The Holocene 14,2 (2004) pp. 185–193

Holocene glacier fluctuations, Waskey Lake, northeastern Ahklun Mountains, southwestern Alaska

Laura B. Levy,[1] Darrell S. Kaufman[1]* and Al Werner[2]

([1]Department of Geology, Northern Arizona University, Flagstaff, AZ 86011, USA; [2]Department of Earth and Environment, Mount Holyoke College, South Hadley, MA 01075, USA)

Received 9 September 2002; revised manuscript accepted 22 November 2002

A HOLOCENE RESEARCH PAPER

Abstract: Lake sediments from Waskey Lake, Ahklun Mountains, southwestern Alaska were studied to decipher the history of upvalley glacier fluctuations during the Holocene. Several indicators of glacier activity were measured including: magnetic susceptibility, organic-matter content, grain-size distribution, bulk-sediment mineralogy and diatom assemblages. Seven radiocarbon ages on macrofossils, along with cross-checks by tephrochronology, provide the chronology of the cores. The results from core WL-1 indicate that glaciers lingered near Waskey Lake until 9100 cal. yr BP, perhaps under conditions of high winter accumulation. Peak organic-matter content occurs at 7400 cal. yr BP, when precipitation might have shifted to summer. The onset of Neoglaciation occurred 3100 cal. yr BP, and glaciers reached their maximum extent ~700 cal. yr BP. This chronology is consistent with the lichenometrically dated moraines from the glacier forefields. Although the ages are tentative, the youngest and most widespread group of moraines was deposited sometime between 650 and 200 cal. yr BP (during the 'Little Ice Age'). Since then, glaciers in the Waskey Lake area have shrunk by ~50% and equilibrium-line altitudes (ELA) have risen by 35 ± 22 m. This rise in ELA is much less than the 100 to 200 m rise observed elsewhere in Alaska and indicates considerable spatial variability in late-Holocene climatic change.

Key words: Glacier variations, lake cores, equilibrium-line altitude, lichenometry, Alaska, late Holocene.

Introduction

One of the most dramatic climatic changes in the Arctic during the Holocene was the widespread cooling that took place during the second half of the epoch. The 'Neoglaciation' of the late Holocene culminated in many places during the 'Little Ice Age' (LIA) of the fourteenth to nineteenth centuries (e.g., Bradley and Jones, 1993), but the synchronicity and intensity of the cooling varied from place to place (e.g., Grove, 1988). Additional records are needed to improve the spatial resolution of the most recent, pre-industrial climatic shift and to lead to an understanding of the spatial and temporal variability of that climatic change, and perhaps those of the future.

This study focuses on the late-Holocene climatic change in an alpine setting of southwestern Alaska. It reconstructs the history of Neoglacial activity of a glaciated basin using both the geomorphic and palaeolimnological records of mountain glaciers. This study follows previous investigations elsewhere in the Arctic that have used distal, proglacial lake sediments to decipher the history of upvalley glacier fluctuations during the Holocene (e.g., Karlén, 1976; Nesje *et al.*, 1991; 2000; 2001; Lemmen *et al.*, 1988; Matthews and Karlén, 1992; Dahl and Nesje, 1994; 1996; Snyder *et al.*, 2000), and lichenometric dating of moraines to determine Neoglacial histories (e.g., Denton and Karlén, 1973; Calkin, 1988). This is the first study from Alaska, however, to combine lacustrine and moraine records, together with reconstructions of equilibrium-line altitudes (ELAs) to document Neoglacial glacier fluctuations.

