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Holocene glacier fluctuations in Alaska

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ABSTRACT

This review summarizes forefield and lacustrine records of glacier fluctuations in Alaska during the Holocene. Following retreat from latest Pleistocene advances, valley glaciers with land-based termini were in retracted positions during the early to middle Holocene. Neoglaciation began in some areas by 4.0 ka and major advances were underway by 3.0 ka, with perhaps two distinct early Neoglacial expansions centered respectively on 3.3–2.9 and 2.2–2.0 ka. Tree-ring cross-dates of glacially killed trees at two termini in southern Alaska show a major advance in the AD 550s–720s. The subsequent Little Ice Age (LIA) expansion was underway in the AD 1180s–1320s and culminated with two advance phases respectively in the 1540s–1710s and in the 1810s–1880s. The LIA advance was the largest Holocene expansion in southern Alaska, although older late Holocene moraines are preserved on many forefields in northern and interior Alaska.

Tidewater glaciers around the rim of the Gulf of Alaska have made major advances throughout the Holocene. Expansions were often asynchronous with neighboring termini and spanned both warm and cool intervals, suggesting that non-climatic factors were important in forcing these advances. However, climatic warming appears to have initiated most rapid iceberg-calving retreats. Large glaciers terminating on the forelands around the Gulf of Alaska may have had tidewater termini early in the Holocene, but have progressively become isolated from the adjacent ocean by the accumulation and subaerial exposure of their own sediments.

1. Introduction

Glacier ice currently covers approximately 74,700 km^2 of Alaska (Kaufman and Manley, 2004). These ice fields and glaciers are distributed among multiple mountain ranges (Fig. 1) with the most extensive ice complexes being in the south around the margins of the Gulf of Alaska. Land-based glaciers in Alaska are sensitive indicators of climate and their fluctuations have been used to assess both contemporary and past climate change (e.g. Wiles et al., 2004, 2008; Dyurgerov and Meier, 2005; Lemke et al., 2007). Moreover, volume changes of the ice complexes of southern Alaska have been sufficiently large in recent centuries to affect both eustatic sea level (Arendt et al., 2002; Meier et al., 2007) and to locally cause meters of glacioisostatic depression and rebound (Motyka, 2003; Larsen et al., 2004; Mann and Streveler, 2008).

The purpose of this paper is to update the review of Holocene glaciation of Alaska by Calkin (1988). Considerable work has been done in the past two decades, with glacier histories being reconstructed in new regions and new analytical and dating methods being applied to previously studied areas. Furthermore, ongoing glacier retreat throughout Alaska and concomitant erosion of forefields is continually providing new exposures of stratigraphy and datable materials. Several other reviews (Mann et al., 1998; Calkin et al., 2001; Wiles et al., 2008) have synthesized some of these new data for specific time intervals or regions of Alaska; herein we emphasize both a statewide and Holocene-length perspective.

Following a summary of late Wisconsinan deglaciation, we divide the glaciers of Alaska into three groups for review of Holocene fluctuations. First, we consider the histories of valley and cirque glaciers that have been land-based throughout the Holocene. These termini occur in all major mountain ranges of Alaska and their fluctuations are most directly controlled by climate change. Secondly, we review the histories of tidewater glaciers in southern Alaska; these large systems have iceberg-calving termini and are significantly affected by non-climatic forcing factors. Lastly, we consider the large foreland glaciers of the Chugach and Saint Elias mountains that terminate on the fringe of the Gulf of Alaska,

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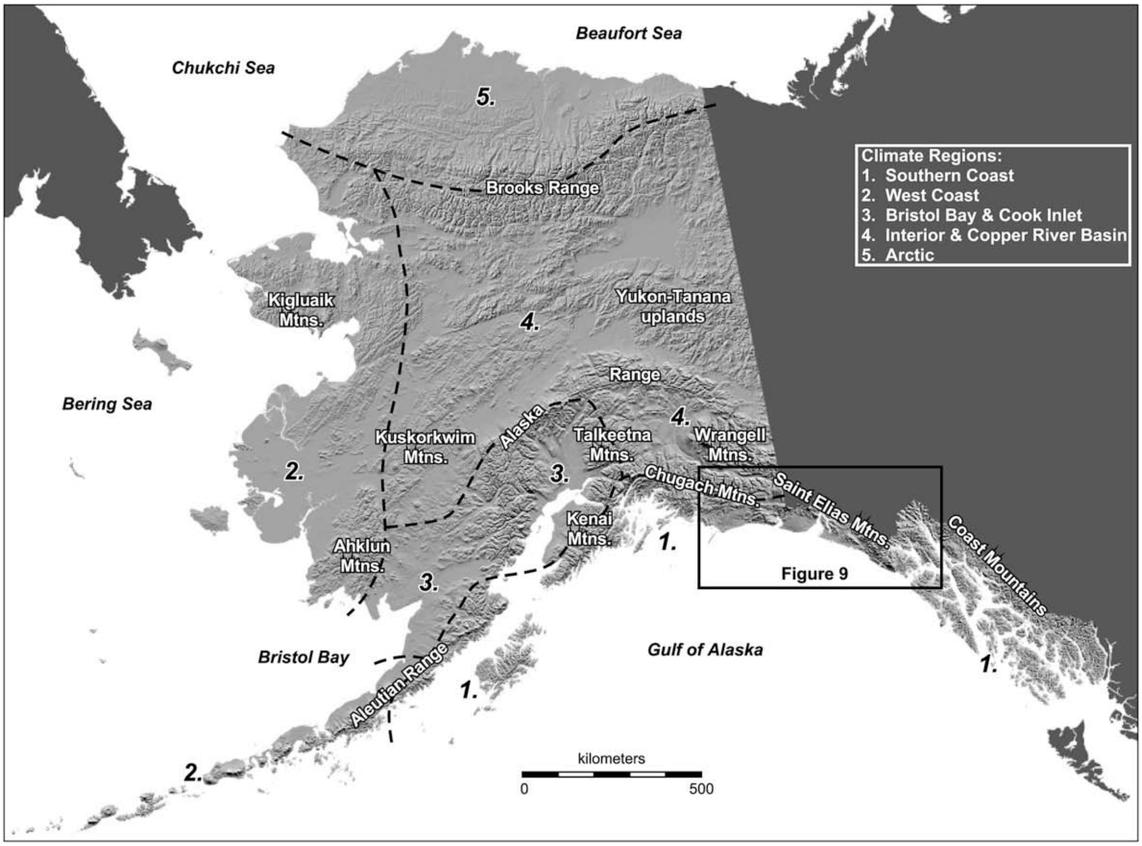

Fig. 1. Shaded relief map showing areas mentioned in text. Dashed lines separate climate regions of Shulski and Wendler (2007).

which may have changed from tidewater-calving to land-based termini during the Holocene.

1.1. Geographic and climatic setting

The largest glaciers and ice fields of Alaska are in the south in the Kenai, Chugach, Saint Elias and Coast mountains (Fig. 1). Large valley glacier systems also occur in the Alaska Range and Wrangell Mountains, whereas smaller glaciers are found in the Brooks and Aleutian ranges, Kigluaik, Ahklun and Talkeetna mountains, and on islands rimming the Gulf of Alaska (Molnia, 2007). Peaks in Alaska's mountain ranges are generally 1000–3000 m above sea level (asl), while a few peaks in the Saint Elias, Wrangell and Chugach mountains and Alaska Range reach 4000–6000 m asl.

This distribution of glaciers reflects the Gulf of Alaska being the main source of precipitation to the region (Shulski and Wendler, 2007). Additional moisture originates from Bristol Bay and the Bering Sea, while low air temperatures and sea ice cover on the Chukchi and Beaufort seas limit moisture transport from the north. Snowlines similarly reflect these moisture sources and rise northwards from the south coast; snowlines in the interior and north also rise eastwards away from the Bering Sea (Péwé, 1975).

Glaciers on the Gulf of Alaska side of the Kenai, Chugach, Saint Elias and Coast mountains are in the Southern Coast climate region (Fig. 1). The adjacent ocean moderates temperatures in all seasons and annual precipitation is very high, with annual water equivalent totals exceeding 5 m in some locations (Shulski and Wendler, 2007). The maritime influence is also strong in the West Coast climate region, although the lack of a continuous mountain barrier reduces the regional orographic influence on precipitation. The Bristol Bay and Cook Inlet climate region is transitional to the more continental Interior and Copper River Basin climate region where winter temperatures and precipitation are significantly lower than in areas to the south and west, while the Brooks Range forms the southern boundary of the Arctic climate region (Shulski and Wendler, 2007). Glacier activities reflect these climate zones, with mass turnovers and flow rates being greatest in southern Alaska.

1.2. Methods

Most of the studies synthesized here are based on forefield mapping of glacial landforms and stratigraphy. Moraines define the limits of advances (Fig. 2) and are dated by historical observations, germination dates of trees, lichenometry, and in a few cases, cosmogenic exposure ages and tephra layers (e.g. Begét, 1994; Wiles and Calkin, 1994; Evison et al., 1996; Viens, 2005). Calkin and Ellis (1980) and Loso and Doak (2006) discuss the application and assumptions of lichenometric ages in Alaska, and Solomina and Calkin (2003) present five revised lichen growth curves for distinct areas within the state. We use these revised growth equations in our data compilation and follow Ellis and Calkin (1984) in assuming an age reliability of $\pm 20\%$ with lichenometric data.

