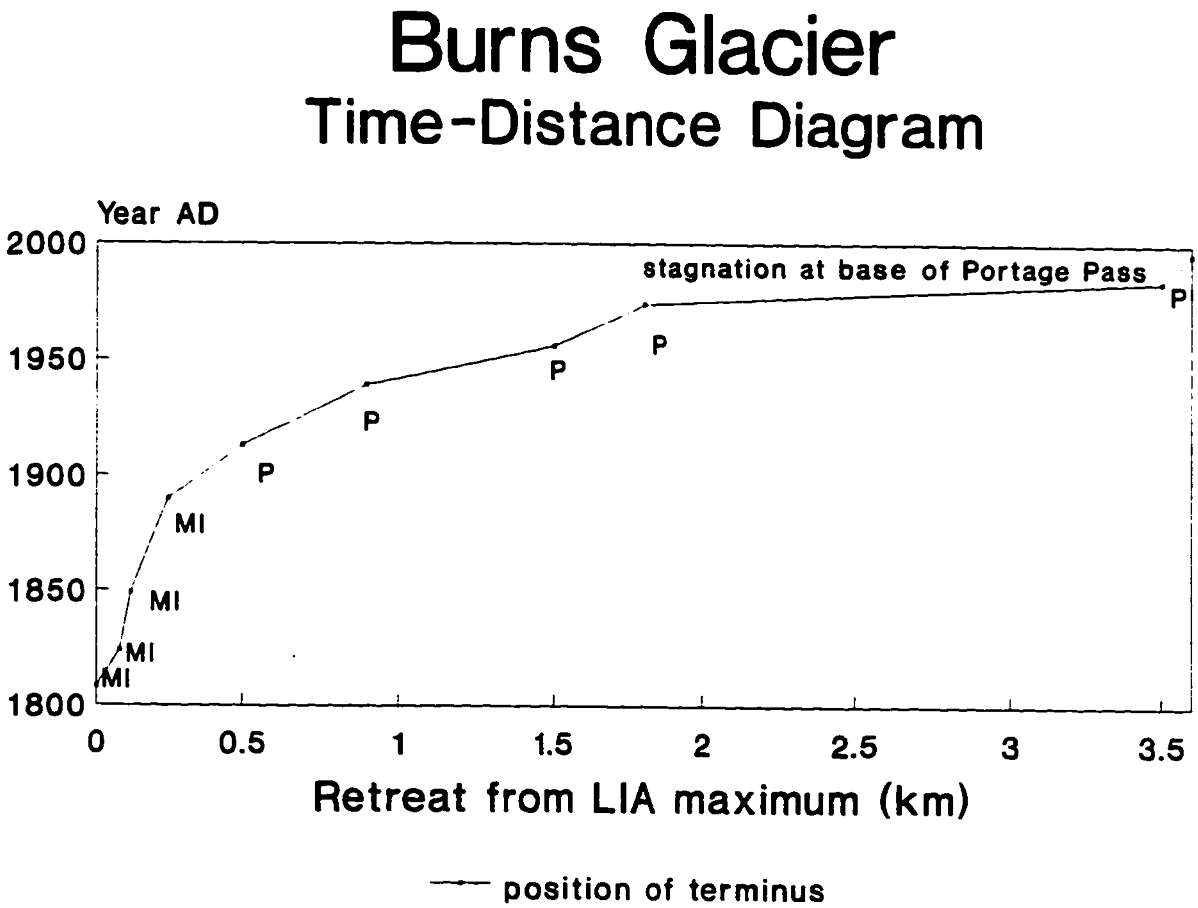

Fig. 4.4 - Time-distance diagram showing LIA retreat of Burns Glacier based on moraines dated by lichenometry (Ml), and photographs (P).

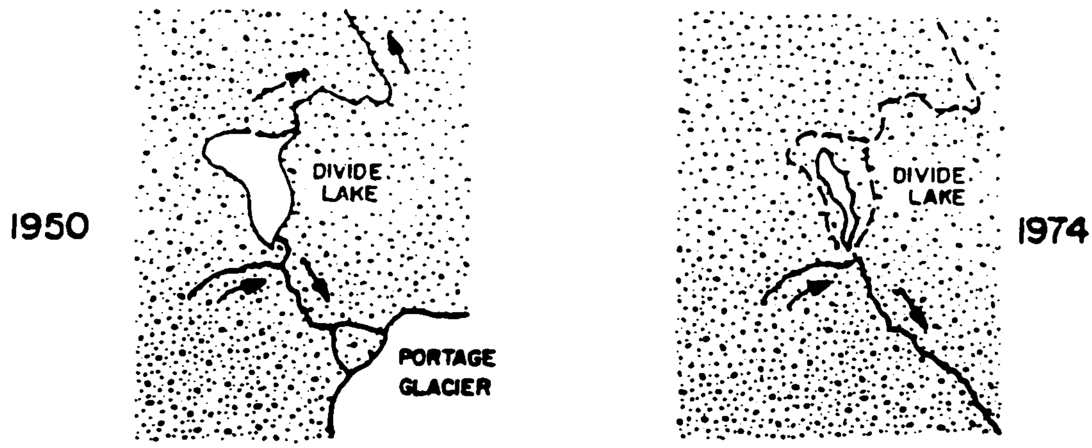

Fig. 4.5 - Drainage reversal of Divide Lake between 1950 and 1974.

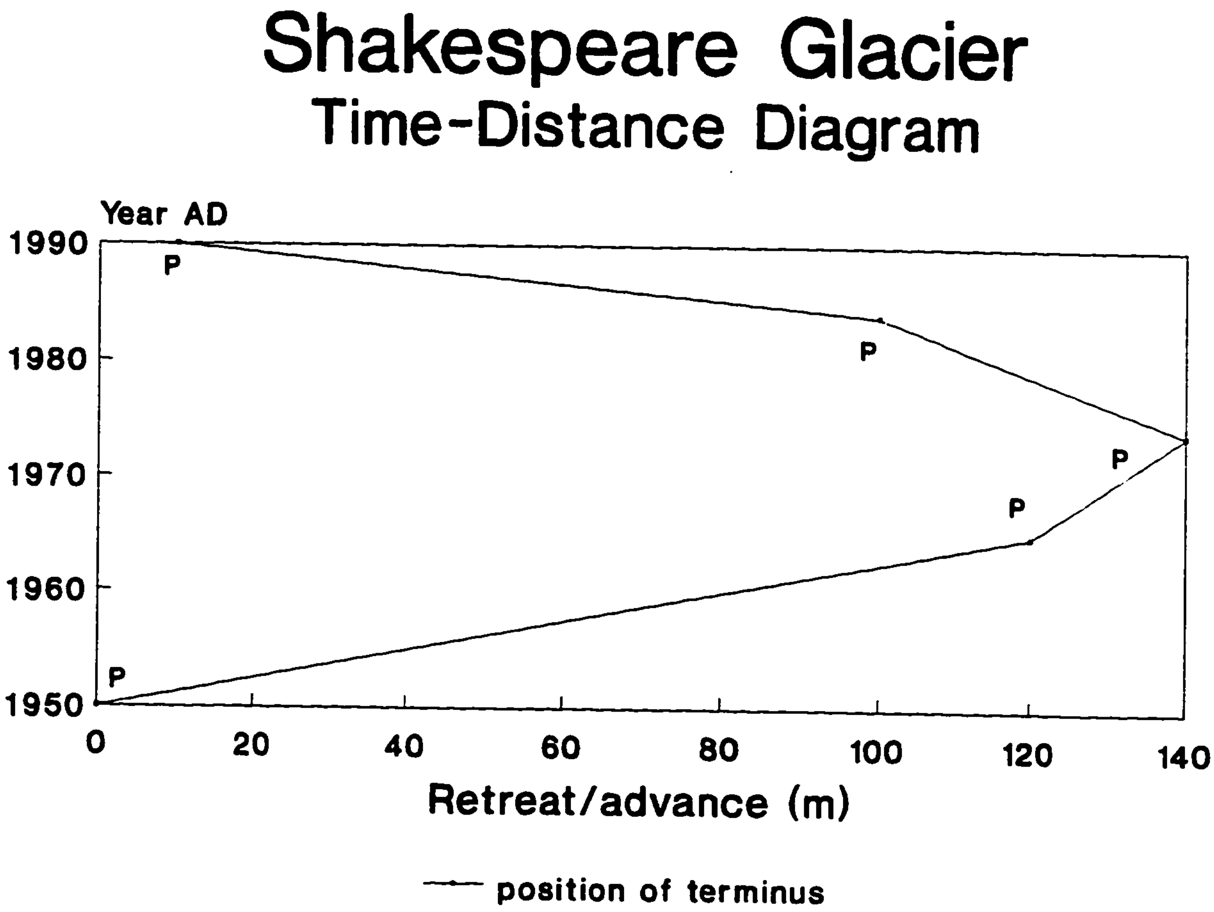

Fig. 4.6 - Time-distance diagram showing 1950-1990 AD fluctuations of Shakespeare Glacier based on photographic evidence (P).

TABLE 4.1 - PORTAGE PASS

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on ctr.	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Portage Pass - lowest area - mature forest - Whittier side								
300	cored	*Tsuga*	1985	328	good	30		1657
Portage Pass - mid elevation - outside glacial limit - Whittier side - central pass - near trail								
333	cored	*Picea*	1985	183	good	0	25	1777
343		*Tsuga*		205	good	5		1780
Portage Pass - outside glacial limit - top of pass - west side								
301	cored	*Tsuga*	1984	163	good	10		1822
302				170	none	25		<1814
Portage Pass - oldest moraine - west & central portions								
T-12	cut	*Tsuga*	1984	119	good			1865
Portage Pass - oldest moraine - east side - over canyon								
309	cored	*Tsuga*	1985	119	good	10		1865
Portage Pass - Dry Lake moraine								
T-14	cut	*Tsuga*	1984	49	good	0		1935
Portage Pass - between Dry Lake & Divide Lake								
T-15	cut	*Picea*	1984	41	good	0		1918
Portage Pass - south side of Divide Lake								
T-16	cut	*Picea*	1984	32	good	0		1927
Portage Pass - area below 17 mm moraine - across from bedrock area								
T-c	cut	*Alnus*	1987	12		0		1975

* Correction factors
25 - years for spruce colonization
0 - years for alder and hemlock colonization
n - years for correction of height cored (spruce only)
(see Table 2.1 for correction calculation)

Data condensed from Appendix C.

TABLE 4.2

CANYON EROSION IN PORTAGE PASS

BETWEEN DIVIDE LAKE AND THE ICE TERMINUS

DATES	HEADWARD EROSION total (m)	HEADWARD EROSION rates (m/yr)	CANYON WIDENING (max) width (m)	CANYON WIDENING (max) rates (m/yr)
1959			60	
1959-65	260	43.3	100	6.7
1965-74	60	6.7	140	13.3
1974-84	40	4.0	220	7.0
1959-84 (25 yr total)	360	14.4	160	6.4

Sources: Appendix A.

CHAPTER 5 - PLACER RIVER VALLEY

Introduction

The Placer River Valley lies south of Portage Valley and drains Spencer, Bartlett, and Skookum glaciers, as well as small unnamed ice masses (Fig. 1.1). Spencer Glacier, the largest glacier (82 km ^{2}), extends 19 km from its source area and terminates at 60 m elevation in the Placer Valley where its 2.5-km wide lobe ends in a small lake. The meltwater supplies much of the runoff feeding the Placer River.

Bartlett Glacier lies 6 km south of Spencer Glacier, in a hanging tributary valley, where its terminus reaches 394 m elevation and its meltwater flows into the Placer River via a narrow canyon. Where this stream enters the lower portion of the Placer Valley, a large delta is forming in the Spencer proglacial lake.

Trail Glacier is located 6 km south of Bartlett Glacier and covers 28 km^2. This glacier marks the southernmost extent of the Spencer-Blackstone Ice Complex, which ends where a high arete separates Trail Glacier from the area farther south.

Skookum Glacier, a 8-km-long ice tongue covering 7 km ^{2}, lies 8 km north of Spencer Glacier. It excluded from this study due to lack of moraines at its terminus.

Spencer Glacier

One major moraine lies approximately 1.5 km beyond Spencer Glacier. This moraine is a wide irregular landform (Fig. 5.1) with no obvious single crest, many ridges and depressions, and inferred meltout and channel features. Numerous moss-covered boulders protrude as much as 0.5 m above the moraine surface. The Placer River crosses the moraine, exposing 15 m of nonstratified sandy diamicton containing numerous graywacke boulders, many of which are striated. The Alaska Railroad (ARR) uses this moraine as its roadbed across the Placer River Valley and the railroad bridge crosses the Placer River where the river cuts the moraine.

Behind this moraine, a large pitted outwash plain occupies most of the deglaciated area northwest of the proglacial lake (Fig. 5.1), while a large delta, fed by the northward-flowing Placer River, progrades into the southern portion of the lake. Fragmentary sections of a young moraine are found approximately 20 m from the northwestern shore of the lake (labeled 1958 on Fig. 5.1). These ridges are 6-8 m high, have narrow undulating sparsely vegetated crests and steep sides (often near the angle of repose), and are composed of sand and gravel. Outwash surrounds the base of the ridges, where it has been eroded into two or more distinct levels.

An obvious trimline is present along both valley sides and numerous bedrock channels occupy former ice-lateral positions along the northern margin of the valley (Fig. 5.1).

Vegetation varies in different sections of the valley. An unusual feature of the large moraine is its cottonwood and alder cover that is devoid of spruce and hemlock trees. This is in sharp contrast to the trimlines and valley walls beyond the moraine that are covered with mature climax forests containing hemlock and spruce. Spruce are even growing inside the upper sparsely vegetated portion of the trimline.

Little Ice Age Landforms

The lack of inset recessional moraines and the wide areas of pitted and unpitted outwash limit the evidence of past ice margins beyond Spencer Glacier. The area outside the terminal moraine is covered by mature stands of hemlock forest growing in thick organic-rich soil with low underbrush. Cored trees were up to 138 years old (Table 5.1), but Viereck (1967) found trees up to 250 years old in this location. The landscape could be thousands of years older than this, based on comparisons with other valleys in the area.

The major moraine (Fig. 5.1) in this valley (traversed by the Alaska Railroad), sustains young trees, all cottonwoods. Table 5.1 lists numerous cores having 60-70 rings. The trees with the largest ring counts (72 and 73) also had rotten centers, so only a minimum age can be inferred. Based on these data, the oldest trees started growing before 1913, assuming a short colonization time for cottonwood. Viereck (1967) located a tree dating to the 1890's on this moraine, in agreement with Tarr and Martin's earlier observations (1912).

Although the railroad builders cut local trees for ties and burned areas adjacent to the tracks, this does not explain the numerous 1890-age trees here. The lack of spruce and youthfulness of the cottonwoods on this moraine cannot unequivocally be attributed to railroad construction along the crest of the moraine. Even more unusual is the absence of small (young) spruce. This is not due to a lack of seed source, as both the area beyond the moraine and the area of the trimline are covered with spruce. However, in some cases, when one species (e.g., cottonwood) colonizes a landform, the ecological and species interactions can cause exclusion of other competing species (G. Jacoby, pers. commun., 1992).

Although the terminal moraine can be dated only tentatively to the 1890's, it is continuous with a trimline along the valley sides. Along the north side of the valley, this trimline dates to ca. 1869 based on the 62-mm lichens found there (Fig. 5.1; App. D). This same feature also supports trees that colonized in 1898 (Table 5.1), so this suggests either that downwasting began by 1869 and predated frontal retreat by ca. 20 years, or that the terminus readvanced between 1869 and 1890.

A lower trimline appears to have been deglaciated by 1907, based on the lichens measured there (Fig. 5.1). This agrees with the historical observations (Tarr and Martin, 1912) of the 1911 ice-marginal bedrock channel along the northern edge of Spencer Glacier (Fig. 5.1).

At least two ice-marginal channels eroded into bedrock along the northern side of the glacier indicate its downwastage into the valley, although it is assumed that the channels were used as subglacial conduits prior to their deglaciation. Although these channels are now choked with alder, they are related to some railroad construction. By 1907 the railroad constructed a dike from the valley wall to the railroad track to divert water flowing through the bedrock gorge north of the dike (Tarr and Martin, 1912). When this failed, the water was diverted to the adjacent channel to the south, which also marked the terminus in 1911 (Fig. 5.1). The rock used to construct these dikes and a dam near the bedrock hill was quarried from the hill at the east end of the dikes. These quarries serve as a reference point for the lichen curve developed for this region (Fig. 2.1)

A log collected from the terminal moraine (Karlstrom, 1964) was likely buried by Spencer Glacier's advance to its

LIA maximum. The reported radiocarbon date of "post AD 1650" (Karlstrom, 1964) is not useful for determining the date of that advance. Wentworth and Ray (1936) reported an advance in the late 1890's and again in 1916, a slight recession between 1890 and 1906, and about 640 m retreat between 1906 and 1931. Karlstrom (1964) recorded small push moraines developed between 1931 and 1951. The American Geographical Society determined that 500 m of recession occurred from 1931 to 1964, and that the glacier terminated 1.2 km from the 1890's moraine in 1964 (Field, 1975).

The chronology constructed for Spencer Glacier is based upon the dendrochronology of the single large moraine (Fig. 5.1), lichenometry (App. D) and dendrochronology (App. C) of the trimlines, maps and photographs from the ARR archives (provided by Ken Liebner, ARR Engineering), and aerial photography (App. B). Figure 5.2 shows the retreat of Spencer Glacier from its late 1890's LIA moraine, with retreat continuing to the present time. It may be assumed that the ice held this LIA maximum position, with little frontal retreat, from prior to 1869 (when downwasting of the margins began) until sometime in the late 1890s. Alternatively, the ice recession began by 1869, but an advance to the outer moraine occurred during the 1890s.

Table 5.2 indicates the average retreat rate from the LIA maximum to the present for this valley glacier. It also indicates that the formation of the modern proglacial lake, although currently quite large, has not substantially affected the retreat rate, and suggests that the lake is not very deep. The small number of icebergs in this lake also suggest a similar conclusion about its depth.

The lack of concentric recessional moraines behind the late 1890s moraine possibly is caused by two factors: 1) the large pitted outwash plain suggests that meltwater processes may have eroded or buried any small moraines, and 2) the modern proglacial lake may cover other small moraines. A small moraine along the northern lakeshore, with one 28-year-old willow growing on it in 1986 (App. C), indicates deglaciation of that region by 1958 (Fig. 5.1). This agrees with photographic data (Appendix B) that show the area to be deglaciated after 1951.

This valley is less productive than others for glacial chronological studies because it has been heavily impacted since the early 1900s by railroad construction, quarrying, construction of dikes, and mining of gravel. In addition, the U.S. Army uses this outwash area to stage glacier training missions, which results in substantial disturbance of vegetation.

Bartlett Glacier

Bartlett Glacier (Fig. 1.1) is a nearly isolated ice mass that extends 7 km down its valley from surrounding peaks and aretes. Only one small area is connected to the higher Spencer-Blackstone Ice Complex, giving the glacier a limited accumulation zone. The glacier currently terminates at about 330 m. My research indicates that the shape and elevation of the terminus are substantially different than shown on the 1951 Seward (C-6) quadrangle map (produced from August, 1950 photography) when compared to the June 25, 1951 vertical photography (App. B).

This glacier was investigated by Karlstrom (1964) and is the type locality of the Tunnel I and Tunnel II moraines of his Alaskan Age events. The Tunnel I advance was dated as AD 550 ± 220 years (Karlstrom, 1964) and is calibrated to 650 AD (Stuiver and Pearson, 1993). The Tunnel II advance is dated to 1500 AD based on correlation with moraines around Tustemena Lake (Karlstrom, 1964).

The modern glacier has a broad terminus that narrows as it extends into a bedrock canyon near the center of the valley. The glacier flows over a series of bedrock strike ridges along the northern side of its terminus, and the higher topography on that side has caused thinning of the ice and an indentation of the terminus. In comparison, a series of small alder-covered moraines dominate the land beyond the southern side of the terminus (Fig. 5.3). Retreat of Bartlett Glacier left a distinct trimline along both sides of the valley, but only the southern edge exhibits a set of end moraines. A thick ice-cored lateral moraine is melting out along the southern edge of the terminus. The northern edge of the glacier is retreating over bedrock strike ridges that are covered by thin debris, without forming distinct moraines. The past advances of this glacier were controlled by a bedrock ridge (now covered with glacial debris) that directed the southern edge of the ice downvalley to the north (Fig. 5.3).

Bartlett Glacier proved to be poorly suited for lichenometric studies: in areas lacking dense alder growth, the lichen cover is sparse on boulders and bedrock ridges. Although the high elevation of this glacier may be a factor influencing lichen colonization, the results from this site are not easily explained.

Little Ice Age Landforms

Figure 5.3 shows the positions of the Bartlett Glacier terminus during the past 100 years based on the dendrochronological (Table 5.3) and photographic evidence (App. B). A distinct trimline separates mature spruce-hemlock forest from alder- and cottonwood-covered moraines. The oldest trees are cottonwoods (similar to the Spencer Valley moraines) and date to >1898 AD (Table 5.3) suggesting the outermost moraine was formed in the later part of the 19th century. Early photographs from 1906-1911 (App. B) show an extensive area of bare ground adjacent to the glacier terminus, and a trimline outside the ice margin (App. B). Post-1950 positions are mapped using vertical aerial photography (App. B).

The landform I have interpreted as the pre-1898 moraine (Fig. 5.3) is Karlstrom's Tunnel I moraine, and my 1906-11 margin is his Tunnel II event. The surface vegetation on what I consider to be the LIA moraine matches late 19th century moraines in adjacent valleys, so I infer glacier retreat beginning before 1898 (Fig. 5.3). Karlstrom considered his radiocarbon date of 550 ± 220 AD (calibrated to 427-883 AD; App. A) to place the age of the Tunnel I advance between 500 and 1500 AD. The dated material was a rooted stump buried by 2 m of bouldery till (Karlstrom, 1964), so it may date the advance to the LIA maximum in this valley. If so, it would require Bartlett Glacier to hold its maximum LIA position for more than 1000 years, a situation that seems unlikely based on other chronologies in the area (see Blackstone Bay and Tebenkov Glacier discussions in subsequent chapters). Alternatively, this may date an earlier Neoglacial advance, or may be an erroneous date.

Karlstrom (1964) based the dating of his Tunnel II advance (1500 AD) on correlations with the Tustemena Glacier moraines east of this area. I have mapped the 1906-1911 positions (Fig. 5.3) from ARR photography and believe that this area is simply a transient location during the retreat from the LIA maximum.

A radiocarbon age of 2140 ± 25 yr BP (2288-2066 cal yr BP; App. A) that I obtained from basal peat just outside the moraine (Fig. 5.3), shows that this valley was deglaciated more than 2000 years prior to the LIA maximum. Karlstrom (1964) also obtained a date of 2370 ± 100 yr BP (2322-2703 cal yr BP with 2 intercepts; App. A) for transported wood below the Tunnel I moraine (Fig. 5.3) and interpreted it to date the Tustemena III advances on the Kenai Peninsula. Because these dates overlap, they may be dating the same buried horizon, and it is difficult to say whether they date a previous advance or just a previously deglaciated surface.

Table 5.4 details the retreat of Bartlett Glacier from its LIA maximum. The unusually rapid rate of retreat at the turn of the century suggests the lobe stagnated in place (Figs. 5.3, 5.4). Retreat slowed to a rather steady average rate in the 20th century (Table 5.4).

The intensive use of the glacial deposits by the railroad have made this location a less than perfect situation to extract lichenometric and dendrochronological data. The railroad does, however, provide excellent access to these glaciers in the Placer River Valley and the quarries used there helped to establish the lichen curve for the region (Fig. 2.1).

<u>Trail Glacier</u>

Trail Glacier is a 11-km-long ice tongue that extends south of the drainage divide at Grandview (an Alaska Railroad sectionhouse), and its meltwater flows south toward Seward instead of north into the Placer River.

The area downvalley (west) of a moraine (LIA?) is covered in thick old-growth forest, and resembles other glacier forelands deglaciated for thousands of years. No radiocarbon samples were recovered from the area, but based on similar landforms in adjacent valleys, I infer that Trail Glacier has not advanced down its valley as far as the Alaska Railroad (ARR) tracks during the past several thousand years.

Historical studies of this glacier show 1200 m of retreat from its LIA maximum by 1931 (Wentworth and Ray, 1936), and an additional 500 m of recession between 1931 and 1957 (Field, 1975).

This project did not include Trail Glacier, because investigations showed no moraines adjacent to the ice margin, and a possible moraine that lies downvalley from the terminus also lies in the path of numerous avalanches that travel down the south-facing slope and impact the trees on that moraine. Most of the trees had been knocked down when I visited this moraine, and none of them were very large. This leads me to conclude that numerous avalanches over the years had decimated the trees growing on the moraine, making this a poor location for a glacial chronology based on dendrochronology. Also the dense vegetation found on the moraine had overgrown the older lichens, making it impossible to determine a lichen-based chronology.

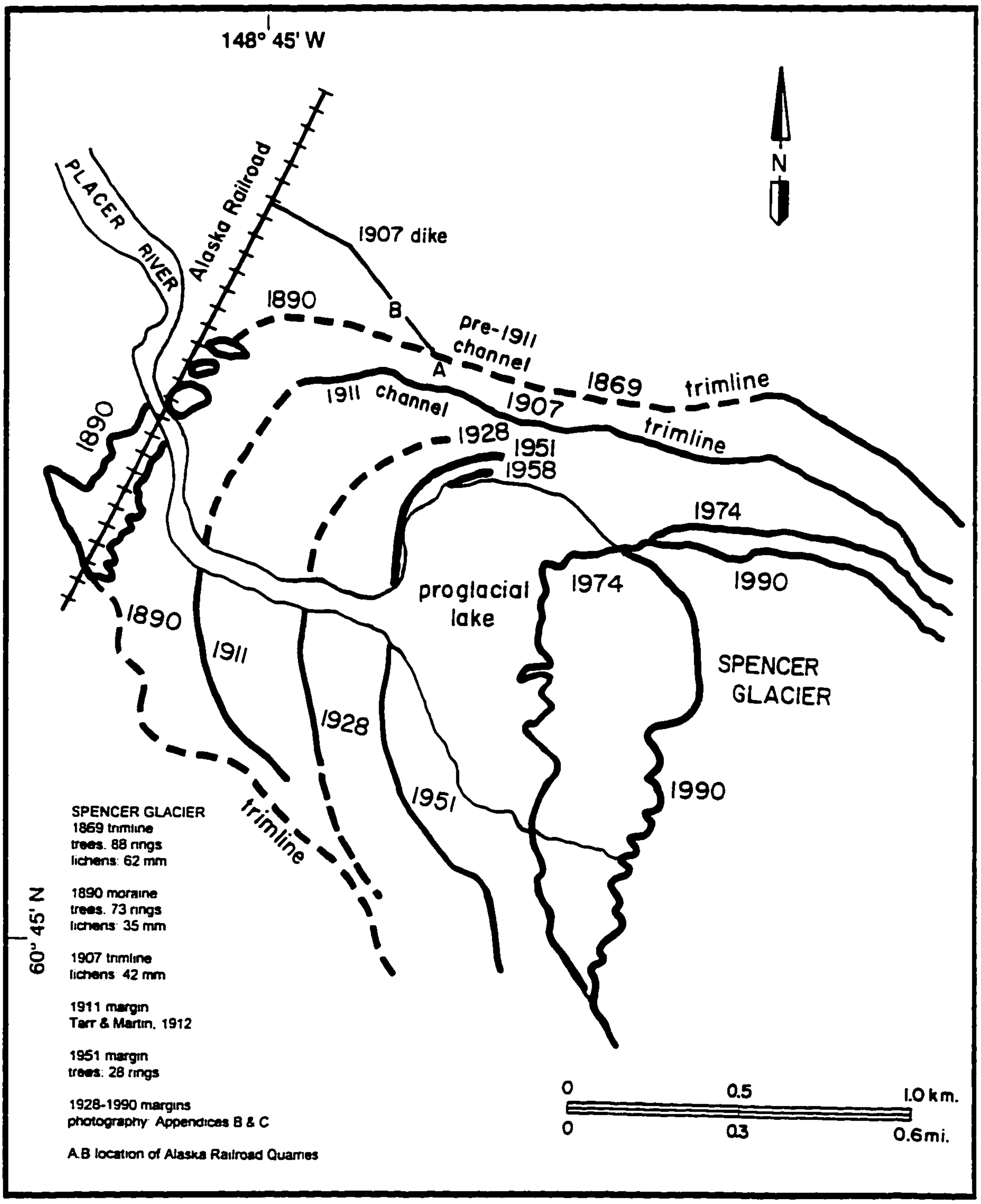

Fig. 5.1 - Map of the Spencer Glacier forelands showing LIA moraines, trimlines, ice-marginal channels, and proglacial lake. Locations A and B indicate the Alaska Railroad quarries and 1907 dike.

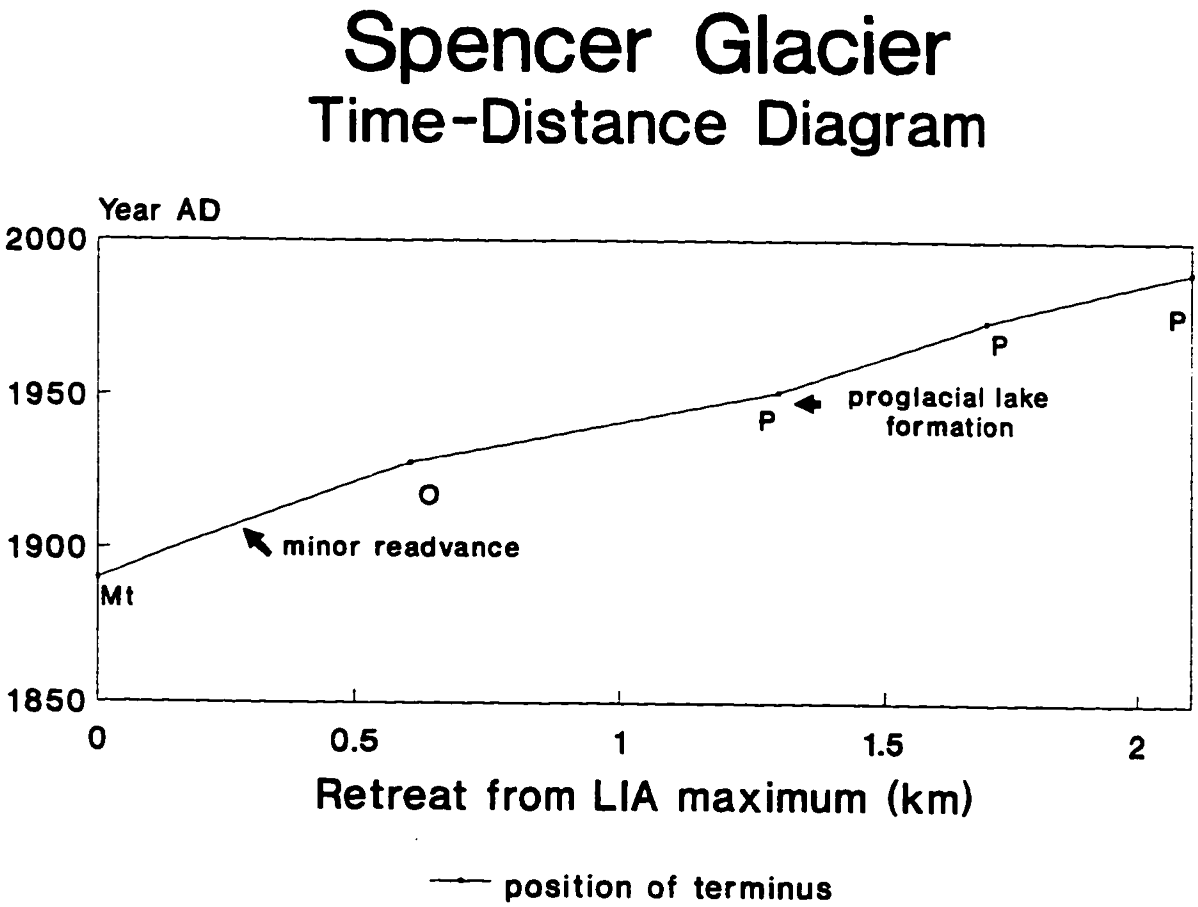

Fig. 5.2 - Time-distance diagram showing LIA retreat of Spencer Glacier based on evidence from moraines dated by tree-rings (Mt), observations from the Alaska Railroad (O), and photographs (P). A minor readvance was noted in 1916 (Wentworth and Ray, 1936).