Study area

The Ahklun Mountains form the most mountainous region in southwestern Alaska, outside of the Aleutian Range (Figure 1). More than 100 modern glaciers are located in the northern part of the range, where summits exceed 1500 m, and hundreds of lakes are impounded by drift throughout this deeply glaciated region. The Waskey Lake valley (Figure 1) was chosen for study because it heads in one of the most extensively glaciated massifs

*Author for correspondence (e-mail: darrell.kaufman@nau.edu)

10.1191/0959683604hl675rp

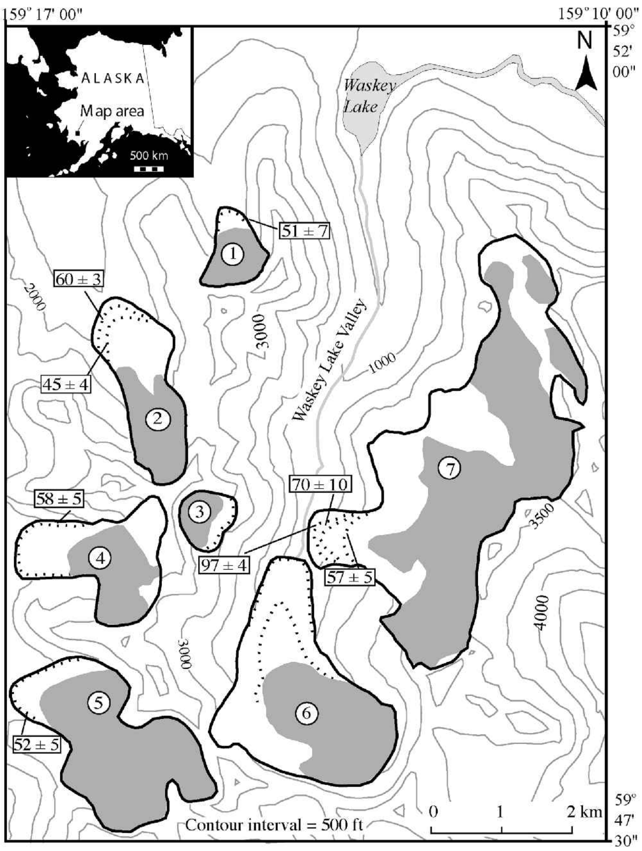

Figure 1 Waskey Lake valley showing the extent of Neoglacial and modern glaciers in the area. Grey areas with numbers are modern glaciers (based on 1979 maps and more recent observations); dark lines delimit the maximum extent of Neoglacial glaciers; dashed lines indicate crests of Neoglacial moraines. Numbers in boxes are lichen diameters (mm; average of the five largest and standard deviation) for moraines on which they were measured.

in the Ahklun Mountains and because the modern glaciers discharge their meltwater into a nearby lake.

Waskey Lake (informal name; 59°52′46″N, 159°12′24″W) is a small (0.2 km^2), relatively shallow (8 m) proglacial lake situated at 150 m above sea level at the mouth of a steep, north–south-trending trough carved ~3 km into a granodiorite pluton. The lake is dammed on the north end by the Waskey Lake moraine, located in the unnamed trunk valley to which the outflow of Waskey Lake is a tributary.

Modern glaciers cover ~20% of the drainage basin of Waskey Lake. The lake is ~6 km downvalley from six active glaciers, each ~1–3 km^2, and several smaller perennial ice masses. Several bouldery, ~2 m high loop moraines delineate recently occupied positions in front of most glaciers that terminate in Waskey Lake valley, and in adjacent valleys that radiate from the same massif.

Glacial geology

Neoglacial moraines are easily identified by their proximity to modern glaciers and by their bouldery surfaces that lack vegetation (Figure 2). All moraines within Waskey Lake valley and two adjacent valleys were mapped on aerial photographs and most were studied in the field. All of the bouldery ridges are within 2 km of all of modern glacier margins, but many lie closer to the glacier termini (~1 km). The moraines delimit former glaciers that averaged ~2.0 km^2 (n = 9), compared to modern glaciers that average ~1.0 km^2 (n = 12), indicating that the late-Holocene glaciers were approximately twice as large as modern glaciers.

