Tree-ring-dated 'Little Ice Age' histories of maritime glaciers from western Prince William Sound, Alaska

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Received 25 June 1997; revised manuscript accepted 10 May 1998

A HOLOCENE RESEARCH PAPER

Abstract: Tree-ring studies at 13 glacier forefields in western Prince William Sound show 'Little Ice Age' glacial fluctuations were strongly synchronous on decadal timescales. Cross-dated glacially overrun trees at eight sites indicate ice margins advanced in the early (late twelfth through thirteenth centuries) and middle (seventeenth to early eighteenth centuries) 'Little Ice Age'. Tree-ring dates of 22 moraines at 13 glaciers show two main periods of stabilization. The earlier of these, in the first decades of the eighteenth century, overlaps with the second period of glaciers overrunning trees and marks culmination of this middle 'Little Ice Age' expansion. Stabilization of moraines on nine of the study forefields in the latter part of the nineteenth century delineates a third interval of 'Little Ice Age' glacial advance. The detailed 'Little Ice Age' record from land-terminating glaciers in western Prince William Sound is consistent on a timescale of decades with four other tree-ring-dated glacial histories from across the northern Gulf of Alaska. This coastal northeastern Pacific glacial record reveals the structure of the 'Little Ice Age' in the region and provides a strong basis for comparison with other proxy climate records spanning the past 1000 years.

Key words: 'Little Ice Age', glacier fluctuations, dendrochronology, glacier foreland, moraine chronology, Alaska, Prince William Sound, late Holocene.

Introduction

Records of glacial fluctuations can provide insights into decadal to centennial scale climate variations and their causes. In particular, such records from the boreal regions of the North American Cordillera potentially represent some of the most precisely dated records because histories can frequently be dated using tree-ring analysis of trees overrun by ice. This technique offers several advantages over radiocarbon ages of forest debris incorporated in glacial deposits. Luckman (1995; 1996) and Wiles and Calkin (1994) have shown that in many cases a calendar date can be assigned to a glacial advance based on dating trees affected by a glacial process. Where many subfossil trees are preserved on a glacial forefield, the strength of this technique lies in the replication of multiple calendar dates on the logs that enable past glacier movements to be reconstructed with great precision (Luckman, 1995; 1996; Barclay *et al*., 1996). In addition to dating glacial changes, tree-ring series derived from subfossil trees can also be used to extend living tree-ring chronologies back in time. These provide an independent records of past climate variation that can be compared with the glacial record (Luckman, 1996).

In this paper we present a late Holocene glacial chronology developed from the forefields of 13 glaciers in the western Prince William Sound area, south-central Alaska (Figure 1). These histories are primarily based on a total of 130 subfossil logs that have been cross-dated to determine times of glacial advance. The dendroclimatic significance of the resulting 1119-year tree-ring-width chronology is discussed in Barclay *et al*. (1999). Synthesis of the dendroclimatic and glacial records is related to ongoing work on multiproxy records of climate history from coastal sites along the northeastern Pacific rim (Wiles *et al*., 1998; Wiles, 1997).

Setting

Much of the Prince William Sound region is an archipelago rimmed by mountains and dissected by a system of radiating fjords. The study area is located in the western portion of the Sound between the towns of Seward and Whittier (Figure 1). These fjords were occupied by ice during the Pleistocene and some still maintain iceberg calving margins. Glacier mass balance variations at these coastal sites integrate North Pacific ocean-atmosphere climate. The climate is maritime with air-sea circulation dominated by the Alaska gyre driven in spring and winter by the Aleutian Low (Wilson and Overland, 1987). Seward (Figure 1) has a mean temperature of −4°C in January and 13.2°C in July with an annual precipitation of over 1700 mm averaged for

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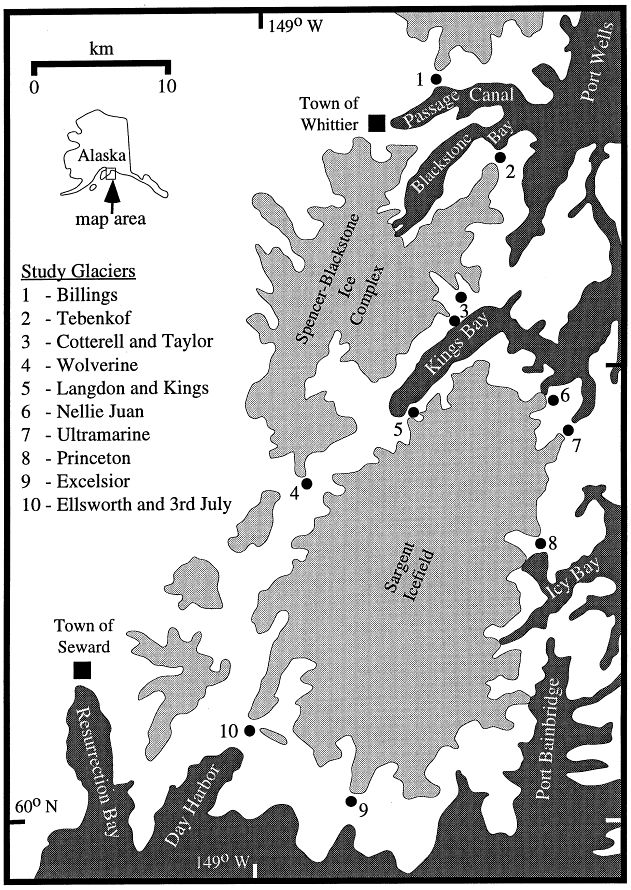

Figure 1 Location map of study glaciers in western Prince William Sound region, Alaska.

the period 1908 to 1997 (unpublished data from Alaska Climate Research Center, 1997). Temperatures at Whittier are similar but with generally higher annual precipitation (Farr and Hard, 1987).

The broad intermontane regions of the northern Kenai and Chugach Mountains coupled with the abundant precipitation present an ideal environment for glaciers to develop and persist. Most of the study glaciers are outlet ice tongues draining the Spencer-Blackstone and Sargent Icefields (Figure 1). With the exception of Princeton, Wolverine and 3rd July glaciers, ice margins presently reach to within 150 m of sea level. Only Nellie Juan Glacier currently has a tidewater calving terminus, both Princeton and Taylor glaciers formerly reached tidewater at their recent maxima. None of the study glaciers are known to have surged.

Firn limits were estimated from 1950 photographs to be between 350 and 500 m for Ultramarine and Nellie Juan glaciers (Mercer, 1960). Field (1975) estimated firn limits from the 1950 photos to generally rise southward from 400 m at Tebenkof Glacier in the north to between 600 and 700 m at Excelsior Glacier and as high as 900 m at Ellsworth Glacier in the south (Figure 1). Firn limit estimates based on 1960s aerial photography by the US Geological Survey (A. Post, unpublished data) show similar levels as in 1950 with some values 100 m higher (Field, 1975).

North Pacific coastal forests fringing the icefields are primarily composed of mountain hemlock (*Tsuga mertensiana*) together with Sitka spruce (*Picea sitchensis*) and western hemlock (*Tsuga heterophylla*). Areas of open parkland with pure stands of mountain hemlock are found at the coast near Nellie Juan Glacier and throughout the region at the altitudinal tree-line (Cooper, 1942; Viereck and Little, 1972). Deglaciated zones are initially colonized by dense thickets of alder and willow before development of conifer forest.

Previous glacial investigations

Late Pleistocene ice filling Prince William Sound wasted from the inner fjords around 10 000 yr BP. The timing of this is inferred from bog-bottom dates of 9440 yr BP from Perry Island located 25 km west of Tebenkof Glacier (Heusser, 1983), 9330 yr BP from Columbia Glacier 70 km east of the study area (Post, 1975), and 10 015 yr BP from Port Wells (Heusser, 1983) in the north-east corner of the study area (Figure 1). Little is known of subsequent Holocene glaciation in western Prince William Sound except for fluctuations during the past century. However, Heusser (1983) notes Neoglacial activity in northern Prince William Sound between 3200 and 2500 yr BP and into recent centuries, whereas Wiles and Calkin (1994) identified episodes of glacial advance at 3600 yr BP, AD 600 and from AD 1300 to 1850 in the Kenai Mountains immediately southwest of the current study area.