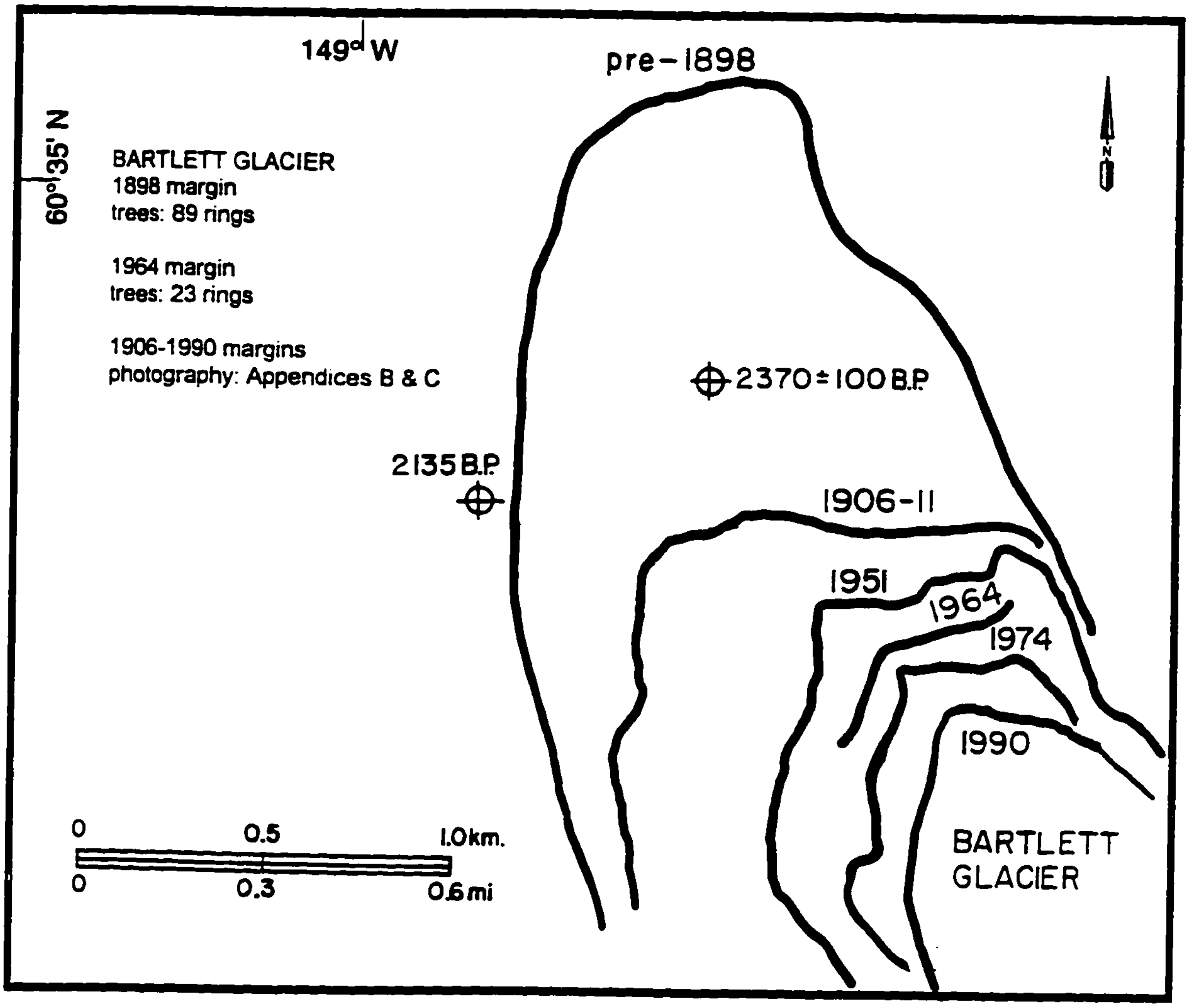

Fig. 5.3 - Map of Bartlett Glacier forelands showing positions of LIA moraines and radiocarbon samples. Karlstrom (1964) designated this area the type locality of the Tunnel I and II moraines of the Alaskan age events, but dendrochronologic and photographic evidence suggest the Tunnel I location is the outermost LIA moraine predating 1898 AD, and the Tunnel II location is a transient position during post-LIA retreat.

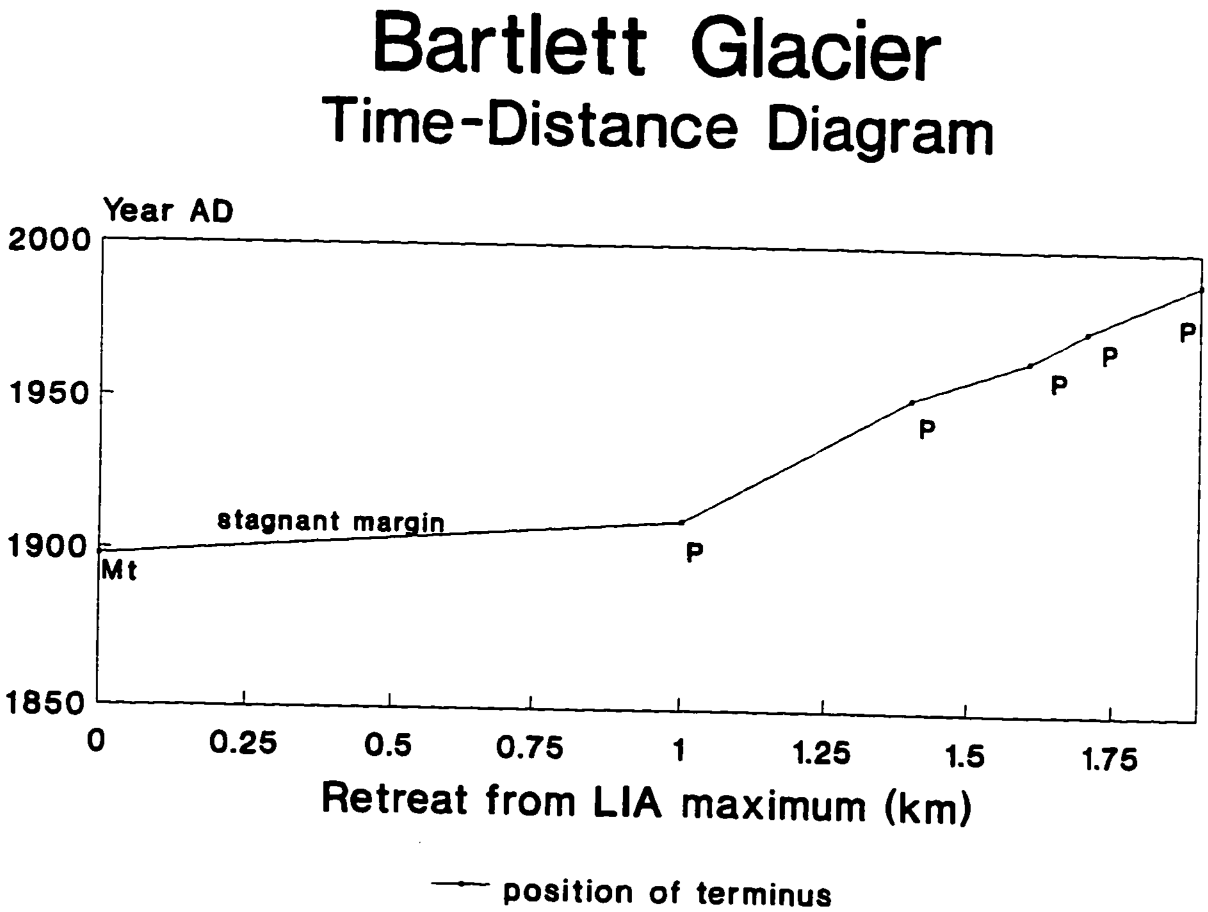

Fig. 5.4 - Time-distance diagram showing LIA retreat of Bartlett Glacier based on evidence from moraines dated by tree-rings (Mt) and photographs (P).

TABLE 5.1 - SPENCER GLACIER

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Spencer-outside glacial limit								
376	cored	*Tsuga*	1986	138	good	20		1848
Spencer-ARR moraine								
359	cored	*Populus*	1986	73	none	20		<1913
Spencer-below trimline								
379	cored	*Picea*	1986	60	good	10	25+3	1898
T-d	cut	*Tsuga*	1986	60				1926
Spencer - lakeshore moraine								
T-1	cut	*Salix*	1986	28	good	0		1958

* Correction factors
- 25 - years for spruce colonization
- 0 - years for alder and hemlock colonization
- n - years for correction of height cored (spruce only) (see Table 2.1 for correction calculation)

Data condensed from Appendix C.

TABLE 5.2

RETREAT OF SPENCER GLACIER

MORAINAL AND PHOTOGRAPHIC EVIDENCE

DATE	DISTANCE of RETREAT (m)	RETREAT RATE (m/yr)
1890-1928	640	21.3
1928-1951	640	27.8
1951-1974	427	18.6
1974-1990	450	28.1
Average		21.6 m/yr
TOTAL RETREAT 1890-1990		2.16 km

TABLE 5.3 - BARTLETT GLACIER

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Bartlett - outside glacial limit								
430	cored	*Tsuga*	1987	189	good	15		1798
Bartlett - RR moraine								
425	cored	*Populus*	1987	82	none	45		1905
436				89	none	35		<1898
Bartlett - older moraine - adjacent to ice - Sample Bag B								
B-a	cut	*Alnus*	1987	23	good			1964
Bartlett - deglaciated ridge - adjacent to ice - Sample Bag A just outside 74 margin on aerial photo								
A-c	cut	*Populus*	1987	19	good			1968
Bartlett Glacier - deglaciated since 1974 photo - southwest edge of ice								
T-19	cut	*Alder*	1987	12				1975

* Correction Factors
25 - years for spruce colonization
0 - years for alder and hemlock colonization
n - years for correction of height cored (spruce only)
(see Table 2.1 for correction calculation)

Data condensed from Appendix C.

TABLE 5.4

RETREAT OF BARTLETT GLACIER

MORAINAL AND PHOTOGRAPHIC EVIDENCE

DATE	DISTANCE OF RETREAT (m)	RETREAT RATE (m/yr)
1898-1910	1,016	84.7
1910-1951	406	9.9
1951-1964	203	15.6
1964-1974	102	10.2
1974-1990	200	12.5
Average (1898-1990)		21.0
TOTAL RETREAT 1898-1990		1.93 km

CHAPTER 6 - BLACKSTONE BAY

Introduction

Blackstone Bay is a 19-km-long embayment along the western side of Prince William Sound (Fig. 1.1). Willard Island, in the center of the bay, is surrounded by five glaciers. Two tidewater glaciers, (Blackstone and Beloit) terminate in steep ice cliffs along the western end of the bay, and three smaller ice tongues (Marquette, Lawrence, and Ripon) extend close to tidewater along the southern side of the fiord.

Willard Island displays a trimline that crosses the island about midway down its axis. On the northeastern side of the trimline, thick peat and old-growth forest mantle the surface, while on the southwestern side, alder thickets containing occasional spruces cover bedrock strike ridges. The trimline is continuous with submarine bars that extend across Blackstone Bay both north and south of the island. These rise precipitously from depths of 96-160 m (Post, 1980) and are covered by angular striated boulders. The shoals act as navigation hazards, trap icebergs, exhibit proximal slopes steeper than distal ones (Post, 1980), and are interpreted as submarine moraines (Grant and Higgins, 1913; Tarr and Martin, 1914).

Blackstone and Beloit Glaciers and Willard Island

The alder-covered area southwest of the trimline (Fig. 6.1) appears young in age. However, many large old spruce trees are interspersed among the alder thickets. In 1910, Tarr and Martin (1914) noted trees 20 cm (12") in diameter. In 1935, Cooper (1942) found a 417-year-old tree growing within 3 km of the termini of Blackstone and Beloit glaciers. My research examined large spruces from both central and western Willard Island, but all trees cored were too large to reach the center rings. The maximum number of rings recorded was 276 (Table 6.1).

In addition, C. Heusser cored four trees on Badger Point (on the mainland west of Willard Island) and allowed me to process the cores. The maximum ring count totaled 265 (Table 6.1), but no cores reached the center of the tree. Both Heusser (1983) and Cooper (1942) recognized alder communities more than 250 years old in other portions of Prince William Sound. These dominate areas for extended periods and likely explain the apparent trimline on Willard Island.

Heusser (1983) and Post (1980) report a radiocarbon date of 580 $\pm$ 55 yrs BP (calibrated age of 531-569 or 592-642 yrs BP) from basal bog peat on Badger Point. This data has led Post (1980) to conclude that the morainal shoals on either side of Willard Island could have been deposited as late as 1350 AD, when Beloit and Blackstone glaciers reached their

Neoglacial maximum in Blackstone Bay (Fig. 6.1).

Lichenometry was unsuccessful on Willard Island due to the age of the substrate. No large *Rhizocarpon* thalli were located among large lichens of other species. The lack of large *Rhizocarpon* thalli is likely caused by spalling of the strike ridges and the thick undergrowth that shades and kills older *Rhizocarpon.*

Few features mark the retreat between the ca. 1350 AD morainal shoal and the modern tidewater glacier termini (Figs. 6.1, 6.2, 6.3). Post (1980) indicates likely locations for the ice margins in 1909, 1935, and 1952 as noted by historical references (Grant and Higgins, 1914; Field, 1937) and aerial photography (Appendix B). In 1909, Blackstone and Beloit glaciers extended ca. 1 km down their respective fiords (Grant and Higgins, 1914), so that only the eastern tip of the bedrock ridge now separating them was exposed (Fig. 6.1). From 1935 to 1950, Beloit Glacier terminated on a sharp-crested shoal about 0.6 km (0.4 mi) in front of its current position (Fig. 6.1).

The current termini of Beloit and Blackstone glaciers appear stable at retracted fiord-head positions (Figs. 6.2, 6.3). Both descend steep slopes to tidewater and discharge small icebergs. Both end in shallow water less than 30 m deep (Post, 1980). Lethcoe (1987) reports that the eastern edge of Blackstone Glacier retreated several feet above high tideline during the early 1980's, but reached tidewater by 1987. Dramatic calving events occur along the northern valley wall adjacent to the Blackstone terminus, where ice breaks off and cascades down 75 m cliffs to tidewater.

Marquette Glacier

Marquette Glacier forms the westernmost of the three ice tongues descending the southern wall of Blackstone Bay (Fig. 6.1). Although a clear trimline separates the alder-covered forelands from the densely forested slopes above, no morainal ridges surround the ice margin. A bedrock ridge along the western edge of the glacier controls the location of the ice margin.

Evidence for the LIA maximum extent is equivocal due to the lack of moraines and a sparse lichen cover. A single large boulder in eastern glacier foreland dates to ca. 1894 based on lichenometry (Fig. 6.1).

Grant and Higgins (1913) reported that the glacier terminus reached high tide in 1909, with outwash prograding down the tidal flats, whereas Tarr and Martin (1914) report the ice terminated 50-65 m (150-200') from the water in 1909-1910 (Fig. 6.4). The American Geographical Society (AGS) 1935 photography (App. B) shows this glacier extending almost to the waterline, with an embayment along the western side where a bedrock ridge underlies the ice. The AGS 1957 oblique photography (App. B) shows bedrock exposed along the western edge of the glacier, with ice close to the water along the eastern side.

Deglaciation of the present intertidal area (submerged during the 1964 earthquake) dates to 1942, based on the 22-year-old rooted alder stumps found there (Table 6.2). Shoreline changes resulting from the 1964 earthquake make it difficult to locate precisely the 1950 ice margin shown on aerial photography. Cut alders on the ice-marginal bedrock ridge (Table 6.2) imply a minimum deglaciation date of 1961 (Fig. 6.1, Table 6.2) .

Lawrence Glacier

Lawrence Glacier is the central ice tongue along the southern wall of Blackstone Bay (Fig. 6.1). Two lichen-covered moraines are found in the glacier foreland: the innermost is discontinuous but found on both sides of the terminus, while the outer moraine is found along the western side only. Deglaciation of the outermost ridge dates to ca. 1877, based on the 59-mm-diameter lichens found there (Fig. 6.1).

The inner ridge was deglaciated by either 1898 or 1913. The western segment is covered by lichens up to 40 mm in diameter, and matches the eastern segment covered with lichens up to 39 mm in diameter, and a bedrock ridge along the water's edge covered with lichens up to 38 mm in diameter. Based on these data, this moraine would date to ca. 1913. However, contradictory evidence is present on a bedrock ridge which lies inside the western portion of this moraine and supports lichens up to 48 mm in diameter (Fig. 6.1). This suggests deglaciation of the ridge by 1898. In any case, the area near the inner moraine was deglaciated between 1898 and 1913.

The northern (distal) sections of the inner moraine are now intertidal and may indicate the location of a tidewater terminus during the LIA maximum. Alternatively, subsidence related to the 1964 earthquake may have submerged the northern moraine segments.

Grant and Higgins (1913) recorded that in 1909 this glacier reached tidewater, with an outwash apron preventing icebergs from calving into Blackstone Bay. In 1910, Tarr and Martin (1914) noted that portions of the Lawrence Glacier terminus were so near the water's edge that ice was cascading down the cliff producing icebergs. They also noticed a small stony push moraine along other portions of the terminus. The AGS photographs (App. B) show Lawrence Glacier terminating in water in 1935, but withdrawn from the shore by 1957.

Tree-ring data (Table 6.3) indicate that the area between the modern high-tide line and the glacier has been deglaciated since at least 1964. In 1966, the American Geographical Society found evidence of small push moraines and a small advance (App. B) along the ice margin (Fig. 6.5).

In 1987, the glacier terminated on top of a steep bedrock cliff that extends to tideline, but no ice was noted cascading down this cliff to produce icebergs. On top of this same cliff, a small push moraine was found adjacent to the western ice margin. Along the eastern part of the terminus, overrun shrubs without bark were flattened, pointed down-ice, and covered by rocks and debris. Twelve alders (8-11 years old) sampled here (Table 6.3) indicate that a small readvance occurred after at least 11 years of recession (Fig.6.5). Small readvances forming push or dump moraines seem to be common features of this glacier (as indicated by the advances recorded in 1966 and 1987).

Ripon Glacier

Ripon Glacier, the easternmost of the three glaciers along the south wall of Blackstone Bay, has a clear trimline, but no end moraines. This is due to retreat across a steep bedrock cliff and erosion by meltwater issuing from the glacier terminus.

In 1909, Grant and Higgins (1913) mapped the terminus of this glacier close to the sea (Fig.6.1). In 1910, Tarr and Martin (1914) indicated that Ripon terminated about 64 m (200 ft) from the sea, ending on bare bedrock ridges and surrounded by an extensive barren zone. They suggested that 20 years of vegetational growth occurred since retreat from the trimline. Field (1937) shows the ice margin 0.6 km (0.4 mi) from the ocean in 1935, and Post (1980) shows the margin 0.9 km (0.55 mi) from the sea at 161 m (500') elevation. (However, consideration must be given to coastline changes after the 1964 earthquake subsidence.)

Ripon Glacier appears to have reached a LIA maximum ca. 1890, with subsequent retreat up a steep bedrock cliff (Fig. 6.6). This glacier exhibits the greatest amount of post-LIA retreat of the three terrestrial glaciers in Blackstone Bay.

Conclusions

The two tidewater glaciers (Blackstone and Beloit) in Blackstone Bay provide a record of Neoglacial advance pre-dating 1350 AD. Assuming only a short residence period on the terminal moraine and using advance rates from modern tidewater glaciers (12-32 m/yr) (Viens, 1995), Blackstone and Beloit Glaciers could have advanced the 7.5 km to the 1350 AD moraine as early as 725 AD or as late as 1116 AD.

The smaller glaciers along the southern side of the bay (Marquette, Lawrence, and Ripon) may also have advanced concurrently, acting as tributaries to the larger ice bodies. Little evidence exists for the location of the ice margins between the time they retreated off the submarine shoals (about 1350 AD) and the historic period. LIA margins of the three small tributary glaciers lie close to tidewater, but the main tidewater glaciers had retreated up-fiord past Lawrence Glacier position when these moraines formed (as evidenced by the moraine geometry).

All glaciers in this system have retreated during the past 100 years. Of the five glaciers previously reaching tidewater, Blackstone and Beloit glaciers now occupy fiord-head positions in shallow water after retreating ca. 1 km since 1909. Marquette and Lawrence glaciers have retreated upslope short distances out of the ocean, while Ripon Glacier has retreated the greatest distance due to its steep slope (Fig. 6.1). Although retreat has been the norm since the LIA maxima, Lawrence Glacier experienced marginal fluctuations, as evidenced by two small-scale readvances in 1966 and 1987 (Fig. 6.5).

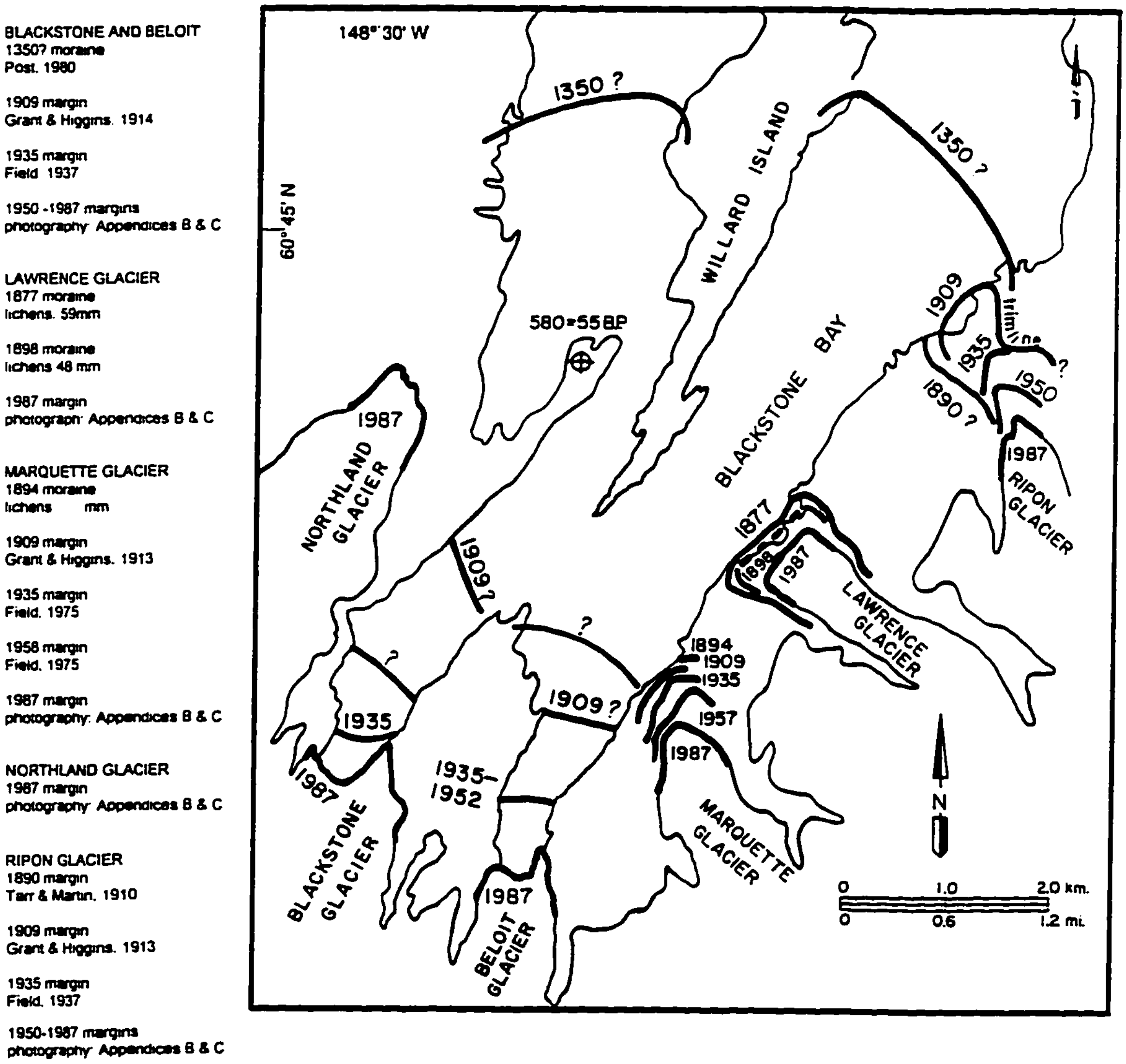

Fig. 6.1 - Map of moraines and ice positions in Blackstone Bay.

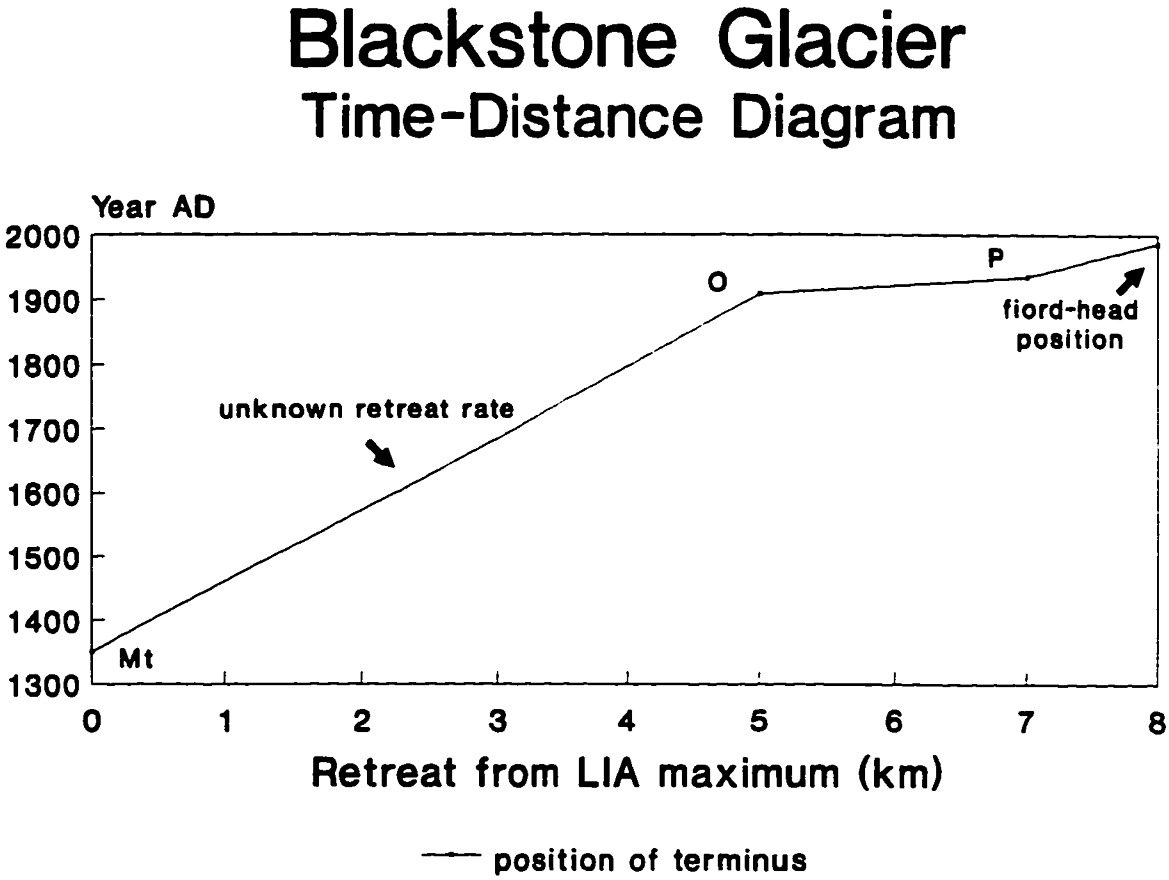

Fig. 6.2 - Time-distance diagram showing LIA retreat of Blackstone Glacier based on evidence from moraines dated by tree-rings (Mt), observations (O), and photographs (P). The glacier currently occupies a stable fiord-head position.

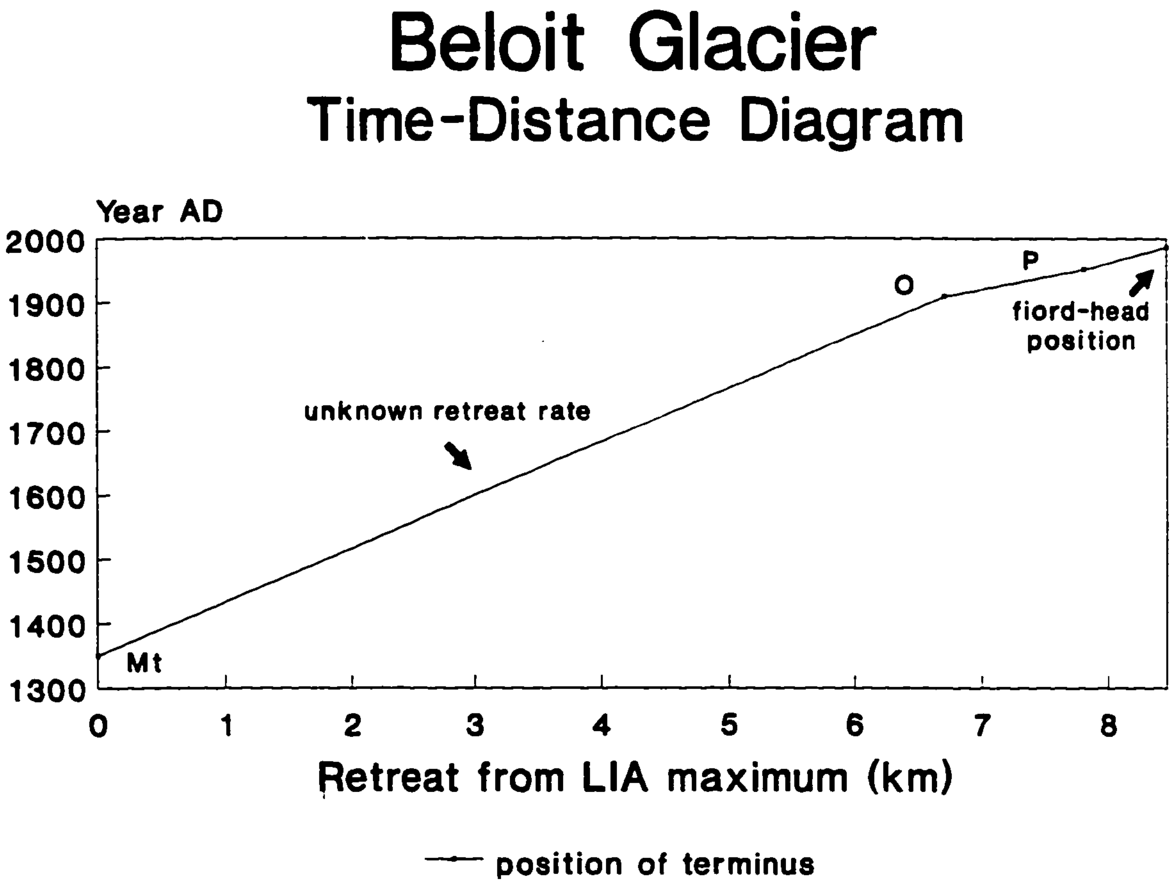

Fig. 6.3 - Time-distance diagram showing LIA retreat of Beloit Glacier based on evidence from moraines dated by tree-rings (Mt), observations (O), and photographs (P). The glacier currently occupies a stable fiord-head position.

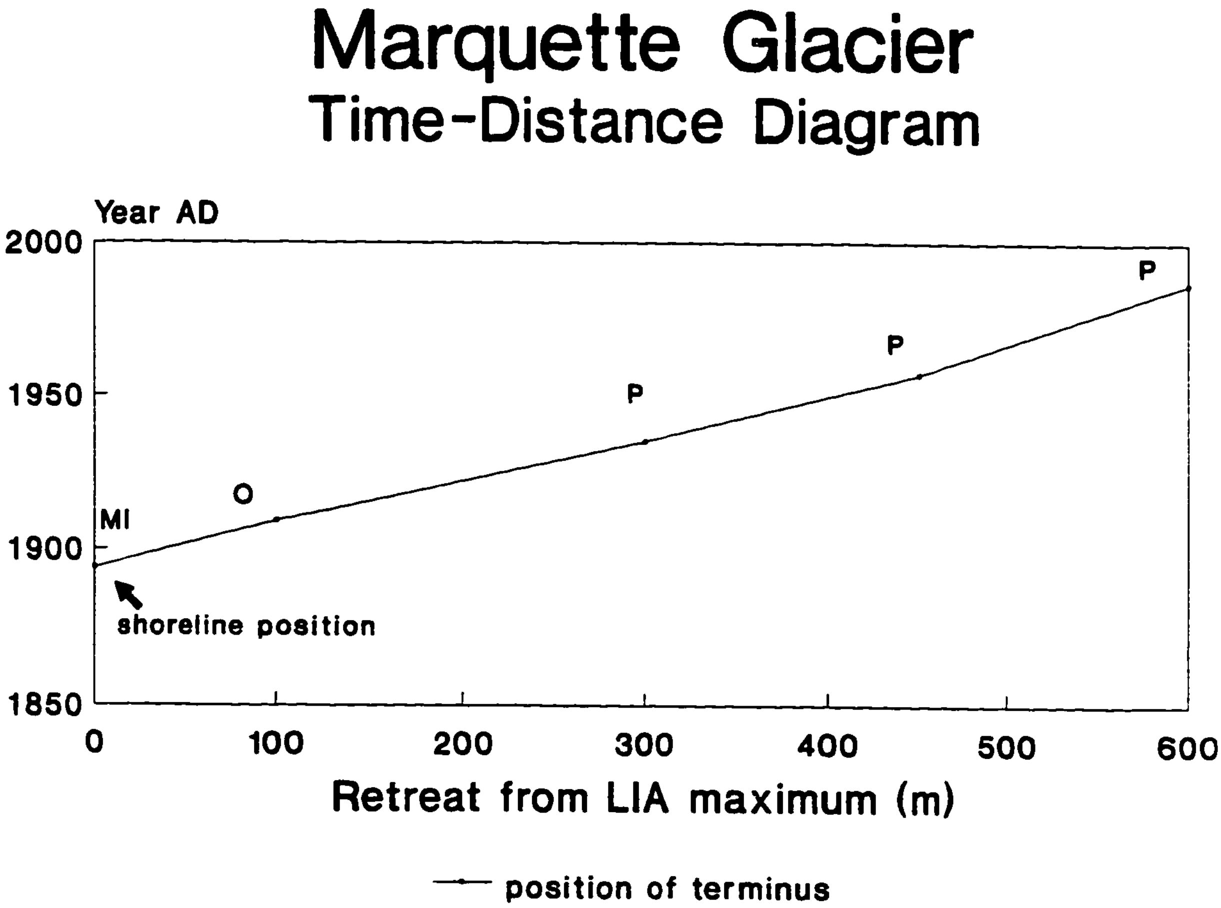

Fig. 6.4 - Time-distance diagram showing LIA retreat of Marquette Glacier based on evidence from moraines dated by lichens (Ml), observations (O), and photographs (P).