Dendrochronologic dates from glacially damaged or killed trees (Fig. 3) are our preferred method for constraining late Holocene ice margin fluctuations in southern Alaska (e.g. Wiles et al., 1999a; Barclay et al., 2006). Composite tree-ring-width chronologies composed of multiple overlapping generations of subfossil trees now extend back almost two millennia (Table 1) and sites such as Glacier Bay have the potential to yield chronologies that will span much of the Holocene (Lawson et al., 2006). Strengths of this method include the annual precision of tree-ring cross-dates and the ability to derive both germination and kill dates for each log

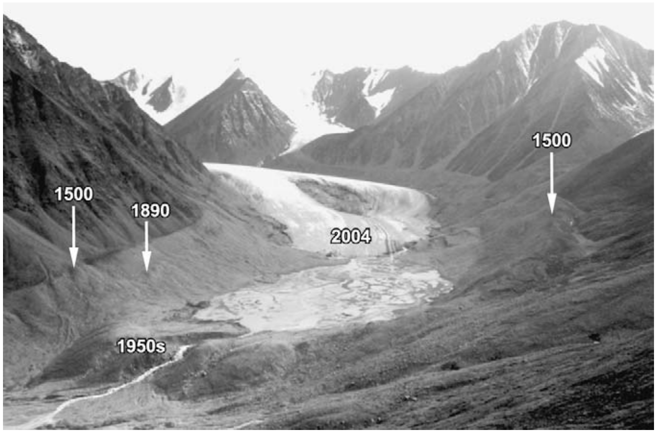

Fig. 2. Okpilak Glacier in northeastern Brooks Range. Arrows indicate lateral moraines lichenometrically dated, respectively, to AD 1500 and 1890 (Evison et al., 1996). Moraine in center of valley was at the glacier terminus from 1947 to 1956. Photograph by Nolan (2004).

Fig. 3. Logs in outwash at Tebenkof Glacier. Cutbank is about 3 m high and logs were collectively alive from AD 338 to 712.

dated that serve as minimum and maximum dates for glacier advances (Fig. 4). Furthermore, the abundance of cross-datable material in southern Alaska means that tens of logs can be readily dated for some forefields, allowing ice margin fluctuations to be reconstructed in considerable temporal and spatial detail.

Radiocarbon ages on logs and wood unsuitable or too old for tree-ring cross dating are also used extensively in this paper. Selected dates were calibrated in CALIB 5.1 (Stuiver and Reimer, 1993) using the IntCal04 calibration curve (Reimer et al., 2004) and are reported as the median of the probability distribution function (Telford et al., 2004) in calibrated radiocarbon years before 1950 (ka) and, for samples from the last two millennia, calibrated radiocarbon years AD (cal. yr AD). The 2σ-age ranges and additional details for all samples cited in this paper are presented in Appendix. More generalized ages based on multiple radiocarbon samples, lichen growth curves that are tied to radiocarbon ages, and cosmogenic isotope exposure ages, are also expressed in ka and, for Holocene ages, are rounded to the nearest 100 years.

A few studies have inferred nearby glacier activity using multiple proxies in sediment cores from meltwater-fed lakes (e.g. Levy et al., 2004; McKay and Kaufman, 2009). Although these are indirect measures of glacier extent, these lacustrine records can provide a Holocene-length perspective that complements the discontinuous nature of forefield deposits.

2. Late Wisconsinan deglaciation

Glaciers covered approximately 727,800 km^2 of Alaska and its adjacent continental shelf at the Last Glacial Maximum (LGM) (Kaufman and Manley, 2004). Most of this ice was in central to southern Alaska and on the adjacent continental shelf, where it formed an inter-connected network of ice caps and piedmont lobes as a northwestern extension of the Cordilleran Ice Sheet (Hamilton and Thorson, 1983). Isolated ice caps and valley glacier systems developed in the Ahklun Mountains, the Brooks Range, and in other upland and mountain areas (Porter et al., 1983). Equilibrium line altitudes (ELAs) were depressed between 700 and 200 m below modern values and show that at the LGM, just as during the Holocene, the North Pacific was the dominant source of moisture to the region (Hamilton and Thorson, 1983; Briner and Kaufman, 2000; Balascio et al., 2005).

Retreat from the LGM was underway by 25 ka in the Brooks Range and by 22–20 ka in central and southern Alaska (Briner and Kaufman, 2008). Minimum ages of 19.4–14.8 ka (Mann and Peteet, 1994; Reger and Pinney, 1996; Stilwell and Kaufman, 1996; Briner and Kaufman, 2008) suggest that deglacation may have been later along the Gulf of Alaska rim. Ice margins throughout Alaska were at or behind their modern positions by the start of the Holocene (Porter et al., 1983; Mann and Hamilton, 1995).

Multiple still-stands and readvances occurred during the recession from the LGM. In particular, Briner et al. (2002) report an end moraine in the Ahklun Mountains that formed during the Younger Dryas event and other moraines have been reported that may also have developed at this time (summarized by Briner and Kaufman, 2008). However, age control is weak for most other post-LGM moraines in Alaska and so the statewide synchrony and significance of these readvances and still-stands cannot currently be assessed.

Table 1
Master tree-ring-width chronologies from southern Alaska.

Chronology	Time span[a]	Total years	Tree species[b]	Number of series	Mean series length	Mean series intercorrelation[c]	Reference
Princeton	949–1235	287	TM	29	162	0.591	Barclay et al. (1999)
Nellie Juan	1069–1999	931	TM	72	357	0.645	Barclay et al. (2003)
Tebenkof	312–725	414	TM	129	174	0.599	Barclay et al. (2009)
	1093–1991	899					
Columbia	583–2002	1420	TM	512	230	0.560	Wilson et al. (2007), Wiles (unpublished)
Sheridan	122–605	484	TM,TH,PS	202	182	0.589	Barclay (unpublished)
	893–1993	1100					

[a] Interval where chronology has ≥2 trees, so is slightly shorter than total duration of forest growth at site.
[b] TM, *Tsuga mertensiana*; TH, *Tsuga heterophylla*; PS, *Picea sitchensis*.
[c] Mean of correlations between series and their respective site chronologies.

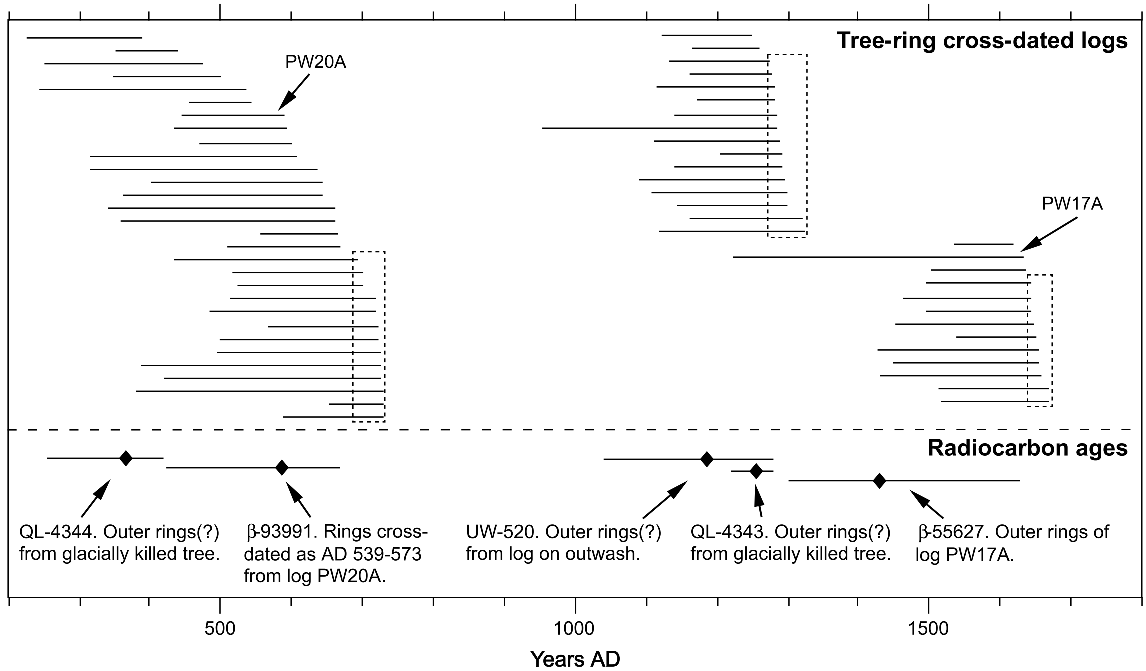

Fig. 4. Tree-ring cross-dates and radiocarbon ages of logs on the forefield of Tebenkof Glacier. Tree-ring cross-dates: horizontal lines show lifespans of each log, and dotted areas indicate inferred intervals when trees were being killed by ice margin advance. Radiocarbon ages: horizontal lines show 2-σ age range of each sample, and filled diamonds are median age estimates. Data from Crossen (1997), Wiles et al. (1999a) and Barclay et al. (2009).

3. Holocene fluctuations of land-terminating glaciers

3.1. Early to middle Holocene

Evidence for advances of land-terminating valley and cirque glaciers in the early to middle Holocene is largely equivocal or only loosely constrained. Calkin (1988) reported four moraines of possible early Holocene age in the Brooks Range, but their lichen-derived ages do not cluster and are based on large extrapolations of the regional lichen growth curve. An early Holocene age estimate for moraines (Crater Mountain drift) in the Kuskokwim Mountains was based on morphology, lichen cover and soil development (Bundtzen, 1980; Kline and Bundtzen, 1986). The suggestion by Schmoll et al. (1999) that end moraines (Winner Creek drift) in the western Chugach Mountains were formed in the early Holocene was based solely on the position of these deposits in their host valleys between moraines of late Pleistocene and late Holocene age.