Numerous observations of glacial movements during the past century have been undertaken in western Prince William Sound. The earliest scientific visits to the area were those in 1908 and 1909 (Grant and Higgins, 1913) and in 1910 by Tarr and Martin (1914). Field (1933; 1937; 1975) made systematic observations of glacier termini in 1931 and 1935, and again in 1956 and 1966. Cooper (1942) accompanied Field in 1935 and suggested a generalized Holocene glacial history for the region. Viereck (1967) cored trees on moraines at Tebenkof, Taylor, Nellie Juan and Ultramarine glaciers to estimate times of recent glacial maxima. Austin Post of the US Geological Survey began photographing the glaciers of coastal Alaska on a semi-annual basis in the early 1960s. He also summarized the recent glacial history of Blackstone and Beloit glaciers in Blackstone Bay (Figure 1) as part of a bathymetric investigation of the fjord (Post, 1980). Lethcoe (1987) presents many photographs in a comprehensive observers guide to the glaciers of Prince William Sound.

Dating periods of glacial advance and retreat

The dating of periods of glacier advance was primarily accomplished using dendrochronology. Mapping of glacial forefields often led to the discovery of extensive subfossil forests that have been revealed during ice retreat over the twentieth century. Those trees that were buried in till against mid-valley bedrock protuberances in near-growth position are considered to give the best estimate for times of glacial advance. Some stumps preserved in growth position may have originally been encased in outwash sediments and thus may have been killed prior to arrival of the ice margin at that site. Cores or discs were removed from the subfossil trees, which consisted primarily of mountain hemlock with minor amounts of Sitka spruce and western hemlock.

Initial cross-dating focused on development of internally consistent floating chronologies of samples collected at each glacier forefield. Ring widths were measured to the nearest 0.001 mm and the XDATE routine (E. Cook, unpublished) used to suggest possible cross-date positions. Confirmation of cross-dating was done with the COFECHA routine (Holmes, 1983) and by visual comparison of marker rings on the cores and log sections (Stokes and Smiley, 1968). The floating chronologies for each site were then cross-dated with each other and with calendar-dated tree-ring chronologies developed from living mountain hemlock growing on the forefields of Tebenkof and Nellie Juan glaciers. Further confirmation of dating was done by comparing these data with existing chronologies from the Gulf of Alaska, including a 716-year living tree-ring chronology from mountain hemlock 350 km to the east at Icy Bay (G.C. Jacoby, unpublished data).

A total of 123 subfossil logs have been cross-dated with living

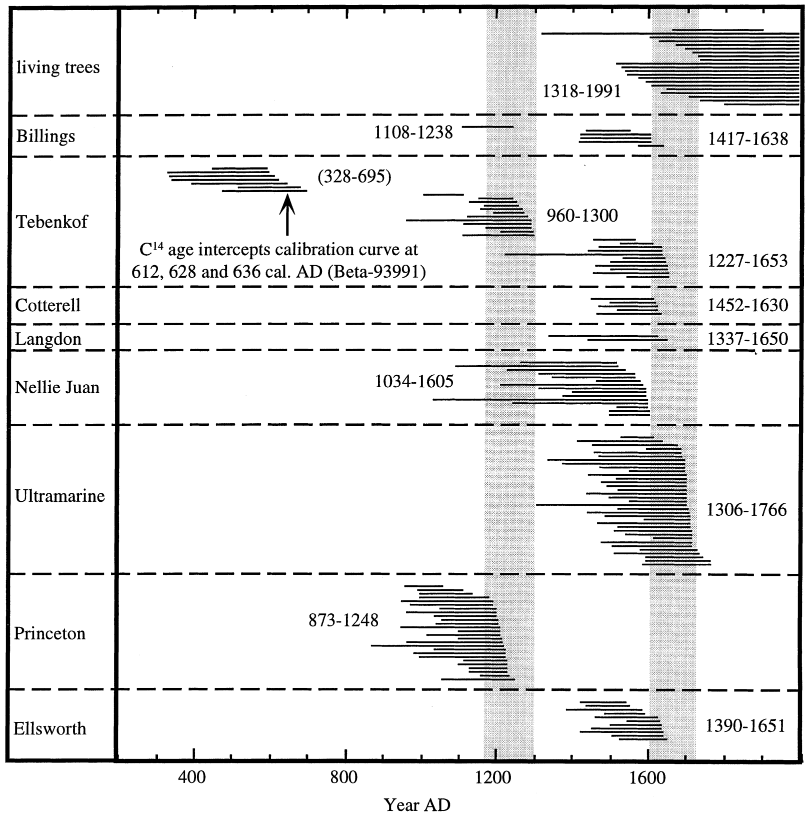

Figure 2 Lifespans of cross-dated, glacially killed subfossil trees from western Prince William Sound. The living trees were two stands of mountain hemlock growing near Tebenkof and Nellie Juan glaciers. Dates indicate the timespans covered by each population of logs. General times of glacier advance in western Prince William Sound are shaded. The early, floating ring-width series from Tebenkof Glacier is fixed in time by a radiocarbon age that intercepts the calibration curve at three points (see Table 1).

trees (Figure 2), providing a composite tree-ring chronology spanning AD 873 to 1991 (Barclay *et al.*, 1999). In addition to this calendar-dated series, seven logs from Tebenkof Glacier form a floating chronology tied to a radiocarbon age showing tree death during the seventh century.

Radiocarbon ages are mostly used in this study as an independent verification of the dates assigned to glacial advances by the tree-ring cross-dating. Wood samples were taken from the outer rings of subfossil logs to date the time of tree death as closely as possible. All radiocarbon data compiled for the area (Table 1) have been calibrated using the program CALIB 3.0 (Stuiver and Reimer, 1993) and ages are discussed below as AD dates.

The ages of living trees growing on moraines and other geomorphic surfaces were used to provide minimum dates of ice retreat (Lawrence, 1950). Our field sampling was carried out to minimize inaccuracies inherent in this technique. See Wiles *et al.* (1996) for a discussion of these potential problems. Trees were cored at the base of the stem to minimize sampling height errors; although no correction was applied if the pith was missed, multiple cores were taken to minimize this uncertainty. An ecesis time (the interval for tree establishment on a recently deglaciated surface) of 15 years was used in this study. This estimate is based on our work from the forefield of Taylor Glacier. It is consistent with the work of Viereck (1967) who found the ecesis time in Prince William Sound varied from six to 20 years, and work by

Table 1 Radiocarbon ages from western Prince William Sound glacier forefields

Laboratory number[a]	Age (in yrs BP)	Calibrated interval[b]		Glacier[c]	Reference
		(years AD)	(years BP)		
***	690 ± 95	1210 (1297) 1440	740 (653) 510	NJ	Field (1975)
UW-518	680 ± 60	1260 (1290) 1410	690 (656) 540	PR	Post (personal communication)
UW-519	860 ± 75	1000 (1210) 1290	950 (740) 660	PR	Post (personal communication)
UW-520	845 ± 70	1020 (1210) 1290	930 (736) 660	TB	Post (personal communication)
BETA-55627	490 ± 70	1320 (1430) 1620	630 (517) 330	TB	This study
BETA-55628	200 ± 70	1520 (1680, 1780, 1800, 1950) 1955	430 (279, 169, 151, 5, 0) 0	UL	This study
BETA-93991	1460 ± 70	430 (612, 628, 636) 690	1520 (1338, 1322, 1314) 1260	TB	This study

[a]Laboratories: UW, University of Washington; BETA, BETA Analytic; ***, laboratory unknown.
[b]Determined from dendrocalibrated data using CALIB 3.0 (Stuiver and Reimer, 1993). Presented as calibration curve intercept(s) with two standard deviation range. Lab error multiplier used = 1.
[c]NJ = Nellie Juan Glacier; PR = Princeton Glacier; TB = Tebenkof Glacier; UL = Ultramarine Glacier.

Wiles and Calkin (1994) in the southern Kenai Mountains immediately southwest of the current study area where an estimate of 15 years was used. All moraine dates in this study (Figure 3) are presented in years AD, and were calculated by subtracting 15 years from the date of the innermost ring of the oldest year.