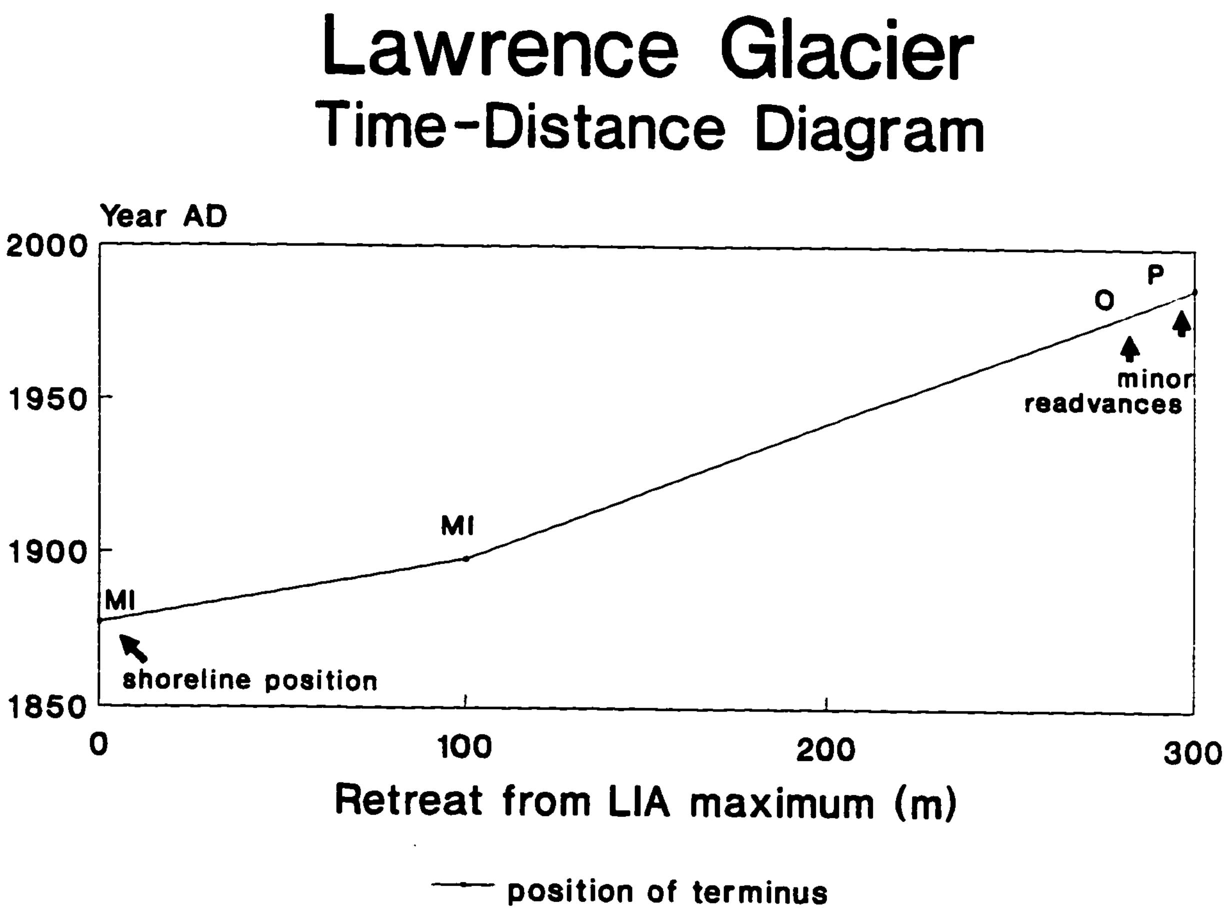

Fig. 6.5 - Time-distance diagram showing LIA retreat of Lawrence Glacier based on evidence from moraines dated by lichens (Ml), observations (O), and photographs (P).

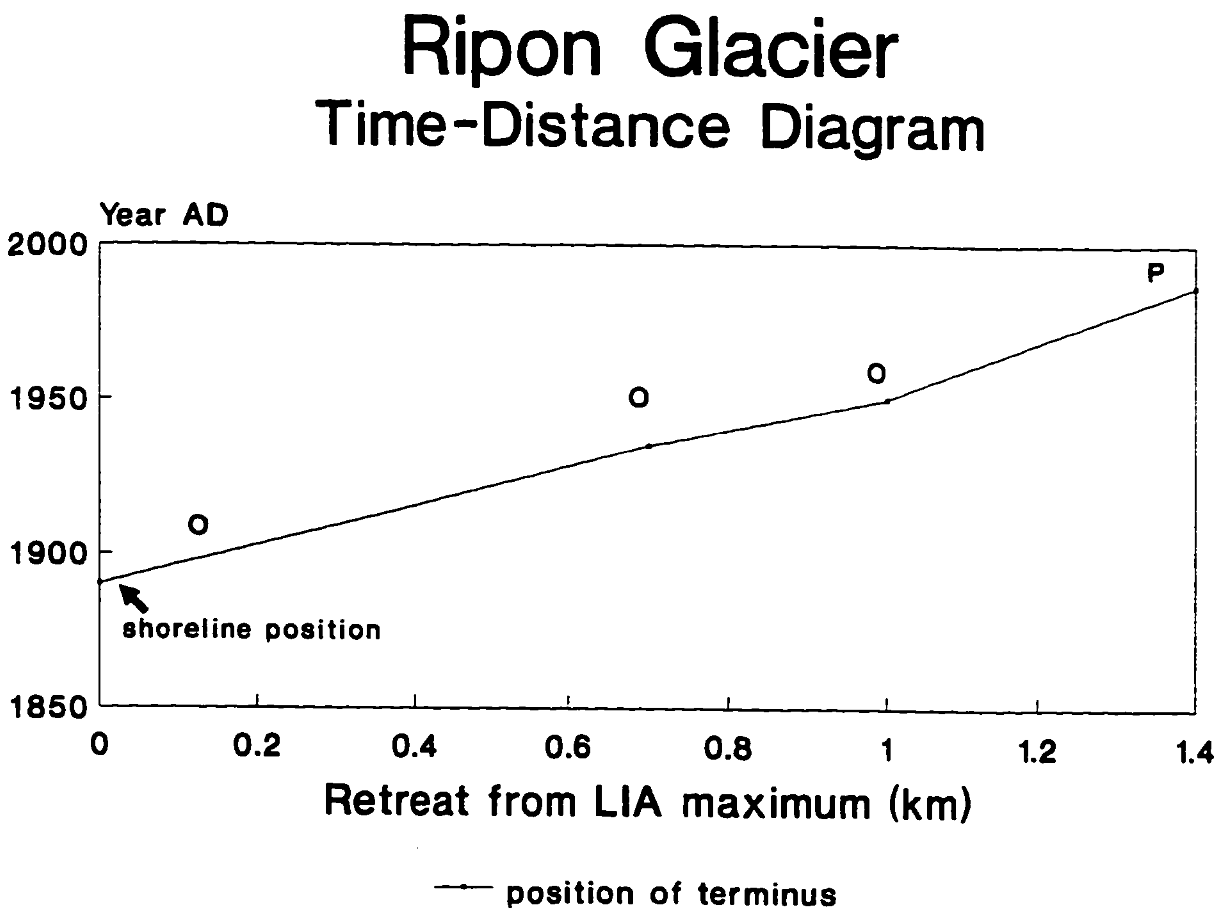

Fig. 6.6 - Time-distance diagram showing LIA retreat of Ripon Glacier based on evidence from observations (O) and photographs (P).

TABLE 6.1 - WILLARD ISLAND, BLACKSTONE BAY

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Willard Island - W peninsula								
133	cored	*Picea*	1984	276	none	100	25+30	<1653
Willard Island - Central - trimline & moraine								
419x	cored	*Picea*	1987	242	good	U	25	1720
Willard Island/Badger Point - C. Heusser's cores								
CH-3	cored	U	U	265	none	U		<1664

U - unknown
* Correction factors
  25 - years for spruce colonization
   0 - years for alder and hemlock colonization
   n - years for correction of height cored (spruce only)
      (see Table 2.1 for correction calculation)

Data condensed from Appendix C.

TABLE 6.2 - MARQUETTE GLACIER, BLACKSTONE BAY

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Marquette Glacier - older, higher area - Sample Bag 1								
e	cut	*Alnus*	1987	16				1961
Marquette Glacier - beach - driftwood								
2-b	cut	*Alnus*	1987	22				1965
Marquette Glacier - cleared bedrock ridge adjacent to ice margin - Sample Bag 1								
1-e	cut	*Alnus*	1987	16				1971
Marquette Glacier - beach area - outwash? -1964 earthquake subsidence - Sample Bag 2 - rooted stumps assumed killed in 1964								
2-b	cut	*Alnus*	1987	22				1942

* Correction factors
25 - years for spruce colonization
0 - years for alder and hemlock colonization
n - years for correction of height cored (spruce only)
(see Table 2.1 for correction calculation)

Data condensed from Appendix C.

TABLE 6.3 - LAWRENCE GLACIER - BLACKSTONE BAY

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Lawrence Glacier - ridge C (waterfall) - Sample Bag A								
h	cut	*Tsuga*		23				1964
Lawrence Glacier - ridge A (adjacent to ice) - Sample Bag B								
T-10	cut	*Alnus*	1987	20				1967
Lawrence Glacier - overrun trees - ridge A (adjacent to ice) - Sample Bag C								
C-h	cut	*Alnus*	1987	11				1976

* Correction factors
  25 - years for spruce colonization
  0 - years for alder and hemlock colonization
  n - years for correction of height cored (spruce only)
  (see Table 2.1 for correction calculation)

Data condensed from Appendix C.

CHAPTER 7 - TEBENKOV BAY AND GLACIER

Introduction

Tebenkov Bay, off northeastern Blackstone Bay, is named for M.K. Teben'kov, former governor of Russian Alaska. The bay is 1 km long, 1.6 km wide, and 100 m deep. South of the bay, the valley contains a glacier that terminates 2.2 km from the coast. Two outwash plains lie on either side of a bedrock strike ridge that divides the valley in half. The meltwater channels cross a moraine 1.4 km beyond the present ice margin and 0.9 km from the coast. The channels pass into a wide muddy, gravelly intertidal area that grades into a marsh lined with standing dead trees (likely related to 1964 earthquake subsidence) at the inland limit. Between the moraine and the shoreline, a small bedrock knoll (labelled C on Fig. 7.1) rises 20-30 m above the marsh on the eastern valley edge. A small island lies 0.6 km offshore along the eastern edge of the bay (Fig. 7.1).

Deglaciation of Tebenkov Bay

The Holocene deglaciation of Blackstone Bay is inferred from evidence found on both the offshore island and the bedrock knob. On the island, a mature spruce and hemlock forest with mossy ground cover drapes the strike ridges, while small peat bogs and lakes occupy the valleys. This landscape indicates considerable time has passed since deglaciation. A basal peat sample from the bottom of a 1 m bog section exposed above the northern beach yielded a radiocarbon date of 4,180 ± 30 yr BP (App. A). This indicates deglaciation of the area by 4,563-4,860 years ago (calibrated ages from Pearson and others, 1986) and may suggest a time when tidewater glacier retreat and subsequent isostatic rebound raised the island above high tide and allowed peat formation.

The bedrock knob lying between the coast and the moraine is also covered with trees up to 90 cm (35") in diameter. All cored trees were older than 200 years, with the oldest sample containing 296 rings in a core taken 1 m above the base (Table 7.1). These data indicate deglaciation of the bedrock knob by 1636 AD, but do not rule out deglaciation as early as the offshore island.

Little Ice Age Chronology

An arcuate moraine ridge, exhibiting 2 m of surface relief, stands 5-6 m above the outwash plains. This ridge is steep-sided and covered with striated graywacke boulders. Aerial photography shows a double ridge with depressions and kettles along the axis. Spruce and hemlock trees forest the ridge, and moss covers both rocks and trees. Alders and salmonberries complete the ground cover.

The cut and cored trees from this moraine ridge date 44-

72 years old (Table 7.1), and indicate ice retreat from this ridge by 1884 AD (Fig. 7.1). These data agree closely with both Viereck's (1967) dating of trees dating as old as 1875-85 AD on the outwash plain adjacent to the LIA moraine, and with D. Maraldo's work (personal communication, 1992) dating the moraine to 1883 AD. These data, in conjunction with the dates from the offshore island and the bedrock knob, indicate that the LIA advance produced the maximum expansion of Tebenkov ice during at least the last 4,500 years.

Post-LIA Retreat

Recession of Tebenkov Glacier during the past 100 years is documented by trimlines, small moraines, and historical records. Evidence for the 1910 position of the glacier is found on a large strike ridge that rises 30-50 m above the eastern outwash channel. Striated bedrock is commonly visible and ice-transported boulders are numerous. The spruce, hemlock, and alder growing here indicate deglaciation by 1910 AD, based on a 47-year-old spruce (Fig. 7.1, Table 7.1). These data agree with the 1909 photographs of Grant and Higgins (1913) as well as the 1910 mapping of Tarr and Martin (1914) who measured 250-350 m of recession from the LIA moraine. In 1935, Field (1937) observed an additional 300 m of recession.

Locations of the terminus from 1950 to 1987 (Fig. 7.1) are documented by aerial photography (App. B) and field reconnaissance. A diamict-covered hill, ice-covered in the 1950 photographs but adjacent to the 1975 terminus, is covered with alders that date deglaciation to 1965 (Table 7.1). This agrees with the 1964 American Geographic Society (Field, 1975) survey that located the terminus 1100 m from the LIA moraine.

Evidence of recession (based on dendrochronology and aerial photography) suggests an average retreat rate of 15 m/yr from 1884-1950, 16 m/yr from 1950-1975, and 17 m/yr from 1975-1987 (Fig. 7.2). This indicates a very uniform rate of retreat following the LIA for this large valley glacier with a terrestrial terminus.

The 1909 maps (Grant and Higgins, 1913) show that approximately 6 km upvalley of the terminus, small tongues of Tebenkov Glacier flowed over the valley sides: one eastward toward Cochrane Bay and one westward toward Blackstone Bay. Neither is shown on the 1951 U.S. Geological Survey maps (Seward C-3 and C-4 quadrangles), based on 1950 aerial photography, or on more recent photography, suggesting these small distributary lobes retreated between 1909 and 1951.

The 1950-75 photography allows strict chronological control on the colonization of different plant species. On the 1975 photographs, the 1950 trimline is a well-defined, poorly vegetated area. In the field, the zone inside this trimline is devoid of lichens, but rocks are moss-covered.

This implies that early moss colonization prevents successful lichen colonization on the same substrate.

Colonization of different tree species is documented here as well (Table 7.1). On a diamict-covered hill, ice-covered in 1950 and adjacent to the ice in 1975, alders dating 16-22 years old in 1987 (Table 7.1) imply deglaciation by at least 1965 and document the rapid colonization of this species. Likewise, a series of small push moraines adjacent to the 1975 ice margin, and 150 m in front of the 1987 margin, are devoid of spruce but colonized by alder as long ago as 1973 and hemlock as long ago as 1967. These data show alder and hemlock as the earliest colonizers (with hemlock colonization almost simutaneous with deglaciation and preceding alder by up to 7 years in this field area), with spruce colonization lagging both these species by 22 years or more.

The terminus of this glacier now lies 2.2 km from the shoreline and has a debris-covered, lobate front. Radial crevasses and shear planes are evident, as well as ice-cored moraines. A bedrock strike ridge protrudes 30-50 m above the outwash plain and exerts topographic control on both the ice and meltwater flow. Two large subglacial and supraglacial meltwater streams flow from the terminus to the ocean, being separated by this ridge. The 1950 photography shows a high-volume discharge in the western channel, whereas portions of the eastern channel appear dry. The 1974 photography shows equal flow in both channels. During the summer of 1987, only the western channel carried water, and field investigations showed a large subglacial stream emanating from the eastern edge of the ice. This 30-m-wide stream ran westward along and under the eastern terminus until reaching a large north-south trending strike ridge in the valley center, where it turned north along the base of the ridge and flowed across the central portion of the valley, following the western channel to the bay.

Neoglacial Fluctuations

Tebenkov Glacier is the one glacier where evidence for pre-LIA Neoglacial fluctuations can be found. Trees incorporated in a diamict at two locations adjacent to the retreating terminus give evidence for past advances that overran and killed the trees. In the first location (A on Fig. 7.1), a 30-m-deep canyon along the eastern margin was exposed in 1987. This channel was ice-covered in the 1950 photo (App. B), and filled with meltwater in the 1975 photo (App. B). Large water-smoothed boulders fill the channel bottom where 10-12 upright, rooted, and supine trees are located. Most trees have broken roots and branches attached, and all lack bark. The largest are up to 15 m long, and the highest ring count is 120 years (field counting methods). One tree has a radiocarbon age of 1,680 ± 25 years (calibrated to 250-428 AD; App. A).

In the second location (B on Fig. 7.1), two rooted overrun trees were found on the southern edge of the bedrock ridge deglaciated between 1950 and 1975. One has a radiocarbon date of 780±20 yr BP (calibrated to 1216-1275 AD; App. A). These ages agree with samples dated to 1210 AD collected from outwash by Austin Post. The Neoglacial chronology (Fig. 7.3) indicates an ice advance around 250-428 AD, with a one or more retreats separating this from another advance dated 1216-1275 AD. All advances reached less than 0.6 km down valley of the LIA moraine (based on the basal peat date from the offshore island). Prior advances that may have extended further downvalley are not evident in either the dendrochronological or radiocarbon record because any older moraines have been destroyed by subsequent advances, outwash deposition, or earthquake subsidence.

A third advance between 1319 and 1430 AD is based on D. Maraldo's tree ring data (pers. comm., 1992) and may represent an early LIA advance of this glacier. The last LIA advance began before the late 1800's, and ice retreat began by 1884 from this valley's most prominent moraine.

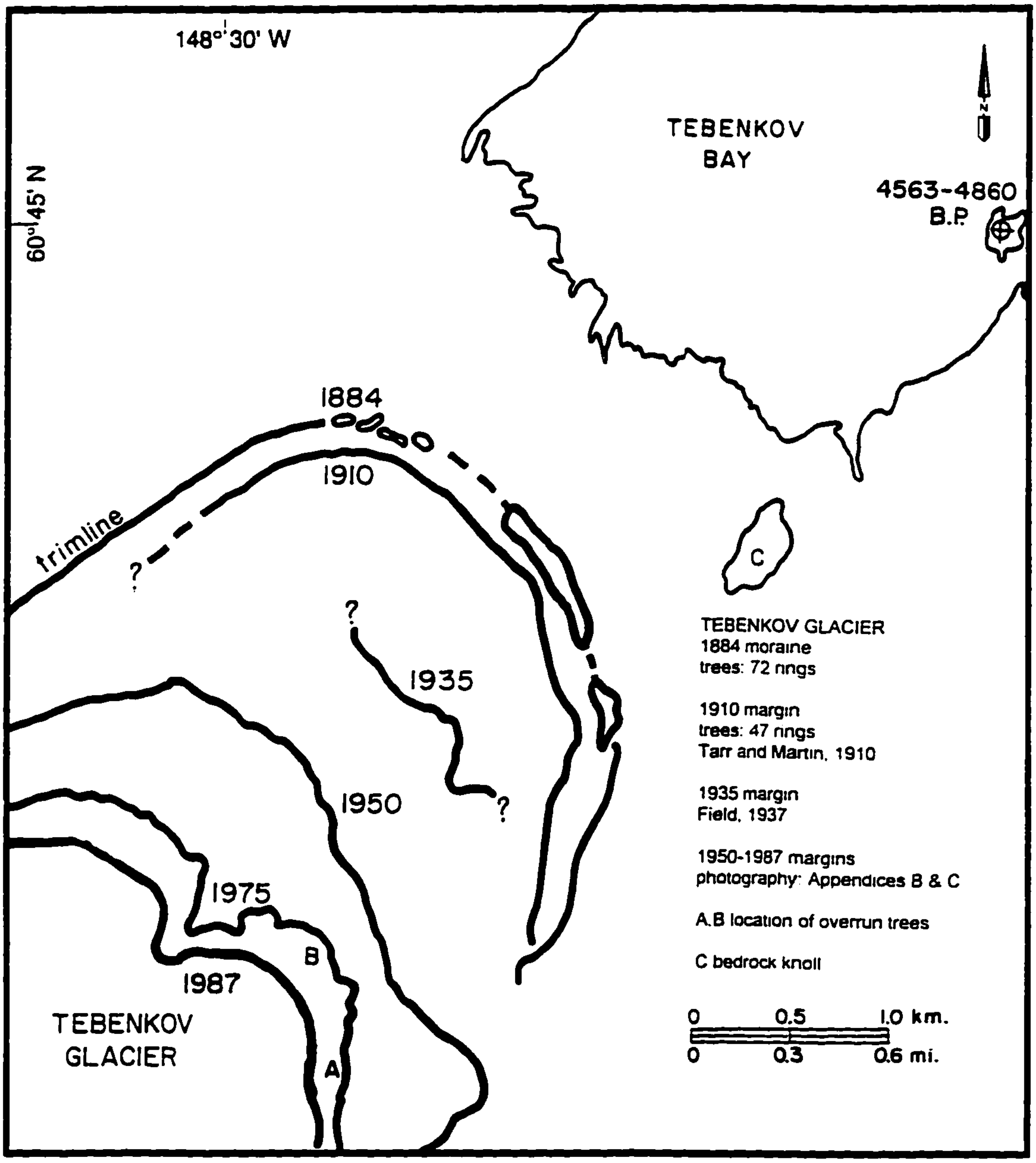

Fig. 7.1 - Map of dated moraines and ice positions of Tebenkov Glacier.

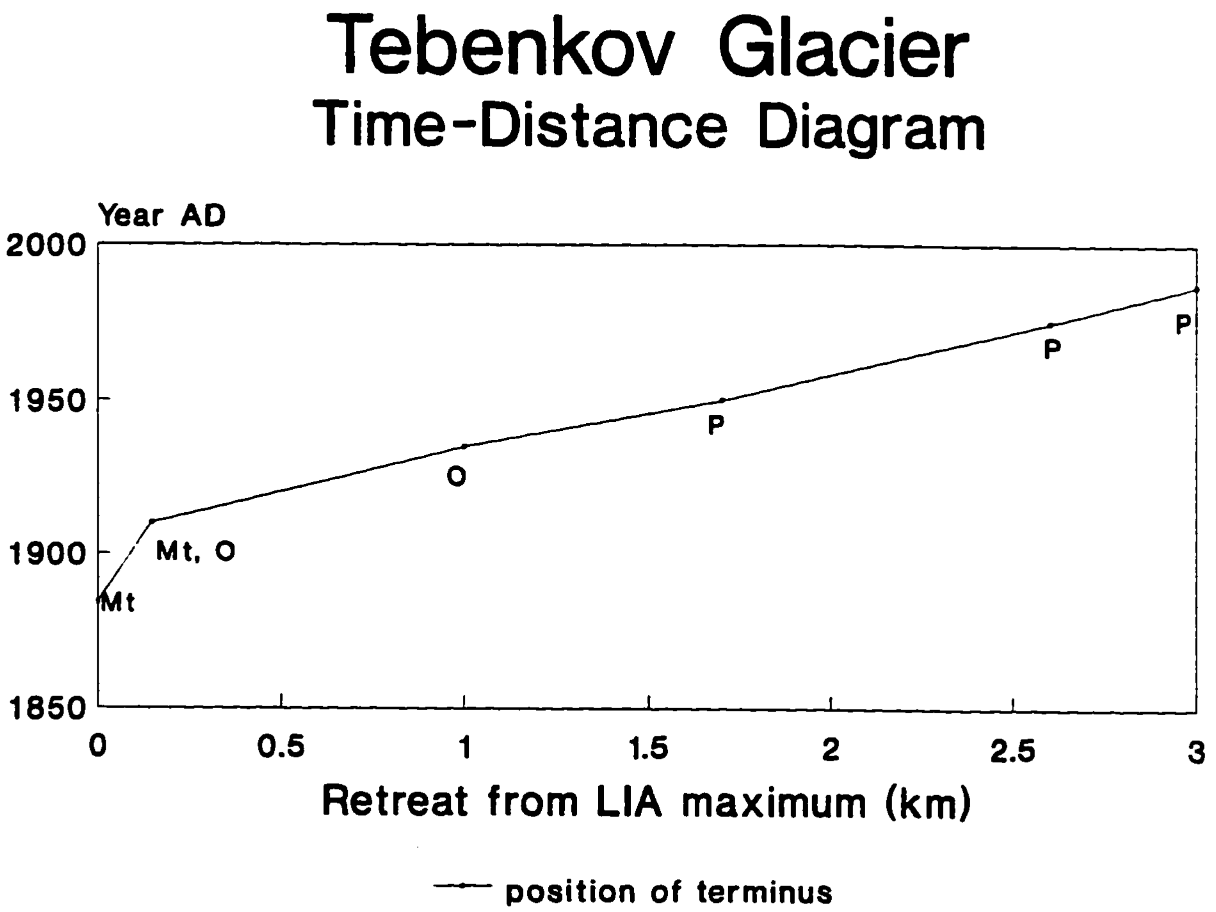

Fig. 7.2 - Time-distance diagram of LIA retreat of Tebenkov Glacier based on evidence from moraines dated by tree-rings (Mt), observations (O), and photographs (P).

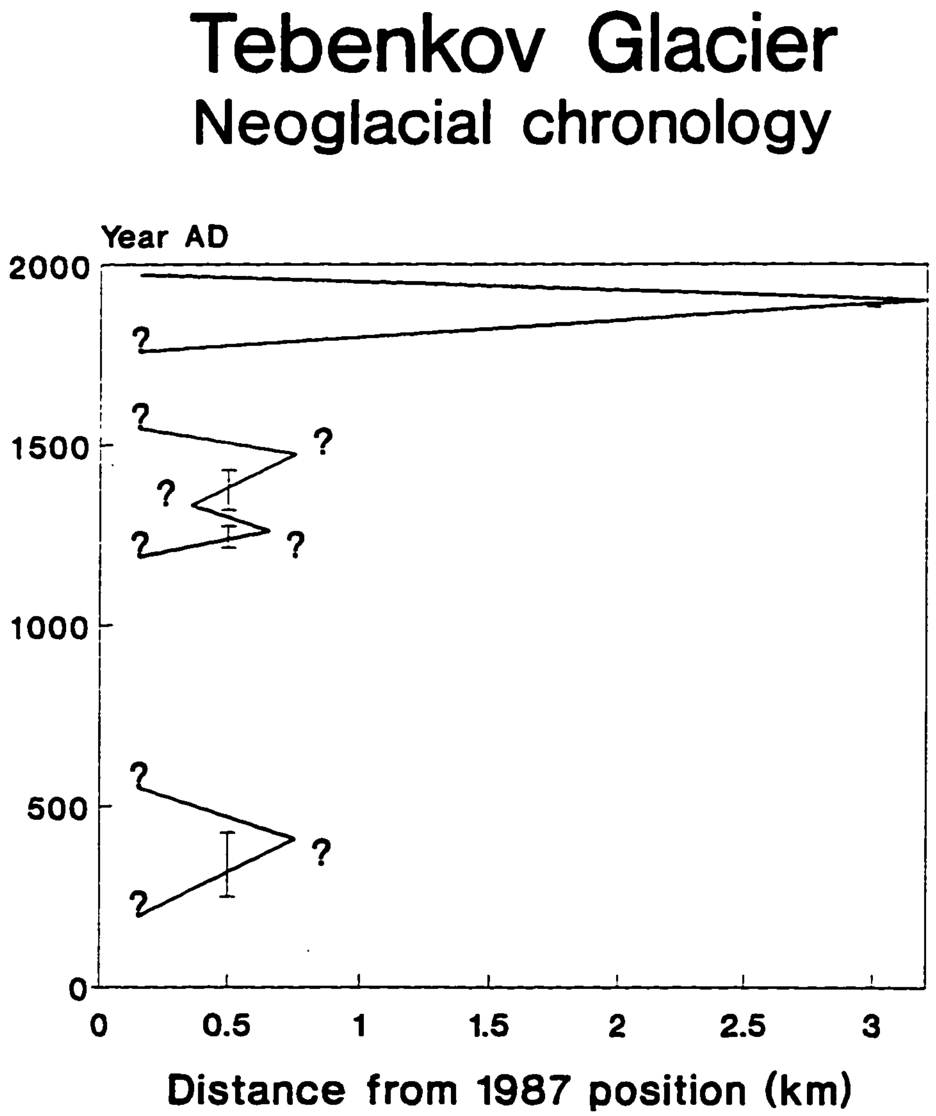

Fig. 7.3 - Neoglacial chronology of Tebenkov Glacier. The first three advances date to 250-428 AD (calibrated date on buried wood), 1216-1275 AD (calibrated date on buried wood), and 1319-1430 AD (tree ring data, D. Maraldo, pers. comm., 1992). Dating is poor on the late-LIA advance, but retreat from the outer moraine began by 1884 AD.

TABLE 7.1 - TEBENKOV VALLEY

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Tebenkov Glacier - beyond glacial limit - bedrock knob - N side of valle								
400	cored	*Picea*	1987	296	none	100	25+30	1636
Tebenkov Glacier - moraine ridge								
405	cored	*Picea*	1987	72	good	20	25+6	1884
Tebenkov Glacier - bedrock strike ridge behind moraine								
414	cored	*Picea*		47	good	15	25+5	1910
Tebenkov Glacier - moraine/hill between channel and ice - E side								
3 alder samples								
b	cut	*Alnus*	1987	22				1965
Tebenkov push moraines - central portion - pulled and cut samples								
6-d	cut	*Tsuga*	1987	14				1973
A-b	cut	*Alnus*	1987	7				1980

* Correction factors
25 - years for spruce colonization
0 - years for alder and hemlock colonization
n - years for correction of height cored (spruce only)
(see Table 2.1 for correction calculation)

Data condensed from Appendix C.

CHAPTER 8 - HOLOCENE GLACIER FLUCTUATIONS

Pleistocene Deglaciation

Ice covered most of Cook Inlet and Prince William Sound (Fig. 1.1) during the LGM (Reger and others, 1996; Mann and Hamilton, 1995; Karlstrom, 1964). In the Cook Inlet area, ice began to retreat ca. 16,000 yr BP (Reger and others, 1996), the Kenai Peninsula lowlands (Fig. 1.1) were deglaciated by 14,000 yr BP (Ager, 1983; Ager and Sims, 1984), and both Turnagain Arm and Kachemak Bay (Fig. 1.1) were ice-free by 10,500 yr BP (Reger and Phinney, 1996; Wiles, 1992). In Prince William Sound, ice retreated from most of the fiords by ca. 14,000 yr BP (Sirkin and Tuthill, 1987; Reger, 1991). A peak in solar radiation at this latitude ca. 10,000 yr BP (Ritchie and others, 1983; Barnosky and others, 1987; Bartlein and others, 1991) accentuated the late Pleistocene warming trend.

Evidence for postglacial deglaciation of the valleys surrounding the Blackstone-Spencer Ice Complex shows that glaciers retreated from areas beyond the LIA margins prior to at least several thousand years ago. Portage Pass was deglaciated sometime before 5645 yr BP (5724-5596 cal yr BP; App. A). Likewise, Tebenkov Bay was deglaciated before 4725 yr BP (4615-4837 cal yr BP; App. A), and the Bartlett Glacier foreland was deglaciated before 2135 yr BP

(2288-2066 cal yr BP; App. A). These are only minimum local dates for deglaciation; many areas likely were deglaciated by the early Holocene. Karlstrom (1964) reported that Hope Valley, to the west, was deglaciated by at least 6800 yr BP, and Reger and others (1996) show that Turnagain Pass was deglaciated by 9800 yr BP. Collectively, the data suggest that the glaciers of the Blackstone-Spencer ice complex likely retreated from their Pleistocene maxima to approximately their present locations by early Holocene time.

Hypsithermal Interval

Pollen data suggest that a period warmer and drier than today occurred during the early Hypithermal interval between ca. 9000 and 7500 yr BP at Icy Cape (Peteet, 1986), whereas in Prince William Sound, the warm interval lasted from ca. 9000 to 6000 yr BP; a precipitation minimum occurred ca. 8000 yr BP (Heusser, 1985). Glaciers in Glacier Bay (Goodwin, 1988) and those in the southern Kenai Mountains (Wiles and Calkin, 1990) were in retracted positions from ca. 10,000 to 6,000 yr BP (Fig. 8.1).

Neoglacial advances

Tidewater glaciers in southeastern Alaska provide the earliest evidence of Neoglacial advances in the Gulf of Alaska region. Hubbard Glacier (near Yakutat) advanced over wood ca. 6000-5000 yr BP (Wiles, 1992), Fairweather Glacier

(Lituya Bay) advanced ca. 6300-4900 yr BP (Mann and Ugolini, 1985; Mann, 1986b), and glaciers in the West Arm of Glacier Bay advanced ca. 5000 yr BP (Goldthwait, 1966). However, at that time, the fiord under Bering Glacier was ice free and burrowing clams lived in a muddy environment there (Molnia, 1996). In the northern Kenai Mountains, peat deposition in Portage Pass shows that area to have been ice-free by ca. 5650 yr BP.

Wetter and cooler climate, as inferred from the pollen record, was established in Prince William Sound by ca. 3300 yr BP (Heusser, 1985; Heusser and others, 1985). Tidewater glaciers known to have advanced during this time include glaciers in McCarty Fjord (southern Kenai Mountains) ca. 3500 yr BP (Wiles, 1992), glaciers of Harriman and College fiords (Prince William Sound) ca. 3500-2500 yr BP (Heusser, 1983), Lituya Bay glaciers (southeastern Alaska) ca. 3200-2600 yr BP (Mann and Ugolini, 1985; Mann, 1986b), and glaciers in Muir Inlet of Glacier Bay ca. 2900-2600 yr BP (Goodwin, 1984). Mason and Jordan (1993) interpret beach features implying increased coastal erosion in northwestern Alaska as indicating times of increased summer storminess between 3300 and 1700 yr BP, and they apparently correlate with growth of cirque glaciers and dune mobilization in northern Alaska (Galloway and Carter, 1993). In the Canadian Cordillera, Neoglacial ice advances began ca. 5000 yr BP, with maximum expansions dating ca. 3300 yr BP (Luckman and others, 1993; Osborn and Luckman, 1988; Denton and Karlén, 1977; Ryder, 1989; Ryder and Thompson, 1986; Clague and Matthews, 1992).

A third period of Neoglacial glacier advance occurred near the beginning of the Christian era. Mendenhall Glacier (Juneau Ice Field) sheared stumps during an advance ca. 200 BC-0 AD (Miller, 1977), the advancing Bering Glacier buried stumps in outwash ca. 200 AD (Molnia, 1996), and both Dinglestadt and Grewingk glaciers (southern Kenai Mountains) advanced ca. 100 AD (Wiles, 1992). There was also a period of glacier expansion in the Canadian Cordillera ca. 100 AD (Luckman and others, 1993; Osborn and Luckman, 1988; Denton and Karlén, 1977; Ryder, 1989; Ryder and Thompson, 1986; Clague and Matthews, 1992).

Several glaciers experienced advances ca. 350-400 AD. Moraines were built in Icy Bay by ca. 400 AD (Porter, 1989). Two terrestrial glaciers in Prince William Sound, Tebenkov (Chapter 7) and Sheridan (Tuthill and others, 1968), over-rode forests ca. 350 AD. The tidewater Northwestern Glacier (southern Kenai Mountains) advanced ca. 410 AD (Wiles and Calkin, 1993).

The period from 550 to 650 AD is another interval when both tidewater and terrestrial glaciers expanded. These include Bering Glacier ca. 550 AD (Wiles, 1992) and five other glaciers in the Kenai Mountains: Dinglestadt and

Grewingk glaciers ca. 620 AD (Wiles, 1992), Northwestern Glacier ca. 650 AD (Wiles, 1992), McCarty Glacier ca. 550 AD (Wiles, 1992), and Bartlett Glacier (Chapter 5) ca. 600 AD.