Radiocarbon ages provide limited temporal control for other deposits. Early to middle Holocene advances of Yanert and Capps glaciers, respectively in the central and southwestern Alaska Range, have been proposed (Ten Brink, 1983; Yehle et al., 1983) but are only constrained by minimum radiocarbon ages of 6.67 and 8.60 ka. In contrast, wood buried by glacifluvial deposits with a median age of 6.52 ka provides a maximum age for advance of one or both of Gulkana and College glaciers in the east-central Alaska Range (Péwé and Reger, 1983; Begét, 1994). Thorson and Hamilton (1986) report that the Russell Creek drift at Cold Bay in the Aleutian Range is constrained by both maximum and minimum ages to have been deposited between 12.33 and 7.57 ka; however, Detterman (1986) correlated these same deposits with the late Wisconsinan Iliuk advance.

A sediment core from Waskey Lake at 150 m asl in the Ahklun Mountains (Fig. 5) suggests that land-terminating glaciers in this area were in retracted positions or perhaps entirely absent during the early to middle Holocene. This lake currently receives meltwater from cirque glaciers upstream and Levy et al. (2004) interpret the low magnetic susceptibility, high organic matter content, and high kaolinite:quartz ratio from 9.1 to 3.1 ka as indicative of reduced glacier activity in the catchment. Similar conclusions of reduced early to middle Holocene glacier activity were derived from analyses of cores from glacier-fed Cascade Lake, 38 km to the north at 335 m asl (Kathan, 2006), and Hallet and Greyling lakes at about 1100 m asl in the central Chugach Mountains (McKay and Kaufman, 2009).

3.2. Early neoglaciation

The Waskey Lake core (Fig. 5) indicates an increase in glacier activity in the Ahklun Mountains after 3.1 ka (Levy et al., 2004) and a similar increase in glacier activity is shown in the nearby Cascade Lake core at about 3.0 ka (Kathan, 2006). Cores from the higher elevation Hallet and Grayling lakes in the central Chugach Mountains show an earlier cooling and transition to increased glacier activity between 4.5 and 4.0 ka (McKay and Kaufman, 2009).

Major glacier advances in the early Neoglacial are suggested by radiocarbon ages of logs and wood in glacial deposits in southern Alaska (Fig. 6). Dates with median age estimates of 3.27–2.93 ka (Tuthill et al., 1968; Denton and Karlén, 1977; Röthlisberger, 1986; Wiles et al., 2002; Viens, 2005; De Simone, pers. comm. 2008) are from wood in or buried by till and suggest advances of six termini in the Saint Elias, Chugach and Coast mountains. These expansions were generally coeval with the early advances of the Tiedemann and Peyto intervals identified, respectively, in the Coast Mountains and Rockies of Canada (Ryder and Thomson, 1986; Luckman et al., 1993; Menounos et al., 2009).

A later interval of early Neoglacial advance is well defined at Nabesna Glacier in the Wrangell Mountains (Wiles et al., 2002) where ice-marginal deltaic sediments accumulated in a side valley during expansion of the terminus. An *in situ* stump below these sediments is dated at 1.99 ka (Fig. 6) and records an advance of the

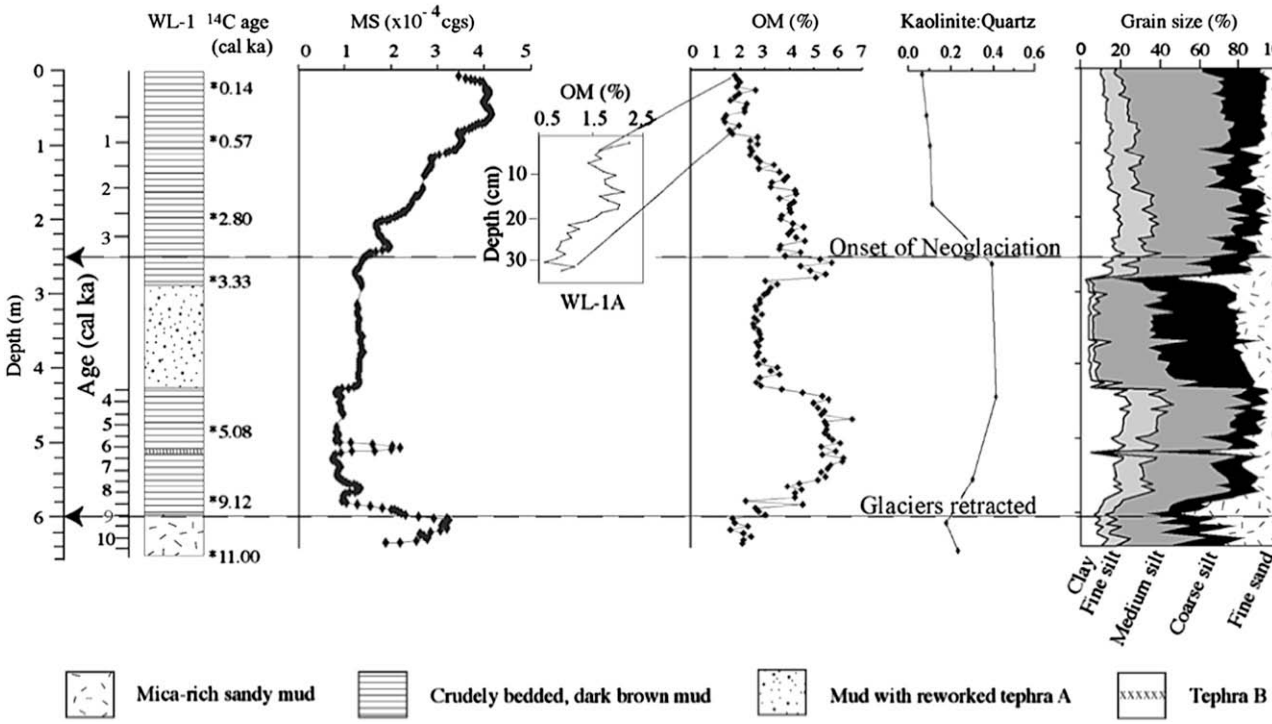

Fig. 5. Lithologic log, radiocarbon ages, magnetic susceptibility (MS), organic matter content (OM), kaolonite:quartz ratio, and grain-size distribution for Waskey Lake core (WL-1), Arklun Mountains, southwestern Alaska. Late Glacial and Neoglacial glacier activity in the upstream drainage basin are indicated by low kaolinite:quartz ratio, low percentages of organic matter, and high magnetic susceptibility values. From Levy et al. (2004). Reprinted by permission of Sage Publications.

ice margin to a point close to the position of its subsequent Little Ice Age maximum. This age is supported by two additional ages from transported wood in the sand and gravel encasing the stump; these are slightly older and are presumably woody fragments from the landsurface over-run and eroded by Nabesna Glacier during this expansion. Ages of 2.21–2.07 ka from Battle, Herbert and Mendenhall glaciers in the Coast Mountains (Röthlisberger, 1986) suggest advances farther south at this same general time (Fig. 6).

Farther north, lichenometric ages have been used to assign extant moraines to the early Neoglacial on the forefields of Foraker, Peters, Yanert, Black Rapids, Castner, Canwell, and Gulkana glaciers in the central Alaska Range (Bijkerk, 1984; Werner, 1984; Reger and Péwé, 1991; Reger et al., 1993; Begét, 1994; Howley and Licciardi, 2008). Similarly aged moraines have also been identified at Seven Sisters Glacier and perhaps other termini in the Wrangell and northern Saint Elias mountains (Denton and Karlén, 1977; Wiles et al., 2002). And in the Brooks Range, lichenometric ages show three clusters of early Neoglacial moraine stabilization (Ellis and Calkin, 1984; Calkin, 1988) that, based on the revised Brooks Range lichen growth curve of Solomina and Calkin (2003), were centered on 4.1, 3.3 and 2.6 ka. Although the estimated errors of ±20% (Ellis and Calkin, 1984) associated with these lichenometric ages are quite large, the field evidence suggests that there are a substantial number of Holocene moraines preserved in interior and northern Alaska that pre-date the pervasive Little Ice Age advances.

3.3. First millennium AD (FMA)

Tree-ring cross-dates of glacially killed logs show that Tebenkof Glacier in the Kenai Mountains advanced into a forefield forest in the AD 710s and 720s (Barclay et al., 2009; Figs. 4, 6 and 7). Similar logs at Sheridan Glacier on the coastal side of the Chugach Mountains cross-date with those at Tebenkof (Barclay, unpublished data) and show that outwash was aggrading on the Sheridan forefield in the AD 550s; trees were being incorporated into till in the 560s and ice margin advance and outwash aggradation continued at Sheridan until at least the 600s (Fig. 6). It is currently unclear whether the approximately 100-year offset between the Sheridan and Tebenkof tree-kill dates reflects multiple distinct phases of expansion in the FMA, or is due to differing dynamic responses of the Sheridan and Tebenkof termini to FMA climate forcing.

Radiocarbon ages of cal. yr AD 570–650 (1.38–1.30 ka) from Beare, Bartlett, Grewingk, Dinglestadt and Herbert glaciers (Karlstrom, 1964; Röthlisberger, 1986; Wiles and Calkin, 1994; Reyes et al., 2006) all indicate advances generally coeval with those of Tebenkof and Sheridan glaciers in the FMA (Fig. 6). Slightly older ages of cal. yr AD 200–320 (1.75–1.63 ka) may indicate an earlier interval of FMA advance at Kuskulana, Eagle, Herbert and Mendenhall glaciers, and this is consistent with the interpretation by Jackson et al. (2008) of two distinct advances within the FMA in the northern Coast Mountains in Canada.