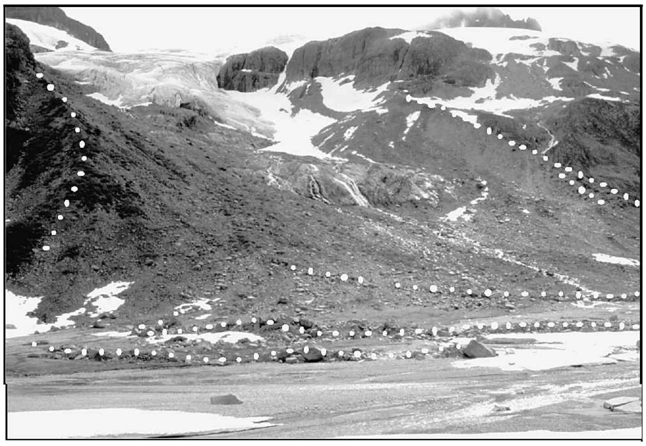

Figure 2 Neoglacial moraines (marked by white dotted line) located in front of westernmost outlet of glacier 7 in Waskey Lake valley (Figure 1). Modern glacier terminus is in upper right corner of view. The distance between the terminus of the modern glacier and the outer moraine loop is ~0.5 km. View is to the east.

Lichenometry was used on the moraines to evaluate the relative timing of their deposition, and possibly the synchronicity of the advances between valleys. Moraine maps were used to reconstruct former glacier extents from which equilibrium-line altitudes were calculated.

Lichenometry

Lichens were measured on eight moraines in front of five glaciers within the Waskey Lake valley and surrounding valleys (Figure 1). Moraines in front of glaciers 3 and 6 were not studied in the field because of their relative inaccessibility. The diameters of at least 25 and as many as 151 individual lichen thalli were measured on each moraine, with more than 600 measured in all (Levy, 2002). Identifying lichens to their species level was not possible in the field (e.g., Werner, 1990), therefore all green-black *Rhizocarpon* lichens were measured and are reported here as *Rhizocarpon geographicum sensu lato* (*s.l.;* Locke *et al.*, 1979). *R. geographicum s.l.* was used because it is distinctive in the field, is the most abundant lichen on the moraines, and it grows slowly. We characterized the lichen size of each moraine using the average of the five largest long-axis thalli (e.g., Innes, 1985; Calkin *et al.*, 1998). Averaging the five largest thalli diameters avoids the effect of a single anomalous lichen, while still using the diameters of lichens that probably were the first to colonize the landform.

Lichen diameters on moraines in the Waskey Lake area fall into three size classes. The largest five lichens (97 ± 4 mm) are found only on one moraine, the outermost Holocene moraine deposited by the western outlet of glacier 7 (Figures 1 and 2). Lichens of intermediate sizes (70 ± 10 mm) were observed on the next younger moraine in the sequence also deposited by glacier 7. This outlet of the amalgamated glacier complex may have responded somewhat differently to subsequent advances if discharge shifted to other outlets; this would explain why younger, more extensive advances experienced by other glaciers did not obliterate these two older moraines. The other six moraines had boulders with average lichen thalli ranging from 45 ± 4 to 60 ± 3 mm, and averaging overall 55 ± 10 mm in diameter. The similar ages of moraines indicated by the lichenometry at five independent glaciers imply a climatic rather than glaciological control to the formation of these moraines.

A lichen growth curve has not been established for the Ahklun Mountains, and therefore absolute ages cannot be assigned to the moraines. However, the lichen data can be compared with lichen growth curves from elsewhere in Alaska to roughly determine the moraine ages. We considered growth curves from four other

Table 1 Temperature and precipitation in the Ahklun Mountains and other areas of Alaska

Location	MAP (mm)	MST (°C)	MAT (°C)
Kenai Mountains	500	17	–1
Seward Peninsula	500	14	–4
Wrangell-St Elias	1000	12	–7
Brooks Range	600	12	–9
Ahklun Mountains	2500	19	–2

Approximate mean annual (MAT) and mean summer temperature (MST), and mean annual precipitation (MAP) taken from maps by the Spatial Climate Analysis Service (2000).