Medieval Period

The Gulf of Alaska tree-ring record shows a warmer period from ca. 650-1100 AD (Wiles and others, 1996; Barclay and others, 1996). At this time, forests formed around both the Juneau Ice Field (Miller, 1977) and McCarty Fjord (Wiles, 1992) and were subsequently overrun by ice. Only two tidewater glaciers are known to have advanced during this period: McCarty ca. 900 AD (Wiles, 1992) and Bering ca. 1030 AD (Wiles, 1992).

Following this mild interval, the period 1100-1200 AD was cooler than average in western Prince William Sound, as indicated by a 1000-year-long tree-ring chronology derived from temperature-sensitive hemlock wood incorporated in tills and cross-dated with living trees (Barclay and others, 1996). Numerous glaciers, both tidewater and terrestrial, expanded at that time. In southeastern Alaska, tidewater glaciers advanced in Glacier Bay ca. 1150 AD (Goodwin, 1988), in Lituya Bay by 1150 AD (Mann and Ugolini, 1985), and in Icy Bay by 1275 AD (Porter, 1989). Three tidewater glaciers in the Kenai Mountains also advanced; Ailak Glacier overran trees by ca. 1125 AD (Wiles, 1992), and calculations for Blackstone Bay (Chapter 7) imply an advance of Blackstone and Beloit glaciers ca. 1120 AD. The positions of Lawrence, Marquette, and Ripon glaciers, along the southern side of Blackstone Bay, suggest that they were tributaries to the expanded tidewater glaciers. Two terrestrial glaciers in the northern Kenai Mountains also advanced into forests: Billings Glacier by ca. 1210 AD (Barclay and others, 1996) and Tebenkov Glacier by ca. 1240 AD (Chapter 7).

Little Ice Age

Early LIA Advances

The LIA in southern Alaska was marked by a cold interval from 1380-1680 AD, according to tree ring records from the Prince William Sound region (Barclay and others, 1996; Wiles and others, 1996). Lower summer temperatures over much of northern North America between 1570 and 1730 AD (Bradley and Jones, 1993) may have helped promote glacial growth at that time.

The LIA began with a well-documented expansion of terrestrial glaciers on both sides of the Kenai Mountains. In the southern Kenai Mountains advances were occurring: ca. 1440 AD for Grewingk Glacier, ca. 1460 AD for Tustumena Glacier, and ca. 1580 AD for Exit Glacier (Wiles, 1992). In the northern Kenai Mountains, Tebenkov Glacier was advancing ca. 1375 AD (Chapter 7), Ultramarine Glacier ca. 1400 AD (Barclay and others, 1996), and Billings and Nellie Juan glaciers ca. 1540 AD (Barclay and others, 1996). At the same time, two tidewater glaciers in the southern Kenai Mountains also expanded: Bear Glacier ca. 1460 AD and Yalik Glacier ca. 1650 AD (Wiles, 1992). Portage Glacier, with a calving lacustrine terminus, advanced sometime between ca. 1380 and 1640 AD (based on calculations in Chapter 3). Davidson Glacier, in the Juneau Icefield, also advanced ca. 1400 AD (Egan, 1971).

However, at least three tidewater glaciers reached their LIA maxima and began retreating during this time. Blackstone and Beloit glaciers in western Prince William Sound retreated from their outer LIA moraines by ca. 1350 AD (Chapter 6), and McCarty Fjord (southern Kenai Mountains) was ice-free during this interval (Wiles, 1992).

Middle LIA Period

The period from 1680 to 1840 AD is interpreted as one of slight warming in northwestern Prince William Sound and the northern Kenai Mountains, according to studies of buried trees in the forelands of Princeton, Billings and Tebenkov glaciers (Barclay and others, 1996). However, in a similar study from the southern Kenai Mountains, Wiles (1992) interpreted 1740-1880 AD as a cold period. In Glacier Bay, ice receded from the outermost LIA position by ca. 1750 AD (Goldthwait, 1966; Field, 1975; Clague and Evans, 1993), and Davidson Glacier and nine other glaciers in the Juneau Ice

Field retreated from their LIA maxima by ca. 1750 AD (Field, 1975). On the Kenai Peninsula, the tidewater Ailak Glacier (southern Kenai Mountains) retreated from its outermost LIA moraine by 1700 AD (Wiles, 1992), and ten terrestrial glaciers in northwestern Prince William Sound began to retreat by ca. 1710 AD (Barclay and others, 1996).

On the other hand, numerous glaciers that advanced during the early LIA did not retreat at this time. These include three terrestrial (Grewingk, Tustumena, and Exit) and two tidewater (Bear and Yalik) glaciers in the southern Kenai Mountains (Wiles, 1992).

Dates of advance are unknown for many other glaciers that reached their LIA maxima before or during the 19th century. Barclay and others (1996) state that even the glaciers known to have retreated during the mid-LIA period readvanced to approximately their earlier positions during the late LIA. This may also be true of Tebenkov Glacier, which exhibits a double-crested terminal moraine (Chapter 7).

Late LIA Advances

The late LIA expansion is well recorded in both historical and morainal records. The extended Portage Glacier acted as a portage across the Kenai Mountains from Prince William Sound to Turnagain Arm (Cook Inlet) for natives, explorers, and gold-seekers (Chapter 3). Other expanded glaciers may also have promoted travel across the Kenai Mountains south of Portage Glacier (Crossen, 1993).

Porter (1981b) indicated that ELAs were lowered 100-200 m worldwide during the LIA. Calkin (1988) cites a 230 m snowline depression for eastern Prince William Sound, and Mercer (1961) calculated a 76 m LIA drop of ELA on Tebenkov Glacier. Estimates of ELA changes for additional glaciers around the Blackstone-Spencer Ice Complex were not definitive because the high plateau area (source for Spencer, Bartlett, Blackstone, Beloit, Portage, and Burns glaciers) provides an unusually large accumulation area resulting in anomalously high accumulation area ratios (AARs).

Jacoby and coworkers (Jacoby and others, 1985; Jacoby and D'Arrigo, 1989; D'Arrigo and Jacoby, 1992) report extended cold intervals in northwestern North America during the 19th century, and Bradley and Jones (1993) note especially low summer temperatures from 1815 to 1870 AD for northern North America. In Prince William Sound, ring widths from buried trees indicate a colder period from 1830-1900 AD (Barclay and others, 1996).

The late LIA left its mark in major moraine-building events in the Gulf of Alaska region. With the exception of the two tidewater glaciers, all of the glaciers around the Blackstone-Spencer Ice Complex produced late Little Ice Age moraines. The juxtaposition of these moraines to mature forests and to peat deposits having ages of several thousands years, suggests that many of the LIA advances culminating in the 18th and 19th centuries were the most extensive advances of Neoglaciation. On Tebenkov Glacier, all of the buried wood indicating older advances is located behind the LIA margin. Luckman's work (1977, 1986, 1988) in the Canadian Rockies shows a similar relationship; LIA maxima all cluster within a 150-year time period and appear to represent the greatest glacier expansion during the past 2,600 years.

Post-LIA Retreat

All of the glaciers around the Spencer-Blackstone Ice Complex retreated at the end of the LIA. Tree rings document the warming trends in northwestern North America at the end of the LIA (Jacoby and D'Arrigo, 1989; D'Arrigo and Jacoby, 1992), especially an increase in summer degree days since 1840 AD (Jacoby and others, 1985). In Prince William Sound, tree-ring widths suggest a warmer climate after 1900 AD (Barclay and others, 1996; Wiles and others, 1996).

All terrestrial glaciers in the Spencer-Blackstone Ice Complex reached maximum positions and began retreating during the 19th century. On the coastal side of the northern Kenai Mountains, Tebenkov Glacier had begun to retreat by 1884 AD. The three tributary glaciers along the southern wall of Blackstone Bay (Marquette, Lawrence, and Ripon) abandoned their shoreline moraines by 1894 AD, 1877 AD, and ca. 1890

AD, respectively (Chapter 6).

On the landward side of the Spencer-Blackstone Ice Complex, all terrestrial glaciers retreated from their outermost moraines during the 19th century. Portage Glacier, retreated along a terrestrial margin from 1799 AD until 1914 AD when the terminus began calving in Portage Lake. Burns Glacier began to retreat by 1809 AD, and Spencer and Bartlett glaciers started receding by 1890 AD and 1898 AD, respectively (Chapters 3-5). Wiles (1992) documented the retreat of terrestrial glaciers on the landward side of the southern Kenai Mountains, including Grewingk Glacier by 1858 AD, Tustumena Glacier by 1864 AD, and Exit Glacier by 1825 AD. Barclay and others (1996) reported the retreat of an additional ten glaciers in western Prince William Sound during this time.

Glaciers with terrestrial margins are found in all locations around the Blackstone Ice Complex. Most have retreated 1.8-2.5 km from their outermost LIA limits (Fig. 8.4). This implies that glaciers dominated by terrestrial melting regimes respond strongly to climate warming.

Tidewater glaciers known to retreat off their LIA maximum moraines during the late 19th century include Bear and Yalik glaciers, southern Kenai Mountains, by 1888 AD and 1890 AD, respectively (Wiles, 1992); glaciers in Icy Bay by 1880 AD (Porter, 1989), glaciers in Lituya Bay by 1900 AD (Mann and

Ugolini, 1985), and Bering Glacier by 1890 AD (Molnia, 1996). Hubbard Glacier, near Yakutat, had retreated by 1790 AD (Molnia, 1986).

Although the major recession of the tidewater glaciers in the Blackstone-Spencer Ice Complex predates the late LIA, the total retreat since ca. 1350 AD for Blackstone and Beloit glaciers measures 7.2 and 7.5 km, respectively (Fig. 8.4). The three small tributary glaciers in Blackstone Bay have retreated far less (0.5-1.0 km) from the LIA moraines deposited in the late 1800's (Fig. 8.4), although accurate measurements are precluded because of the partial submergence of these moraines (Chapter 6). The two main tidewater tongues have receded ca. 1.0 km since 1909 AD (Post, 1909).

Portage Glacier is the only glacier in the Blackstone-Spencer Ice Complex with a calving margin ending in a deep lake. This glacier has receded 5 km from its LIA limit (Fig. 8.4), it had a terrestrial margin from its maximum position before 1799 AD until 1914 AD when it retreated into the lake basin. Whereas only 0.5 km of retreat occurred between ca. 1799 AD and 1914 AD, 4.5 km of retreat has occurred from 1914 AD to 1996 AD (Fig. 8.4).

Recent Advances

Although the general trend during the 20th century has been continuing glacier recession, a few tidewater glaciers in southern Alaska are advancing. These include Harriman and Harvard glaciers in the Harriman and College fiords of northwest Prince William Sound (Lethcoe, 1987), Taku Glacier in the Juneau area (Motyka and others, 1992), Hubbard Glacier near Yakutat (Motyka and others, 1992), and Brady, Johns Hopkins, and Grand Pacific glaciers in Glacier Bay (Bengston, 1962). Most of these glaciers have been in an advancing mode for ca. 100 years, but those in Glacier Bay began advancing ca. 1960 AD (Bengston, 1962).

A minor readvance between 1974 and 1990 AD is well documented for Shakespeare Glacier, a small terrestrial glacier in Portage Pass, and correlates with increased temperature and precipitation during this period (Chapter 4).

Causes for Glacier Fluctuations

Several hypotheses have been proposed to explain Neoglacial, and especially LIA, glacier fluctuations. Porter (1986) has interpreted regional northern hemisphere glacier fluctuations as being generally synchronous and attributed this to decadal-scale climate changes, particularly cooling, which typically coincide with increased snow accumulation (e.g. Orombelli and Porter, 1982). Bickerton and Matthews (1993) also see a high degree of correlation among glacier fluctuations in Scandinavia. Around the Spencer-Blackstone Ice Complex, terrestrial glaciers are most likely to show a strong response to temperature, probably because it is a major control on ablation.

A related hypothesis, proposed by Wiles and Calkin (1994), suggests that glacier response is regulated not only by temperature and precipitation, but also by geographic location. Wiles' (1992) dendroclimatological work, based on cross-dating of living and buried trees, suggests that glacier advances on the landward flanks of the southern Kenai Mountains correlate with cool summer temperatures, whereas advances on the maritime side of those same mountains correlate with increased winter precipitation (Fig. 8.2). Recent climate records from southcentral Alaska clearly show that increased temperature correlates with increased precipitation, resulting in positive glacier mass balance during the interval 1980-1990 (Mayo and March, 1990). Fitzharris and others (1992) suggest that positive mass balance of New Zealand glaciers is correlated with increased cloudiness, decreased ablation, and increased precipitation.

Wiles' hypothesis could also explain variations of the maritime and tidewater glaciers on the seaward side of the Blackstone-Spencer Ice Complex. If warm temperatures correlate with increased precipitation, then warm intervals may have produced positive mass balance and increased the AAR (accumulation area ratio) to ≥ 0.8 (as argued by Motyka and others, 1992). This could allow tidewater and tributary glaciers in Blackstone Bay, as well as the adjacent Tebenkov

Glacier, to advance.

Tidewater glaciers present a special case (Meier and Post, 1987; Mann, 1986a; Viens, 1995). Tidewater glaciers fluctuate cyclically in a fashion that is dependent on water depth at the terminus and fiord geometry, with climatic forcing dominating only at sensitive times in the cycle.

Evidence from Icy Bay (Porter, 1989) and Unakwik Inlet (Viens and Post, 1995) indicates that tidewater glaciers retreat at least three times more rapidly than they advance. Meier and Post (1987) suggest that glaciers ending in deep water and regulated by a calving regime can advance only when a terminal shoal in the form of a moraine, delta, or subaqueous fan protects the ice front from extensive calving (Fig. 8.5). Thus, the advance-retreat history of the glacier may be partially controlled by the sediment supply to the terminus (Powell, 1980; Alley, 1991; Trabant and others, 1991). Although, it is not possible to estimate the sedimentation rate in Blackstone Bay without repeated soundings and suspended sediment sampling, both Blackstone and Beloit glaciers appear to be devoid of supra- and englacial debris.

Tidewater glacier theory also supports a direct correlation between water depth at the terminus and rate of retreat due to calving (Brown and others, 1982; Meier and Post, 1987). Water depths in Blackstone Bay and Portage Lake reach 400 and 200 m, respectively, and calving is the dominant form of ablation. Retreat rates decrease only when water depth decreases over subaqueous shoals, at the heads of fiords, or at the end of lake basins (Fig.8.5). Tidewater glaciers may be sensitive to changes in climate that raise the ELA and initiate the retreat from a stable extended position (Viens, 1995).

Additional sources of variability lie in the differences in length, slope, and orientation among glaciers. Karlén and Matthews (1992) interpret their nonsynchronous Scandinavian data as reflecting individual variation among terrestrial glaciers.

Discussion

Any application of the previous hypotheses to Gulf of Alaska glacier chronologies must bear in mind that early Neoglacial events are less-well documented than LIA events. The larger data base of buried trees (used for dendro-climatology) and the surface expression of LIA moraines permits a better understanding of the younger events.

Hypothesis for general synchrony

The hypothesis that regional glacier fluctuations are broadly synchronous is well-supported by the Gulf of Alaska glacier chronologies. Early Neoglacial advances occurred ca. 0-200, ca. 350-400, and 550-650 AD. Although evidence of these early advances is based on only 4-6 glaciers in each case, it shows that at least a few tidewater and terrestrial glaciers advanced more-or-less synchronously.

A milder period from ca. 650-1100 AD partly overlaps the warm Medieval Maximum and is evidenced by forest expansion onto glacier forelands around the Gulf of Alaska. Only two glaciers recorded advances during this period.

A late Medieval cold period (ca. 1100-1200 AD) is documented in the tree-ring record and correlates with glacier expansion around the Gulf of Alaska. Five tidewater glaciers, including two in Blackstone Bay, and two terrestrial glaciers advanced during this interval.

A middle LIA cold period during the 15th and 16th centuries correlates with the advance of three tidewater and eight terrestrial glaciers, including Portage Glacier. Two tidewater and eleven terrestrial glaciers retreated during the mild 18th century, although many did not.

Six tidewater and nineteen terrestrial glaciers built moraines during the late LIA cold interval 1840-1900 AD. These moraines commonly mark the maximum extent of glaciers in the region, and indicate that the LIA was a period of unusual cold and extensive glacier growth. Although two glaciers around the Spencer-Blackstone began retreat ca. 1799-1809 AD, their termini remained close to the outermost moraine for the first 100 years of retreat. All glaciers around this ice complex retreated as climate warmed at the end of the 19th century, including the two tidewater glaciers in Blackstone Bay that previously retreated from their Neoglacial maxima. In this region, six tidewater and nineteen terrestrial glaciers show documented retreat during the past century.

This analysis strongly supports the relationship between glacier fluctuations and climate changes. Gulf of Alaska glaciers show broadly synchronous behavior that correlates with LIA temperature fluctuations.

Hypothesis for asynchrony between maritime and inland regions

Wiles' (1992) hypothesis that maritime glaciers advance during warmer and wetter conditions, while colder conditions allow inland glaciers to advance, is based on his interpretation of tree rings in the southern Kenai Mountains. The minor readvance of Shakespeare Glacier during a wet, warm period from 1970-1990 AD supports this hypothesis (Chapter 4).

However, the same periods that Wiles considered warm have been recently reinterpreted as cold in the Prince William Sound region by Barclay and others (1996). The more recent dendroclimatological interpretation for the northern Kenai Mountains allows me to correlate glacier advances with colder intervals. Were it not for these new data, the Spencer-

Blackstone Ice Complex information could be interpreted as supporting Wiles' hypothesis. Whether the climate data really show a difference between the northern and southern Kenai Mountains needs to be resolved by dendroclimatologists working in both areas.

Hypothesis for individual glacier variability

The hypothesis that glaciers respond individually to climate change based on distinctive differences in length, slope, orientation, or other factors (Karlén and Matthews, 1992) is not supported by the Kenai Mountains' data. Figures 8.6 and 8.7 compare the pattern of glacier variations in the southern Kenai Mountains (Wiles, 1992) with those in the northern Kenai Mountains. The onset of individual glacier retreat shows no obvious relation either to the altitude of the glacier basins or the orientation of the ice tongues. Most glaciers in the Kenai Mountains, regardless of size, orientation, terminal altitude or characteristics, began retreating during the 19th century.

Tidewater glacier hypothesis

The hypothesis that tidewater glaciers cycle independently of other glaciers in the region, is only partly supported by this analysis.

Some tidewater glaciers clearly fluctuate independently of the regional trend in glacier behavior and the documented climate signals. Examples include the advances of McCarty and Bering Glaciers ca. 900-1000 AD when most glaciers were in retracted positions and when the tree-ring record points to an interval of warm climate. A second example concerns Blackstone, Beloit, and McCarty tidewater glaciers which retreated during the cold early LIA interval when eleven other glaciers were advancing. Lastly, the advance of seven tidewater glaciers around the Gulf of Alaska during the warm 20th century argues for a dynamic rather than a climatic explanation for their apparently anomalous behavior with respect to the regional trend.

On the other hand, the data show that in many cases, Neoglacial tidewater glaciers were responding in synchrony with terrestrial glaciers. The best supporting data are from the end of the LIA when warming temperatures and rising ELAs caused the retreat of 19 terrestrial and 6 tidewater glaciers around the Gulf of Alaska. The rise in ELA could affect extended tidewater glaciers by decreasing the AAR, causing destabilization and recession of the calving terminus, and initiating nonreversible calving retreat (Viens, 1995). The strongest evidence, noted by Viens (1995), shows that 80-90% of Alaskan tidewater glaciers have retreated during the past 100-200 years. According to Viens, if the tidewater glacier cycle was entirely independent of climate, only 10% of Alaskan glaciers should be in retreat at any one time. The retreat of numerous Alaskan tidewater glaciers during a time of post-LIA warming suggests that climate may play an important role in the tidewater glacier cycle.

Conclusions

The Blackstone-Spencer Ice Complex is unique in having a wide array of glaciers within a small area. The glaciers there display three types of margins, including terrestrial, tidewater, and calving lacustrine termini. In addition, steep temperature and precipitation gradients exist across the ice complex from the maritime environment of Prince William Sound to the colder, drier interior. Dating of glacier fluctuations can help determine if variations are based on individual glacier characteristics or more regional climate patterns.

Following Pleistocene deglaciation and the mild Holocene Hypsithermal interval, Neoglacial advances around the Gulf of Alaska occurred ca. 6000-5000 yr BP, ca. 3200-2600 yr BP, ca. 0-200 AD, ca. 350-400 AD, and ca. 550-650 AD. A warm interval corresponding to the Medieval period coincided with forest expansion, but colder temperatures and glacier expansion returned to the region ca. 1100-1200 AD, when Tebenkov Glacier and the two tidewater glaciers in Blackstone Bay began to advance.

The earliest LIA cold climate in Prince William Sound is recorded in tree rings, and dates to a time (1380-1680 AD) when numerous glaciers, including Tebenkov, advanced. However, the tidewater glaciers in Blackstone Bay retreated by 1350 AD, apparently asynchronously with respect to the regional climate signal. The subsequent mild mid-LIA interval (1680-1840 AD) coincided with retreat of some, but not all, of Prince William Sound glaciers.

The late LIA is the most recent period of glacier expansion in Prince William Sound, with tree-rings indicating cold temperatures between 1830 and 1900 AD. With the exception of the two tidewater glaciers, all the glaciers around the Blackstone-Spencer Ice Complex produced late-LIA moraines. In many cases, these moraines mark the greatest expansion of ice during Neoglaciation. In valleys surrounding this ice cap, mature forest stands and bog peats adjacent to LIA margins yield dates as old as 5,600 yrs BP.

Post-LIA retreat has involved all the glaciers of the Spencer-Blackstone Ice Complex, including those with terrestrial, tidewater, and lacustrine margins. In the Kenai Mountains, all glaciers, no matter their size, altitude, or orientation, retreated at the end of the LIA. Terrestrial glaciers surrounding the Blackstone-Spencer Ice Complex retreated from moraines built between 1799 and 1898 AD. The climate signal, especially temperature, appears to be the strongest control on glacier behavior during the past millennium. The tidewater glacier record suggests that climate may exert a first-order control on behavior, but water depth, calving regime, and terminal shoals become important once retreat begins.

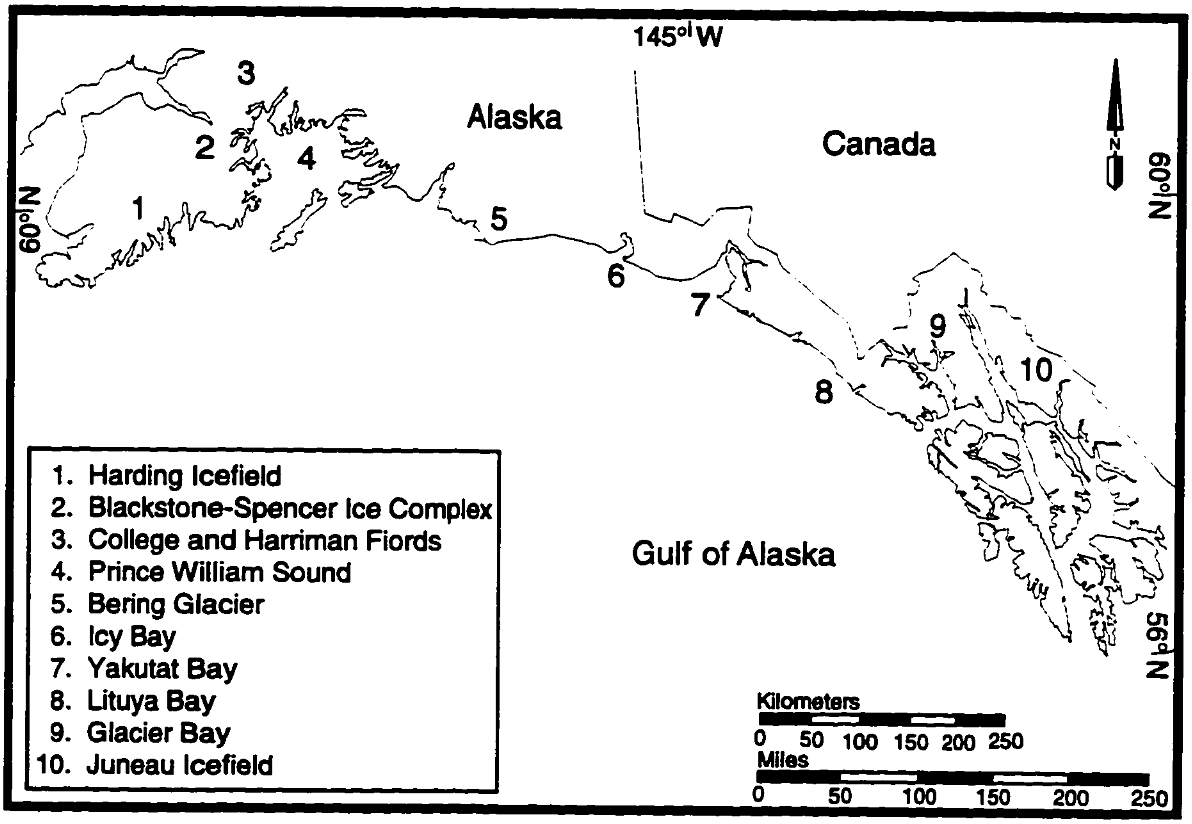

Fig. 8.1 - Index map showing location of study area and adjacent glaciated regions along the Gulf of Alaska.

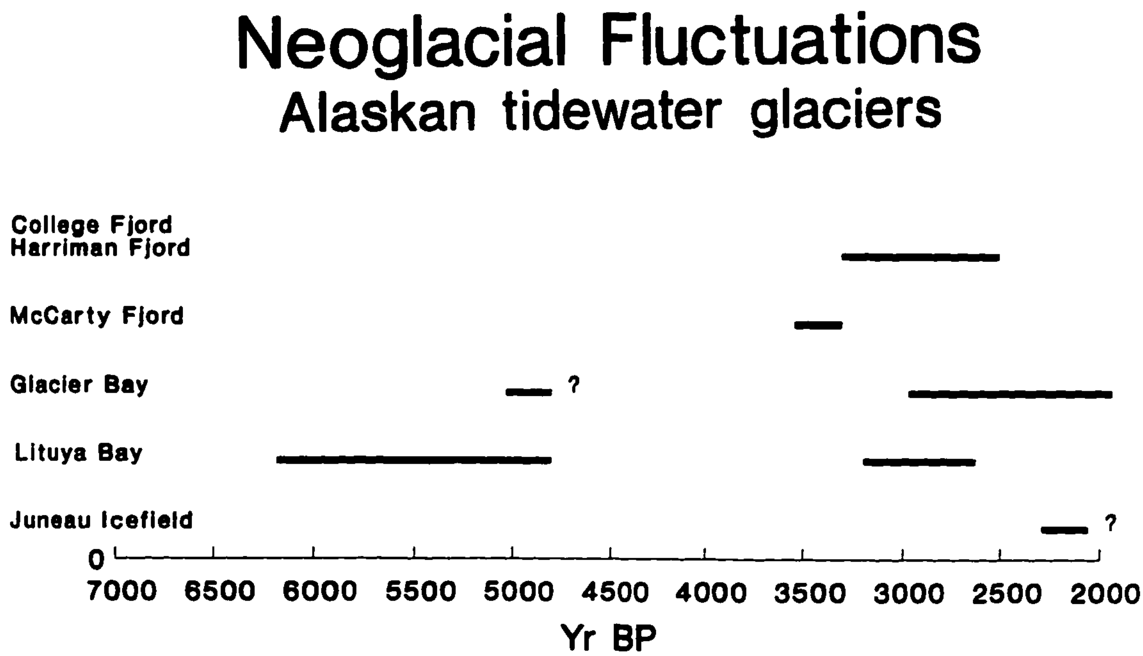

Fig. 8.2 - Neoglacial fluctuations of Alaskan tidewater glaciers, 7000-2000 BP. Sources cited in text.

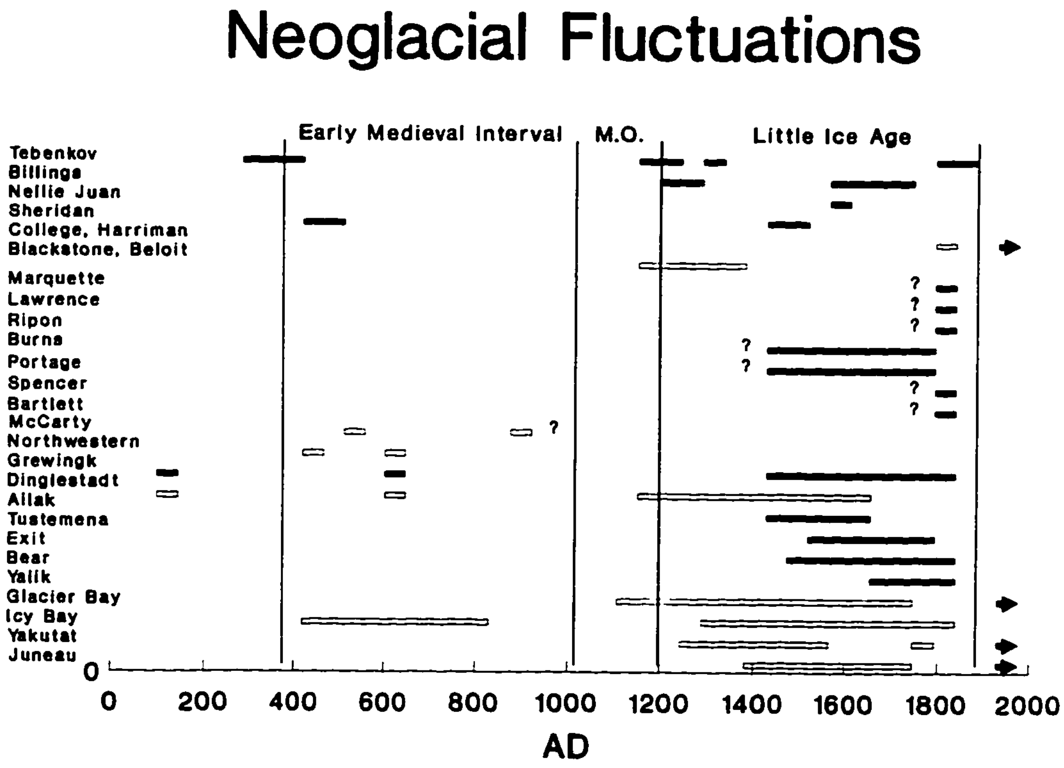

Fig. 8.3 - Neoglacial fluctuations of southern Alaska glaciers, 0-2000 AD. M.O. refers to Medieval Optimum. Question marks indicate lack of exact dates. Open bars represent tidewater glaciers. Arrows represent currently advancing tidewater glaciers. Sources cited in text.

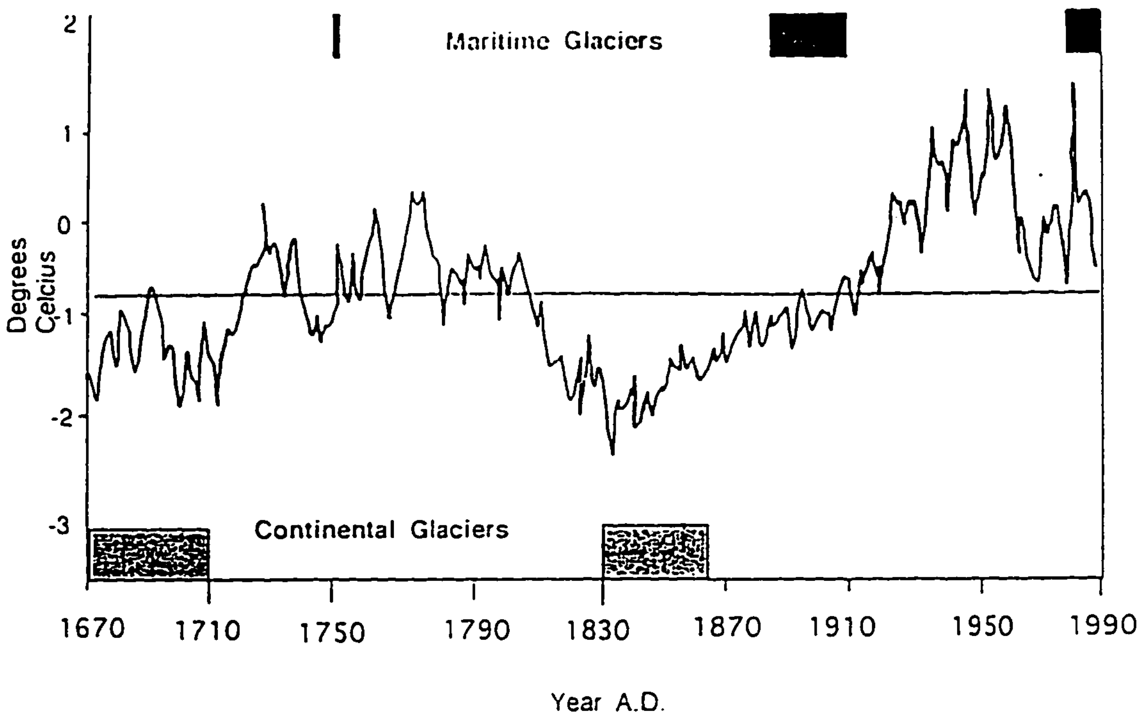

Fig. 8.4 - Wiles' (1992) model for the southern Kenai Mountains, comparing temperature fluctuations to advances of maritime and continental glaciers. Temperature values derived from tree-ring proxy data of Jacoby and D'Arrigo. Source: Wiles (1992).