Lichen ages show that moraine building was occurring in the Brooks Range through most of the first millennium AD (Fig. 8). Many of these are terminal moraines and so the magnitude of many advances in this interval must have been comparable to those of the subsequent Little Ice Age. Lichen ages also suggest moraine building at Copper, Muldrow, Peters and Yanert glaciers (Bijkerk, 1984; Werner, 1984; Wiles et al., 2002) in the Wrangell Mountains and Alaska Range during the FMA.

3.4. Medieval warm period (MWP)

Trees were germinating on the Tebenkof, Sheridan and Princeton glacier forefields by the AD 950s or earlier (Figs. 6 and 7), showing ice margin recession in the Medieval Warm Period (MWP) following the FMA advance. Similarly aged forefield forests also

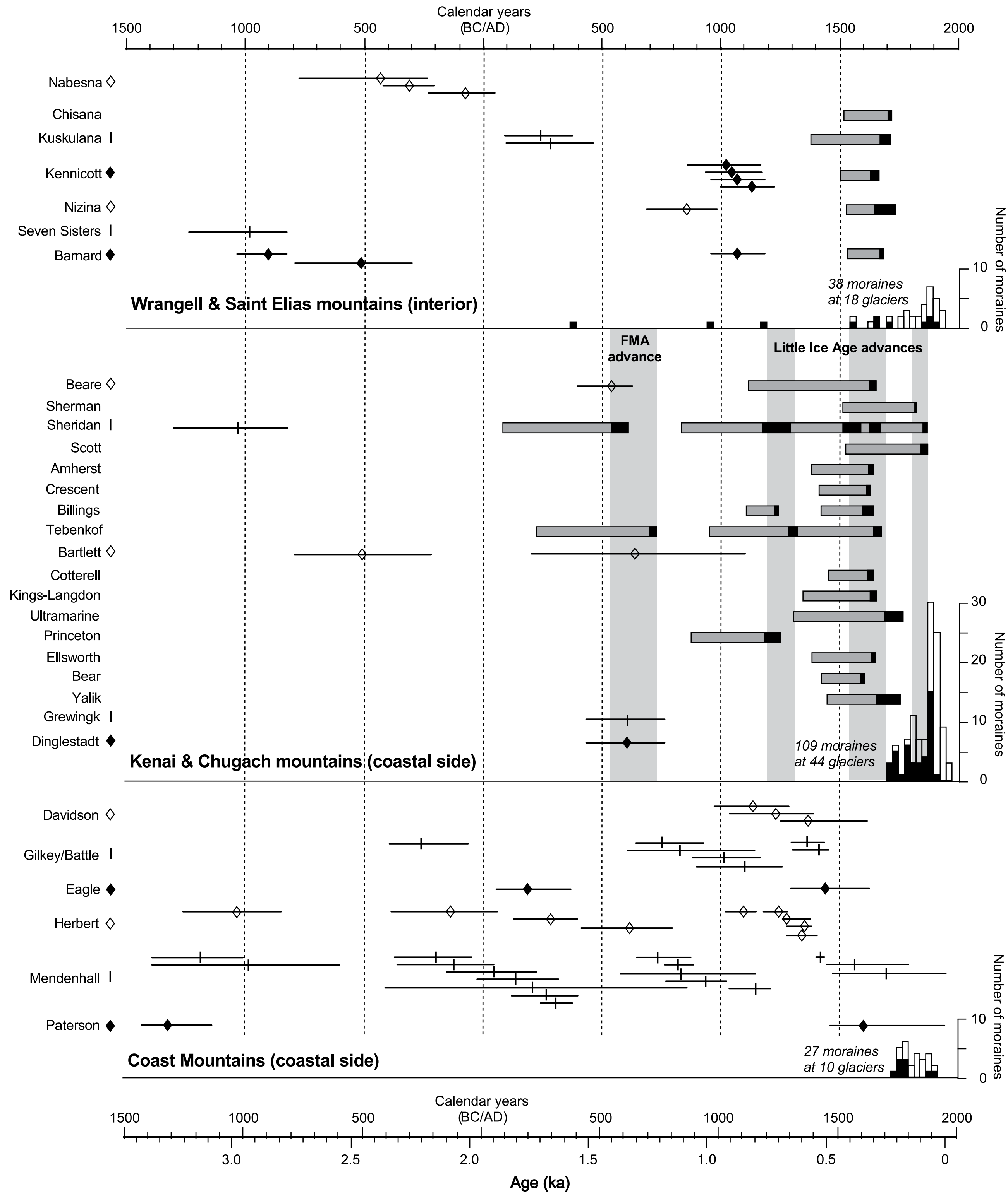

Fig. 6. Forefield wood and moraine dates from land-terminating glaciers. Radiocarbon ages: horizontal lines show 2-σ age range and symbols are median age estimates; different symbols (cross-lines, filled and open diamonds) are used to distinguish samples from different forefields. Tree-ring cross-dates: horizontal grey bars are total periods of forefield forest growth with black areas indicating inferred intervals when ice margins were killing trees. Moraines: dated using tree-ring minimum ages, lichenometry and direct observations; filled columns are terminal moraines and open columns are recessional moraines. Sources of radiocarbon ages are listed in Table A2, tree-ring and moraine data are from Lawrence (1950), Heusser and Marcus (1964), Vierek (1967), Field (1975), Miller (1976), Denton and Karlén (1977), Wiles and Calkin (1994), Crossen (1997), Wiles et al. (1999a, 2002, 2008), Viens (2005), Daigle and Kaufman (2009), Barclay (unpublished data), and lichen growth curves from Solomina and Calkin (2003).

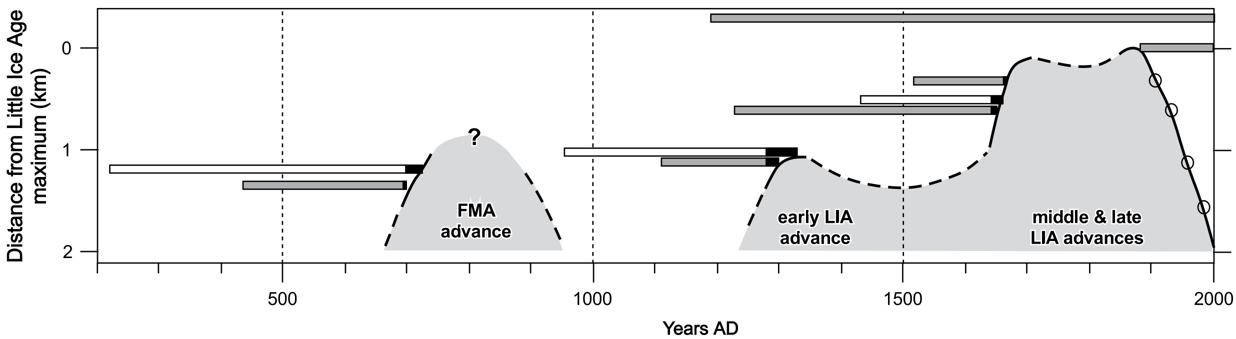

Fig. 7. Time–distance diagram for Tebenkof Glacier. Horizontal bars show durations of forest growth on different areas of the forefield based on tree-ring cross-dates; grey bars are for *in situ* trees, open bars for transported logs, and black areas indicate inferred intervals when trees were being killed. Circles indicate ice margin positions based on direct observations. From Barclay et al. (2009).

developed at Beare, Billings, Davidson, Gilkey, Herbert, Mendenhall, Kennicott, Nizina, and Barnard glaciers (Crane and Griffen, 1968; Röthlisberger, 1986; Lacher, 1999; Wiles et al., 1999a, 2002; Weber, 2006). Remnants of many of these MWP forests are still being revealed by ice margin retreat today, suggesting that glacier recession in southern Alaska during the MWP was comparable to that which had occurred by the late the 20th century.

On the basis of a glacially reworked tephra and minimum radiocarbon ages from overlying materials, Denton and Karlén (1977) suggested that Guerin, Giffin, Russell and an un-named glacier in the northern Saint Elias Mountains advanced between cal. yr AD 780 and 980 (1.17–0.97 ka). However, Rampton (1978) noted that the correlative advance of Natazhat Glacier, just 10 km away in Yukon Territory, was several hundred years later in the Little Ice Age, and Wiles et al. (2008) concluded that there was no clear evidence of a MWP advance in the Wrangell-Saint Elias region.

3.5. Little Ice Age (LIA)

The forefield records from the Kenai, Chugach and Coast mountains (Figs. 6 and 7) show a consistent pattern of ice margin advances during the LIA. Tree-ring cross-dated logs indicate that the early advances were from about AD 1180s to 1320s, the middle LIA advances from about 1540s to 1710s, and the late advances about 1810s–1880s. Numerous radiocarbon ages (e.g. Crane and Griffen, 1968; Röthlisberger, 1986; Wiles and Calkin, 1994) support these tree-ring data but are too imprecise to differentiate between the distinct phases of LIA advance. Many terminal moraines (Fig. 6) were built at maxima of the middle LIA expansions, and terminal moraines or large recessional moraines were deposited on almost every forefield at the culmination of the late LIA advance.