locations: (1) Kenai Mountains (Wiles and Calkin, 1994); (2) Seward Peninsula (Calkin *et al.*, 1998); (3) Wrangell-St Elias Mountains (Denton and Karlén, 1973); and (4) central Brooks Range (Calkin and Ellis, 1980). Climate data (Spatial Climate Analysis Service, 2000) indicate that the mean annual precipitation in the Ahklun Mountains is two to four times higher than the other four areas (Table 1). Temperature in the Kenai Mountains is most similar to the Ahklun Mountains, and the other three areas are colder. Lichen growth rate is dependent on climate (Innes, 1985). We determined the range of possible ages from the fastest growth rate (Kenai Mountains) to the slowest growth rate (Brooks Range) for each lichen-size class (Table 2). Although the climate of Ahklun Mountains is most similar to the Kenai Mountains, the lichen growth curve for the Kenai Mountains is based on faster-growing *Rhizocarpon alpicola* (see Solomina and Calkin, 2003, for a review). The ages derived from the Seward Peninsula and Wrangell-St Elias growth curves are our preferred ages for the moraines because of the relatively similar climates compared to the Ahklun Mountains and because their lichen growth curves are based on the taxa that we measured in the Ahklun Mountains.

On the basis of these two growth curves, the moraine with the largest lichens probably formed sometime between 900 and 1800 yr BP and records a Neoglacial advance prior to the LIA (where BP = before 1950). Although the age range is broad, this relatively small moraine was probably deposited over a short timespan. The moraine with the intermediate lichen diameters was probably deposited sometime between 350 and 1050 yr BP. The youngest and largest number of moraines (n = 6) were probably stabilized sometime between 200 and 650 yr BP. These three moraine ages might correspond with advances at 1150 ± 300, 800 ± 200 and 390 ± 90 yr ago in the central Brooks Range (Ellis and Calkin, 1984) and elsewhere in Alaska (Calkin, 1988). The ages of the youngest moraines overlap with LIA advances in coastal Alaska of the seventeenth and nineteenth centuries (Calkin *et al.*, 2001).

The ages of the three lichen-size classes vary greatly depending on which lichen growth curve is used (Table 2). We cannot rule out the possibility that the moraines are significantly older or younger than the age ranges cited. If the lichen growth rate is as fast as in the Kenai Mountains, then even the oldest moraines in the Ahklun Mountains may date to the LIA. If the rate is as slow as in the Brooks Range, then the oldest moraine may be as old as 3000 cal. BP.

Table 2 Ages of lichen-size classes based on lichen-growth curves from other areas of Alaska (ages in cal. yr BP)

Growth curve location	Ref.*	Taxon†	Oldest (97 mm)	Intermediate (70 mm)	Youngest (55 ± 10 mm)
Kenai Mountains	1	Ra	200	100	100
Seward Peninsula	2	Rg s.l.	900	350	200
Wrangell-St Elias	3	Rg s.l.	1800	1050	650
Brooks Range	4	Rg s.l.	2950	2000	1500

Lichen diameters are shown in Figure 1.
*(1) Wiles and Calkin, 1994; (2) Calkin *et al.*, 1998; (3) Denton and Karlén, 1973; (4) Calkin and Ellis, 1980.
†Ra = *Rhizocarpon alpicola*; Rg s.l. = *Rhizocarpon geographicum s.l.*

Equilibrium-line altitudes (ELAs)