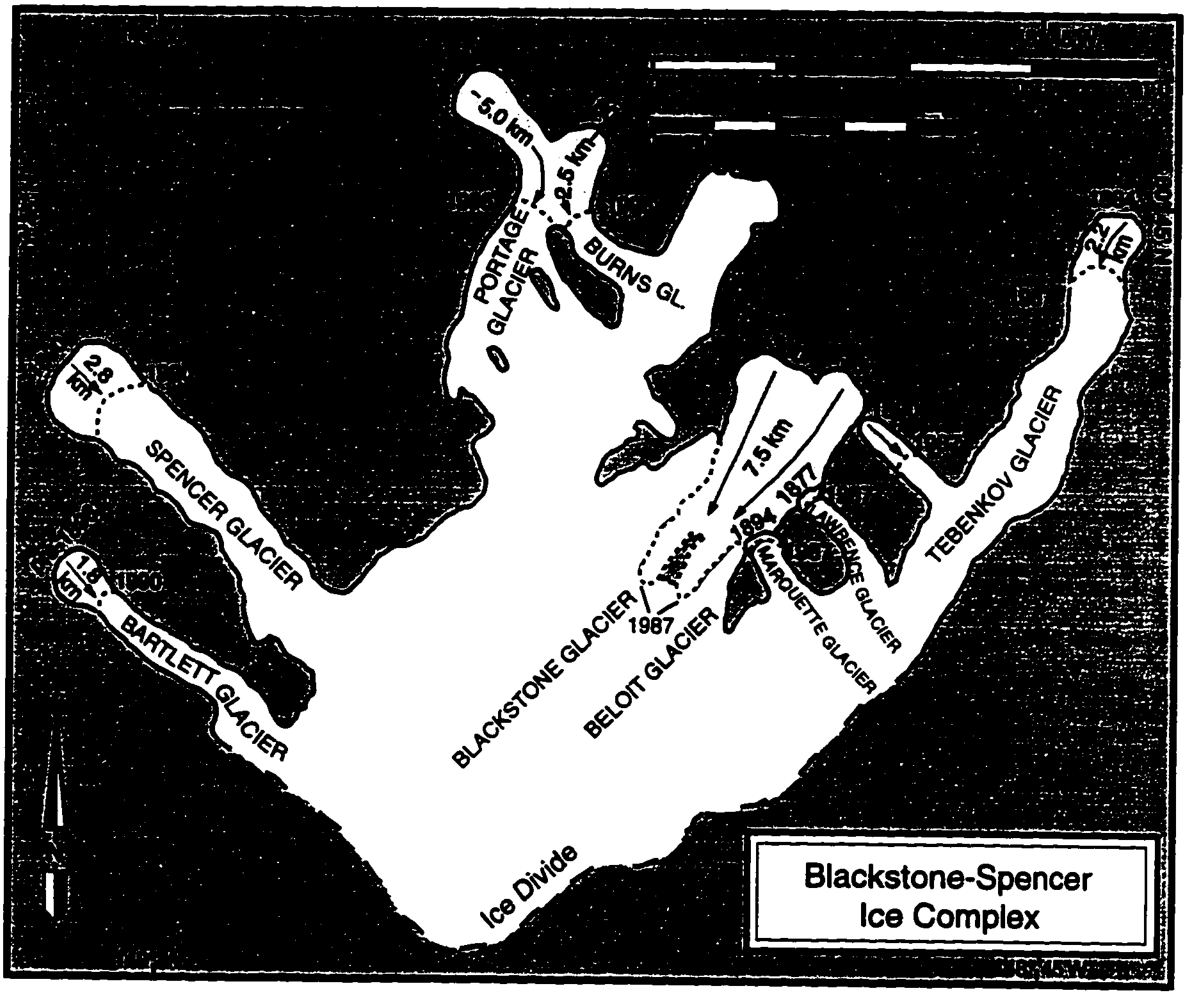

Fig. 8.5 - Distance of retreat (km) from outermost moraines for glaciers of the Blackstone-Spencer Ice Complex.

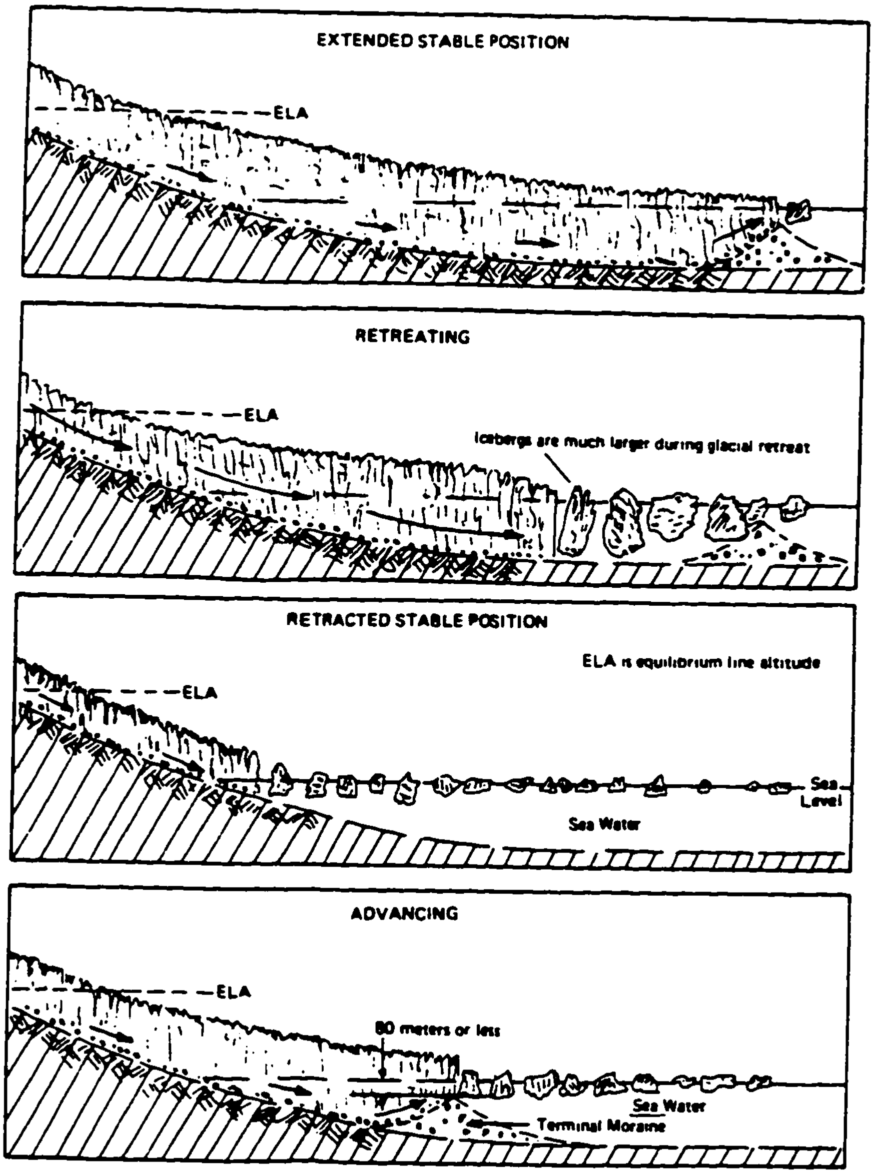

Fig. 8.6 - Tidewater glacier cycle (from Motyka and others, 1992).

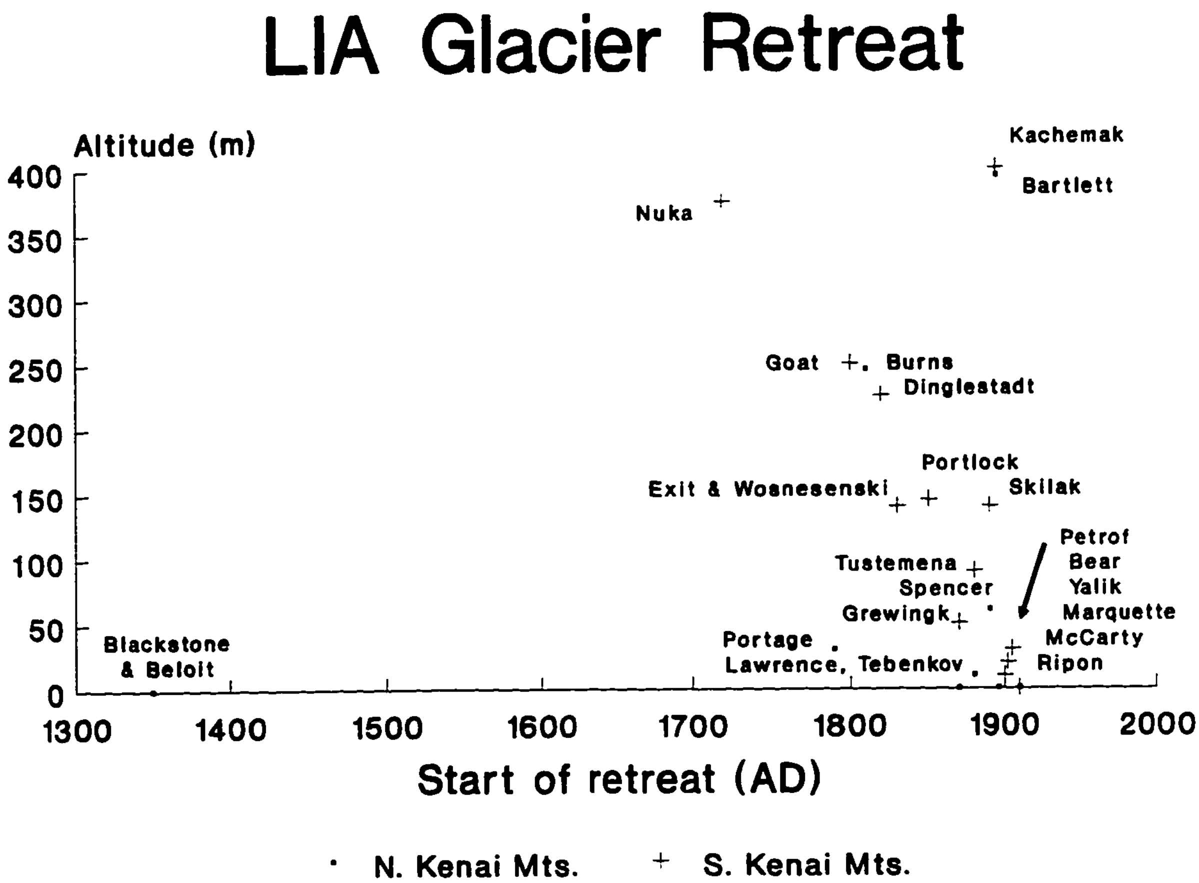

Fig. 8.7 - Comparison of retreat dates and termini elevation of glaciers in the Kenai Mountains, Alaska. Sources: Wiles (1992) and this work.

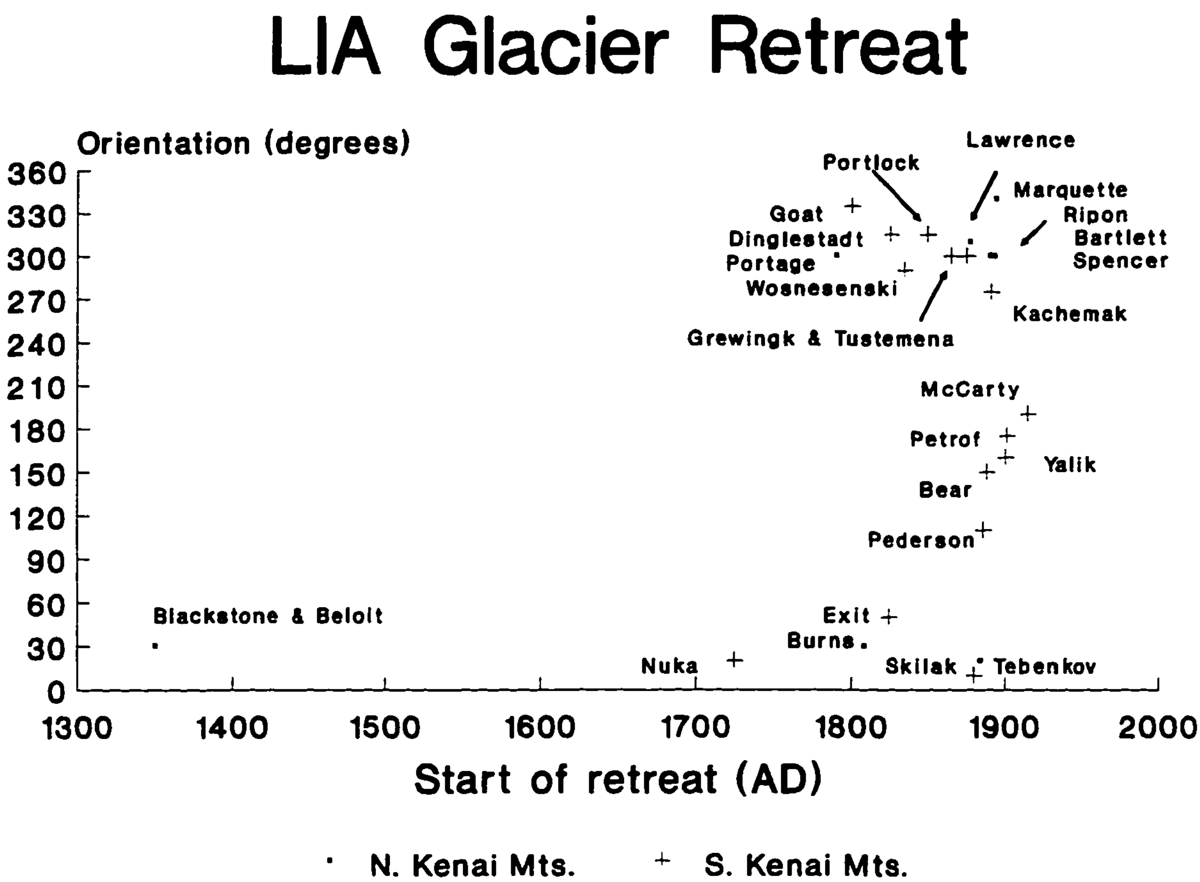

Fig. 8.8 - Comparison of retreat dates and orientation of glaciers in the Kenai Mountains, Alaska. Sources: Wiles (1992) and this work.

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APPENDIX A

RADIOCARBON DATES

ALL FIELD LOCATIONS

APPENDIX A - RADIOCARBON DATES

Sample	Location	Radiocarbon Date (BP)	Calibrated Date (BP)* (age ranges)	Importance	Reference
QL-4345	Bartlett Glacier	2140±25	2135 (2288-2066)	deglaciation of valley	this work
W-318	Bartlett Glacier	1400±220	1296 (1523-1067)	Tunnel I advance	Karlstrom, 1964
W-78	Bartlett Glacier	2370±100	2351 (2703-2647) (2483-2322)	Tustemena III advance	"
UW-537	Blackstone Bay - Badger Pt.	580±55	550 (642-592) (569-531)	deglaciation of bay	Heusser, 1983
QL-4346	Blackstone Bay - Willard Is.	290±25	311 (430-369) (326-300)	deglaciation of trimline	this work
QL-4347	Blackstone Bay - Willard Is.	330±25	431 363 327 (468-312)	deglaciation of trimline	" "
QL-4348	Portage Pass	4900±25	5646 (5724-5697) (5656-5596)	deglaciation of pass (N)	" "
QL-4349	Portage Pass	2770±25	2864 (2934-2905) (2892-2846)	deglaciation of pass (N)	" "
L-163G	Spencer Glacier	"post 1650 AD"		Tunnel II advance	Karlstrom, 1964

QL-4342	Tebenkov Bay	4180±30	4824 4751 4727 4664 4655 (4837-4812) (4762-4741) (4737-4689) (4676-4645) (4627-4615)	deglaciation of bay	this work
QL-4343	Tebenkov Glacier	780±20	691 (721-683)	advance of glacier	" "
QL-4344	Tebenkov Glacier	1680±25	1569 (1685-1670) (1617-1539)	advance of glacier	" "

* Calibration: Pearson and others, 1986
Pearson and Stuiver, 1986
Stuiver and Reimer, 1986
Stuiver and Pearson, 1986
Stuiver and Reimer, 1993

APPENDIX B

AERIAL AND GROUND PHOTOGRAPHY

ALL FIELD LOCATIONS

APPENDIX B1 - VERTICAL AERIAL PHOTOGRAPHY

DATE	SOURCE	FLIGHT LINE	PHOTO NO.	FORMAT	COMMENTS
Portage Valley and Portage Pass - Portage, Burns, and Shakespeare glaciers					
8-5-50	USFS	91 RTS M371 915RW	331-333VV	B & W	High Alt.
6-25-51	USAF	VV 55RTS M-509 338 SRS	0429-0432	B & W	High Alt.
8-16-59	Baumann Skyline	1	1-18	B & W	Low Alt.
	(for USFS)	2	1-10		Scale= 1:1,200
8-19-65	BLM	1	1-7	B & W	Low Alt.
		2	1-7		
10-2-67	BLM	U	2,5	B & W	10,000'
8-19-68	BLM?	U	U	B & W	10,000'
6-23-74	USFS	F16CN 0028 02020 574-10	6-78	COLOR	Scale= 1:18,867
10-84	USFS	U	32-34	COLOR	High alt.
8-14-90	USFS	32 16 6100 41 A 290	14-17	COLOR	
		32 15 6100 41 A 290	50-52	COLOR	
Placer Valley - Skookum, Spencer, Bartlett, and Trail glaciers					
6-25-51	USAF	VV 55RTS M-509 338 SRS	433-435	B & W	High Alt.
7-28-74	USFS	F16CN 0023	155-168	COLOR	Scale= 1:18,867
8-14-90	USFS	32 13 6100 41 A 290	102-108	COLOR	
Blackstone Bay - Blackstone, Beloit, Marquette, Lawrence, and Ripon glaciers, and Willard Island					
8-7-50	USAF	371	364 403-404	B & W	High Alt.
6-23-74	USFS	31	89-93	COLOR	Scale= 1:18,867

Tebenkov Glacier and Tebenkov Bay					
8- 7-50	USAF	371	6	B & W	20,000'
8-11-50		383	36		
8-28-75	USFS	25	213-6	COLOR	Scale= 1:18,867

<u>KEY</u> BLM: Bureau of Land Management
B&W: Black and white photography
USAF: U.S. Air Force
USFS: U.S.D.A. Forest Service, Chugach National Forest
USGS: U.S. Geological Survey
U: Unknown

APPENDIX B2 - OBLIQUE AND GROUND PHOTOGRAPHY

LOCATION	DATE	PHOTOGRAPHER	PHOTO NO.	REPOSITORY
Portage Valley				
Bear Valley	1914	R.J. Weir	B 17:15	Anchorage Hist.
	.		B 17:17	Museum
	also	F. Barnes	259,260	USGS-Boulder
	1915?	P.S.Hunt	G4192	Anchorage Hist.
			B62.1A.420	Museum
			BL79.2.3492	
Portage Glacier	1910-5	W. Guerin	none	Anchorage Hist. Museum
Portage Lake	1939	F. Barnes	124	USGS-Boulder
			125	" "
			195	" "
			196	" "
			197	" "
			199	" "
	1957	U.S. Navy	019	World Data Ctr.
			020	Glaciology A
	1972	Air Oblique	72M2-20	Anchorage Hist. Museum
	1984-96	K. Crossen		Univ. Alaska
Portage Pass				
Burns Glacier	1913	B. Johnson	492	USGS-Boulder
			493	" "
			496	" "
			497	" "
	1914	R. Weir		
	(also	F. Barnes)	261	USGS-Boulder
	1939	F. Barnes	122	USGS-Boulder
			123	" "
			126	" "
			127	" "
			128	" "
	1984-94	K. Crossen		Univ. Alaska
Shakespeare Glacier	1913	B. Johnson	490	USGS-Boulder
	1984-95	K. Crossen		Univ. Alaska

Placer Valley				
Spencer Glacier	1910-5	W.Guerin	B62.1A.422 BL79.2.2433	Anchorage Hist. Museum
		E.Andrews	239	World Data Ctr. Glaciology A
	1927	P.Smith	U	USGS-Boulder
	1957	Goodwin	V24-30 V248	World Data Ctr. Glaciology A
		Viereck	B22-27	"
		IGY	SI 258-260	"
		US Navy	021	"
			024	"
			025	"
	1972	Air Oblique	72M4-88	"
	1973	Air Oblique	73M3-43	"
			73M3-196	"
	1986	K. Crossen		Univ. Alaska
Bartlett Glacier	1906	S. Paige	804	USGS-Boulder
	1911	Martin	488	" "
			559	" "
	1910-5	W. Guerin	BL65.18.746 BL79.2.2707	Anchorage Hist. Museum
			BL79.2.2451	"
			BL79.2.2452	"
			BL79.2.2453	"
			BL79.2.2456	"
			BL79.2.4816	"
	1935	AGS	F-35-934	World Data Ctr.
	1957	L. Viereck	B 31-34	"
		IGY	264-271	"
		Goodwin	V38-41	"
	1987	K. Crossen		Univ. Alaska
Trail Glacier	1911	Martin	489	USGS-Boulder
	1973	Air Oblique	73M3-194	Anchorage Hist. Museum
Blackstone Bay	1935	AGS	F35-650-665 F35-678-681 F35-699-702 F35-736-741	World Data Ctr. Glaciology A
	1937	Washburn	104	"
	1957	Millett	SG95-114	"
		Goodwin	V255-263	
		Morrison	H102-108	
	1972	U	72M2-21	Anchorage Hist. Museum
	1984-87	K. Crossen		Univ. Alaska

Tebenkof Glacier	1935	W. Field	F-35-751-753 F-35-762-764	World Data Ctr Glaciology A
	1937	B. Washburn	105	World Data Ctr Glaciology A
	1957	Millett	SG 78-92	"
			V 245-252	"
			H 89,97	"
		Viereck	LV-B-211	"
	1980	Air Oblique	80M4-104	Anchorage Hist. Museum
	1987	K. Crossen		Univ. Alaska

KEY BLM: Bureau of Land Management
B&W: Black and white photography
USAF: U.S. Air Force
USFS: U.S.D.A. Forest Service, Chugach National Forest
USGS: U.S. Geological Survey
U: Unknown

APPENDIX C

CUT AND CORED TREE SAMPLES

ALL FIELD LOCATIONS

<u>KEY</u>

U - unknown
* Correction factors:
  25 - years for spruce colonization
  0 - years for alder and hemlock colonization
  n - years for correction of height cored (spruce only)
    (see Table 2.1 for correction calculation)

APPENDIX C

CUT AND CORED TREE SAMPLES

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Portage Valley - oldest moraine - west of VC moraine								
23	cored	*Picea*	1984	69	none	10		
25				60	good	branch		
26				45	wide	U		
27				130	none	100	25+30	<1799
61				52	good	20		
62				54	wide	150		
63				64	none	20		
95				51	wide	15		
100				90	good	20		
101				117	wide	30	25+9	<1833
102				72	none	30		
103				91	wide	30		
106				104	good	20		
135				51	good	10		
136				75	wide	10		
137				36	wide	15		
139				90	good	5		
140				115	good	20	25+6	1838
143				72	wide	20		
145				68	wide	45		
148				64	none	20		
149				65	wide	10		
152				90	none	10		
163				80	good	15		
164				120	wide	15	25+5	<1834
Portage Valley - middle moraine - between VC and oldest moraine								
29	cored	*Picea*	1984	102	good	U		
31				88	wide	25		
35				90	good	15		
37				46	wide	25		
40				114	wide	U		
41				149	good	U	25	1810
42				94	none	U		
43				70	none	U		
45				101	wide	20		
46				75	wide	10		
47				99	good	30		
50				102	none	35		
51				112	wide	50		
53				88	wide	35		
54				62	good	10		
55				122	good	20		

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Portage Valley (con't)								
56				86	wide	30		
58				87	good	25		
59		*Tsuga*		58	good	40		
60				80	good	30		
159		*Picea*		112	none	10		
Portage Valley - VC moraine - west shore								
176	cored	*Picea*	1985	35	good	20		
177				33	wide	0		
178				48	none	0		
179				57	good	15	25+5	1898
180				32	wide	10		
181				32	wide	10		
Portage Valley - VC moraine - south side - near Byron (1980?)								
16	cored	*Picea*	1984	33	good	25		
17				66	good	U	25	1893
18				41	wide	U		
19				53	good	U		
21				54	good	U		
22				54	good	U		
155				56	none	10		
156				45	good	15		
157				57	good	15		
Portage Valley - VC moraine - central portion - near VC								
T-18	cut	*Picea*	1984	100	good	0	25	1859
T-19				40	good	0		
T-21				73	good	0		
T-22				78	good	0		
T-23				70	good	0		
Slab A	cut	*Picea*	1985	72	good	U		
B				73	good	U		
C				73	good	U		
D				73	good	U		
158	cored	*Picea*	1984	79	good	10	25+3	1877
64				70	good	20		
66				75	none	20		
67				53	good	20		
68				53	good	10		
69				67	good	30		
70				50	wide	30		
71				49	wide	25		
72				58	good	20		
74				77	good	0		
86				72	good	20		
87				71	wide	30		
88				72	good	30		

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Portage Valley - VC moraine - Bear Valley								
107	cored	*Picea*	1984	56	good	20		
109				56	wide	20		
110				73	good	20		
111				58	wide	20		
112				58	good	10		
113				63	good	15		
114				58	wide	20		
115				48	good	10		
116				67	good	25		
118				76	good	10		
119				75	good	15		
120				81	good	10		
121				45	wide	20		
122				83	good	10	25+3	1873
123				56	good	10		
125				78	good	15		
126				53	wide	10		
127				45	wide	10		
128				76	good	10		
129				56	wide	10		
130				60	good	10		
Portage Valley - trimline area								
186	cored	*Picea*	1985	47	wide	10		
187				51	good	0		
189				60	good	10		
191				59	good	0		
192				62	good	0	25	1898
193				41	wide	20		
194				43	good	10		
Portage Valley - lakeshore - between Byron channels								
165	cored	*Picea*	1985	40	good	10		
166				46	good	10		
169				36	wide	5		
195				46	good	15		
196				35	good	10		
198				26	wide	10		
199				25	wide	10		
200				36	good	15		
201				30	wide	10		
203				48	good	10	25+3	1909
204				39	good	10		
205				46	good	0		
206				31	good	5		

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Portage Valley - lakeshore - Byron delta - near boulder with X								
305	cored	*Picea*	1985	29	good	10		
306				35	good	10	25+3	1922
307				33	good	10		
Portage Valley - lakeshore - zigzag moraine								
T-29	cut	*Salix*	1985	30	good	0		1955
174	cored	*Picea*	1985	23	good	U	25	1937
Portage Pass - lowest area - mature forest - Whittier side								
300	cored	*Tsuga*	1985	328	good	30		1657
316				114	wide	5		
317				127	good	0		
318				212	good	10		
319				234	good	10		
320				229	wide	10		
321				197	good	10		
322				224	wide	0		
323				119	wide	30		
324				213	good	30		
325				169	good	100		
326				137	good	20		
327				164	wide	15		
328				126	wide	10		
329				186	good	25		
Portage Pass - mid elevation - outside glacial limit - Whittier side								
330	cored	*Picea*	1985	65	good	0		
331				60	none	0		
333				183	good	0	25	1777
335				101	wide	5		
336				129	good	15		
337		*Tsuga*		91	good	40		
339		*Picea*		144	good	0		
340				75	wide	5		
343		*Tsuga*		205	good	5		1780
Portage Pass - outside glacial limit - top of pass - W side								
75	cored	*Tsuga*	1984	46	wide	U		
301				163	good	10		1822
302				170	none	25		<1814
T-a	cut	*Tsuga*	1984	49	wide			
Portage Pass - oldest moraine - west & central portions								
77	cored	*Tsuga*	1984	70	good	50		
79				94	good	10		
81				97	good	20		
82				92	good	10		
T-11	cut	*Picea*	1984	40	good			

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Portage Pass (con't)								
T-12		*Tsuga*		119	good			1865
T-13		*Picea*		76	good			
T-b				44	wide			
Portage Pass - oldest moraine - E side - over canyon								
308	cored	*Tsuga*	1985	110	wide	10		
309				119	good	10		1865
Portage Pass - Dry Lake moraine								
T-14	cut	*Tsuga*	1984	49	good	0		1935
Portage Pass - between Dry Lake & Divide Lake								
T-15	cut	*Picea*	1984	41	good	0		1918
Portage Pass - S side of Divide Lake								
T-16	cut	*Picea*	1984	32	good	0		1927
Portage Pass - area below 17 mm moraine - across from bedrock area								
T-a	cut	*Alnus*	1987	11				
b				11				
c				12		0		1975
d				9				
e				11				
Spencer-outside glacial limit								
375	cored	*Tsuga*	1986	92	wide	U		
376				138	good	20		1848
377				118	wide	25		
Spencer-ARR moraine								
350	cored	*Populus*	1986	61	good	10		
352				63	wide	15		
353				62	wide	10		
354				57	wide	10		
355				46	wide	10		
356				52	wide	15		
359				73	none	20		<1913
360				72	none	10		
361				67	good	15		
362				51	none	15		
363				62	good	10		
364				49	wide	10		
367				34	wide	25		
368				57	none	20		
369				58	wide	20		
370				59	none	20		
372				61	none	10		
374				69	wide	U		

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Spencer-trimline area								
378	cored	*Picea*	1986	55	good	20		
379				60	good	10	25+3	1898
380				38	wide	15		
T-a	cut	*Tsuga*	1986	56				
b				55				
c				53				
d				60				1926
Spencer - lakeshore moraine								
T-1	cut	*Salix*	1986	28	good	0		1958
Bartlett - outside glacial limit								
430	cored	*Tsuga*	1987	189	good	15		1798
431				121	2 ctr	0		
Bartlett - ARR moraine								
D	cut	*Populus*	1987	60	good	U		
420	cored	*Populus*	1987	28	good	10		
421				20	wide	10		
422				25	good	10		
423				19	good	10		
424				15	none	10		
425				82	none	45		1905
426				57	wide	35		
427				57	wide	30		
428				58	wide	30		
429				55	wide	U		
434				61	none	10		
435				72	wide	10		
436				89	none	35		<1898
437				57	none	100		
Bartlett - older moraine - adjacent to ice - Sample Bag B								
B-a	cut	*Alnus*	1987	23	good			1964
b				20	good			
c				22	good			
d				19	good			
e				14	good			
f				10	good			
g		*Tsuga*		16	good			
Bartlett - deglaciated ridge - adjacent to ice - Sample Bag A just outside 74 margin on photo								
A-a	cut	*Populus*	1987	10	good			
b				14	good			
c				19	good			1968
d				15	good			
e				10	good			

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Bartlett (con't)								
f				12	good			
g				8	good			
h				9	good			
i				13	good			
Bartlett Glacier - deglaciated since 1974 AP - adjacent to SW margin								
T-17	cut	*Alder*	1987	7				
18				7				
19				12				1975
20				6				
21				5				
22				5				
23				6				
24				5				
25				7				
26				5				
Willard Island - W peninsula								
133	cored	*Picea*	1984	276	none	100	25+30	<1653
134				215	none	100		
Willard Island - Central - trimline & moraine								
419x	cored	*Picea*	1987	242	good	U	25	1720
recorded	cut	*Picea*	1987	140				
Willard Island/Badger Point - C. Heusser's cores								
CH-1	cored	U	U	133	none	U		
CH-2				254	none	U		
CH-3				265	none	U		
CH-4				145	none	U		
Marquette Glacier - older, higher area - Sample Bag 1								
1-a	cut	*Alnus*	1987	10				
b				9				
c				9				
d				12				
e				16				1961
Marquette Glacier - beach - driftwood								
2-a	cut	*Alnus*	1987	20				
b				22				1965
c				18				
d				13				
e				19				
f				18				

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Marquette Glacier - cleared bedrock ridge adjacent to ice margin - Sample Bag 1								
1-a	cut	*Alnus*	1987	10				
b				9				
c				9				
d				12				
e				16				1971
Marquette Glacier - beach area - outwash? -1964 earthquake subsidence - Sample Bag 2 - rooted stumps assumed killed in 1964								
2-a	cut	*Alnus*	1987	20				
b				22				1942
c				18				
d				13				
e				19				
f				18				
Lawrence Glacier - ridge C (waterfall) - Sample Bag A								
A-a	cut	*Alnus*	1987	18				
b				10				
c				19				
d				11				
e				22				
f				15				
g		*Tsuga*		16				
h				23				1964
Lawrence Glacier - ridge A (adjacent to ice) - Sample Bag B								
T-10	cut	*Alnus*	1987	20				
T-11				17				
T-12				11				
T-13		*Tsuga*		18				
Lawrence Glacier - overrun trees - ridge A (adjacent to ice) - Sample Bag C								
C-a	cut	*Alnus*	1987	10				
b				8				
c				9				
d				9				
e				8				
f				8				
g				8				
h				11				
i				8				
j				8				
k				9				

Sample No.	Cut/ Cored	Genus	Sample Year	Ring Count	Comments on center	Height cored (cm)	Corr. Factor (yr)*	Oldest Possible Year AD
Tebenkov Glacier - beyond glacial limit - bedrock knob - C on Fig. 7.1								
400	cored	*Picea*	1987	296	none	100	25+30	1636
401				290	good	15		
403				274	good	5		
404				221	none	20		
Tebenkov Glacier - moraine ridge								
405	cored	*Picea*	1987	72	good	20	25+6	1884
406				36	good	10		
407				40	wide	15		
408				46	good	10		
409		*Tsuga*		53	good	10		
410				69	good	10		
411				47	good	10		
412				48	good	5		
413				44	good	25		
T-X	cut	*Tsuga*		53	good			
Tebenkov Glacier - bedrock strike ridge behind moraine								
414	cored	*Picea*		47	good	15	25+5	1910
415		*Tsuga*		37	good	20		
417		*Picea*		26	wide	15		
418				25	good	10		
419				36	good	15		
Tebenkov Glacier - moraine/hill between channel and ice - E side								
3 alder samples								
a	cut	*Alnus*	1987	21				
b				22				1965
c				16				
Tebenkov push moraines - central portion - pulled hemlock & cut alder								
H-a	cut	*Tsuga*	1987	5				
b				6				
c				7				
d				14				1973
e				10				
f				11				
A-a	cut	*Alnus*	1987	5				
b				7				1980
c				4				
d				5				
e				5				
f				4				
g				6				
h				5				
i				4				
j				4				

APPENDIX D

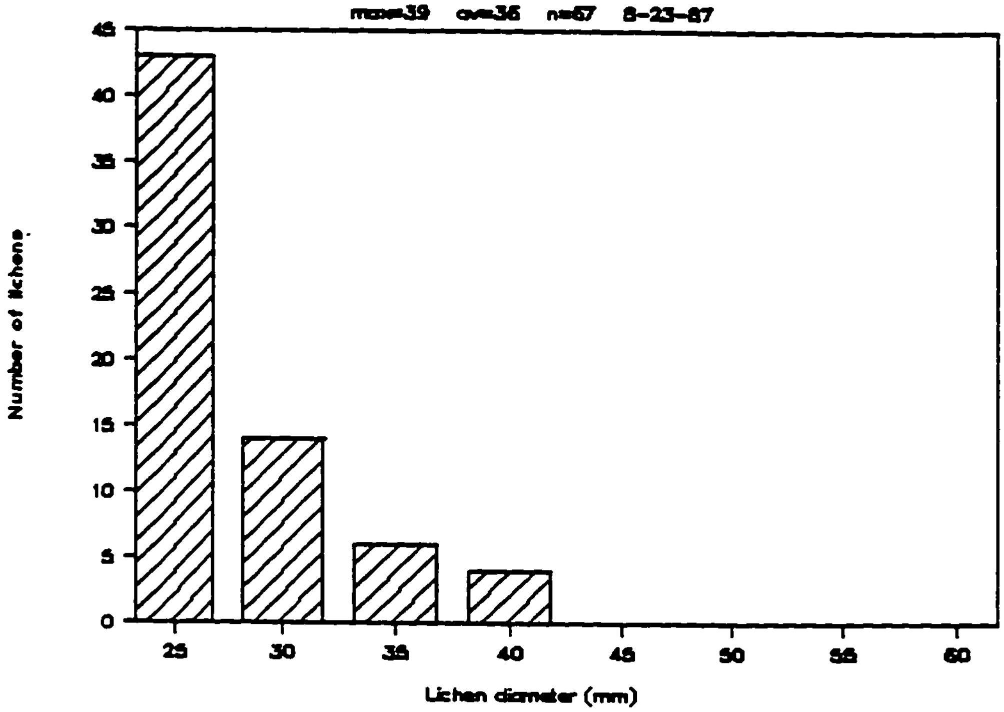
LICHEN THALLI MEASUREMENTS

RHIZOCARPON SP.