Farther north, tree-ring cross-dates at five termini in the Wrangell and Saint Elias mountains show that ice margins advanced and were killing trees between about AD 1630 and 1720 (Fig. 6). Some terminal moraines were deposited following these middle LIA advances, with a larger cluster of moraine dates showing the maximum of the late LIA expansion. The Brooks Range moraine data (Fig. 8) cluster into three LIA groups, centered respectively on AD 1210, 1560 and 1840. Moraines of the *c.*1560 cluster are found on nearly all forefields and are supported by several radiocarbon ages on glacially over-run wood and moss (Ellis and Calkin, 1984; Calkin, 1988).

On the west coast, end moraines at Grand Union Glacier in the Kigluiak Mountains (Fig. 1) were dated with lichen to AD 1645 and 1895, while terminal moraines of two nearby glaciers respectively date to 1675 and 1825 (Calkin et al., 1998). Levy et al. (2004) and Kathan (2006) also used lichen diameters to assign moraines at seven glaciers in the Ahklun Mountains to the LIA. And in the Aleutian Range, preliminary studies have suggested late Holocene ages for moraines within 1 or 2 km of modern termini (Detterman, 1986; Thorson and Hamilton, 1986).

LIA advances have also been noted at glaciers in the central interior of Alaska. Vierek (1967) used tree-rings to date a moraine of Tazlina Glacier on the north side of the Chugach Mountains to *c.*1805, whereas terminal and recessional moraines of nearby cirque glaciers were dated to the LIA with lichen (McKay and Kaufman, 2009). Similar middle to late LIA ages were found for end moraines of Gulkana and College glaciers in the east-central Alaska Range (Péwé and Reger, 1983; Begét, 1994). Farther west in the Alaska Range, Ten Brink (1983) suggested two ice advances in the Mount McKinley area in the past millennium, and ice-cored, un-vegetated moraines of probable late Holocene age were reported in the nearby Farewell area (Kline and Bundtzen, 1986). However, moraines in the Yukon-Tanana Upland (Ramshorn drift) that were

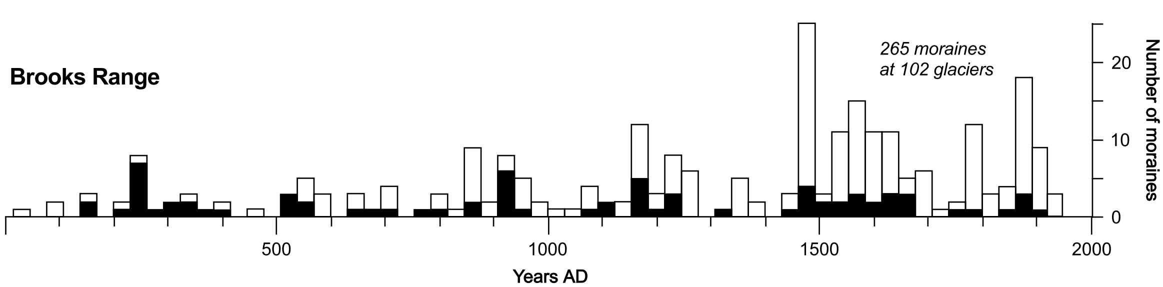

Fig. 8. Moraine stabilization dates for past 2000 years in the Brooks Range based on lichenometry. Filled columns are terminal moraines and open columns are recessional moraines. Data from Haworth (1988) and Sikorski et al. (2009), and lichen growth curve from Solomina and Calkin (2003).

previously suggested as Neoglacial in age (Weber, 1986) have been re-assigned to the Late Wisconsinan on the basis of cosmogenic isotope ages (Briner et al., 2005).

Retreat and thinning of land-based termini has dominated in all areas of Alaska during the 20th and 21st centuries and is documented by Field (1975, 1990) and Molnia (2007, 2008). Cirque glaciers in the Ahklun Mountains and Brooks Range (Levy et al., 2004; Sikorski et al., 2009) show areal ice losses of 25–50% between LIA maxima and the mid to late 20th century. Volumetric ice losses in recent decades have been quantified by comparing digital elevation models, aerial and satellite imagery, and airborne laser profiles (e.g. Adalgeirsdóttir et al., 1998; Sapiano et al., 1998; Arendt et al., 2002, 2006; Hall et al., 2005; Larsen et al., 2007) and show that substantial ice losses have continued to the present.

3.6. Climatic and other controls

The foregoing records show that multiple intervals of broadly synchronous glacier advance occurred in coastal and interior southern Alaska during the early Neoglacial, FMA and LIA. The coherence of these glacier histories supports the inference that regional climate change is the primary control on these terminus fluctuations.

Direct comparisons of temperature sensitive tree-ring chronologies and high-resolution terminus records in southern Alaska show a strong correspondence between multi-decadal cool intervals and phases of LIA glacier advance (Barclay et al., 2003; Davi et al., 2003). Similarly, broader intervals of advance (i.e. LIA, FMA and Neoglaciation) correspond with cool intervals identified in other Alaskan proxy climate records (Heusser et al., 1985; Peteet, 1986; Hansen and Engstrom, 1996; Hu et al., 2001; Kaufman et al., 2004; Loso et al., 2006). Thus, the Alaskan land-based glacier record is interpreted as primarily reflecting changes in regional temperature (Wiles et al., 2004, 2008).

Studies in the Brooks Range and Kigluaik and Kenai mountains have suggested that equilibrium line altitudes (ELAs) were depressed 100–200 m below modern levels during Neoglacial maxima (Ellis and Calkin, 1984; Padginton, 1993; Calkin et al., 1998). In contrast, more recent work in these same areas, based on slightly different methods and reference periods (Levy et al., 2004; Kathan, 2006; Daigle and Kaufman, 2009; McKay and Kaufman, 2009; Sikorski et al., 2009), suggest much smaller ELA depressions of 22–83 m. These latter studies suggest that reductions in precipitation during cool intervals may account for the relatively limited ELA lowering in their study areas.

It is perhaps surprising that LIA advances, which in southern Alaska were invariably the largest of the Holocene, did not destroy early Neoglacial moraines in interior and northern Alaska. Denton and Karlén (1977) suggested that the heavy debris content of some glaciers in the Wrangell and Saint Elias mountains may have resulted in early Neoglacial moraines that were too massive to be over-run by later advances of similar or slightly larger magnitude. If so, then the spatial extent of these early Neoglacial expansions may not mean that climatic forcing of these advances was stronger than during the LIA.

Reger et al. (1993) interpreted transported blocks of unconsolidated sediment in the outermost Holocene moraine of Black Rapids Glacier (central Alaska Range) as evidence for a surge origin of the deposit. In addition, many other glaciers in the Alaska Range, eastern Wrangell Mountains and northern Saint Elias Mountains are known to surge or may have surged in the past (Post, 1969). Although surging activity means that the spatial extent of past advances may not directly reflect the level of climatic forcing, the histories of these surging glaciers are generally similar on century and longer timescales to those of non-surging glaciers and so appear to be broadly reflective of climatic forcing.

Some glaciers developed iceberg-calving margins in proglacial lakes during retreat from LIA maxima. In the Kenai Mountains, such glaciers receded farther and faster than nearby non-calving termini (Wiles and Calkin, 1994; Wiles et al., 1999a), and studies at

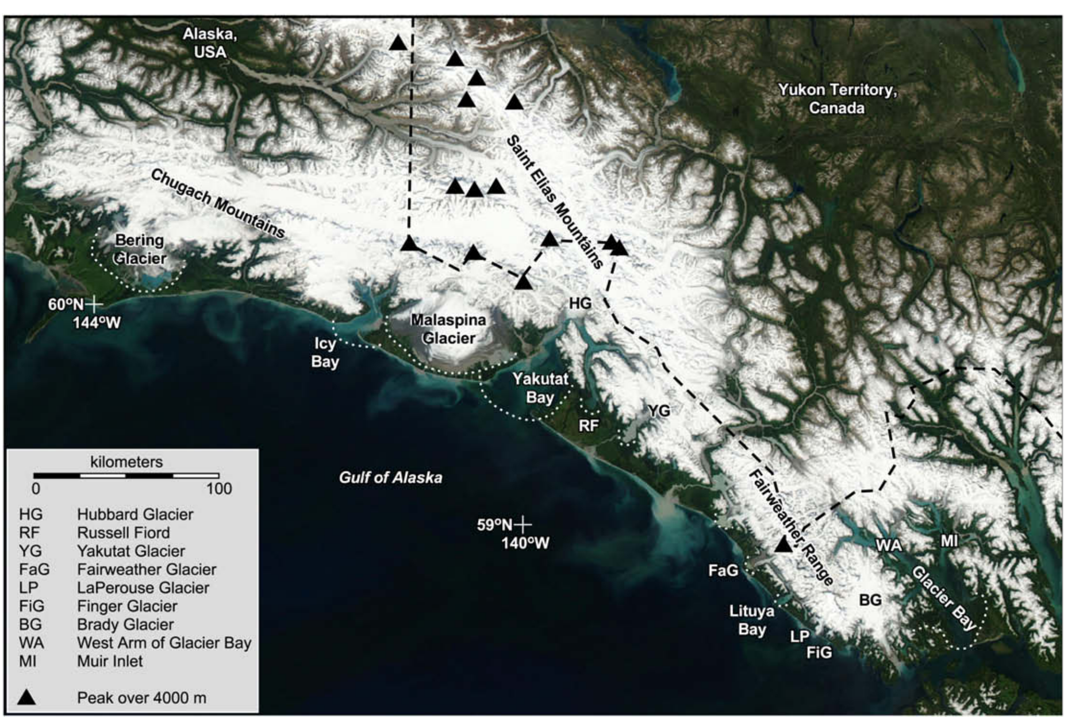

Fig. 9. Map of glacier extents in eastern Chugach and southern Saint Elias mountains. Dotted lines delineate late Holocene maxima for selected termini. MODIS image 2004/171 06/19.