Individual glacier histories

The record presented here focuses on the ‘Little Ice Age’ (LIA) history which we broadly define based on our glacial geologic evidence as the twelfth through nineteenth centuries. The study sites are discussed in a north to south sequence (see Figure 1).

Billings Glacier

Billings Glacier, 7 km northeast of the town of Whittier (Figure 1), is the most northerly of the study glaciers. The earliest evidence of glacial advance comes from six detrital logs found partially encased in modern alluvium within the limit of the most recent LIA advance (location 1, Figure 4). Cross-dating with the regional master chronology shows that one of the logs died in AD 1238 and four others died in the first half of the seventeenth century (Figure 2). The sixth log is badly abraded and so its last discernible ring of AD 1549 is an unreliable estimate of when the tree died. Because these tree kill dates are from detrital logs and their place of origin is uncertain they cannot stand alone as times of glacial advance. However, when considered together with the timing of glacial advances at other sites in the region, the dates are suggestive of increased activity of Billings Glacier in the thirteenth and seventeenth centuries.

A large bedrock knob splits the forefield of Billings Glacier with an active outwash channel to the west and an abandoned channel to the east (Figure 4). A set of nested terminal moraines, although discontinuous and difficult to correlate, records the most recent fluctuations of the terminus. The oldest tree found on the outermost ridge in the western pass was 235 years old which, with an ecesis estimate of 15 years, yields a date of AD 1742 (Figure 3). This moraine can be traced to a trimline on the west valley wall above which mountain hemlock have been growing since at least AD 1587. Trees growing on the outer moraine to the east of the bedrock knob are generally younger than those to the west. The next two ridges nested within the outermost moraine are dated at AD 1818 and AD 1850, whereas the innermost ridge of note, a feature locally attaining 10 m in relief, stabilized by AD 1894 (Figures 3 and 4).

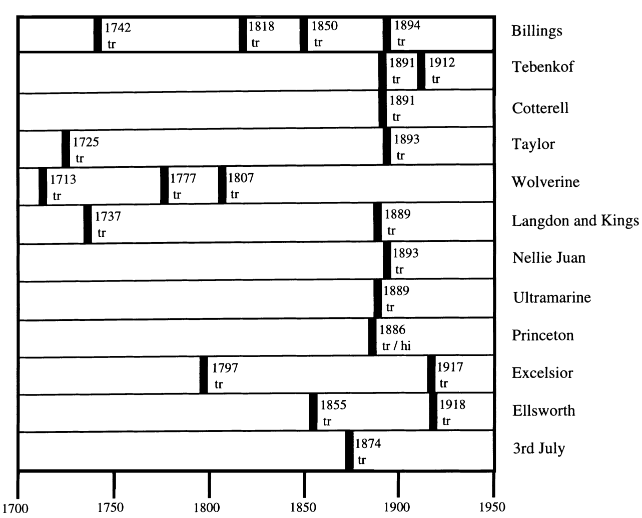

Figure 3 Dates of moraine stabilization for western Prince William Sound. Dates were determined using tree-ring (tr) and historical (hi) data.

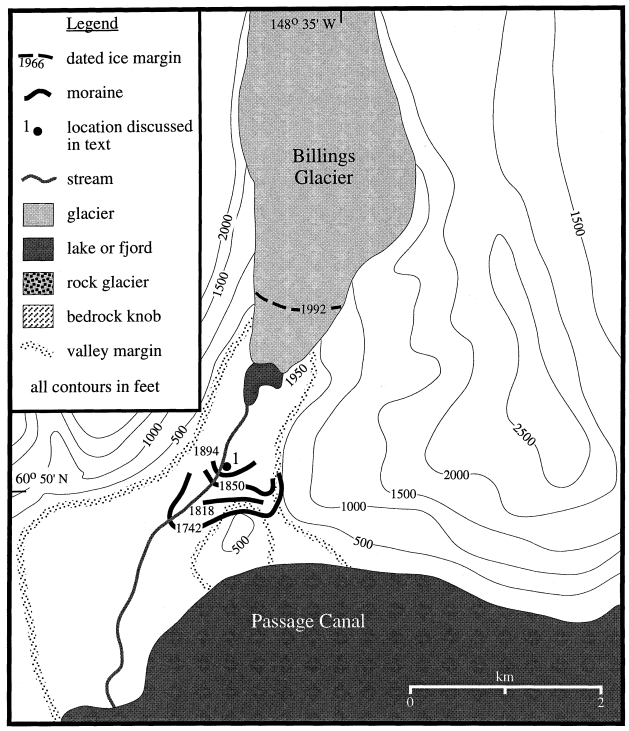

Figure 4 Billings Glacier. Base map is US Geological Survey 1:63 360 series, Seward (D-5), 1951, minor revisions 1966.

Field (1975) estimated that Billings Glacier had retreated more than 1.5 km from its position in 1910 when observed by Tarr and Martin (1914). Aerial photographs show that a proglacial ice-contact lake had developed by 1972. Our visit in 1992 confirms that retreat has continued with a granite ledge, reported as emerging from the wasting ice tongue by Field (1975) and Lethcoe (1987), now almost fully revealed. Retreat from the 1894 position has amounted to about 2.2 km by 1992.

Tebenkof Glacier

Tebenkof Glacier is a 12.5-km-long valley glacier draining northeastwards towards Blackstone Bay (Figures 1 and 5). Subfossil trees were first noted here in 1935 by Field (1937) and Cooper (1942). These logs are scattered across the forefield in till and outwash deposits, glacially pushed into bedrock knobs, and reworked into modern stream channels. Differentiation of episodes of glacial advance based on stratigraphy alone cannot be achieved; however, the strong clustering of tree kill dates into three age groups (Figure 2) suggests three distinct times of advance.

The oldest group of seven logs were found between locations 2 and 3 (Figure 5). These cross-date together and, on the basis of a radiocarbon age of AD 628 (Beta-93991, Table 1), grew between AD 328 and 695 (Figure 2). Outer rings range in age from AD 590 to 695 but are generally rotted and abraded; therefore they only provide minimum dates of death by glacial activity.

Evidence of a second advance is derived from 12 logs that have been cross-dated with living trees as part of the regional composite chronology. These grew between AD 960 and 1300 (Figure 2) and presumably correspond with the calibrated radiocarbon age of AD 1210 (UW-520, A. Post, personal communication, 1993; Table 1) from a log collected on the surface of the outwash plain. The outer rings of five of the better preserved logs date between AD 1289 and 1300, suggesting death by glacial overriding at this time. Six of the logs were found in a boulder deposit (location 1, Figure 5) which may be the eroded remnant of a terminal moraine. That location 1 is the approximate limit of this early LIA advance is supported by a log found at location 2 (Figure 5); this

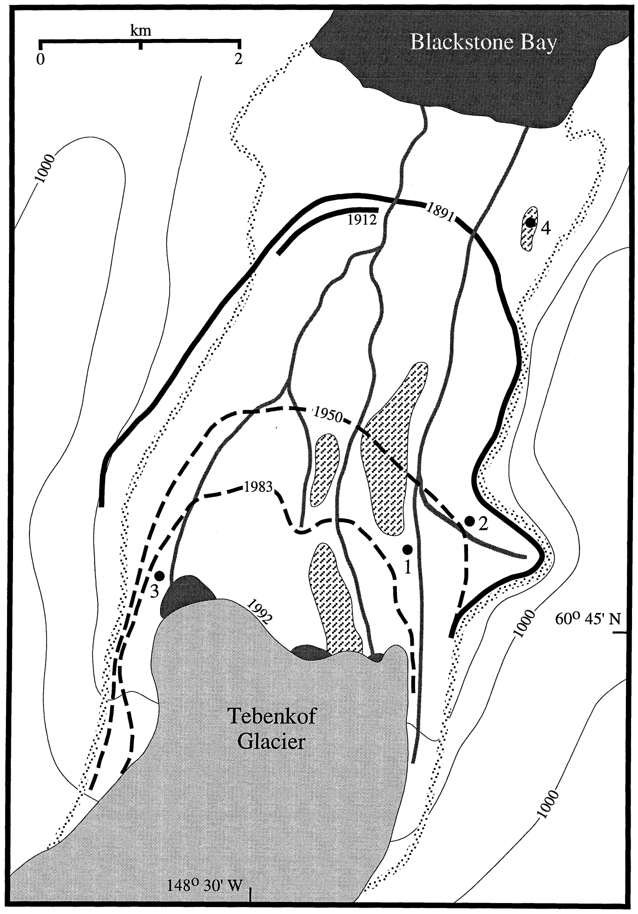

Figure 5 Tebenkof Glacier. Base maps are US Geological Survey 1:63 360 series, Seward (C-4, C-5, D-4), 1951, minor revisions 1983. See Figure 4 for legend.

germinated by AD 1227, over 60 years before the apparent culmination of this thirteenth-century advance. Subsequent recession after the AD 1290 advance is inferred from a log found at location 3 (Figure 5) which was growing between AD 1501 and 1644.