ALL FIELD LOCATIONS

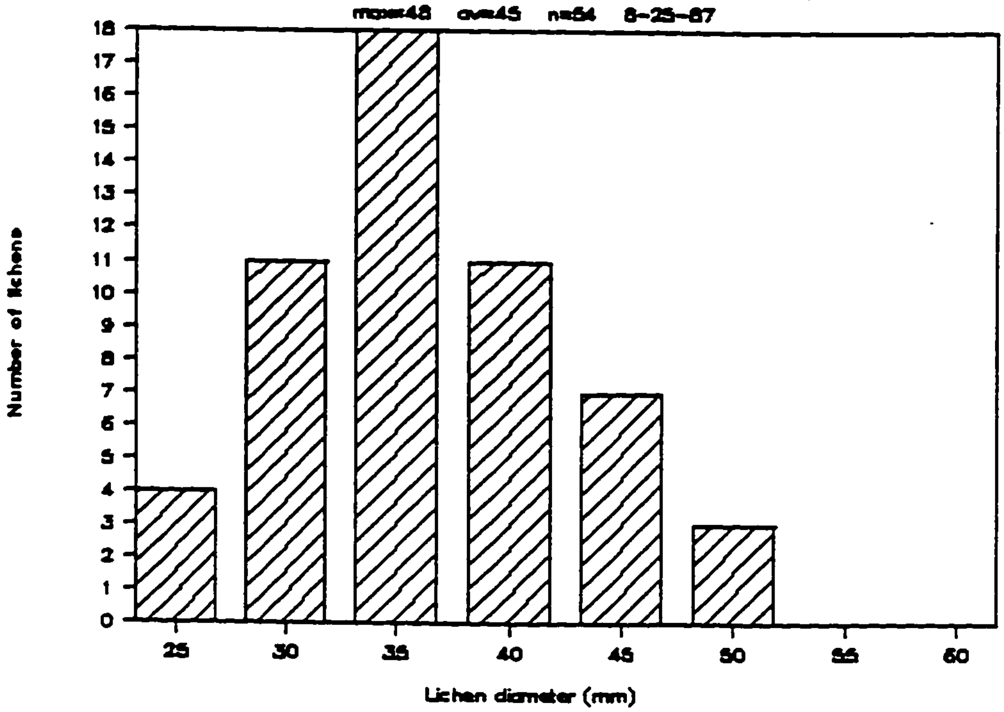
Lawrence Glacier—boulder ridge—E side

max=39 av=36 n=67 8-23-87

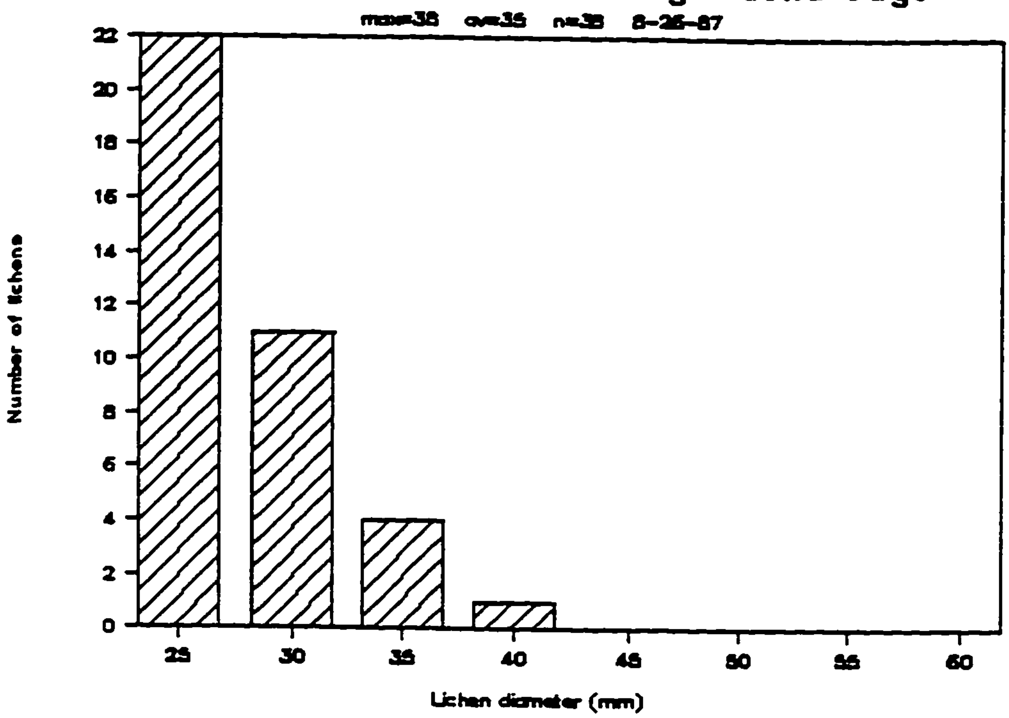


Lawrence Glac—inside boulder ridge—W

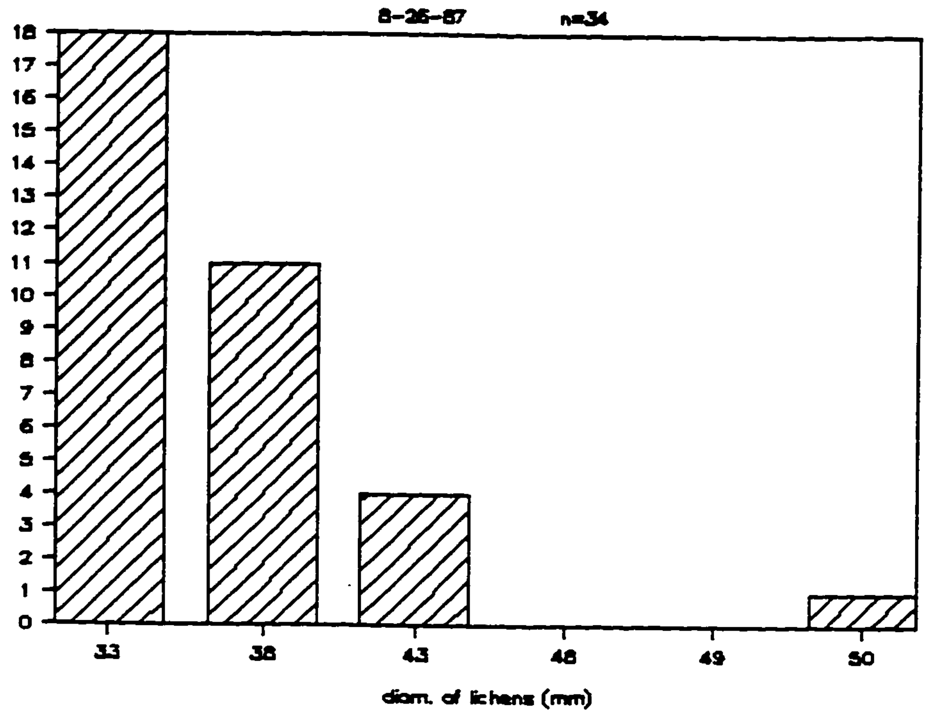
max=48 av=45 n=54 8-25-87

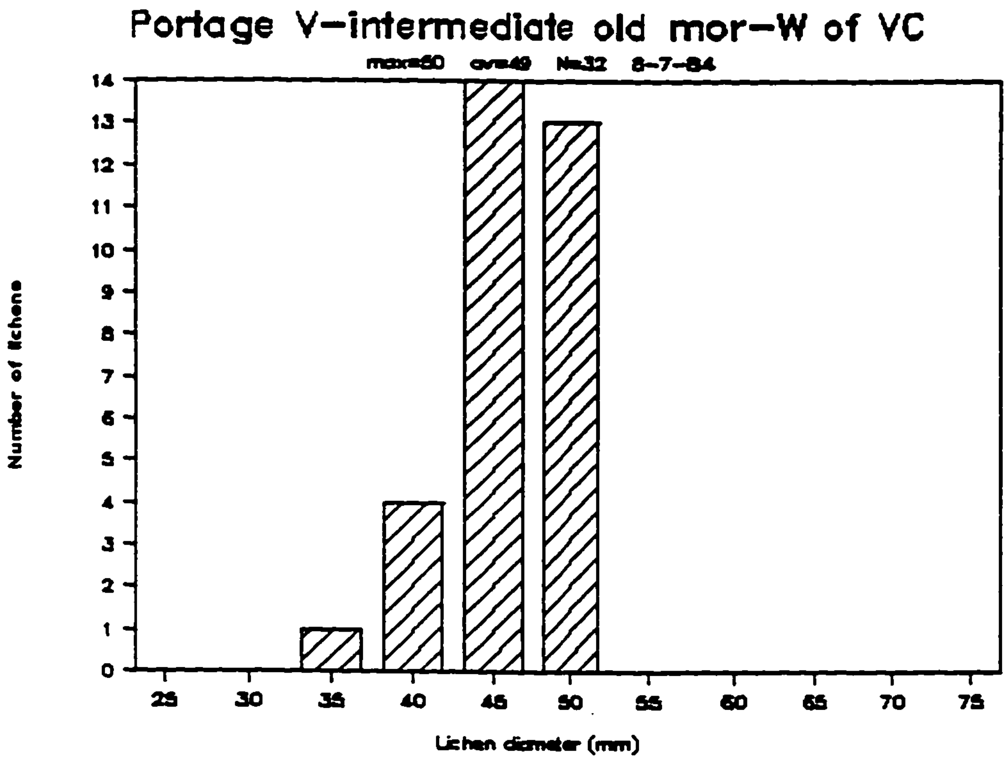


Lawrence Glac—bedrock ridge—delta edge

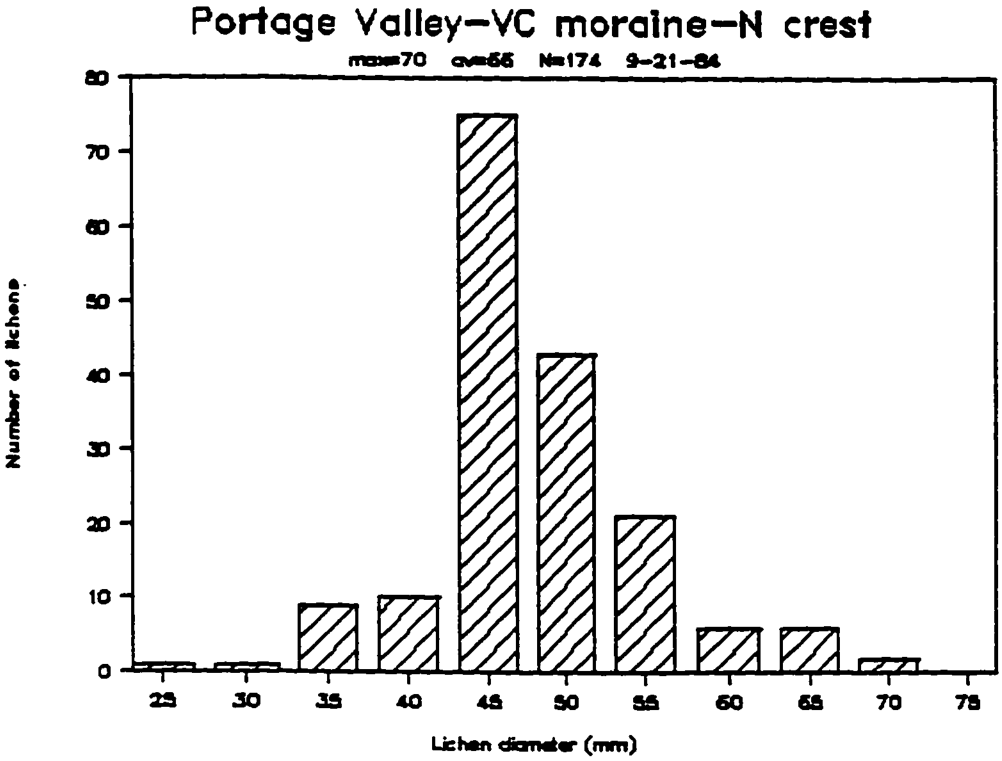


Marquette Glacier—boulder

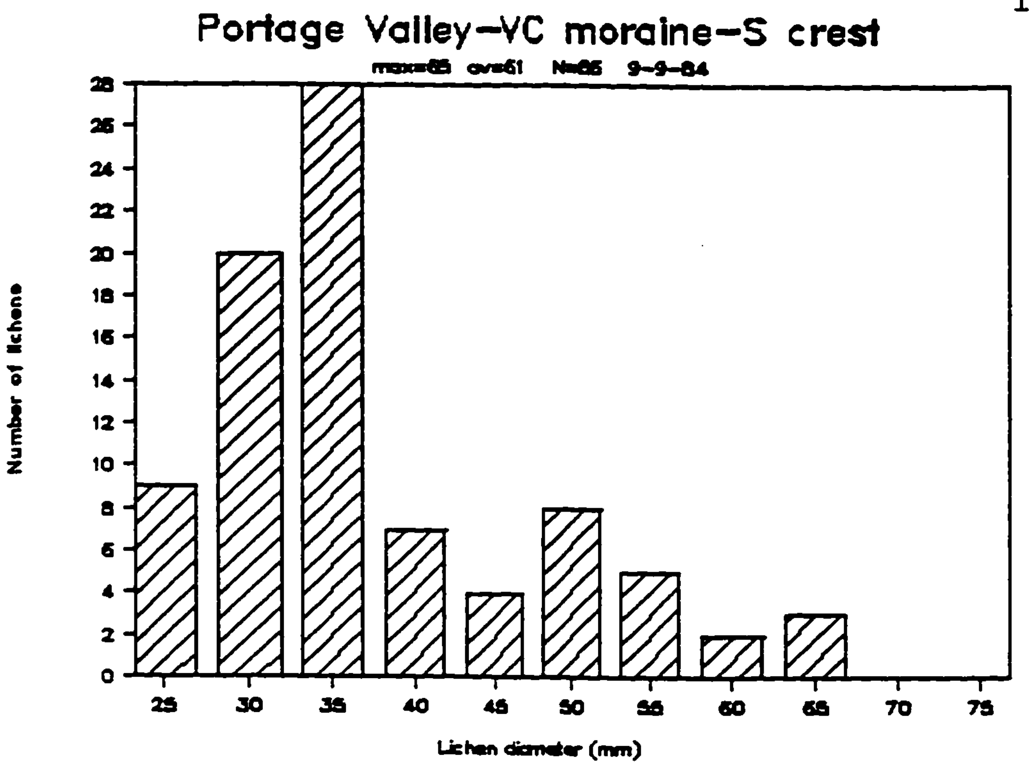


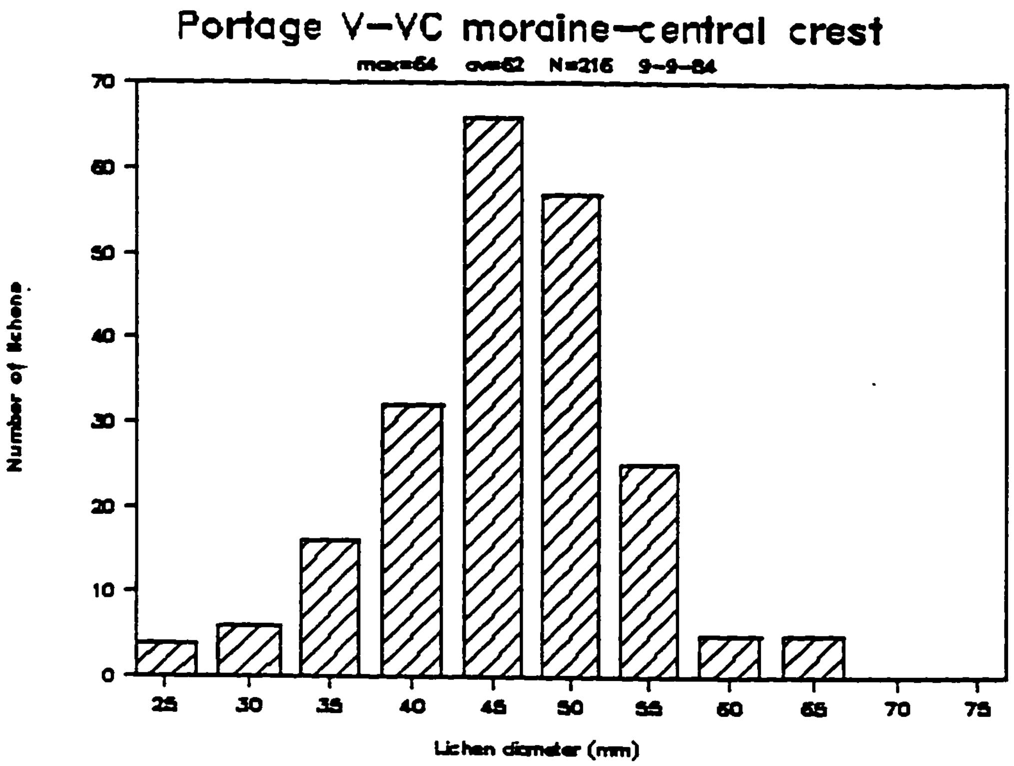


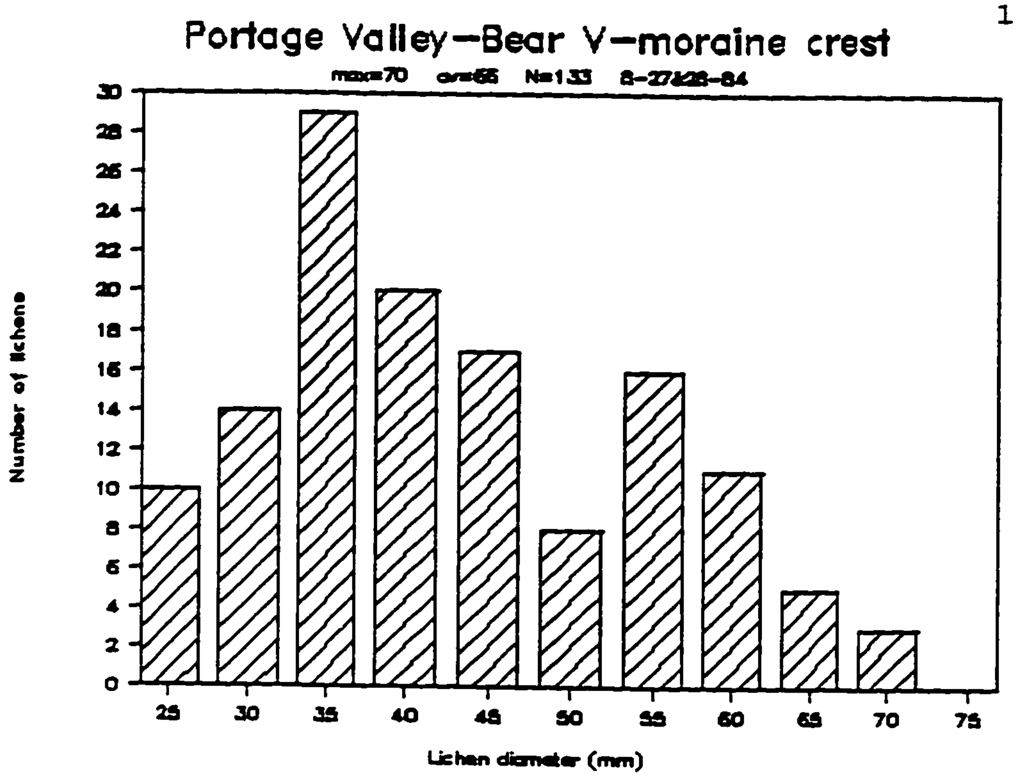
Portage V-intermediate old mor-W of VC
max=60 av=49 N=32 8-7-84
Number of lichens
Lichen diameter (mm)



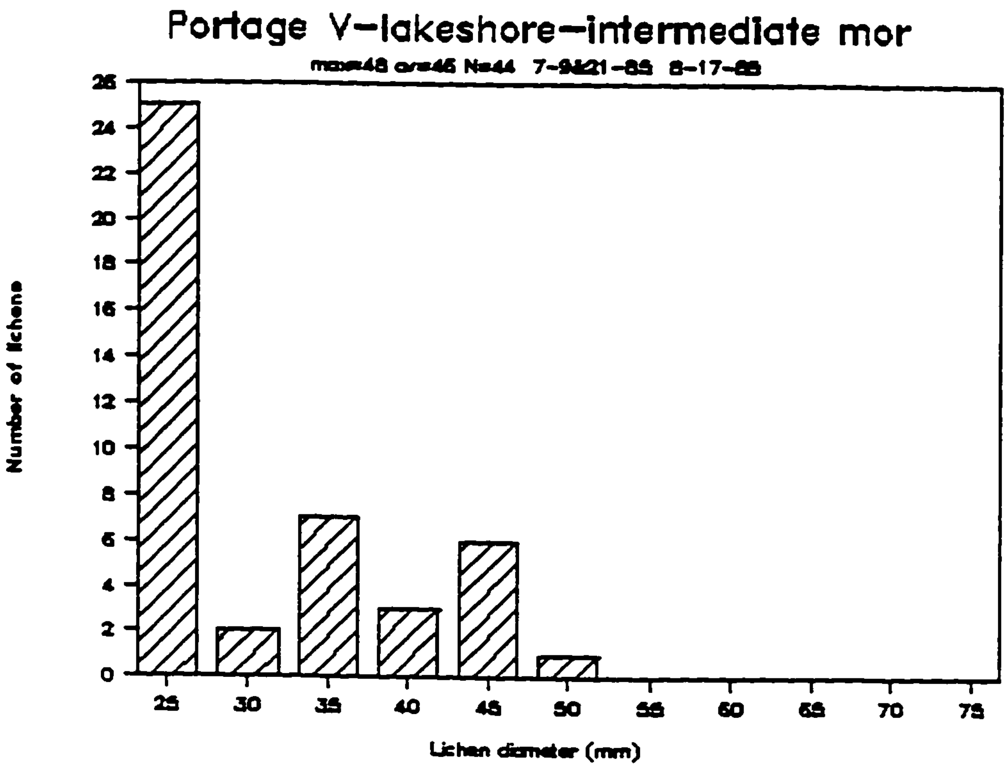
Portage Valley-VC moraine-N crest
max=70 av=55 N=174 9-21-84
Number of lichens
Lichen diameter (mm)


Portage Valley-VC moraine-S crest
max=65 av=61 N=85 9-9-84
Number of lichens
Lichen diameter (mm)
0
2
4
6
8
10
12
14
16
18
20
22
24
26
28
25
30
35
40
45
50
55
60
65
70
75


Portage V-VC moraine-central crest
max=64 av=52 N=216 9-9-84
Number of lichens
Lichen diameter (mm)
0
10
20
30
40
50
60
70
25
30
35
40
45
50
55
60
65
70
75

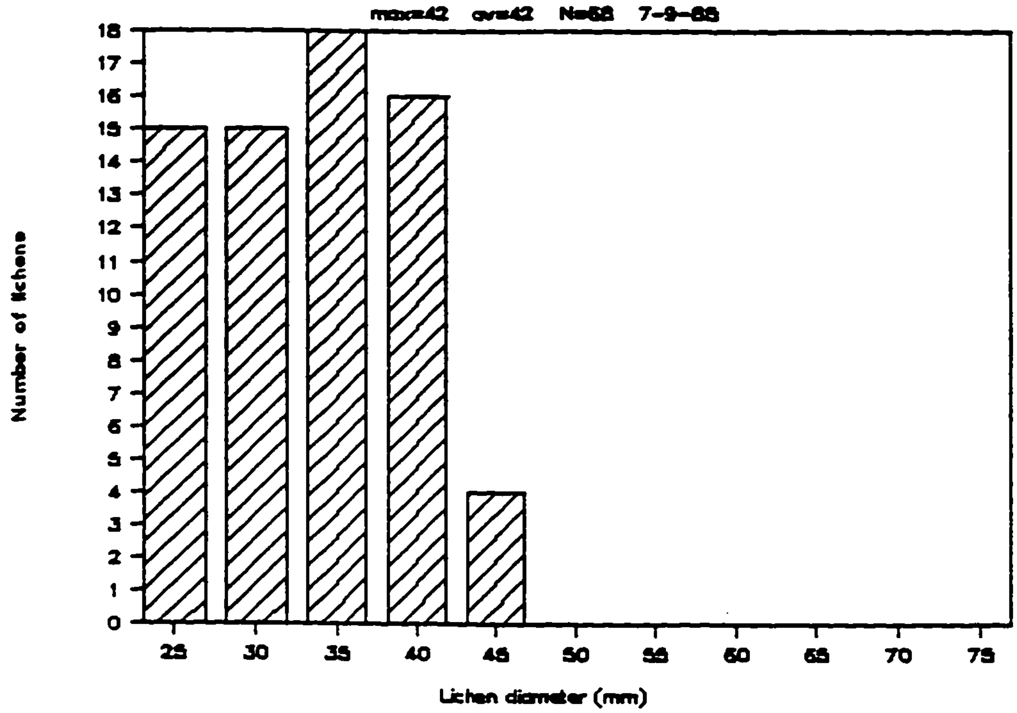


Portage Valley—Bear V—moraine crest
max=70 av=65 N=133 8-27&28-84
Number of lichens
0
2
4
6
8
10
12
14
16
18
20
22
24
26
28
30
25
30
35
40
45
50
55
60
65
70
75
Lichen diameter (mm)

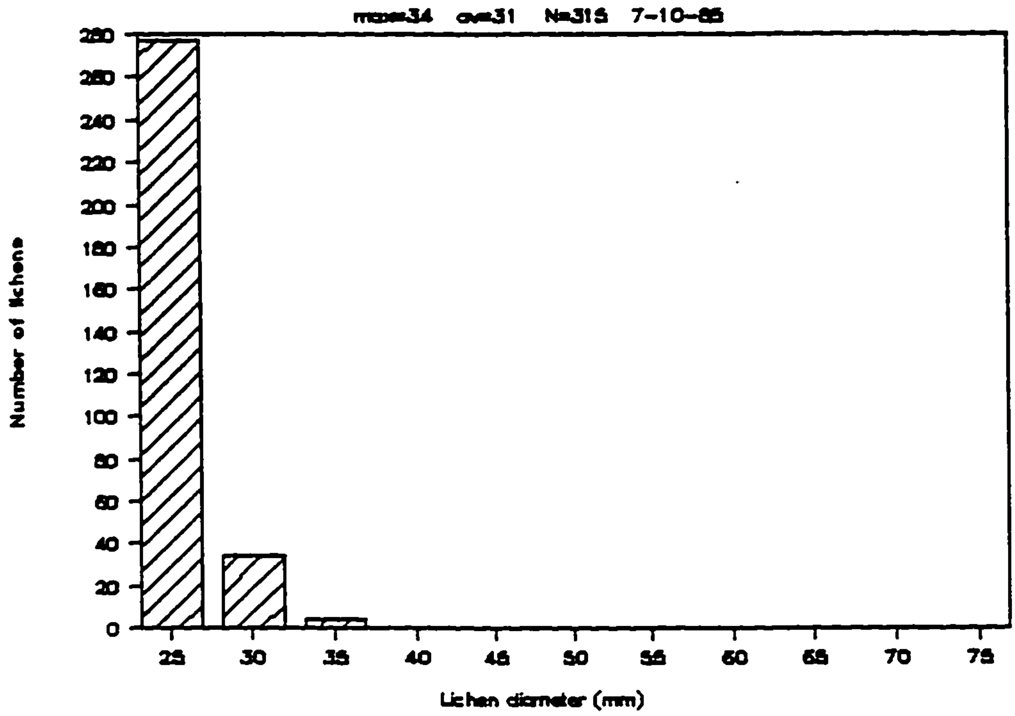


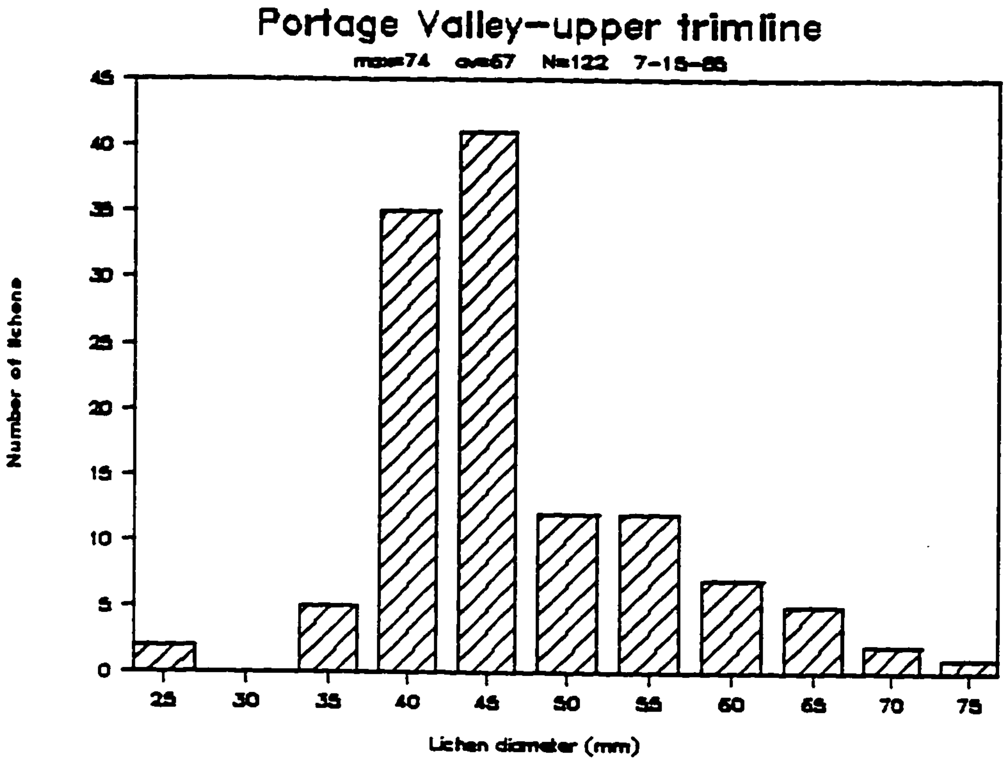
Portage V—lakeshore—intermediate mor
max=48 av=46 N=44 7-9&21-85 8-17-85
Number of lichens
0
2
4
6
8
10
12
14
16
18
20
22
24
26
25
30
35
40
45
50
55
60
65
70
75
Lichen diameter (mm)

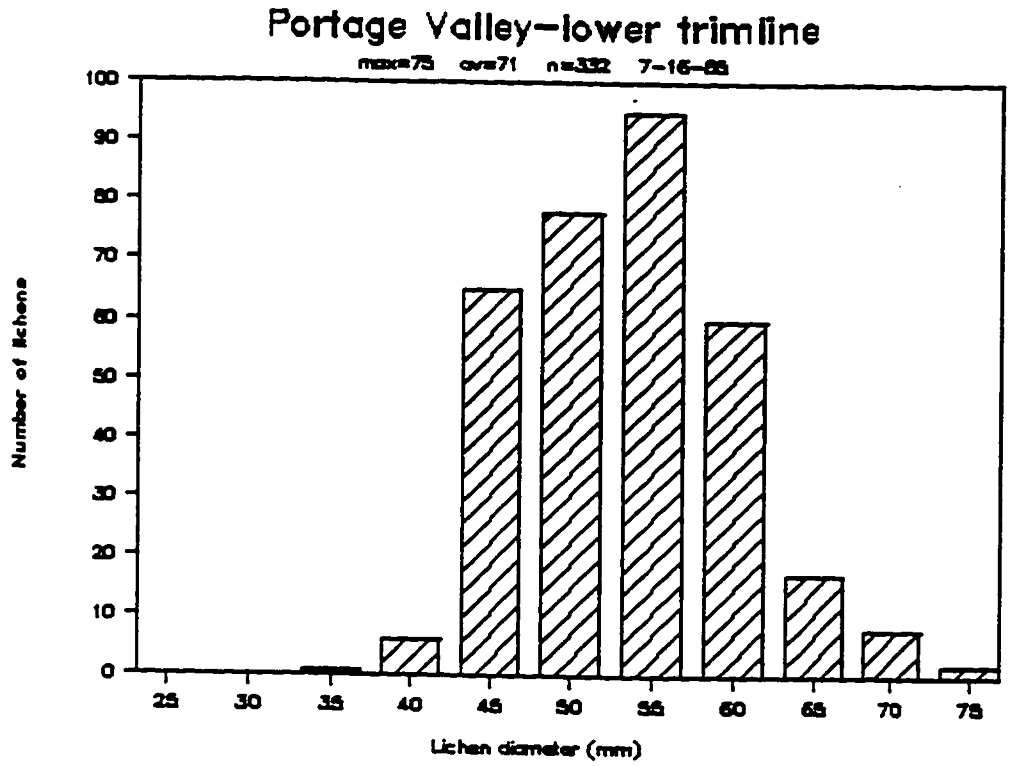
Portage Valley-lakeshore-SW corner



Portage V-lakeshore-youngest moraine(Z)