ELAs were estimated for all six reconstructed Neoglacial glaciers where lichen diameters were measured, and for two other Holocene glaciers within adjacent valleys. Only the ELA for the maximum Neoglacial advance was determined. These are probably somewhat different in age for each glacier, but the ELAs do not differ significantly (within ~10 m) between advances. The perimeter of the reconstructed Holocene glaciers was inferred from the position of the moraines and other topographical features mapped on the aerial photographs. ELAs were estimated using the accumulation area ratio method (AAR; Porter, 2000). Torsnes *et al.* (1993) compared different methods of calculating the ELAs of modern and LIA glaciers and found that this method with an assumed AAR of 0.6 ± 0.05 was the most reliable. We too assumed an AAR of 0.6, and employed a geographic information system program linked with a high-resolution digital elevation model to iteratively solve for this AAR (Manley, 2000). Because no glacier mass-balance data are available for the Ahklun Mountains, we used the same method to estimate the ELAs for modern glaciers. ELAs were calculated for the 12 modern glaciers in the area based on their extent shown on the 1:63 360 scale maps from 1979 and modified based on more recent aerial photography and field observations. These modern ELAs were used to determine the amount of Neoglacial ELA lowering.

The ELAs for both modern and reconstructed Neoglacial glaciers included in this study vary widely (Table 3). The reconstructed Neoglacial glaciers located around and in the Waskey Lake valley have ELAs ranging from 680 to 1080 m. This range of ELAs might be attributed to the varying microclimatic effects involving exposure to solar radiation and wind-blown snow. The average ELA in the Waskey Lake region during the Neoglacial maximum was 850 ± 150 m (n = 7), which is 35 ± 22 m below the modern ELA. The largest ELA depression (100 m) is exhibited at glacier 2, where the outermost Neoglacial moraine is ~2 km in front of modern glacial margin. The ELA depression of this glacier was not included in the average because it falls beyond ± 2 σ of the mean of the others.

The Neoglacial ELA depression in the Waskey Lake area is considerably less than that reported for other locations in Alaska. In the Brooks Range, Seward Peninsula and Kenai Mountains, Neoglacial ELAs ranged from 100 to 200 m below modern (Ellis and Calkin, 1984; Calkin *et al.*, 1985; 1998), or three to five times

Table 3 Equilibrium-line altitudes (ELA) for Neoglacial and modern glaciers

Glacier (see Figure 1)	Neoglacial ELA (m)	Modern ELA (m)	ELA lowering (m)
1	680	720	40
2	680	780	100*
3	1080	1090	10
4	860	930	70
5	940	960	20
6	770	790	20
7	970	920	50
Average	850	885	35

*Excluded from average.

more lowering than other locations in Alaska. Other high-latitude regions register similarly dramatic lowerings of ELA during the LIA (e.g., Dahl and Nesje, 1996; Nesje *et al.*, 2001).

Lake cores

Bathymetry and stratigraphy

A bathymetric map of Waskey Lake was constructed and used to locate two coring sites (Figure 3). Acoustic profiling was conducted at the lake but the signal was degraded by abundant trapped gas within the sediments. The bathymetry of Waskey Lake is irregular. The southern end is dominated by an inflow delta creating a shallow platform (< 2 m deep). Two 8 m deep bathymetric lows form the deepest parts of the lake. We recovered two long (~6 m) sediment cores using a percussion corer operated from a floating platform and four companion surface cores. Site WL-1 is on the distal slope of the main bathymetric low and site WL-2 is adjacent to the deepest point (Figure 3).

Core WL-1 is 6.6 m long and comprises four main lithologic units (Figure 4): (1) 6.6 to 6.0 m is mica-rich mud; (2) 6.0 to 4.3 m is crudely bedded dark-brown mud with a 1 cm thick mafic tephra at 512 cm; (3) 4.3 to 2.9 m contains tephra mixed with non-volcanogenic lake sediment; and (4) 2.9 m to the top of the core is weakly bedded mud with intermittent organic-rich layers. The stratigraphy of core WL-2 is similar to WL-1, but contains a shorter record; we therefore focused on WL-1 and its companion surface core, WL-1A.