Nine of the last group of 11 cross-dated logs have kill dates of between AD 1633 and 1653 (Figure 2) suggesting another glacial advance at this time. Two of these logs are from location 2 (Figure 5) where they were pushed by advancing ice into a till-veneered bedrock knob. A calibrated radiocarbon age of AD 1430 from one of these logs (Beta-55267, Table 1) is consistent with the cross-dating that indicates the tree was alive between AD 1227 and 1639.

Two end moraines were identified on the Tebenkof forefield (Figures 3 and 5). The oldest tree sampled on the outer moraine suggests that ice retreated from this surface before AD 1891. This is consistent with the conclusions drawn by Viereck (1967) who suggested that the outwash beyond the moraine had stabilized between AD 1875 and 1885. An inner moraine tree-ring dated to AD 1912 indicates readvance or stillstand early in this century. The site of living mountain hemlock used to develop the 1119-year composite tree-ring chronology is a bedrock knob (location 4, Figure 5) just outside the outermost moraine. One tree at this site with severely suppressed growth had at least 803 rings, suggesting that the AD 1891 maximum was the greatest extent of Tebenkof Glacier since at least AD 1189.

Our moraine dates are consistent with observations by Tarr and Martin (1914) who in 1910 counted rings in willows and alder on the terminal moraines and found the oldest to be only 18 years old. In 1909 the ice terminus was within 150 m of the terminal moraine (Grant and Higgins, 1913); it was at essentially the same position in 1910 (Tarr and Martin, 1914), had retreated 300 m by 1935 (Field, 1937) and another 500 m by 1964 (Field, 1975). In 1992 the terminus was situated about 2.7 km from the AD 1891 terminal moraine.

Cotterell and Taylor Glaciers

Cotterell and Taylor Glaciers occupy parallel valleys draining southeast towards Kings Bay (Figures 1 and 6). At Cotterell, five transported logs buried in till and outwash within the LIA limit (location 1, Figure 6) suggest glacial advance into forest between AD 1611 and 1630 (Figure 2). The LIA maximum of Cotterell Glacier is recorded by a terminal moraine 3.1 km inland from Kings Bay on which the oldest tree cored indicates a moraine age of AD 1891 (Figure 3).

Two advances of Taylor Glacier are inferred from arcuate terminal moraines that loop from prominent lateral moraines and trimlines along the valley walls to maximum downvalley positions in shallow fjord waters at the valley centre (Figure 6). Trees cored on the outer moraine suggest stabilization by AD 1725, whereas the inner moraine is dated to AD 1893 (Figure 3). Viereck (1967) had previously suggested a single LIA maximum of Taylor Glacier between AD 1865 and 1895. Our results are compatible with the description by Grant and Higgins (1913) who in 1909 observed the terminus reaching to tidewater at the valley axis and a bare zone on the valley walls indicating about 0.4 km of retreat since the most recent maximum. A further 0.8 km of recession had occurred by 1964 (Field, 1975) and in 1993 the terminus of Taylor Glacier was located about 2 km behind the LIA maximum position.

Wolverine Glacier

Wolverine Glacier (Figure 1) is the highest altitude valley glacier studied with its terminus currently lying just above treeline at 400

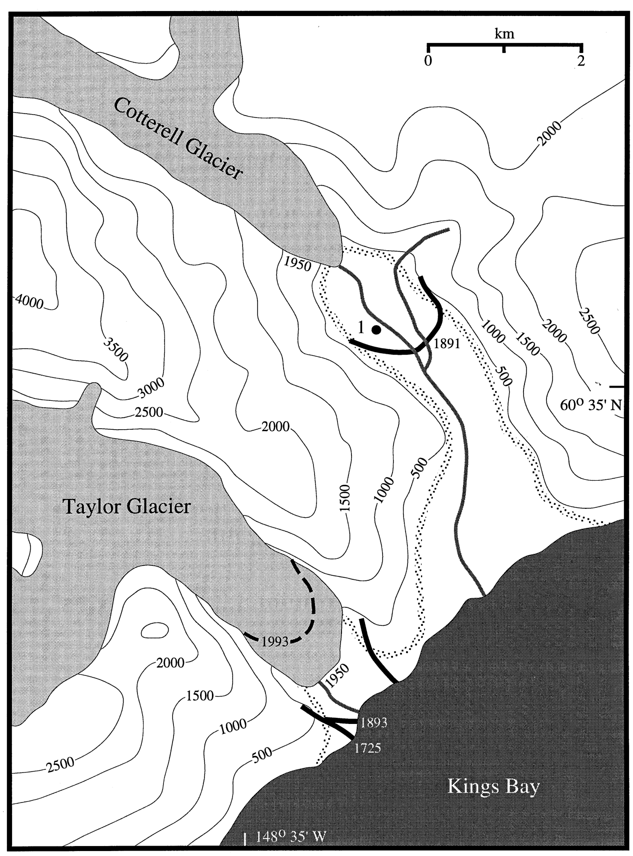

Figure 6 Cotterell and Taylor Glaciers. Base map is US Geological Survey 1:63 360 series, Seward (C-5), 1951, minor revisions 1970. See Figure 4 for legend.

m elevation (Figure 7). Only the southwestern side of the forefield was visited during the 1992 field season.

Two discontinuous lateral moraines were recognized between 750 and 550 m elevation. A better defined moraine complex in front of the terminus (Figure 7) is below tree-line and so can be dated using tree-rings. Tree-ring ages suggest the outer pair of ridges stabilized in AD 1713 and 1777 respectively while the third, innermost ridge was dated to AD 1807 (Figure 3).

Mass balance studies (Mayo and Trabant, 1984; Mayo *et al.*, 1985; Mayo and March, 1990) have been undertaken on Wolverine Glacier by the US Geological Survey. Their analyses show a positive balance following an interval of increased winter temperature and corresponding increase in winter precipitation beginning in the mid-1970s. Little change was noted in the position of the glacier snout which continued its general retreat.

Langdon and Kings Glaciers

Langdon Glacier and its former tributary, Kings Glacier, drain northward towards the southern shore of Kings Bay (Figures 1 and 8). Two subfossil logs from within the LIA glacial limit were cross-dated with the regional master chronology (Figure 2). The first log (location 1, Figure 8) was rooted in place in the modern outwash stream whereas the second log (location 2, Figure 8) had been reworked into modern alluvium. The respective outer ring dates of AD 1624 and 1650 suggest advance of Langdon and/or Kings Glaciers during the seventeenth century.

Outside the LIA moraines of Langdon and Kings Glaciers is open parkland where trees up to 450 years old were cored. The outer moraine of Langdon Glacier has up to 30 m of relief and hosts trees that indicate stabilization by AD 1737. Nested within this ridge is a smaller moraine with a tree-ring age of AD 1889 (Figure 3).

Langdon Glacier has retreated at least 2 km from the AD 1889 moraine. Alders estimated to be 10 to 20 years old were observed being overrun in 1992, suggesting recent thickening and advance of the terminus.

Nellie Juan Glacier

Nellie Juan Glacier flows east–northeast from the northern end of the Sargent Icefield (Figure 1) to terminate in a marine lagoon (Figure 9). Nellie Juan is the only glacier in this study which presently reaches to tidewater as an iceberg-calving terminus. Glaciers of this type have previously been shown to undergo non-climatically induced fluctuations in addition to the climatically driven fluctuations that typify land-terminating, non-surging glaciers (Post, 1975; Mann, 1986; Wiles *et al.*, 1995).