Mendenhall Glacier in the Coast Mountains similarly show increased retreat rates in the 20th century due to iceberg-calving (Motyka et al., 2002; Boyce et al., 2007). Nonetheless, dates for maxima and inception of retreat at these glaciers are no different to those of non-calving termini, because these proglacial lakes and calving margins did not develop until termini had begun to retreat into glacially over-deepened areas behind moraines.

4. Tidewater glaciers

The tidewater glaciers of Alaska are located on the seaward side of the Kenai, Chugach, Saint Elias and Coast mountains (Figs. 1 and 9). There are currently about 50 tidewater termini, although the exact number varies during advance and retreat as termini coalesce, separate or move into or out of tidewater. Almost all termini today are withdrawn into inner fjord areas, but a number of systems have advanced far enough on occasion in the Holocene to calve icebergs directly into the Gulf of Alaska.

Hundreds of tree-ring cross-dates have been used to constrain the most recent advance of Columbia Glacier in the Chugach Mountains (Beckwith-Laube et al., 2004; Wiles, unpublished data). The ice margin was expanding by AD 1020, slowed down through a fjord widening from 1450 to the late 1700s, and then accelerated to reach its maximum stand at *c*.1810. Barclay et al. (2003) used similar tree-ring data to reconstruct the advance of Nellie Juan Glacier in the Kenai Mountains between AD 1539 and 1893. Although both of these expansions occurred largely during the LIA, some of the specific times when these two termini were advancing into forest appear anomalous when compared with records of nearby land-based glaciers. Similarly, Taku Glacier in the Coast Mountains has expanded at times of both advance and retreat of nearby non-tidewater termini (Motyka and Begét, 1996).

This limited synchrony with the land-terminating glacier record is also shown by the histories of the largest tidewater glacier systems (Fig. 10). Tens of radiocarbon ages show that at least three termini made major advances during the early to middle Holocene (Mann and Ugolini, 1985; Barclay et al., 2001; Lawson et al., 2007). Many tidewater glacier expansions lasted hundreds to thousands of years and spanned intervals of both regional ice advance (e.g. FMA and LIA) and regional ice margin retreat (e.g. MWP). There is also limited synchrony between individual tidewater termini, with some glaciers expanding and others retreating or retracted at any given time.

This lack of coherence in tidewater glacier histories suggests that many of these Holocene expansions were primarily compensatory readvances as part of the tidewater glacier cycle (Meier and Post, 1987; Post and Motyka, 1995). This occurs when retreat from a morainal bank after a maximum stand places the ice margin in deep water. Iceberg-calving is faster in deep water (Post, 1975; Van der Veen, 2002; Benn et al., 2007) and so recession continues until the ice margin reaches shallow water, often at the head of the fjord, where a high ice flux to the margin balances the high calving flux. Readvance can then occur with a new morainal bank beneath the terminus (Powell, 1982, 1991), and maxima are usually attained at fjord mouths or junctions where iceberg-calving losses locally increase (Mercer, 1961; Heusser, 1983). Climate change is relatively unimportant in driving these expansions (Mann, 1986a; Warren, 1992).

The history of Hubbard Glacier during the last millennium exemplifies this type of compensatory readvance. About 1000 years ago the Hubbard ice margin stood at the mouth of Yakutat Bay (Fig. 9); recession began before AD 1308 and by 1791 Hubbard Glacier had withdrawn at least 70 km to the head of its fjord (Barclay et al., 2001). This retreat removed almost the entire ablation area while minimally changing the high-elevation

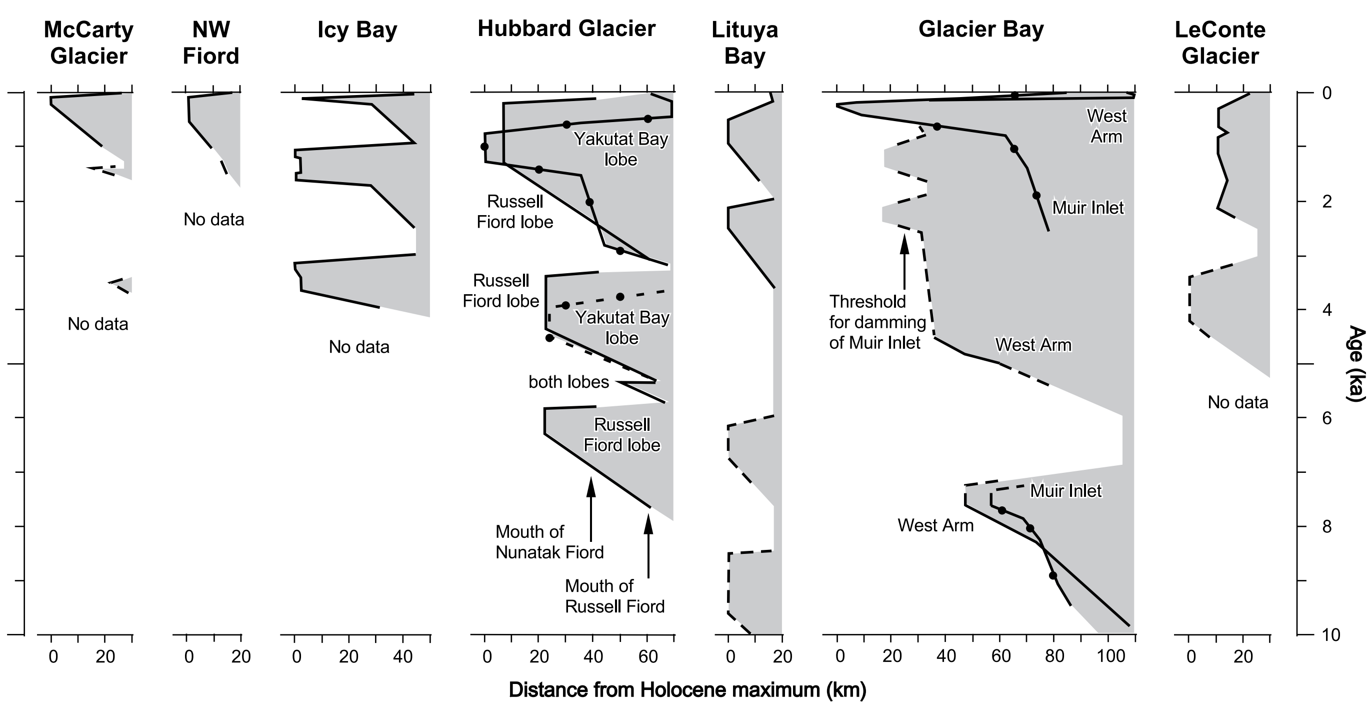

Fig. 10. Time–distance diagrams for large tidewater glacier systems in southern Alaska. McCarty and Northwestern glaciers are in the Kenai Mountains, LeConte is in the Coast Mountains, and the other systems are shown in Fig. 9. McCarty and Northwestern diagrams from Wiles et al. (1995), Icy Bay from Barclay et al. (2006), Hubbard Glacier from Barclay et al. (2001), Lituya Bay based on data in Goldthwait et al. (1963) and Mann and Ugolini (1985), Glacier Bay based on data in Goodwin (1988), Lawson et al. (2007), Monteith et al. (2007) and Molnia (2008), and LeConte based on data in Post and Motyka (1995) and Viens (2005).

Table 2
Moraine and vegetative trimline dates for the most recent maxima of Alaskan tidewater glaciers.

Glacier	Dates[a] (AD)	Reference
McCarty	~1859	Grant and Higgins (1913)
Northwestern	1894–1899	Grant and Higgins (1913)
Aialik	<1700	Wiles and Calkin (1994)
Chenega	1857–1882	Field (1975)
Nellie Juan	1842–1893	Barclay et al. (2003)
Blackstone	>0.60 ka[b]	Heusser (1983)
Harriman	>2.26 ka[b]	Heusser (1983)
Surprise	~1880s	Field (1975)
Barry	1898	Tarr and Martin (1914)
Harvard	>2.42 ka[b]	Heusser (1983)
Yale	1807–1827	Field (1975)
Columbia	1917–1922	Post (1975)
Icy Bay	1886–1904	Tarr and Martin (1914)
Hubbard	<1308	Barclay et al. (2001)
Lituya	~1600	Post and Streveler (1976)
Glacier Bay	1735–1785	Cooper (1937)
Taku	1750	Post and Motyka (1995)
LeConte	1800	Post and Motyka (1995)

[a] Based on direct observations and/or tree-ring minimum dates unless otherwise noted.
[b] Minimum age based on calibrated radiocarbon age of peat.

accumulation area, and so readvance has been continuing for over 100 years as the system moves back towards an extended position. The Hubbard ice margin can probably readvance to near its Holocene maximum position under current (early 21st century) climate conditions (Barclay et al., 2001; Motyka and Truffer, 2007); in the interim the minimal mass loss to surface melting over the small ablation area is offset by very high calving losses (Trabant et al., 1991, 2003).

Climate change does appear to be important in controlling the timing of maxima and in initiating retreat of tidewater glaciers. Moraine dates for tidewater termini in southern Alaska (Table 2) cluster in the same decades as moraine dates for nearby land-based termini (Fig. 6), supporting a hypsometric argument for climatic sensitivity of extended tidewater glaciers (Mann, 1986a; Wiles et al., 1995). However, land-terminating glaciers have receded in a relatively proportional response to climate change, whereas slight retreat of tidewater glaciers has initiated rapid calving retreat leading to tens of kilometers of ice margin recession.