Geochronology

Radiocarbon

The age of sediment in WL-1 was determined by AMS ^{14}C dating of vegetation macrofossils. The core was sampled at ~75 cm intervals for macrofossils, each generally 2 cm thick. Seven samples contained sufficient material for AMS ^{14}C dating (Table 4). The macrofossils were dominantly floral fragments and woody debris >150 μm diameter, presumably from terrestrial plants, but some samples included a mixture of terrestrial and chitinous aquatic material. Targets were prepared for AMS analysis at the University of Colorado. Radiocarbon ages were calibrated using the midpoint ± 1/2 of the 1 σ calibrated-age range from the online version 4.2 of CALIB (Stuiver and Reimer, 1993). Ages are reported in calibrated years BP (where BP = before 1950; Table 4).

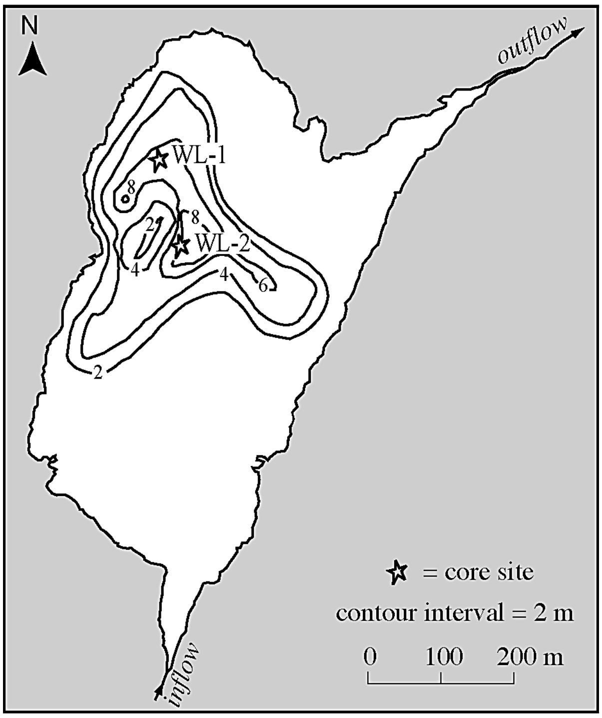

Figure 3 Bathymetry of Waskey Lake with locations of coring sites. The main inflow is from the south. See Figure 1 for location of lake.

The ^{14}C chronology from WL-1 contains no downcore reversals (Figure 5). The basal age of 11 000 ± 200 cal. yr BP indicates that the base of WL-1 penetrated to the latest Pleistocene. This age is in agreement with new cosmogenic surface exposure ages of 12 400 to 11 000 cal. yr BP on boulders from the Waskey Lake moraine that encloses the lake (Briner *et al.*, 2002). Apparently, Waskey Lake became ice-free following a readvance of Younger Dryas age.

An age model was constructed for WL-1 using all seven calibrated ^{14}C ages. The model is divided into three segments, with two linear phases separated by the reworked mud that contains tephra A. The three ^{14}C ages from the lower segment of the core (650 to 440 cm) were fit with a least-squares linear regression ($r^2 = 0.995$). The average sedimentation rate over this interval is 0.29 mm yr^{-1}. The second segment of the age model spans the mud containing tephra A (430–290 cm). We assumed that this unit was deposited instantaneously, consistent with its sharp upper and lower contacts and the lack of internal bedding. The age model for the upper segment of the core, above 290 cm, is based on a linear regression of four ^{14}C ages ($r^2 = 0.937$). In addition, the y-intercept was fixed at the assumed age for the core top (10 yr), which was based on correlation with the surface core from the same site. The correlation of organic-matter fluctuations (see below) indicates that 5 cm of sediment from the top of core WL-1 was lost during recovery. Applying a sedimentation rate of 0.82 mm yr^{-1} (based on regression) indicates that the 5 cm of lost sediment represents ~60 yr, and that the top of the core dates to 10 yr BP.

Tephrochronology

The inferred ages of the two tephras in WL-1 were used as cross-checks on the ^{14}C age model. This is the first attempt to use tephrochronology for lake cores in the Ahklun Mountains. Although the sources of the tephras are not known, they probably originated from Aleutian Arc volcanoes. Major-element geochemical composition, grain size, thickness and stratigraphic order were used to correlate the two tephras with tephras that have previously been dated.