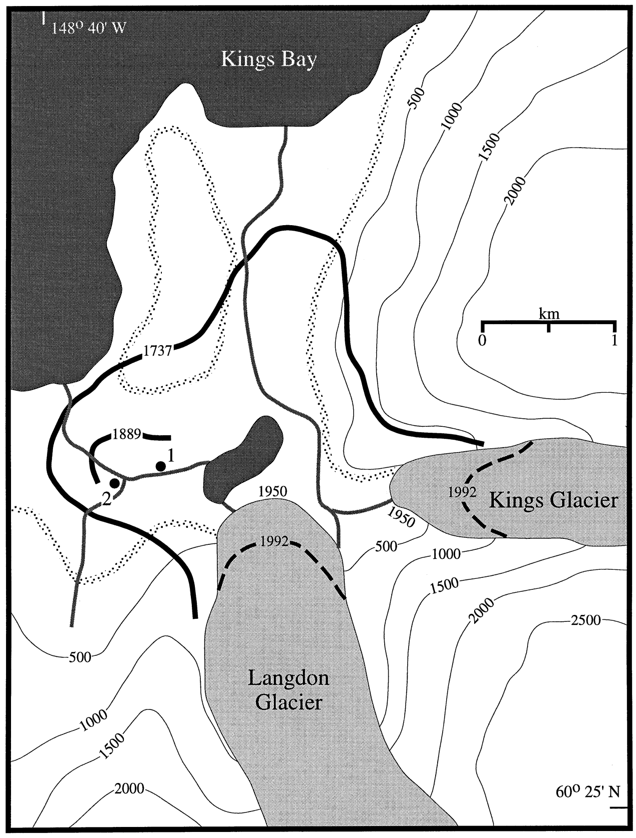

Figure 8 Langdon and Kings Glaciers. Base map is US Geological Survey 1:63 360 series, Seward (B-5), 1951, minor revisions 1973. See Figure 4 for legend.

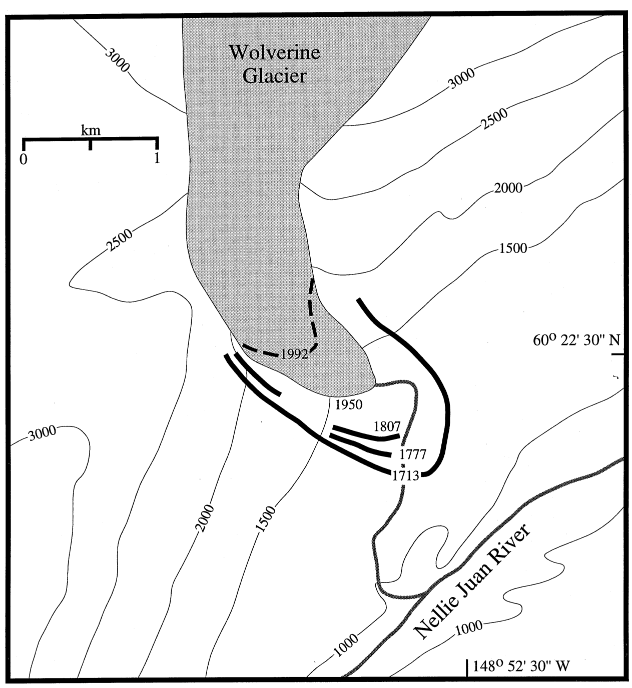

Figure 7 Wolverine Glacier. Base map is US Geological Survey 1:63 360 series, Seward (B-6), 1951, minor revisions 1973. See Figure 4 for legend.

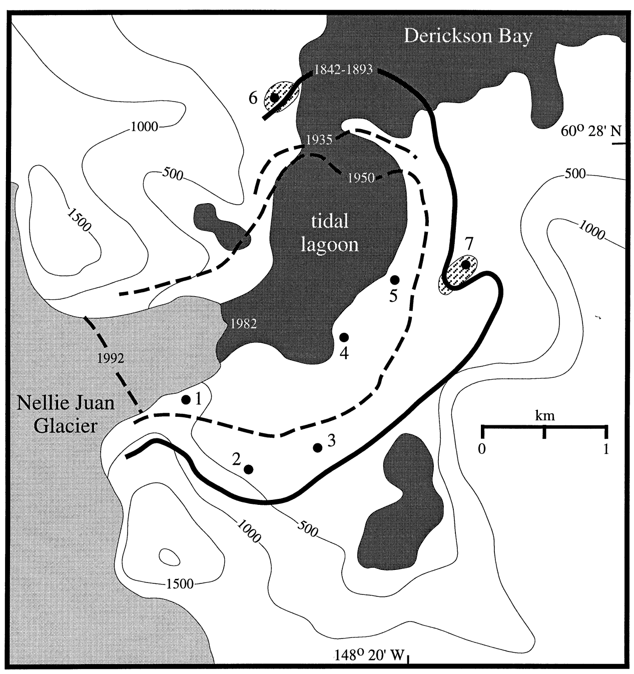

Figure 9 Nellie Juan Glacier. Base map is US Geological Survey 1:63 360 series, Seward (B-4), 1951, minor revisions 1988. See Figure 4 for legend. 1935 ice margin from Field (1937).

Fifteen glacially overrun trees found along the southern shore of Nellie Juan Lagoon have been cross-dated with the regional master chronology to constrain glacial fluctuations (Figure 2). Most were stumps preserved *in situ* or as complete trunks pushed into bedrock knobs with little transport from the site of growth. At location 1 (Figure 9) trees were growing between AD 1092 to 1539 requiring the ice margin to be at or behind this site during that period. The ring-width record at location 1 shows strong suppression immediately prior to the last year of growth in AD 1539, indicating advance of Nellie Juan Glacier was adversely affecting growth before overrunning the growth site. Lateral spreading and continued seaward advance of the ice margin is shown by the near simultaneous death of trees between AD 1594 and 1599 at locations 2 to 4 (Figure 9) and by logs found at location 5 that died after AD 1605.

Living trees on and around a bedrock knob on the west shore of Derickson Bay (location 6, Figure 9) constrain the LIA maximum of Nellie Juan Glacier. We identified several moraines in this area and the oldest tree growing on the outer ridge indicates stabilization by AD 1842. Viereck (1967) identified two trees nearby that had been pushed over by the glacier in AD 1863 and 1880, respectively. In 1992 we relocated one of these trees and concur with Viereck's interpretations. In addition, we found a tree that germinated by AD 1720 beyond the moraine that was damaged by boulders rolling from the moraine about AD 1890. Trees on the innermost ridge on the bedrock knob indicate stabilization by AD 1893 (Figure 9). Considered together these results suggest that Nellie Juan Glacier reached its late LIA maximum by AD 1842 and oscillated at this position until general retreat began around AD 1893.

Ring-width series from living trees used in developing the composite tree-ring chronology were taken from mountain hemlock growing on a bedrock knob on the east side of Nellie Juan lagoon (location 7, Figure 9). The oldest tree found had an inner ring date of AD 1323 which, when considered with the subfossil trees at location 1, indicates that the 1842 to 1893 maximum was the most extensive advance of Nellie Juan Glacier in at least the last 900 years.

In 1908 and 1910 the terminus of Nellie Juan Glacier was grounded on a shoal in Derickson Bay and a bare zone 30 to 150 m wide extended along the shore to the recent maximum position (Grant and Higgins, 1913; Tarr and Martin, 1914). Subsequent retreat was slow but speeded up after 1925; recession by 1935 from the 1910 position varied between 150 and 580 m along the terminus (Field, 1937). The greatest retreat occurred at the western side of the terminus where recession was in part facilitated by development of a calving embayment. In 1935 the eastern end of the terminus rested on the modern lagoon-mouth spit (Figure 9) but soon after retreated into deeper water. With the whole terminus now calving icebergs the rate of recession has increased (Figure 9) and by 1992 was located about 3.8 km behind the lagoon-mouth spit. Photographs taken by A. Post since the 1960s document this retreat; some of these photographs are reproduced in Lethcoe (1987) together with additional observations and commentary.