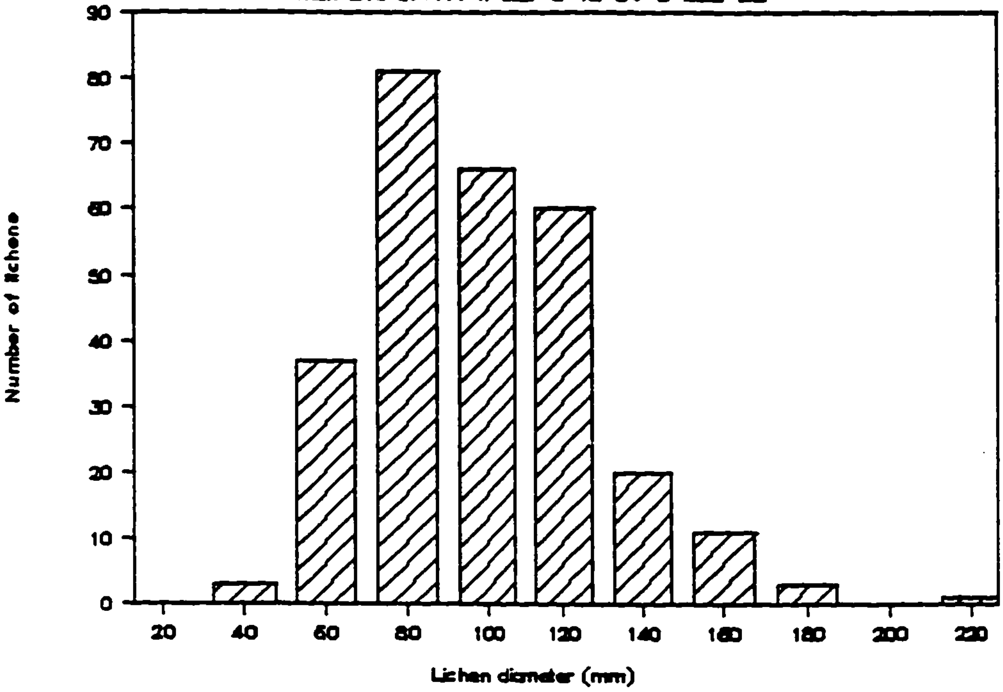



Portage Valley—upper trimline
max=74 av=67 N=122 7-15-88
Number of lichens
Lichen diameter (mm)
0
5
10
15
20
25
30
35
40
45
25
30
35
40
45
50
55
60
65
70
75


Portage Valley—lower trimline
max=75 av=71 n=332 7-16-88
Number of lichens
Lichen diameter (mm)
0
10
20
30
40
50
60
70
80
90
100
25
30
35
40
45
50
55
60
65
70
75

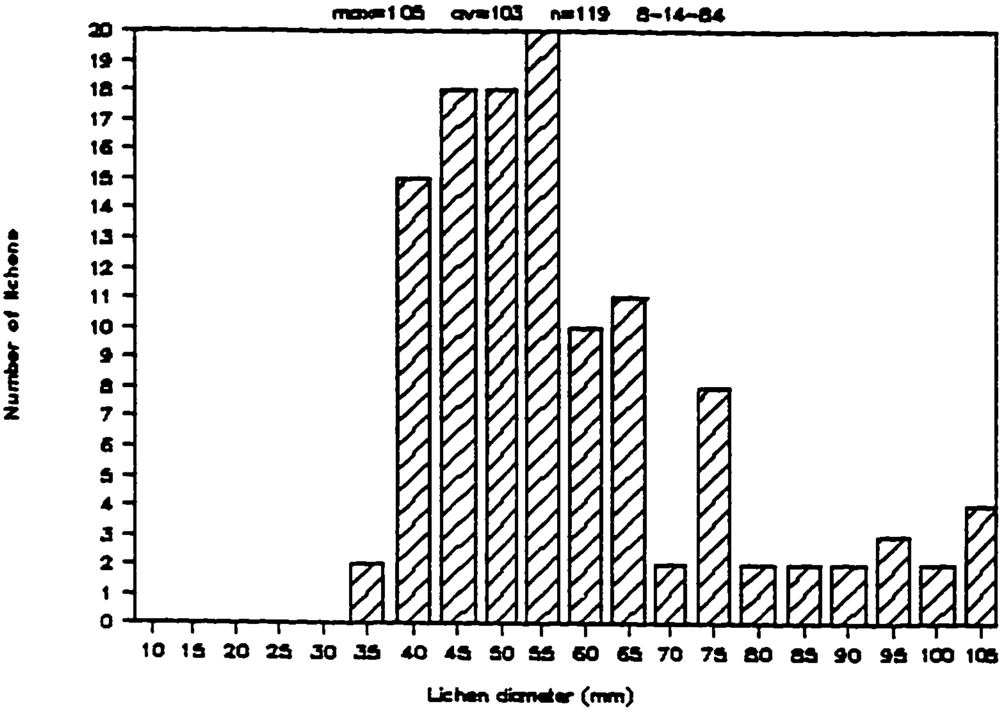
Portage Pass–bedrock outside moraines

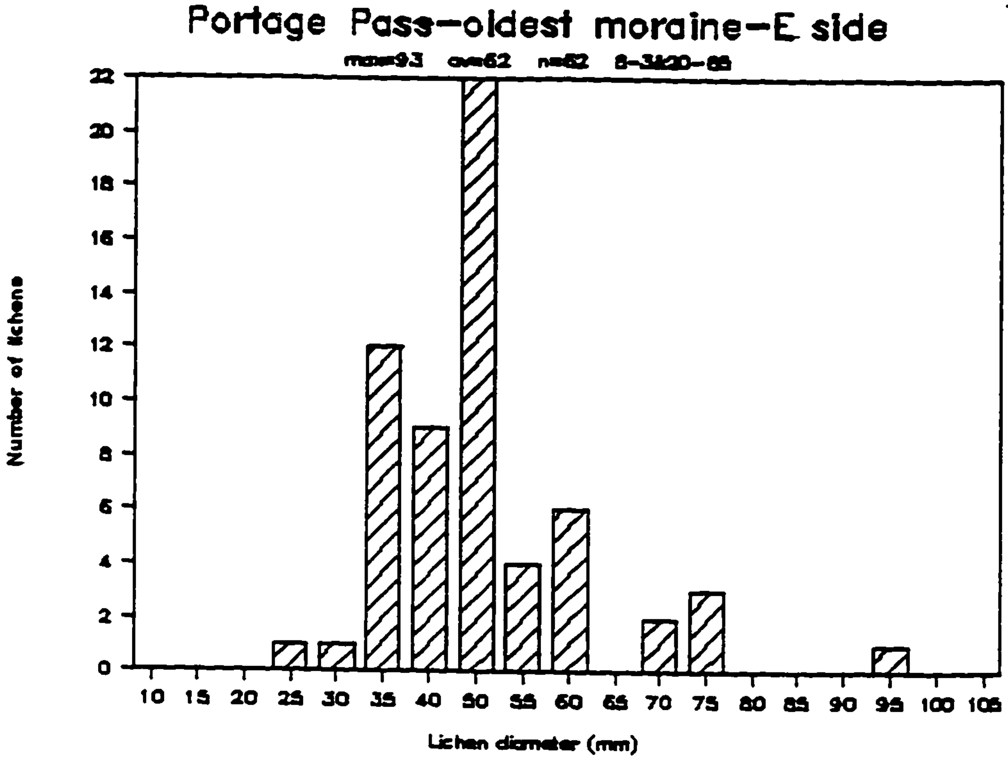
max=210 ave=177 n=282 8-15-84 8-2&3-85

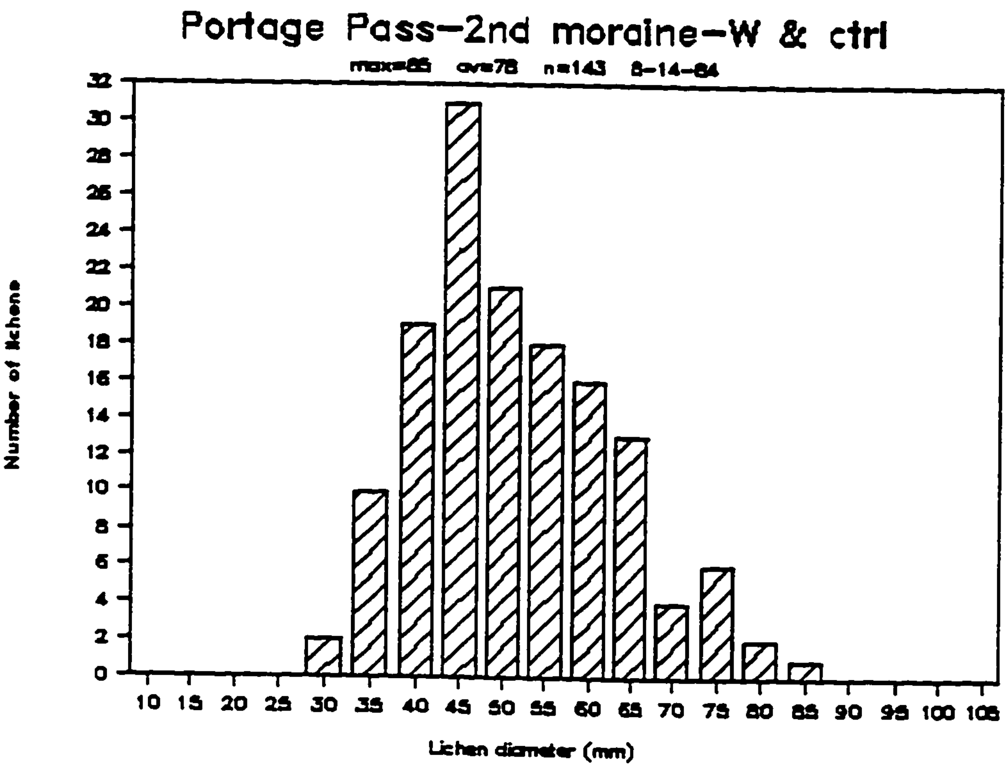


Portage Pass–oldest moraine–W & ctrl

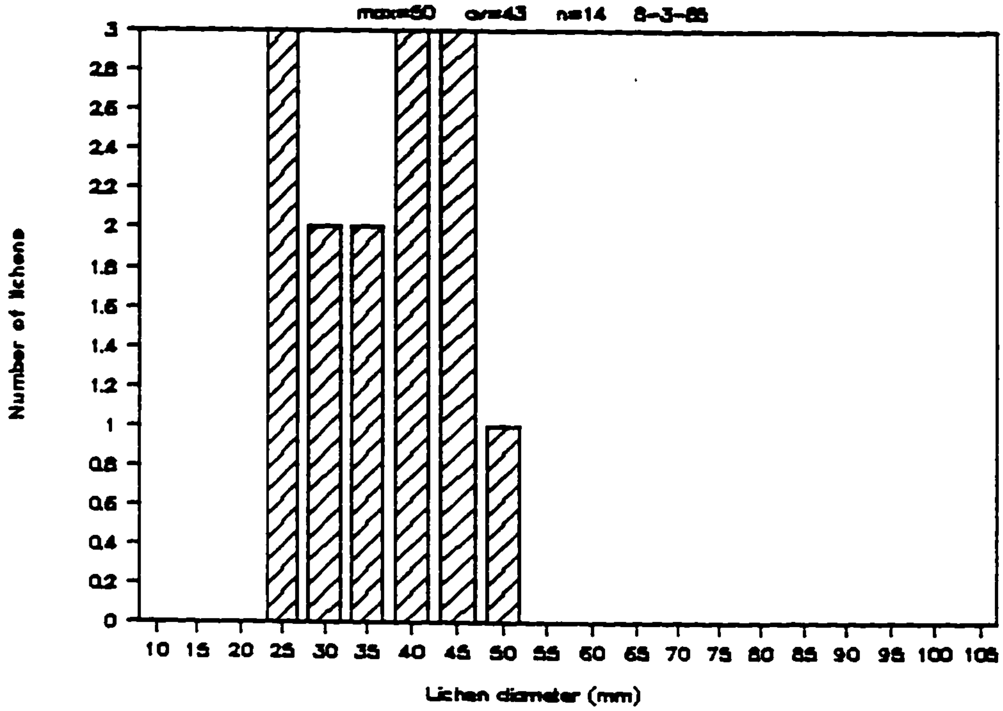
max=105 ave=103 n=119 8-14-84



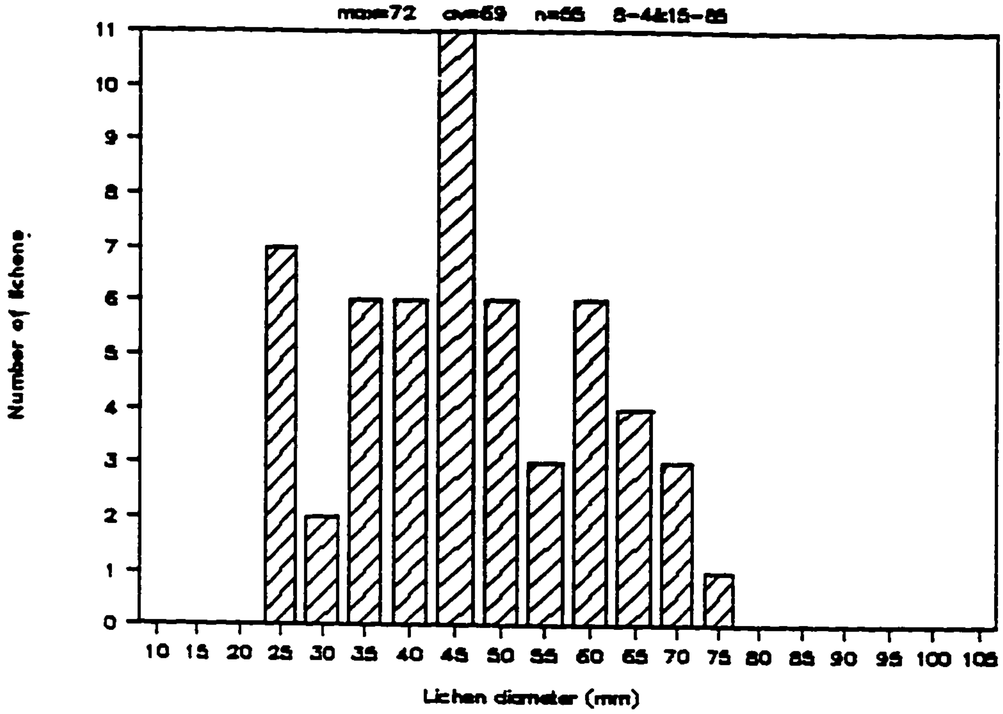

Portage Pass-oldest moraine-E side
max=93 av=62 n=62
Number of lichens
Lichen diameter (mm)
0
2
4
6
8
10
12
14
16
18
20
22
10 15 20 25 30 35 40 50 55 60 65 70 75 80 85 90 95 100 105


Portage Pass-2nd moraine-W & ctrl
n=143 8-14-84
Number of lichens
Lichen diameter (mm)
0
2
4
6
8
10
12
14
16
18
20
22
24
26
28
30
32
10 15 20 25 30 35 40 45 50 55 60 65 70 75 80 85 90 95 100 105

Portage Pass—2nd moraine—E side

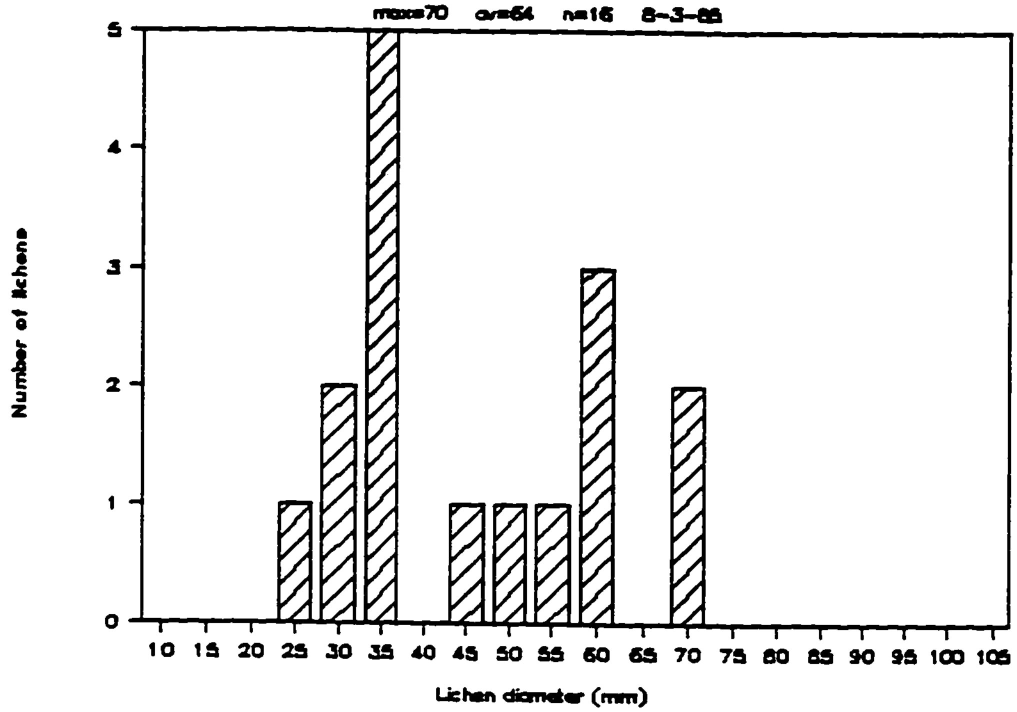


Portage Pass—3rd moraine—W & ctrl



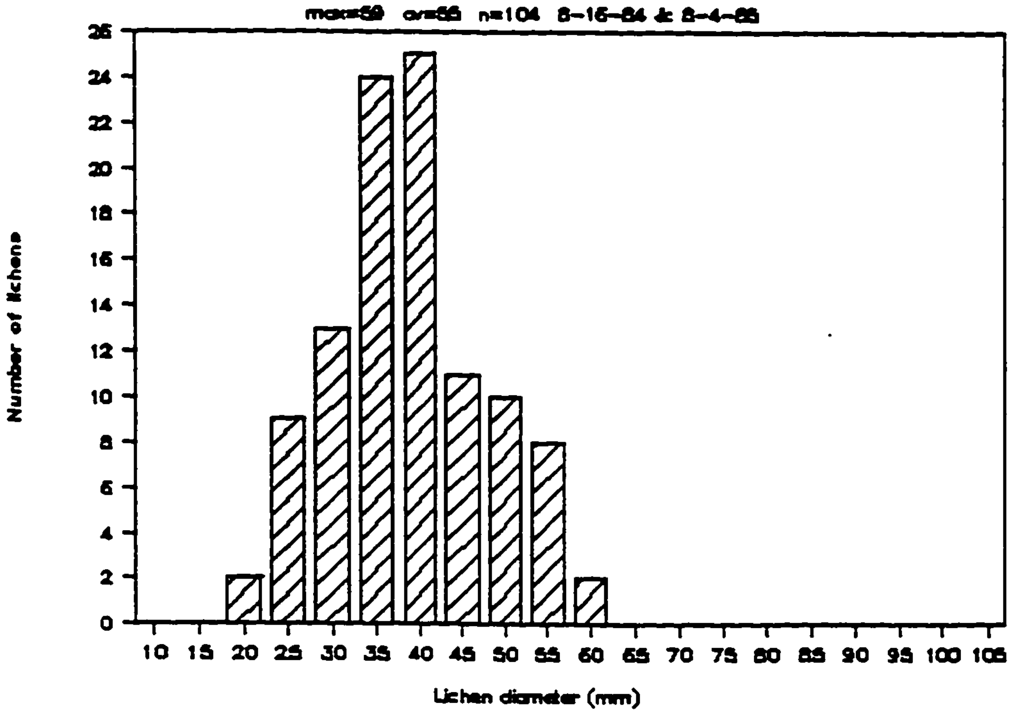
Portage Pass-3rd moraine-E side

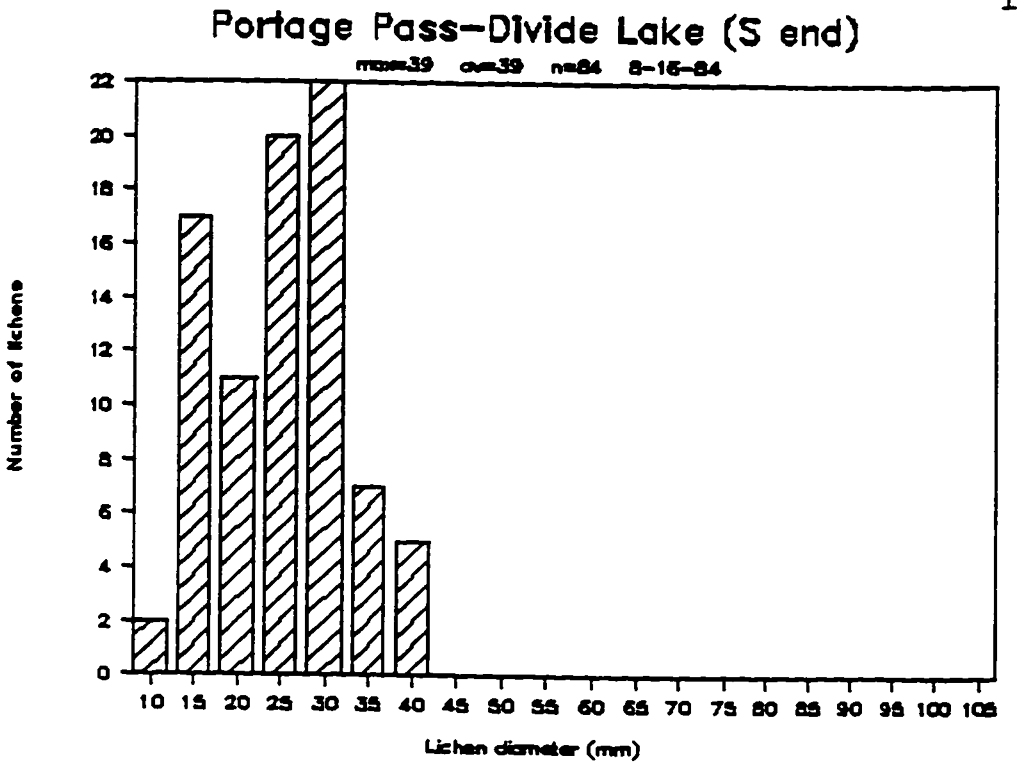
max=70 av=64 n=16 8-3-85



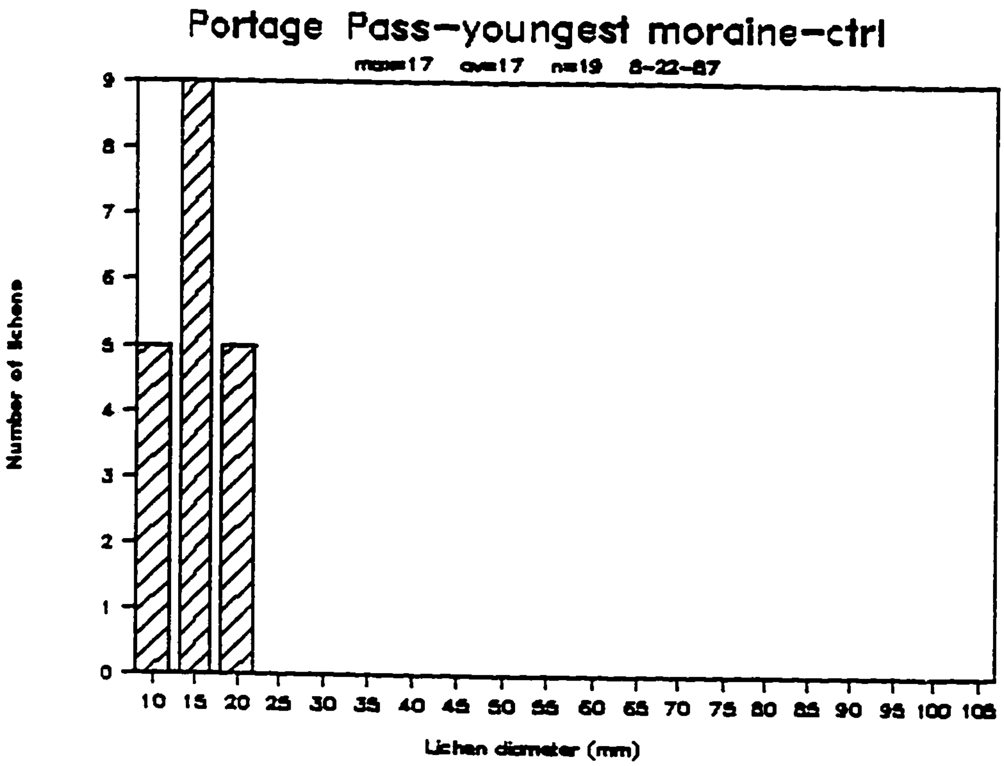
Portage Pass-4th mor-W & ctrl-Dry Lk

max=59 av=55 n=104 8-16-84 & 8-4-85

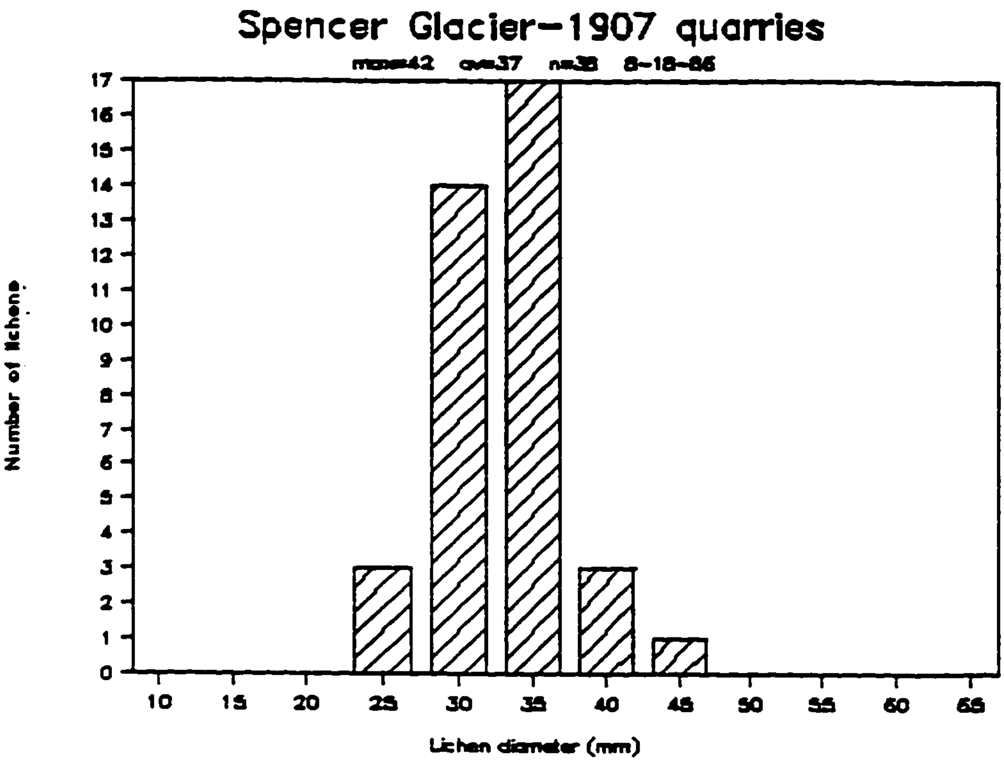


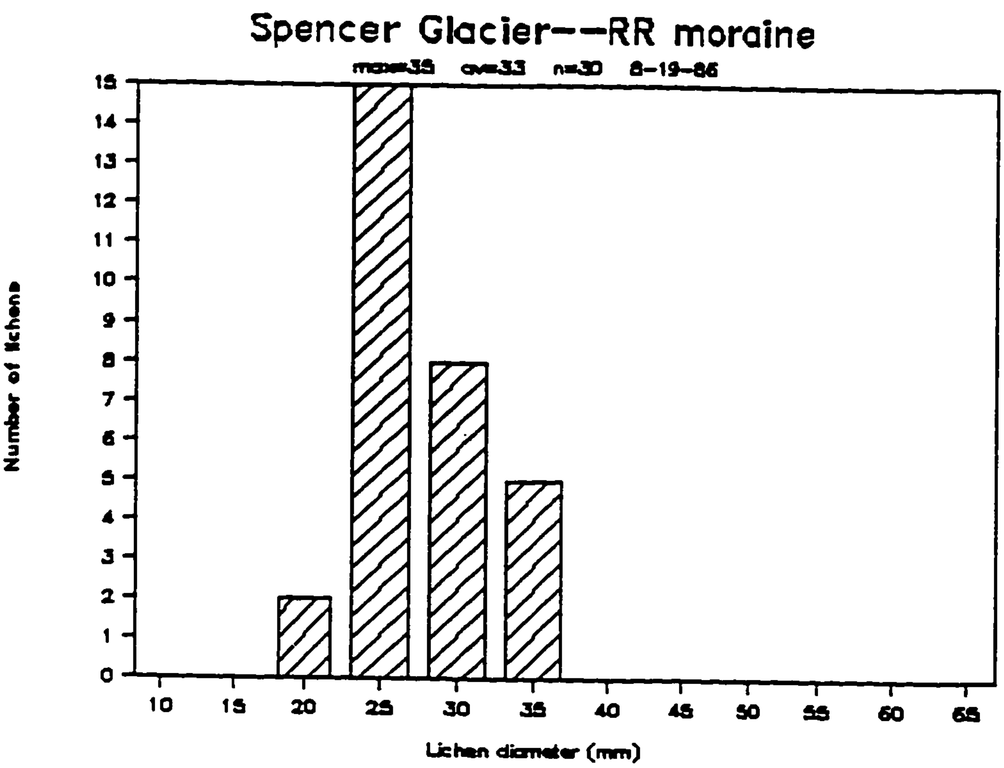


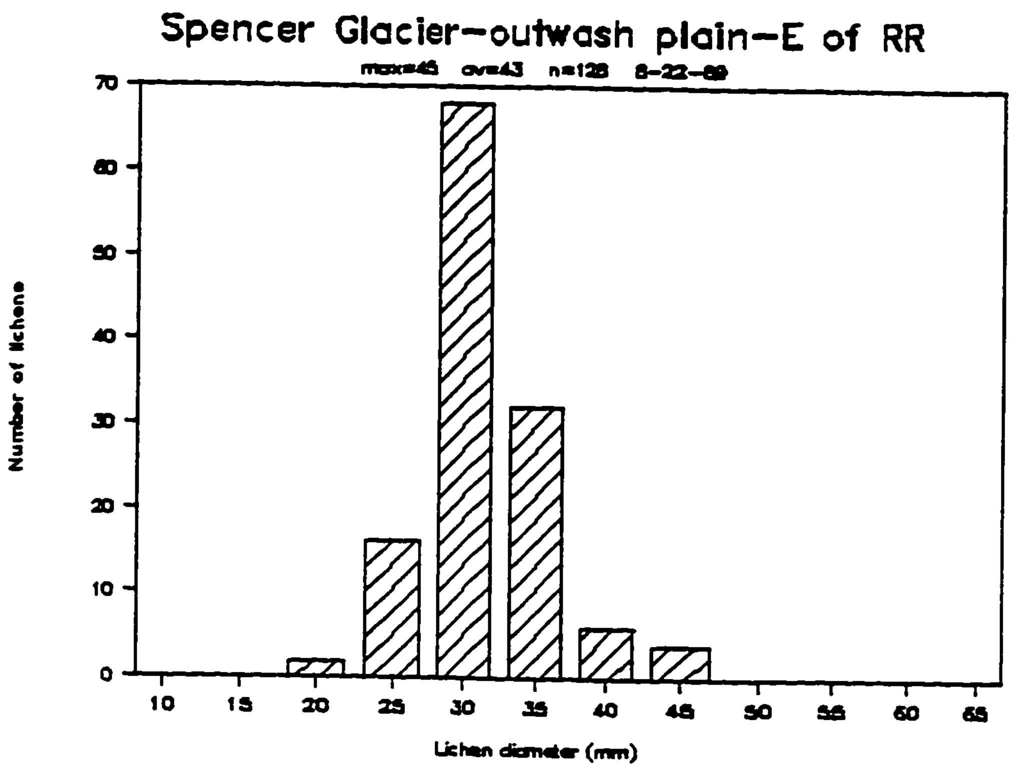
Portage Pass–Divide Lake (S end)
max=39 av=39 n=84 8-16-84
Number of lichens
Lichen diameter (mm)
0
2
4
6
8
10
12
14
16
18
20
22
10 15 20 25 30 35 40 45 50 55 60 65 70 75 80 85 90 95 100 105

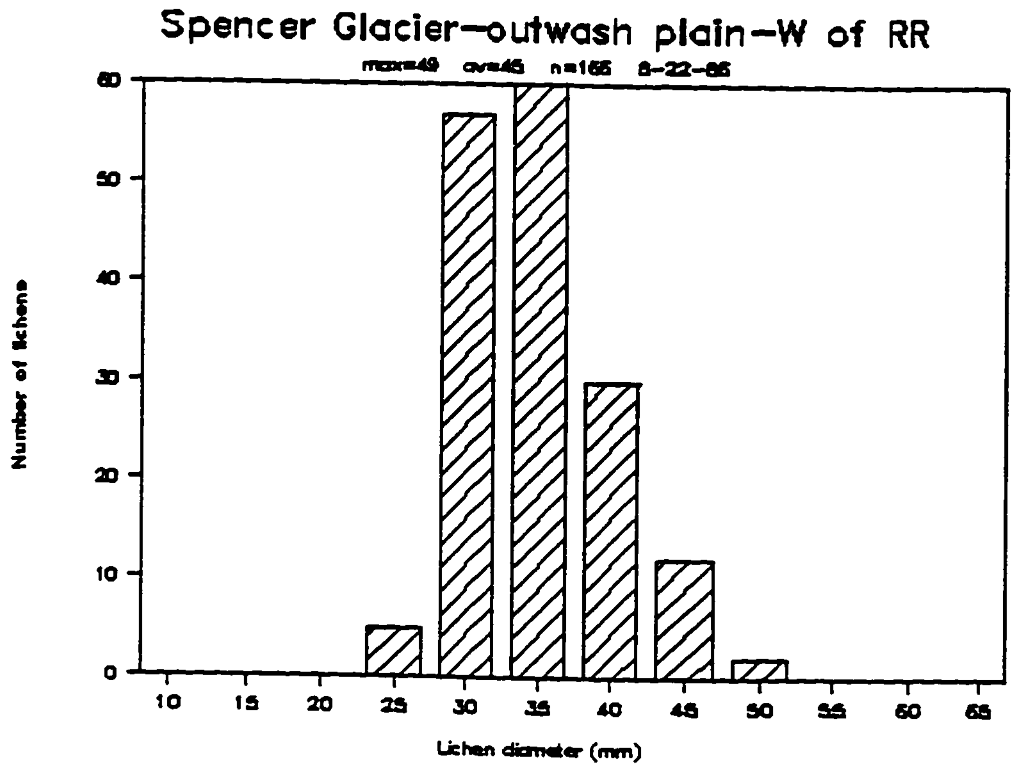


Portage Pass–youngest moraine–ctrl
max=17 av=17 n=19 8-22-87
Number of lichens
Lichen diameter (mm)
0
1
2
3
4
5
6
7
8
9
10 15 20 25 30 35 40 45 50 55 60 65 70 75 80 85 90 95 100 105