Tephras were pretreated with 2% NaOCl to remove the organic matter, then sonicated several times and rinsed. Separation of volcanic glass from mineral grains was not attempted with heavy liquids. For each sample, 20–25 grains were analysed on a Cameca MBX electron microprobe at Northern Arizona University. One point on each grain was analysed. Standards were analysed between samples and the machine was recalibrated when necessary. Analyses with oxides totalling <98 and >102 wt % were discarded and then total oxides were normalized to 100%.

The upper tephra in WL-1 (tephra A) is contained within a mixture of tephra and lake sediment (Figure 4). The layer of reworked tephra is 140 cm thick (430 to 290 m depth), lacks distinct bedding and is bounded by sharp contacts. The grain size of the unit is dominated by fine sand and is prominent in the grain-size record (see below). Glass shards of tephra A are weakly phenocrystic, brown or colourless. The shards are platy with cylindrical, conical and equant vesicles, and some exhibit bubble-wall junctions. Two distinct geochemical populations are present, one with ~72% SiO_2 and the other with ~59% SiO_2 (Table 5). These populations closely coincide with the two populations of a reference sample from the Aniakchak caldera supplied to us by S.T. Dreher (Alaska Volcano Observatory, Fairbanks, Alaska).

The correlation of tephra A in WL-1 with an eruption of Aniak-

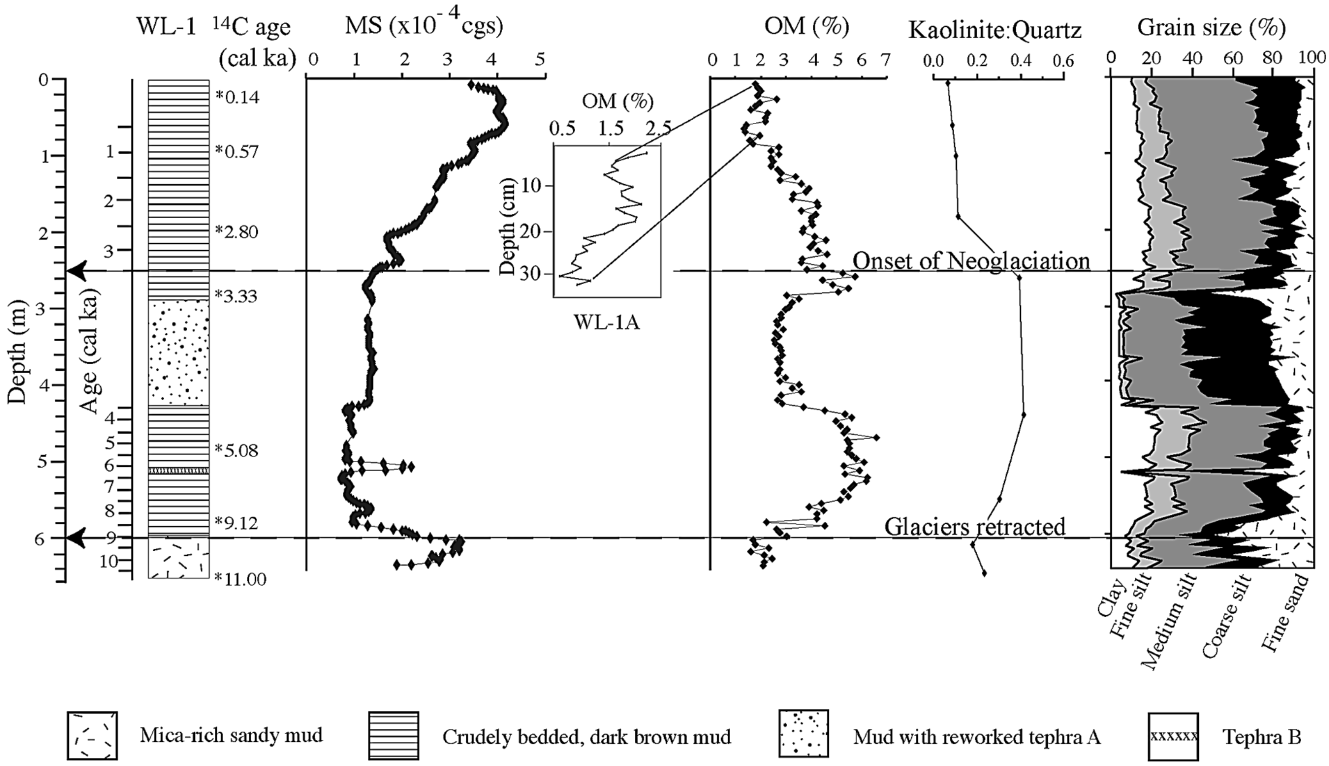

Figure 4 Lithologic log, magnetic susceptibility (MS), organic-matter content (OM), grain-size distribution and kaolinite:quartz for Waskey Lake core 1 (WL-1). Asterisks denote the position of calibrated ^{14}C ages (cal. ka). Age axis is based on model shown in Figure 5. Grain sizes are as follows: clay = <2 μm; fine silt = 2–4 μm; medium silt = 4–20 μm; coarse silt = 20–50 μm; fine sand = 50–250 μm.

Table 4 Radiocarbon dates from Waskey Lake core 1 (WL-1)

Depth (cm)	Lab. I.D. (NSRL)	Radiocarbon age (yr BP)	Cal. age (yr BP)*	Dated material†
16–20	11769	170 ± 35	140 ± 140	Mixed
88–90	11770	550 ± 40	570 ± 50	Vegetation
193–195	11771	2700 ± 35	2800 ± 40	Vegetation