Ultramarine Glacier

Ultramarine Glacier flows north–northeast from the northern end of the Sargent Icefield between steep valley walls to within 3 km of Blue Fiord (Figures 1 and 10). There is a dramatic difference between the landscapes within and beyond the Neoglacial limit. Outside the late Holocene maximum is open parkland where mountain hemlock as old as 450 years grow, whereas inside the moraine is a barren landscape dominated by ice-sculpted bedrock knobs. The LIA history of Ultramarine Glacier is largely inferred from the recently deglaciated area where logs from an extensive subfossil forest are found in thin till deposits and strewn across the land surface.

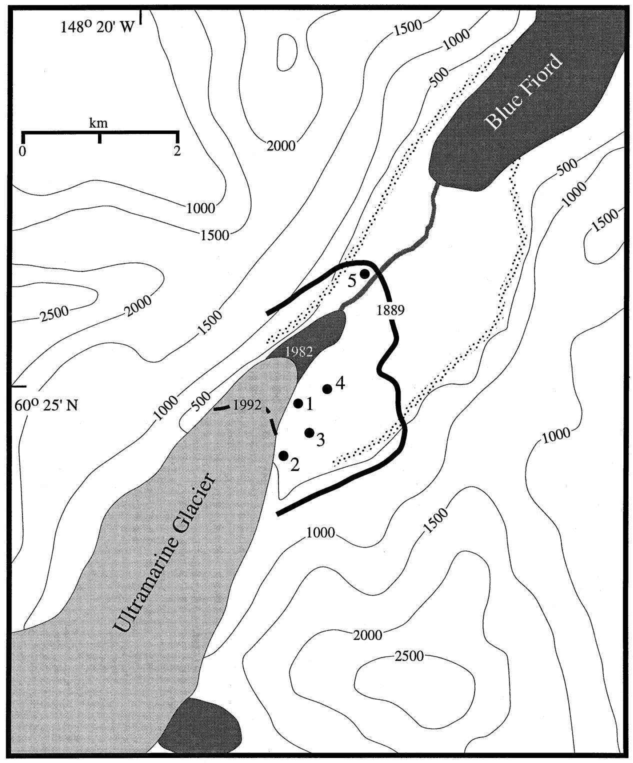

Figure 10 Ultramarine Glacier. Base map is US Geological Survey 1:63 360 series, Seward (B-4), 1951, minor revisions 1988. See Figure 4 for legend.

Twenty-nine subfossil logs from the overrun forest have been cross-dated with the regional master chronology (Figure 2). The earliest record of growth is from a log at location 1 (Figure 10) that was growing by AD 1306; this requires the terminus of Ultramarine Glacier to be upvalley of this position at that time. The earliest reliable kill dates of trees that can be attributed to glacial advance are from locations 2 and 3 (Figure 10) where four trees were killed between AD 1692 and 1697. Thereafter a down-valley sequential progression of kill dates occurs with 10 trees at location 1 being overrun between AD 1701 and 1703 and 12 trees at location 4 being overrun between AD 1711 and 1715. The remaining three logs dated were in modern alluvium and so cannot be used to constrain progression of the ice margin.

An additional six subfossil stumps in growth position at the Holocene terminal moraine (location 5, Figure 10) died between AD 1695 and 1766 (Figure 3). However, these appear to have been overwhelmed by outwash aggradation and so only constrain the arrival of Ultramarine Glacier at this position after AD 1766. Viereck (1967) determined that Ultramarine Glacier retreated from its LIA maximum between AD 1880 and 1890; our date of stabilization of AD 1889 (Figure 3) from a tree growing on the terminal moraine is consistent with this conclusion.

Ultramarine Glacier had retreated over 300 m from its LIA maximum when observed in 1935 (Field, 1937) and at least a further 300 m by 1957 (Field, 1975). Continued recession has left the terminus in 1992 about 1.2 km from the outer moraine.

Princeton Glacier

Princeton Glacier originates from the northern end of the Sargent Icefield and drains southwards towards Nassau Fiord, a northwestern extension of Icy Bay (Figures 1 and 11). The recently deglaciated valley is currently devoid of living spruce or hemlock older than a few decades; however, abundant subfossil hemlock logs

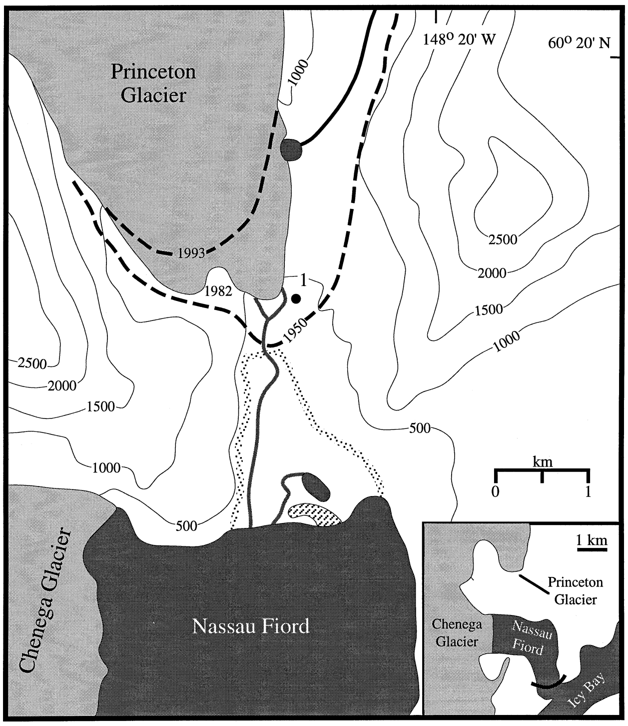

Figure 11 Princeton Glacier. Base map is US Geological Survey 1:63 360 series, Seward (B-4), 1951, minor revisions 1988. See Figure 4 for legend.

within a kilometre of the 1993 terminus attest to a previously extensive forest on the forefield.

Twenty-six detrital logs have been successfully cross-dated with the regional master chronology (Figure 2). The earliest year of growth AD 873 is from a log found at location 1 (Figure 11) and requires the terminus of Princeton Glacier to be upvalley of this position at this time. The first reliable kill dates from unrotted or undamaged trees suggest the ice margin was overriding forest near the 1993 terminus between AD 1183 and 1192. Thereafter, between four and six trees were killed in each decade from AD 1200 to 1240 with the last tree dying in AD 1248 (Figure 2). A calibrated radiocarbon date of AD 1210 (UW-519, Table 1) from a log collected near the terminus (A. Post, personal communication, 1993) is consistent with the cross-dating results.

Timing of events after AD 1248 is unclear. Detrital wood found along the outwash stream with a calibrated radiocarbon age of AD 1290 (UW-518, Table 1) is more recent than the cross-dated wood near the 1993 terminus; however, it is close enough in age to possibly come from the same log population. Ultimately, Princeton Glacier advanced to the fjord shore before uniting with Chenega and Tigertail Glaciers to reach a LIA terminal position at the mouth of Nassau Fiord (Figure 11, inset).

In 1957, Viereck cored trees beyond the terminal moraine at the mouth of Nassau Fiord and determined that this position had not been exceeded since at least AD 1675. Furthermore, he concluded that the coalesced Princeton-Chenega-Tigertail system remained at the mouth of Nassau Fiord until about AD 1857 to 1882. These conclusions are supported by the observations and maps by Vancouver in 1794, Seton-Karr in 1886 and Glenn in 1898 (all summarized in Tarr and Martin, 1914).

By 1908 the three glaciers filling Nassau Fiord had retreated into their respective valleys and the terminus of Princeton Glacier only reached to tidewater at its western end (Grant and Higgins, 1913). Two years later the terminus had retreated completely from tidewater (Tarr and Martin, 1914) and by 1950 had retreated an additional 1500 m (Field, 1975). The terminus in 1993 was 2.9 km from Nassau Fiord at about 180 m elevation.

Excelsior Glacier

Excelsior Glacier is one of the primary outlet tongues of the Sargent Icefield and drains southwards towards the Gulf of Alaska coastline (Figures 1 and 12). At the terminus is a large ice-contact proglacial lake that is bordered on its southern shore by a moraine complex. Only the western margin of the forefield was examined in the 1991 field season.