Advance rates derived from the histories of tidewater glaciers show that iceberg-calving margins generally advanced at less than 50 m a^{-1}. This supports rates observed at termini through the 20th century and the inference that advance rates are controlled by the rate at which the morainal bank can be recycled and advanced seaward (Powell, 1982; Meier and Post, 1987). However, well-constrained histories of Taku Glacier and Icy Bay show that ice margins can advance much faster on occasion (Motyka and Post, 1995; Post and Motyka, 1995; Barclay et al., 2006), and records of Hubbard Glacier and Glacier Bay, although less well constrained, also appear to show large and rapid advances (Barclay et al., 2001; Lawson et al., 2007). These large and rapid advances, which entail advance rates of over 100 m a^{-1} sustained for decades, all appear to correspond with expansion into areas of shallow water.

The Hubbard and Glacier Bay systems were linked, respectively, to Yakutat and Brady glaciers (Fig. 9) during the late Holocene (Barclay et al., 2001; Larsen et al., 2004). These two low-elevation ice fields were expanding in the Neoglacial (Derksen, 1976; Wiles et al., 2008) and likely helped make the late Holocene ice expansions of Hubbard Glacier and Glacier Bay more extensive than earlier Holocene advances (Fig. 10). Subsequent retreat of these tidewater glaciers and the post LIA rise in equilibrium lines has made the now-disconnected Yakutat and Brady glaciers two of the fastest down-wasting ice fields in the region (Larsen et al., 2007).

5. Gulf of Alaska foreland glaciers

A number of large glaciers from Bering Glacier to the Lituya Bay area (Fig. 9) flow southwards from the Chugach and Saint Elias mountains to terminate at or close to sea level on low-elevation forelands. These glaciers are fed by high-elevation névés and form broad piedmont lobes that are separated from the Gulf of Alaska by a fringe of Quaternary glacial, glacifluvial and coastal deposits. Radar and seismic data show that Malaspina and Bering glaciers occupy troughs that extend hundreds of meters below sea level (Sharp, 1958a; Molnia and Post, 1995), and offshore data show that these valleys, presumably eroded during the Pleistocene, continue across the continental shelf (Molnia, 1986).

Radiocarbon ages from low-elevation stratigraphic sections and peat cores (Mann, 1986b; Molnia and Post, 1995) show that portions of these Gulf of Alaska forelands were above sea level in the early through middle Holocene. However, ages from marine shells found within late Holocene ice limits at Bering, Malaspina and Yakutat glaciers (Spiker et al., 1978; Pardi and Newman, 1980; Molnia and Post, 1995) indicate that inter-tidal areas extended inland of the current coastline, at least occasionally, during this same interval. The tidewater glacier histories (Fig. 10) show that there was sufficient ice to form large glaciers in these southern Alaskan mountains in the early to middle Holocene, and so it appears that some of these foreland glaciers may have terminated in marine embayments, perhaps with tidewater-calving margins similar to Icy, Yakutat and Lituya bays, during this time.

Subaerial outwash fans may have fronted Bering Glacier by the mid Holocene and enabled an advance around 4.87 ka (Muller and Fleisher, 1995). Farther east, Mann and Ugolini (1985) suggest that Fairweather Glacier formed an outer moraine before 5.75 ka, and Plafker et al. (1978) suggest that Finger Glacier advanced between 6.49 and 4.02 ka. More widespread evidence suggests significant advances of Finger, Fairweather, Yakutat, Malaspina and Bering glaciers after *c.*3.5 ka (Plafker et al., 1978; Pardi and Newman, 1980; Mann and Ugolini, 1985; Heusser, 1995; Molnia and Post, 1995; Muller and Fleisher, 1995; A. Post, unpublished data). Continued growth of the forelands from glacial and coastal sedimentation (Molnia, 1986), along with Neoglacial lowering of the ELA, likely facilitated these expansions.

Tree-ring cross-dates show that both Bering Glacier and its non-surging Stellar Lobe reached their Holocene maximum positions during the LIA (Wiles et al., 1999b). Advances and maxima also occurred at Malaspina, Yakutat, Fairweather, La Perouse and Finger glaciers during the last several hundred years (Sharp, 1958b; Mann and Ugolini, 1985; Barclay et al., 2006; Wiles et al., 2008). However, exact dates for these maxima and the onset of retreat show some variability, reflecting surging activity and the differing response times of these large lowland glacier lobes. Retreat has continued to the present and has been accelerated at some termini by the development of large ice-contact lakes (Muskett et al., 2003; Larsen et al., 2007).

6. Conclusions

(1) Retreat from the Last Glacial Maximum was underway in northern Alaska by 25 ka and by 22–20 ka in southern Alaska. Termini were at or behind modern positions by the start of the Holocene.
(2) Termini of land-based valley glaciers were in retracted positions during the early to middle Holocene. A few studies in the

literature have suggested large advances during this interval but their age control is equivocal.

(3) Neoglaciation was underway in some areas by 4.5–4.0 ka and major advances of land-based termini occurred by 3.0 ka. Two expansions, centered respectively on 3.3–2.9 and 2.2–2.0 ka, may correspond with advance intervals identified in nearby Canada. Early Neoglacial moraines are preserved on some forefields in interior and northern Alaska, but not on forefields in coastal southern Alaska.

(4) A major advance occurred in coastal southern Alaska between the AD 550s and 720s. This first millennium AD (FMA) expansion is also recognized in interior and northern Alaska, and may have encompassed several distinct intervals of advance.

(5) Little Ice Age (LIA) advances were underway in coastal southern Alaska by the AD 1180s–1320s and culminated in two advance phases, respectively in the 1540s–1710s and in the 1810s–1880s. Moraines of these middle and late LIA maxima are invariably the Holocene maxima in coastal southern Alaska. LIA advances are also recognized as major expansions in all glacierized mountain ranges in Alaska, although massive early Neoglacial moraines may have prevented LIA expansions from being the largest of the Holocene on some forefields in interior and northern areas of the state.

(6) Land-based termini throughout Alaska have been retreating since the last phase of LIA advance.

(7) Holocene fluctuations of Alaskan land-terminating glaciers have primarily been forced by multi-decadal and longer timescale changes in temperature. Iceberg-calving into proglacial lakes has accelerated the recent retreat of some termini.

(8) Tidewater glaciers in southern Alaska have fluctuated greatly throughout the Holocene. These advances were often asynchronous with those of adjacent termini, spanned both multidecadal to century length cold and warm intervals, and appear to have been significantly controlled by non-climatic factors. However, rapid iceberg-calving retreats from maxima have largely been initiated during intervals of climatic warming.

(9) Large glaciers terminating on the forelands around the Gulf of Alaska overlie fjord systems and may have had tidewater termini in the early to middle Holocene. Accumulation of glacigenic and coastal deposits has progressively isolated these termini from the adjacent ocean and they have made major advances in response to climate change during the late Holocene.

(10) Tree-ring cross-dates of glacially killed logs provide the most precise basis for reconstructing late Holocene glacier fluctuations in southern Alaska. Continued development and extension of tree-ring chronologies will enable improved assessment of the regional synchrony of glacier advances.

(11) Relatively few studies to date have used either cores from glacier-fed lakes or surface exposure ages to constrain Holocene histories of Alaskan glaciers. Increased application of both methods will improve Holocene-length estimates of glacier activity and improve age control for moraines in interior and northern Alaska.

Acknowledgements

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Appendix

Table A1
Selected radiocarbon ages constraining Holocene advances of Alaskan glaciers.

Glacier	Laboratory number	Uncalibrated age (^{14}C BP)	Calibrated age[a] (ka)	Material dated	Reference
Central Alaska Range:					
Gulkana	I-11,864	5700 ± 260	7.16 (6.52) 5.94	Wood in outwash.	Péwé and Reger (1983), Begét (1994)
Yanert	A-2147	5850 ± 130	6.99 (6.67) 6.32	Peat in Yanert Fork valley.	Ten Brink (1983); Ten Brink and Waythomas (1984)
Southwestern Alaska Range:					
Capps	W-4056	7750 ± 200	9.12 (8.60) 8.17	Base of bog on lateral moraine.	Yehle et al. (1983)
Aleutian Range:					
Cold Bay	GX-2789	10,625 ± 550	13.57 (12.33) 10.79	Peat below tephra.	Thorson and Hamilton (1986)
Cold Bay	GX-2788	6700 ± 330	8.19 (7.57) 6.80	Basal peat over drift.	Thorson and Hamilton (1986)
Northern Saint Elias Mountains:					
White River	I-6231	1255 ± 90	1.31 (1.17) 0.97	Tree killed by White River tephra.	Denton and Karlén (1977)
Russell	Y-2506	1050 ± 100	1.22 (0.97) 0.74	Organic matter overlying moraine.	Denton and Karlén (1977)
Northern Kenai Mountains:					
Blackstone	UW-537	580 ± 55	0.66 (0.60) 0.52	Basal peat.	Heusser (1983)
Coastal Chugach Mountains:					
Harriman	UW-531	2280 ± 60	2.45 (2.26) 2.13	Basal peat.	Heusser (1983)
Harvard	UW-536	2360 ± 65	2.70 (2.42) 2.17	Basal peat.	Heusser (1983)
Bering	PITT-1090	4295 ± 75	5.26 (4.87) 4.58	Transported wood and peat.	Heusser (1995), Muller and Fleisher (1995)
Coastal Saint Elias Mountains:					
Fairweather	β-6776	5000 ± 80	5.91 (5.75) 5.60	Basal organic sediments on moraine.	Mann and Ugolini (1985)
Finger	W-3312	5670 ± 250	7.16 (6.49) 5.92	Wood between tills.	Plafker et al. (1978)
Finger	W-3310	3670 ± 200	4.53 (4.02) 3.47	Peat between tills.	Plafker et al. (1978)

[a] Median age in parentheses together with maximum and minimum ages of 2σ range.