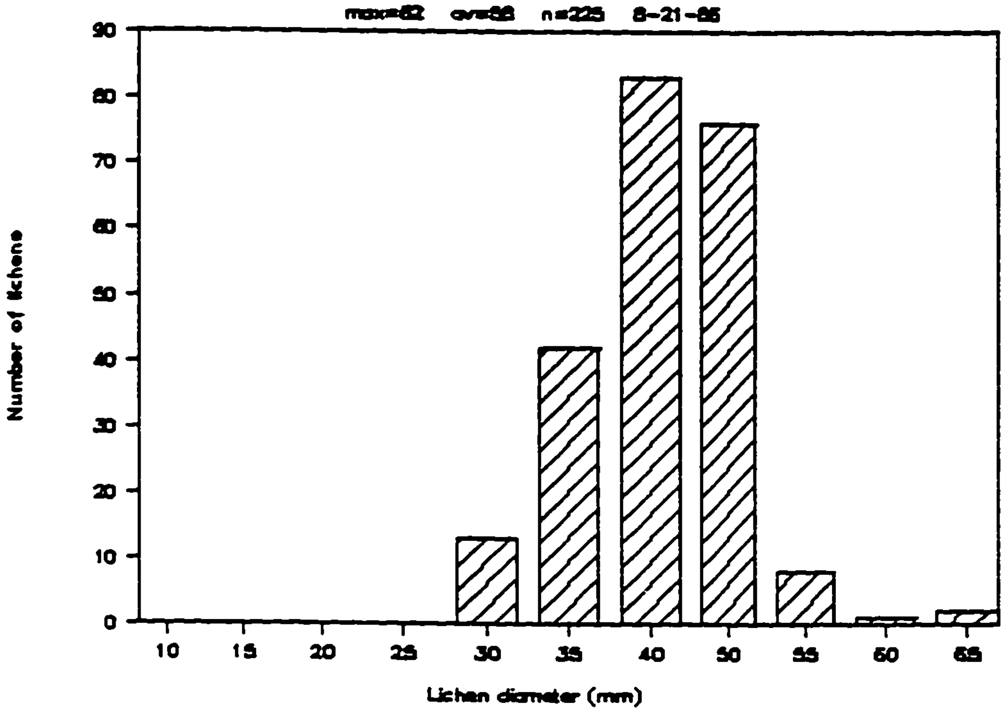

Spencer Glacier—1907 quarries
max=42 av=37 n=38 8-18-86
Number of lichens
Lichen diameter (mm)
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
10
15
20
25
30
35
40
45
50
55
60
65


Spencer Glacier——RR moraine
max=35 av=33 n=30 8-19-86
Number of lichens
Lichen diameter (mm)
0
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
10
15
20
25
30
35
40
45
50
55
60
65

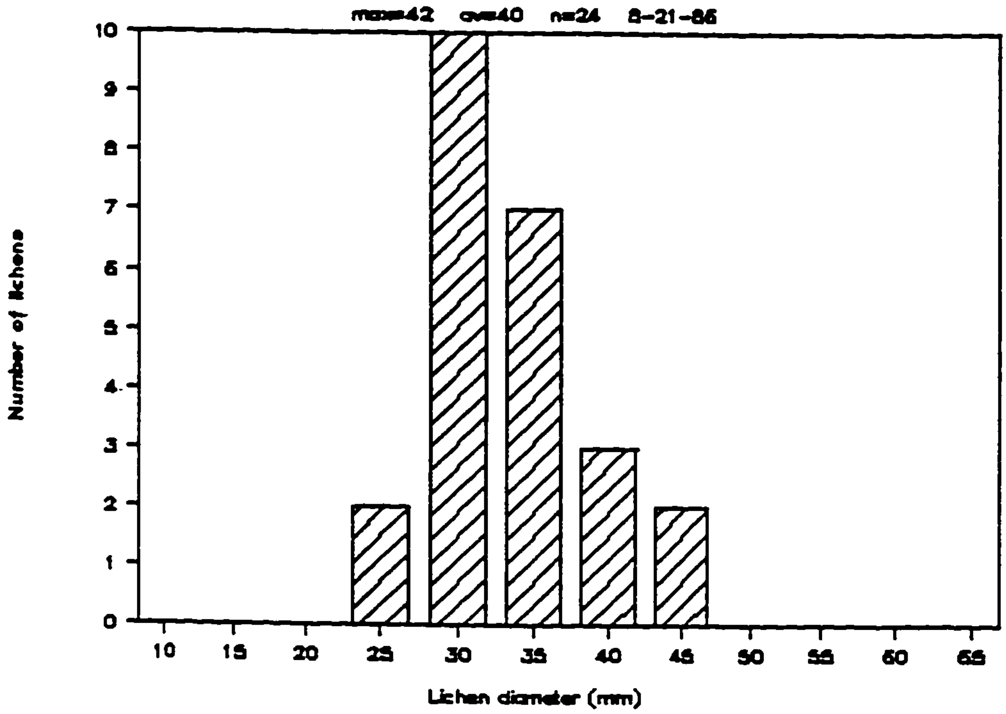

Spencer Glacier—outwash plain—E of RR
max=46 av=43 n=128 8-22-89
Number of lichens
0
10
20
30
40
50
60
70
10
15
20
25
30
35
40
45
50
55
60
65
Lichen diameter (mm)

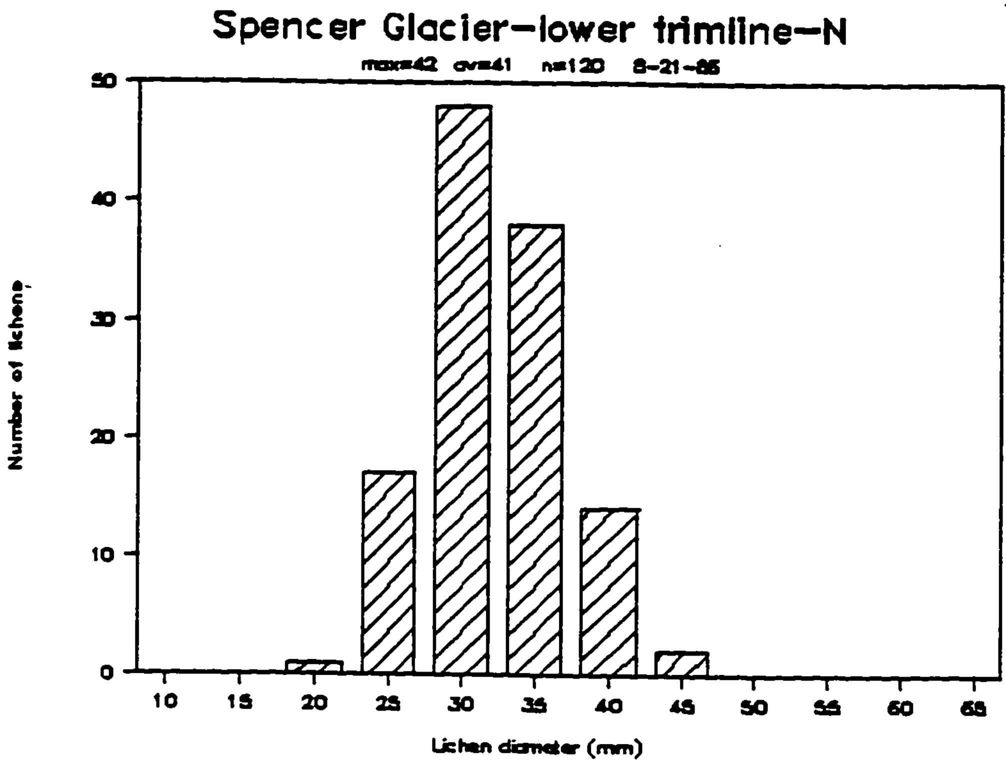

Spencer Glacier—outwash plain—W of RR
max=49 av=45 n=165 8-22-89
Number of lichens
0
10
20
30
40
50
60
10
15
20
25
30
35
40
45
50
55
60
65
Lichen diameter (mm)

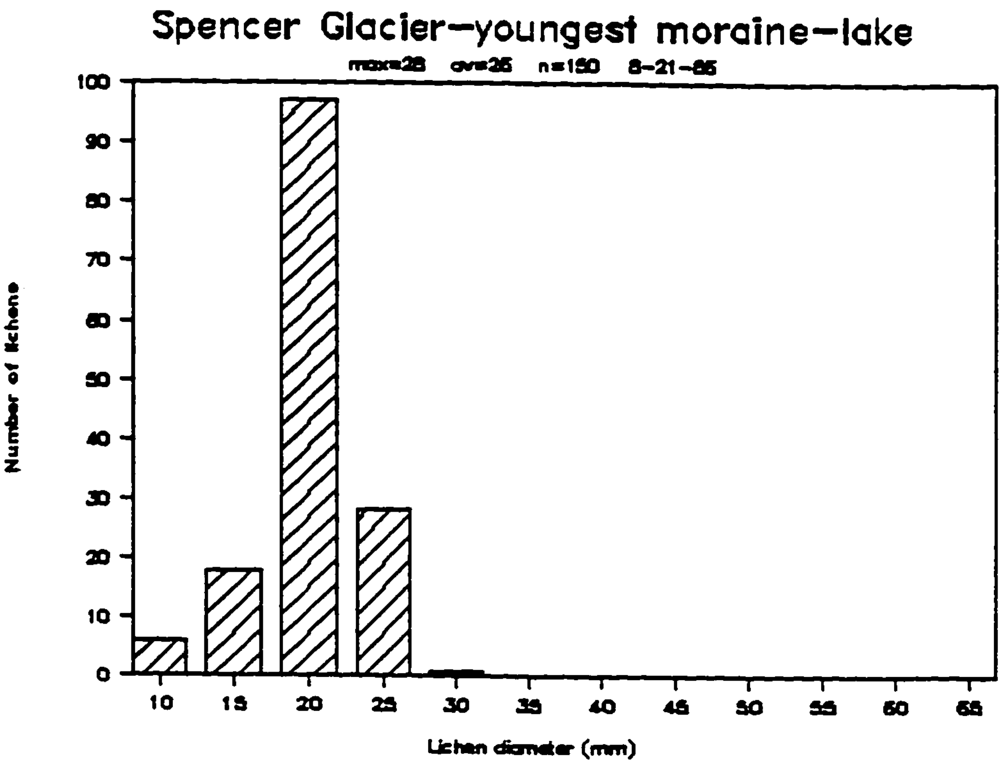
Spencer Glacier–upper trimline–N



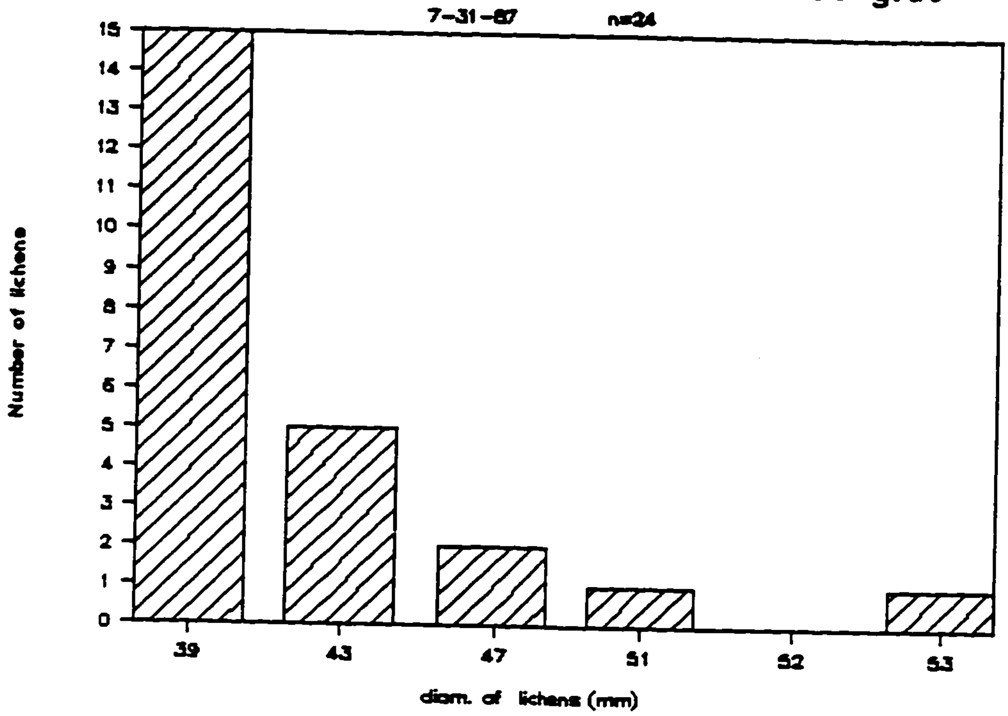
Spencer Glacier–middle trimline–N



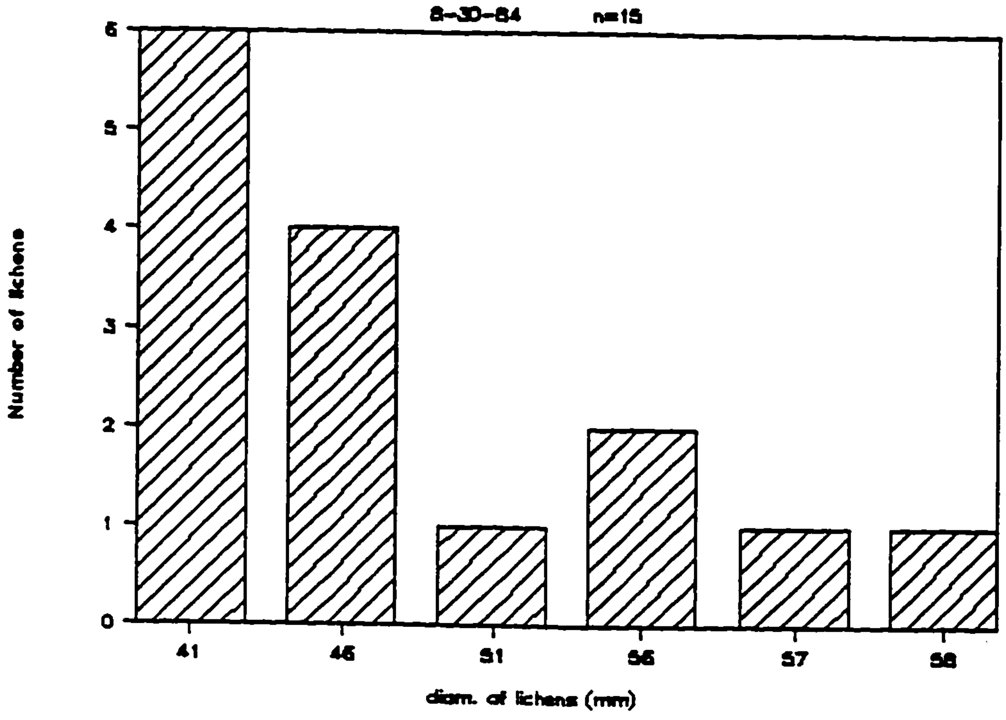

Spencer Glacier—lower trimline—N
max=42 av=41 n=120 8-21-85
Number of lichens
0
10
20
30
40
50
10
15
20
25
30
35
40
45
50
55
60
65
Lichen diameter (mm)


Spencer Glacier—youngest moraine—lake
max=28 av=26 n=150 8-21-85
Number of lichens
0
10
20
30
40
50
60
70
80
90
100
10
15
20
25
30
35
40
45
50
55
60
65
Lichen diameter (mm)

Tebenkov Glac-bedrock knob-outside glac



Willard Is.-bedrx inside limit



KRISTINE J. CROSSEN

CURRICULUM VITAE

1996

BORN: Rhinelander, Wisconsin; June 2, 1949

EDUCATION

Degree	Date	Institution
A.B.D.	Ph.D. expected 1996	University of Washington Geological Sciences
M.S.	1985	University of Maine Quaternary Studies
B.A.	1979	University of Southern Maine Double Major: Geology and Anthropology
	1975-77	University of Alaska, Anchorage
	1967-69	University of Wisconsin, Madison

PROFESSIONAL EXPERIENCE

1988-96	Assistant Professor	University of Alaska Anchorage teaching Physical, Historical, Alaskan geology; Geomorphology; Sedimentation; Permafrost; Field Studies
1985-88	Instructor	University of Alaska Anchorage
1983-85	Teaching Assistant	University of Washington: teaching Physical Geology, Geology of the Northwest, Photogeology; T.A. coordinator
1984	Teacher	Anchorage Community College: teaching Geology 100 (Introductory lecture and lab) and Geology 111 (Physical lecture and lab)
1983	Instructor	University of Southern Maine: teaching Geology 111 (Physical lecture) and Geology 112 (Physical lab)
1982-83	Geologist	Maine Geological Survey: mapping glaciomarine deposits, interpreting Holocene tectonics, mapping lacustrine deposits and synthesizing lake histories

1979-81	Teaching Assistant	University of Maine: teaching Physical, Historical, and Environmental Geology; T.A. coordinator
1980	Research Assistant	University of Maine: analysis of Norwegian Quaternary marine fossils, preparation of radiocarbon samples
1978	Field Assistant	U.S. Geological Survey, Alaska Branch: surficial geology and field cooking, Brooks Range, Alaska
1977	Research Assistant	University of Southern Maine: computer analysis of geological data, King Mountain, South Carolina
1976-77	Field Assistant	University of Alaska, Fairbanks: excavation and field cooking, Dry Creek Archaeological Project, Healy, Alaska

RESEARCH AND FIELDWORK

1993-96	Catastrophic retreat of Portage Glacier, Alaska
1990-95	Geoarchaeology of the Broken Mammoth Site: An early site in the Tanana Valley, Interior Alaska
1990	Sediment Analysis of archaeological samples from SEL-188, for National Parks Service, Exxon Valdez Oil Spill Response Team
1984-92	Holocene Glacial History of the Spencer-Blackstone Ice Complex, Kenai Mountains, Alaska, Ph. D. Research
1987-91	Archaeology of the Northern Beagle Channel, Tierra del Fuego, Argentina, Co-Principal Investigator
1988	Site Formation Processes at the Fox Farm Site, Kachemak Bay, Alaska
1982-83	Holocene Tectonics of Coastal Maine, Glaciomarine Sediment Mapping, Glacial Lake History, Maine Geological Survey
1980-82	Glaciomarine Deltas in southwestern Maine: Formation and Tectonic Movements, M.S. Thesis
1981	Neoglacial chronology and paleoclimatology of Swedish and Norwegian Lapland
1979	Paleoecology of Pleistocene Marine Fossils, Presumpscot Fm., Cumberland, Maine, Senior Thesis

1979-81	Teaching Assistant	University of Maine: teaching Physical, Historical, and Environmental Geology; T.A. coordinator
1980	Research Assistant	University of Maine: analysis of Norwegian Quaternary marine fossils, preparation of radiocarbon samples
1978	Field Assistant	U.S. Geological Survey, Alaska Branch: surficial geology and field cooking, Brooks Range, Alaska
1977	Research Assistant	University of Southern Maine: computer analysis of geological data, King Mountain, South Carolina
1976-77	Field Assistant	University of Alaska, Fairbanks: excavation and field cooking, Dry Creek Archaeological Project, Healy, Alaska

RESEARCH AND FIELDWORK

1993-96	Catastrophic retreat of Portage Glacier, Alaska
1990-95	Geoarchaeology of the Broken Mammoth Site: An early site in the Tanana Valley, Interior Alaska
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1984-92	Holocene Glacial History of the Spencer-Blackstone Ice Complex, Kenai Mountains, Alaska, Ph. D. Research
1987-91	Archaeology of the Northern Beagle Channel, Tierra del Fuego, Argentina, Co-Principal Investigator
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1980-82	Glaciomarine Deltas in southwestern Maine: Formation and Tectonic Movements, M.S. Thesis
1981	Neoglacial chronology and paleoclimatology of Swedish and Norwegian Lapland
1979	Paleoecology of Pleistocene Marine Fossils, Presumpscot Fm., Cumberland, Maine, Senior Thesis

GRANTS, AWARDS, and HONORS

1994	Who's Who Among America's Teachers
1991-3	National Geographic Society grant, Archaeology and Paleo-ecology of the Broken Mammoth Site (Co-P.I. with D. Yesner)
1990-2	National Science Foundation grant, Sediment Analysis and Geoarchaeology of Broken Mammoth site (under RSA with Office of History and Archaeology) (P.I.)
1988	Faculty Senate Research grant, University of Alaska Anchorage, Neoglacial history of the Kenai Mountains
1988	Earthwatch Center for Field Research grant, Archaeology of Northern Beagle Channel, Tierra del Fuego, Argentina (Co-P.I. with D. Yesner)
1988	Faculty Senate Travel Grant, University of Alaska Anchorage
1985-87	Alaska Natural History Association, Neoglacial research, Kenai Mountains, Alaska
1986	Arctic Institute of North America, Neoglacial research, Kenai Mountains, Alaska
1986	Geological Society of America, Neoglacial research, Kenai Mountains, Alaska
1984-85	Ferrel Scholarships, University of Washington, Department of Geological Sciences
1984-85	Mazamas, Neoglacial research, Kenai Mountains, Alaska
1984-85	Corporation Fund, University of Washington, Neoglacial research, Kenai Mountains, Alaska
1984	Sigma Xi, Neoglacial research, Kenai Mountains, Alaska
1981	Scandinavian Exchange Program, University of Maine, Tarfala Research Station, University of Stockholm, Sweden
1981	Faculty Senate Research Fund, University of Southern Maine, Archaeological reconnaissance, Port Heiden, Alaska Peninsula
1980	Graduate Student Board, University of Maine, Lab Training at Smithsonian Radiocarbon Laboratory, Washington, D.C.
1979	Election to *Who's Who Among Students at American Colleges and Universities*
1979	Election to Phi Kappa Phi Honor Society
1977	*Alaska* Magazine Creativity Scholarship Award for Inter-disciplinary Study, University of Alaska, Anchorage

PUBLICATIONS

REVIEWED ARTICLES

Yesner, D.R., and Crossen, K.J., in prep., The Broken Mammoth Site and the Peopling of the Americas, National Geographic Society Research Reports.

Yesner, D.R., and Crossen, K.J., 1996, Pleistocene/Holocene Transition in Interior Alaska, in Meehan, R., ed., Bridges of Science, National Biological Survey Technical Reports.

Yesner, D.R., and Crossen, K.J., 1994, Prehistoric People of Alaska's Interior, in Rennick, J. ed., Prehistoric Alaska, Alaska Geographic, v.21, no. 4, p. 90-93.

Yesner, D.R., Holmes, C.E., and Crossen, K.J., 1993, Archaeology and Paleoecology of the Broken Mammoth Site, Central Tanana Valley, Interior Alaska, USA, Current Research in the Pleistocene, v. 9, p.53-57.

Crossen, K.J., 1991, Structural Control of Deposition by Pleistocene Tidewater Glaciers, Gulf of Maine, In Glacial-Marine Sedimentation; Paleoclimatic Significance, J.G. Anderson and G. Ashley, eds., Geological Society of America Special Paper 261, p. 127-135.

Crossen, K.J. and Banks, A.E., 1991, Sediment Analysis of Archaeological samples from SEL-188, In Site Protection and Oil Spill Treatment at SEL-188, R.C. Betts, C.B. Wooley, C.M. Mobley, J.C. Haggarty, and A. Crowell, A., eds., Exxon Company USA, Anchorage, p. 185-189.

Thompson, W.B., Crossen, K.J., Borns, H.W., Jr., and Andersen, B.G., 1989, Glaciomarine Deltas of Maine and their relation to late Pleistocene-Holocene Crustal Movements, In Neotectonics of Maine - Studies in Seismicity, Crustal Warping, and Sea Level Changes, W.A. Andersen, H.W. Borns, Jr., W.B. Thompson, and J.T. Kelly, eds., Maine Geological Survey Bulletin 40, p. 43-68.

Crossen, K.J., 1987, Portage Lake and Glacier. In Geologic Hazards of Anchorage and Turnagain Arm, Alaska, J.M. Brown, K.J. Crossen, and J. Holtzman, eds., Alaska Geologic Society, Anchorage, p.44-49.

Anderson, W.A., Kelley, J.T., Thompson, W.B., Borns, H.W., Sanger, D., Smith, D.C., Tyler, D.A., Anderson, R.S., Bridges, A.E., Crossen, K.J., Ladd, J.W., Andersen, B.G., and Lee, F.T., 1984, Crustal Warping in Coastal Maine, Geology, v. 12, p. 677-680.

Crossen, K.J., 1983, Glaciomarine Deltas and Moraines, Sebago Lake Region, Maine, Geological Society of Maine Bulletin, No. 3, p. 1-19.

PUBLISHED ABSTRACTS

Crossen, K.J., 1995, Neoglacial dynamics of terrestrial, tide-water, and lacustrine-calving glaciers, Spencer-Blackstone Glacier Complex, Kenai Mountains, Abstracts with Programs, Geological Society of America, v. 27.

Dilley, T.D., Yesner, D.R., Crossen, K.J., and Holmes, C.E., 1995, Pleistocene-Holocene Transition in Interior Alaska, Abstracts with Programs, Geological Society of America, v. 27.

Holmes, C.E., Yesner, D.R., and Crossen, K.J., 1993, Broken Mammoth Archaeological Project: Recent research results, Abstracts with Programs, Society for American Archaeology.

Crossen, K.J., 1993, Little Ice Age history of Portage Glacier and effects of prehistoric travel, Abstracts with Programs, Alaska Anthropological Association, v. 20.

Crossen, K.J., Yesner, D.R., Holmes, C.E., 1992, Late Quaternary Environmental Change and Human Occupation of the Broken Mammoth Site, East-Central Alaska, American Quaternary Association, Abstracts with Programs, v. 12, p. 9.

Crossen, K.J., and Dilley, T., 1992, Late Quaternary Stratigraphy of the Broken Mammoth Site, Tanana Valley, Alaska, Abstracts with Programs, Alaska Anthropological Association, v. 19, p. 34.

Crossen, K.J., and Yesner, D.R., 1991, Stratigraphy and Fossil Assemblages from Broken Mammoth: An 11,500 BP archaeological site in Interior Alaska, Abstracts with Programs, Geological Society of America, v. 23, no. 5, p. 253.

Crossen, K. J., 1990, Neoglacial History of Portage Glacier, Kenai Mountains, Alaska, Abstracts with Programs, Geological Society of America, v. 22, no. 7, p. 177.

Crossen, K.J., Dilley, L., Yesner, D.R., and Workman, W., 1988, Beaches, Volcanoes, and Earthquakes: Processes of site formation at the Fox Farm site, Kachemak Bay, Abstracts with Programs, Alaska Anthropological Association, v. 15.

Crossen, K.J., 1988, Structural Control of Deposition by Pleistocene Tidewater Glaciers, Gulf of Maine, Abstracts with Programs, Geological Society of America, v. 20, no. 7, p. 134.

Thompson, W.B., and Crossen, K.J., 1987, Glaciomarine Deltas in Coastal Maine: Inferences about Neotectonic Events, Abstract Volume, International Quaternary Association Meeting, Ottawa, Canada.

Lowell, T.V., and Crossen, K.J., 1983, Seboomook Lake Area Became Ice Free; But How? In New England Intercollegiate Geologic Conference Guidebook for Greenville - Millinocket, North Central Maine, D.W. Caldwell and L.S. Hanson, eds., Boston, p. 147-174.

Thompson, W.B. and Crossen, K.J., 1981, Preliminary Report on Glacial-Marine Deltas and late Pleistocene-Holocene Crustal Movement in Southwestern Maine, In New England Seismotectonic Studies in Maine During Fiscal Year 1981, (W.B. Thompson, ed.), Augusta; Maine Geological Survey, Nuclear Regulatory Commission Report, pp. 97-107.

MONOGRAPHS

Crossen, K.J., in prep., Neoglacial dynamics of terrestrial, tidewater, and calving lacustrine glaciers, Blackstone-Spencer Ice Complex, Kenai Mountains, Alaska, Ph. D. dissertation, Geological Sciences, University of Washington.

Crossen, K.J., 1995, Little Ice Age landforms and glacial dynamics in Portage Valley and Portage Pass, Alaska Geological Society, Anchorage, 46 p. (2nd ed.)

Crossen, K.J., 1992, Little Ice Age landforms and glacial dynamics in Portage Valley and Portage Pass, Alaska Geological Society, Anchorage, 46 p.

Brown, J.M., Crossen, K.J., and Holtzman, J., eds., 1987, Geologic Hazards of Anchorage and Turnagain Arm, Alaska, Alaska Geologic Society, Anchorage, 62 p.

Crossen, K.J., 1986, Holocene Glacial Variations, Kenai Mountains, Alaska, Report to Alaska Natural History Association, Anchorage, 10 p.

Crossen, K. J., 1985, Glaciomarine Deltas in Southwestern Maine: Formation and Vertical Movements, M.S. thesis, University of Maine, Orono, 124 p.

Thompson, W.B., Crossen, K.J., Borns, H.W., Jr., and Andersen, B.J., 1983, Glacial-Marine Deltas and Late Pleistocene-Holocene Crustal Movements in Southern Maine, Maine Geological Survey Open File Report 83-3, Augusta, 18 p. and map.

Crossen, K. J., 1983, Paleoecology of Pleistocene marine fossils, Presumpscot Fm., Cumberland, Maine, Abstracts with Programs, Geological Society of America, v. 15, no. 4.

Crossen, K.J., 1982, Deglaciation of Southwestern Maine; Inferences from glaciomarine Deltas, Abstracts with Programs, Geological Society of America, v. 14. no. 3.

Thompson, W.B. and Crossen, K.J., 1981, Glacial-Marine Deltas as Indicators of Late-glacial Sea Level in Central and Southwestern Maine, Abstracts with Programs, Geological Society of America, v. 13, no. 3, p. 180.

UNIVERSITY AND PROFESSIONAL SERVICE

1995-96	Vice-President, Alaska Geological Society
1993-96	Chair, Geology Department, University of Alaska Anchorage
1991-96	Board of Directors, Southcentral Alaska Museum of Natural History
1992-96	Chair, Field Trip Committee, Alaska Geological Society
1990-96	Continuing Education Committee, Alaska Geological Society
1992-96	Faculty Advisor, Geology Student Club
1994-95	Cooperative Education Advisory Committee
1992-93	Chair, Board of Directors, Southcentral Alaska Museum of Natural History
1994-96	Guest Lecturer, Eagle River Visitors Center, Chugach State Park
1993-94	Guest Lecturer, Alaska Women in Science, Susitna Girl Scout Council
1993	Guest Lecturer, Alaska Archaeology Week, National Park Service
1992	Field Trip Leader, International Conference on Arctic Margins, Portage Glacial Dynamics, Portage, Alaska
1991-2	Board of Directors, Alaska Geological Society
1991-2	Community Advisory Committee, Geology Dept., University of Alaska Anchorage
1990-1	Natural Science degree committee, CAS, University of Alaska Anchorage

1990-2	Guest Lecturer, Alaska Geological Society, Anchorage
1990	Guest Lecturer, Alaska Cornell Club, Portage
1988-89	Chair, Geology Department, University of Alaska Anchorage
1989-90	Consultant, Begich-Boggs Visitor Center, Portage, Moraine Trail exhibits
1988	Member, Merger Committee on Support Personnel, University of Alaska Anchorage
1988	Member, CAS Curriculum Committee, University of Alaska Anchorage
1988	Candidate, Board of Councilors, American Quaternary Association
1987-88	Co-Chair, Geology Department, University of Alaska Anchorage
1985-95	Organizer and facilitator, Geology Guest Lecture Series, National and International Speakers Program, University of Alaska Anchorage
1985-95	Guest Lecturer, Begich-Boggs Visitor Center, Portage, U.S. Forest Service Interpreter Training Program
1985-95	Academic Advising, Geology Department, University of Alaska Anchorage and Anchorage Community College
1988	Guest Lecturer, Alaska Native Plant Society, Anchorage
1987	Exhibit Organizer, Begich-Boggs Visitor Center, Portage, Lichenometry and Dendrochronology of Glacial Deposits
1987-95	Guest Lecturer, Anchorage School District classes
1986	Field Trip Leader, Alaska Geological Society, Geologic Hazards of Anchorage and Turnagain Arm, Alaska
1985-86	Natural Science Representative, Learning Center Advisory Committee, Anchorage Community College
1984-85	Exhibit Consultant on Glacial Phenomena, Begich-Boggs Visitor Center, Portage, U.S. Forest Service
1983	Field Trip Leader, Geological Society of Maine, Glaciomarine Deposits, Sebago Lake, Maine
1983	Field Trip Leader, New England Intercollegiate Geolgical Conference, Seboomook Lake, Maine

PROFESSIONAL MEMBERSHIPS

Geological Society of America
American Quaternary Association
American Geophysical Union
Alaskan Geological Society
Alaska Ground Water Association
American Women in Geosciences
Honor Society of Phi Kappa Phi

PROFESSIONAL REFERENCES

Stephen C. Porter, Quaternary Research Center, University of Washington

Stanley C. Chernicoff, Geological Sciences, University of Washington

Linda Brubaker, Forest Resources, University of Washington

Harold W. Borns, Jr., National Science Foundation, Polar Programs, Washington, D.C.

Thomas B. Kellogg, Institute for Quaternary Studies, University of Maine

Anne Pasch, Geology Department, University of Alaska Anchorage

Gail Ashley, Geology Department, Rutgers University

Thomas Hamilton, U.S. Geological Survey, Alaska Branch, Anchorage