The outer moraine can be traced to prominent trimlines along both valley walls. Tree ages of stabilization of this ridge range from AD 1892 on the distal slope near the valley axis to AD 1797 near the western valley wall. An inner ridge, which can be traced to a lower trimline on the eastern valley wall, was dated at AD 1917. Retreat of Excelsior Glacier since this maximum has amounted to 4.6 km by 1991; this rapid retreat has been facilitated by iceberg calving into the proglacial lake.

Ellsworth and 3rd July Glaciers

Ellsworth Glacier lies at the head of Day Harbor and extends over 25 km from its ice divide to terminate as an iceberg-calving cliff in a large proglacial lake (Figures 1 and 13). The lake is dammed behind a LIA moraine complex which, together with subfossil wood from a valley on the west side of the terminus, records the late-Holocene fluctuations of the Ellsworth Glacier tongue. In addition, the glacial record of the adjacent 3rd July Glacier (unofficial name) on the east side of the main valley (Figure 13) is presented here.

Numerous features within the Ellsworth area attest to an active Pleistocene or early-Holocene glacial history. These include a lateral moraine on the east wall of the main valley beyond/above the LIA limit, several inactive rock glaciers and a southwest-facing cirque with no evidence of LIA activity. The extensive

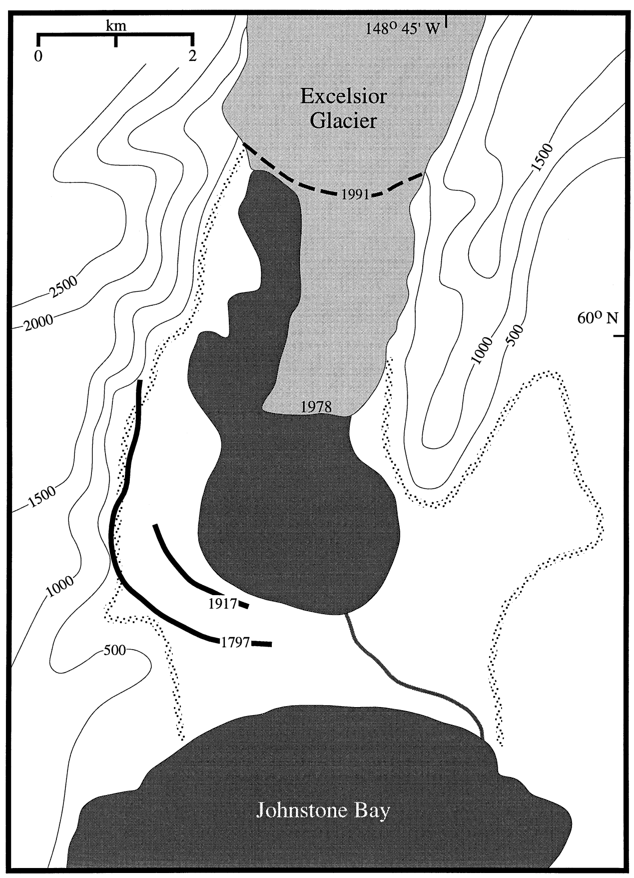

Figure 12 Excelsior Glacier. Base maps are US Geological Survey 1:63 360 series, Seward (A-5) and Blying Sound (D-5), 1951 and 1952, minor revisions 1988. See Figure 4 for legend.

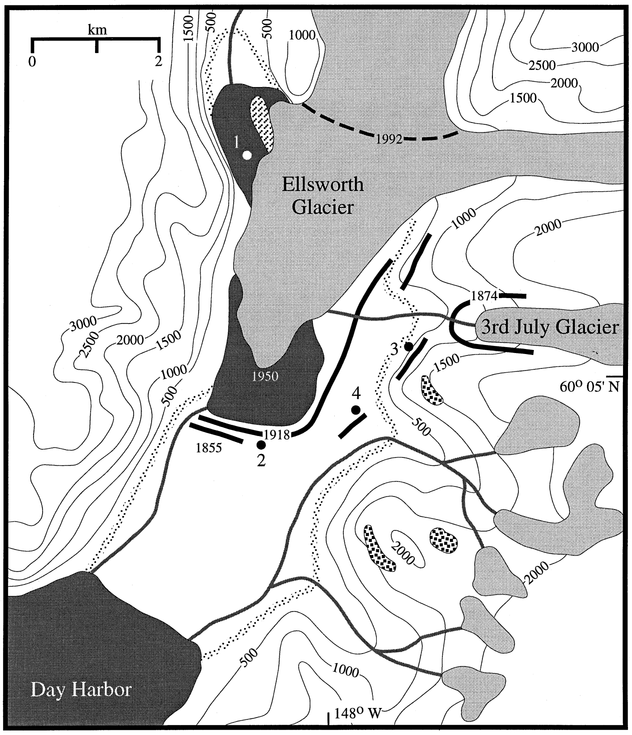

Figure 13 Ellsworth and 3rd July (unofficial name) Glaciers. Base map is US Geological Survey 1:63 360 series, Seward (A-6), 1951, minor revisions 1963. See Figure 4 for legend.

lichen cover on boulders at these sites suggest the moraines and rock glaciers are probably of late Wisconsin age.

Neoglacial advance of Ellsworth Glacier is inferred from 11 subfossil logs from a former ice-dammed lake basin (Stanton Lake) on the western margin of the main valley (location 1, Figure 13). This valley drained about 1972 by retreat of Ellsworth revealing lake strandlines, lacustrine and deltaic deposits. Cores were taken from logs buried in deltaic sediments and lying on the surface of the drained lake bed. Cross-dating with the regional master chronology (Figure 2) gave outer-ring ages from AD 1542 to 1651; however, all logs had been transported from their site of growth and many were badly battered. In general, the better-preserved logs have later outer-ring ages and so we believe AD 1651 is the best date for damming of this valley by advance of Ellsworth Glacier.

More recent fluctuations of the terminus of Ellsworth are suggested by the ages of living trees growing on morainal deposits. Tree ages on a pitted, bouldery, ice-contact deposit beyond a well-developed end-moraine at the south end of the proglacial lake indicate ice retreat by AD 1855 (location 2, Figure 13). This date for retreat is consistent with tree ages of AD 1826 and 1861 on discontinuous lateral moraine segments at locations 3 and 4 (Figure 13) respectively. Considered together, these three sites suggest thinning of the Ellsworth Glacier tongue at location 3 prior to ice retreat from the outer position in the valley floor by AD 1855. A second moraine of 10 to 15 m relief bounds the south and east shore of the proglacial lake and hosts trees that indicate ice-free conditions by AD 1918.

The glacier in 3rd July Valley (Figure 13) is fronted by a single moraine with a tree-ring age of AD 1874. Cirque glaciers in side valleys to the south are fronted by fresh-appearing, presumably LIA moraines. However, these features are above the modern tree-line and cannot be accurately dated.

In general, the termini of the small glaciers along the east side of Ellsworth Valley show little change between the 1950 aerial photographs and our observations in the summers of 1991 and 1992. This may be due to reduced ablation of these glaciers as a consequence of their extensive debris covers. In contrast, Ellsworth Glacier has retreated over 3 km from the moraine bordering the proglacial lake.

A regional glacial chronology

A regional chronology of glacial fluctuations can be developed using cross-dated subfossil wood from glacier forefields (Figure 2) and times of moraine stabilization (Figure 3). There have been three main periods of ‘Little Ice Age’ (LIA) glacial advance in the western Prince William Sound region during the past 1000 years, and for convenience we term these as early, middle and late LIA. The only dated evidence of Holocene glacial activity prior to the last millennium is seven internally cross-dated logs at Tebenkof Glacier tied to a radiocarbon age of AD 628.

The early LIA advance was under way at Princeton Glacier by AD 1190 where ice expansion continued until at least AD 1248. Trees were killed by Tebenkof Glacier between AD 1240 and 1300, whereas a single transported log suggests that Billings Glacier may also have been advancing and burying trees at this time (Figure 2). When the early LIA glacial expansion culminated is uncertain. However, the inner-ring date of a log found within the area covered by Tebenkof Glacier during this early advance suggests that some retreat from the early LIA advance had occurred before AD 1501.