Table A2
Radiocarbon ages of wood on forefields of land-terminating Alaskan glaciers shown in Fig. 6.

Glacier	Laboratory number	Uncalibrated age (^{14}C BP)	Calibrated age[a] (ka)	Sample context	Reference
Wrangell & Saint Elias mountains (interior):					
Nabesna	β-147,584	2330 ± 60	2.69 (2.35) 2.15	Wood in gravel in ice-marginal delta.	Wiles et al. (2002)
Nabesna	β-122,425	2220 ± 50	2.34 (2.23) 2.12	Transported wood in sand and gravel.	Wiles et al. (2002)
Nabesna	β-122,424	2030 ± 60	2.15 (1.99) 1.87	*In situ* stump below ice-marginal delta.	Wiles et al. (2002)
Kuskulana	β-133,815	1760 ± 60	1.82 (1.68) 1.54	*In situ* stump under till and outwash.	Wiles et al. (2002)
Kuskulana	β-133,814	1720 ± 60	1.82 (1.63) 1.45	*In situ* stump under till and outwash.	Wiles et al. (2002)
Kennicott	β-133,808	990 ± 60	1.05 (0.89) 0.75	*In situ* stump.	Wiles et al. (2002)
Kennicott	β-133,810	970 ± 60	0.98 (0.87) 0.74	Roots in soil between tills.	Wiles et al. (2002)
Kennicott	β-133,809	930 ± 60	0.95 (0.84) 0.73	*In situ* stump between tills.	Wiles et al. (2002)
Kennicott	β-83,680	860 ± 60	0.91 (0.78) 0.69	*In situ* stump.	Wiles et al. (2002)
Nizina	β-122,974	1140 ± 60	1.23 (1.06) 0.93	Transported wood between tills.	Wiles et al. (2002)
Seven Sisters	I-6490-C	2780 ± 90	3.16 (2.90) 2.75	Wood below outermost moraine.	Denton and Karlén (1977)
Barnard	β-147,586	2700 ± 60	2.95 (2.81) 2.74	Wood in till.	Wiles et al. (2002)
Barnard	β-133,804	2370 ± 60	2.71 (2.43) 2.21	Wood in till.	Wiles et al. (2002)
Barnard	β-133,803	930 ± 60	0.95 (0.84) 0.73	Transported wood in lateral moraine.	Wiles et al. (2002)
Kenai & Chugach mountains (coastal side):					
Beare	β-95,990	1480 ± 70	1.52 (1.38) 1.29	*In situ* stump below till.	Reyes et al. (2006)
Sheridan	I-1985	2830 ± 100	3.24 (2.97) 2.75	Log in lake sediment.	Tuthill et al. (1968)
Bartlett	W-78	2370 ± 100	2.72 (2.45) 2.16	Transported log in till.	Karlstrom (1964)
Bartlett	W-318	1385 ± 200	1.73 (1.30) 0.83	*In situ* stump between tills.	Karlstrom (1964)
Grewingk	BGS-1278	1440 ± 70	1.52 (1.35) 1.19	*In situ* stump on forefield.	Wiles & Calkin (1994)
Dinglestadt	BGS-1271	1440 ± 70	1.52 (1.35) 1.19	*In situ* stump between tills.	Wiles and Calkin (1994)
Coast Mountains (coastal side):					
Davidson	M-1924	880 ± 100	0.97 (0.81) 0.66	Wood under till and outwash.	Crane and Griffen (1968)
Davidson	M-1922	760 ± 100	0.91 (0.71) 0.55	Tree trunk under gravel.	Crane and Griffen (1968)
Davidson	M-1923	560 ± 100	0.69 (0.58) 0.33	Root in forest bed.	Crane and Griffen (1968)
Gilkey/Battle	Hv-11,306	2190 ± 65	2.34 (2.21) 2.01	*In situ* stump in lateral moraine.	Röthlisberger (1986)
Gilkey/Battle	Hv-11,302	1260 ± 65	1.30 (1.19) 1.01	*In situ* log in lateral moraine.	Röthlisberger (1986)
Gilkey/Battle	Hv-12,095	1195 ± 120	1.34 (1.12) 0.81	Transported log in lateral moraine.	Röthlisberger (1986)
Gilkey/Battle	Hv-11,308	1020 ± 65	1.06 (0.93) 0.78	Transported log in lateral moraine.	Röthlisberger (1986)
Gilkey/Battle	Hv-11,299	930 ± 85	1.04 (0.84) 0.68	*In situ* wood in lateral moraine.	Röthlisberger (1986)
Gilkey/Battle	Hv-11,301	565 ± 55	0.65 (0.59) 0.52	Log in soil in lateral moraine.	Röthlisberger (1986)
Gilkey/Battle	Hv-12,094	515 ± 50	0.65 (0.54) 0.49	Transported log in lateral moraine.	Röthlisberger (1986)
Eagle	Hv-11,293	1815 ± 65	1.88 (1.75) 1.57	Transported logs in lateral moraine.	Röthlisberger (1986)
Eagle	Hv-11,290	470 ± 85	0.65 (0.50) 0.32	Transported wood.	Röthlisberger (1986)
Herbert	β-147,202	2850 ± 70	3.20 (2.98) 2.79	*In situ* stump on forefield.	De Simone (personal comm. 2008)
Herbert	Hv-11,295	2095 ± 95	2.32 (2.08) 1.88	Transported log.	Röthlisberger (1986)
Herbert	Hv-11,296	1745 ± 55	1.81 (1.66) 1.54	Transported wood.	Röthlisberger (1986)
Herbert	Hv-11-294	1410 ± 90	1.53 (1.33) 1.14	Transported wood.	Röthlisberger (1986)
Herbert	HER G3.2	945 ± 24	0.92 (0.85) 0.80	Tree trunk.	Lacher (1999), Weber (2006)
Herbert	GX-24424LS	770 ± 40	0.76 (0.70) 0.66	Tree in soil on forefield.	De Simone (personal comm. 2008)
Herbert	HER G4.5	693 ± 21	0.68 (0.66) 0.57	Transported log.	Lacher (1999), Weber (2006)
Herbert	HER R1.7	647 ± 21	0.67 (0.59) 0.56	*In situ* stump in river.	Weber (2006)
Herbert	β-12,756	610 ± 50	0.66 (0.60) 0.54	*In situ* trunk on forefield.	Motyka and Begét (1996)
Mendenhall	Hv-11,281	2960 ± 65	3.34 (3.13) 2.95	Transported log.	Röthlisberger (1986)
Mendenhall	Y-132-80	2790 ± 130	3.33 (2.93) 2.55	Transported log.	Preston et al. (1955)
Mendenhall	Hv-11,286	2145 ± 60	2.32 (2.14) 1.99	*In situ* tree trunks below till.	Röthlisberger (1986)
Mendenhall	Hv-11,283	2090 ± 60	2.30 (2.07) 1.90	*In situ* tree trunk in soil.	Röthlisberger (1986)
Mendenhall	Hv-11,289	1950 ± 70	2.10 (1.90) 1.72	*In situ* wood below till.	Röthlisberger (1986)
Mendenhall	Hv-11,282	1870 ± 65	1.97 (1.81) 1.62	*In situ* tree trunk.	Röthlisberger (1986)
Mendenhall	L-106C	1790 ± 285	2.36 (1.74) 1.08	Wood under gravel.	Kulp et al. (1951)
Mendenhall	Hv-11,287	1765 ± 60	1.82 (1.68) 1.54	*In situ* wood below till.	Röthlisberger (1986)
Mendenhall	MEN G1.4	1727 ± 23	1.70 (1.64) 1.57	*In situ* stump.	Lacher (1999); Weber (2006)
Mendenhall	Hv-11,284	1280 ± 60	1.30 (1.21) 1.07	Transported wood.	Röthlisberger (1986)
Mendenhall	MEN G1.7	1198 ± 22	1.18 (1.12) 1.06	Transported log.	Lacher (1999), Weber (2006)
Mendenhall	AU-100	1195 ± 150	1.37 (1.11) 0.79	Wood in recent moraine.	Reeburgh and Young (1976)
Mendenhall	Y-132-84	1090 ± 60	1.17 (1.01) 0.92	*In situ* stump in outwash.	Preston et al. (1955)
Mendenhall	MEN R2.6	890 ± 23	0.91 (0.80) 0.74	Wood in river.	Lacher (1999), Weber (2006)
Mendenhall	Unknown	496 ± 22	0.54 (0.52) 0.51	Wood in river.	Weber (2006)
Mendenhall	Hv-11,288	305 ± 55	0.50 (0.38) 0.15	Transported log.	Röthlisberger (1986)
Mendenhall	M-1921	230 ± 100	0.47 (0.25) 0.00	Wood in glacifluvial gravel.	Crane and Griffen (1968)
Paterson	GCW-1	3050 ± 50	3.38 (3.27) 3.08	Transported log in till.	Viens (2005)
Paterson	BF-3	270 ± 50	0.48 (0.34) 0.00	*In situ* stump in outwash.	Viens (2005)

[a] Median age in parentheses together with maximum and minimum ages of 2σ range.

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