The earliest reliable evidence of land-terminating glaciers advancing into forest in the middle LIA is from Cotterell Glacier where trees were being killed by AD 1611. A total of seven glaciers advanced during the seventeenth century and Ultramarine Glacier continued its expansion until at least AD 1715 (Figure 2). Kill dates of trees only indicate when glaciers were expanding and not when those advances terminated. However, the overlap between the last period of glacial expansion into forest (Figure 2) and the first interval of moraine formation (Figure 3) suggests that the mid-LIA glacial advance culminated in the first half of the eighteenth century. Moraines dating to this time at Billings, Taylor, Langdon-Kings and Wolverine glaciers are all downvalley of subsequent moraines; this implies ice retreat punctuated by temporary stillstands or minor readvances dominated until the late LIA.

The late LIA advance resulted in moraine building at nine of the study glaciers between AD 1874 and 1895. These moraines mark the greatest known extent reached by six of the study glaciers; moraines of earlier less extensive advances were presumably destroyed during this late LIA advance. At Billings, Taylor and Langdon-Kings Glaciers the *c.* 1890 moraine is the largest or only ridge nested within the early eighteenth-century moraine (Figure 3). Ice margins were generally close to the late LIA maximum position at the time of the first visits of scientific parties around the turn of the century. Near continuous ice retreat with minor advances has occurred since that time.

Discussion

While there is compelling evidence that LIA glacial activity in western Prince William Sound occurred as three distinct pulses, no single forefield contains a complete record of the entire sequence of advances and retreats. Tebenkof Glacier comes closest to a complete record, lacking only morphological evidence of the 1700s event. The incomplete nature of the glacial geologic record underscores the importance of studying multiple glaciers when drawing inferences about past regional activity.

The advance records of the tidewater-calving margins of Nellie Juan Glacier and the combined Chenega-Princeton-Tigertail system filling Nassau Fiord show little difference from the records of land-terminating glaciers in the region. This suggests that these tidewater-calving glaciers were able to maintain protective terminal shoals during their late-LIA advance and maxima (Post, 1975; Mann, 1986; Powell, 1991). However, once recession began, these tidewater termini retreated farther and faster in general than land-terminating glaciers from the region. The Chenega-Princeton-Tigertail system retreated at 207 m a^{-1} between 1886 and 1908, whereas Nellie Juan Glacier has retreated at an average rate of 59 m a^{-1} since 1935. Two other glaciers in the study area, Excelsior and Ellsworth, calve into freshwater lakes, which have formed during their retreat; their respective rates of retreat over the past 75 years are 62 and 41 m a^{-1}. These rates contrast with the mean retreat rate of 21 m a^{-1} for the other land-terminating valley glaciers in the region.

The moraine dates of Excelsior and Ellsworth glaciers appear anomalous when compared with other glaciers in the region (Figure 3). Specifically, the outermost ridges at these two sites pre-date the late LIA but did not stabilize in the first half of the eighteenth century, unlike at other forefields. Furthermore, stabilization of the last moraine of note occurred about 20 years later at Excelsior and Ellsworth Glaciers than at other glaciers. These two glaciers appear to have remained at their maxima longer than others elsewhere in the region, perhaps due to their southern location on the outer coast and south-facing aspect (Figure 1). This location and orientation would be favourable for intercepting precipitation from the Gulf of Alaska.

The glacial record of western Prince William Sound over the past 1000 years is remarkably similar on timescales of decades to other glacial histories around the Gulf of Alaska (Figure 14), as well as broadly consistent with more continental sites on century timescales (see Calkin, 1998). From along the Gulf of Alaska, early LIA advances have been identified by cross-dating trees from glacially-overrun forests at the nonsurging Steller Lobe of the Bering Glacier, and near Cordova at Sheridan Glacier (Yager *et al*., 1998). Wood buried by advances during the seventeenth through the early eighteenth centuries have been tree-ring cross-dated at Yalik Glacier in the Kenai Peninsula (Wiles and Calkin, 1994), and at Beare Glacier near Icy Bay (Frank *et al*., 1996). All these glaciers are land-terminating and are not known to surge. In addition to cross-dates from overrun trees, the moraine record combined for the Kenai Peninsula and Prince William Sound shows the strong peak in moraine-building during the latter half of the nineteenth century (Figure 14). Earlier moraine-building may be associated with middle LIA advances.

Conclusions

Cross-dating of glacially overrun trees with living trees has been completed at eight glacier forefields in the western Prince William Sound region. The results demonstrate both the timing and regional synchrony of early (twelfth and thirteenth centuries) and middle (seventeenth and early eighteenth centuries) LIA glacial advances. Tree-ring dates of 22 moraines at 13 glaciers delineate two main periods of formation. The earlier interval (early eighteenth century) overlaps with the later time of glacial advance identified from overrun forests and so dates culmination of the mid LIA advances. The later period of moraine formation (late nineteenth century) dates a late LIA advance when moraines were built on nine of the 13 forefields studied.

The Neoglacial record of tidewater-calving glaciers in western Prince William Sound differs from that of land-terminating glaciers primarily in the distance and rate of twentieth-century retreat. Freshwater-calving glaciers have also retreated farther and faster than land-terminating glaciers in the region.

The three-phase LIA glacial record of western Prince William Sound is consistent on a timescale of decades with four similarly dated histories from across the northern Gulf of Alaska (Figure 14). The tight age control using tree-ring cross-dating allows for the recognition of these synchronous glacial advances and provides a strong basis to compare with other climate proxy data from coastal sites along the Northeast Pacific.

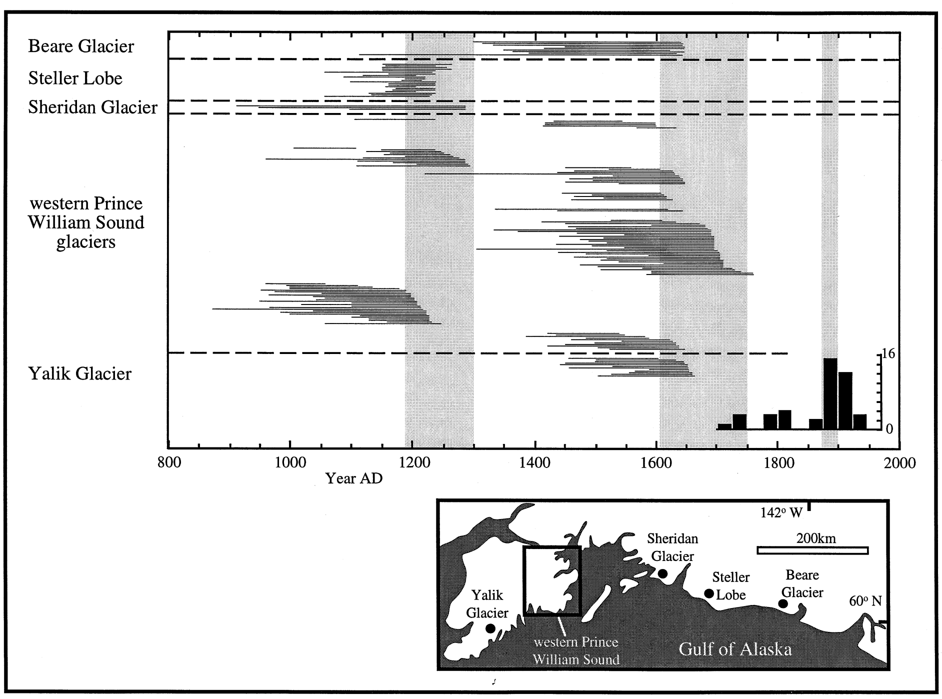

Figure 14 Summary of calendar dates from glacially overrun trees at land-terminating glaciers along the Gulf of Alaska. Horizontal bars depict lifespan of each cross-dated log. Shaded areas show the three times of 'Little Ice Age' glacial advance in western Prince William Sound. A histogram summarizing moraine dates of glaciers around the Gulf of Alaska is also shown.

Acknowledgements

We thank Austin Post for unpublished data and suggesting many of the study sites. A comprehensive review by B.H. Luckman is greatly appreciated. This work was supported by the National Science Foundation grants DPP-8922696 and OPP-9321213. Luigi Cremis and Dino Maraldo assisted in the field. Climate data were provided by the Alaska Climate Research Center (http://climate.gi.alaska.edu).

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