280–282	11772	3140 ± 35	3330 ± 60	Vegetation
478–480	11773	4450 ± 60	5080 ± 200	Vegetation
576–579	11774	8120 ± 70	9120 ± 120	Mixed
650–651	11058	9710 ± 90	11 000 ± 200	Mixed

*Calibrated ages are the midpoint ± 1/2 of 1 σ range from CALIB (Stuiver and Reimer, 1993).
†Macrofossils are either vegetation or mixed (vegetation and chitin).

chak Volcano is consistent with the ^{14}C chronology of WL-1 and with the previously documented widespread distribution of the Aniakchak tephra across western Alaska (Reihle *et al.*, 1987; Beget *et al.*, 1992). Waythomas and Neal (1998) reported a mean age of 3550 ± 30 cal. yr BP for this tephra based on the average of 11 ^{14}C ages. This age agrees with the age model for the upper segment of WL-1. Extrapolating the regression equation for the upper segment downward to the top of the reworked mud with tephra A results in an age of 3560 cal. yr BP. Extrapolating the sediment rate for the lower segment upward to the lowest occurrence of tephra A results in an age of 3300 cal. yr BP, about 250 younger than the tephra. These ages indicate that the reworking of tephra A and the incorporation of ambient lake sediment took place shortly after, if not concurrently with, the eruption of the tephra and that little, if any, of the underlying lake sediment was removed during the reworking.

Tephra B is a pure, 1 cm thick, dark tephra at 512 cm depth in WL-1, ~80 cm below the mud with tephra A (Figure 4). Its grain size is dominantly fine sand to coarse silt. The glass shards are platy, brown, phenocrystic and microlitic. Vesicles are conical and equant, and bubble-wall junctions are present. Tephra B is geochemically homogeneous, varying by only ~4% SiO_2 and ~2% $Na_2O + K_2O$ among individual grains (Table 5). A correlative tephra with similar macro- and microscopic features, and in a similar stratigraphic position below the presumed Aniakchak tephra, is present in four cores taken from Arolik Lake, 115 km southwest of Waskey Lake (Kaufman *et al.*, 2003). In Arolik Lake, the correlative tephra has been dated at 6100 cal. yr BP, which agrees with the age of 6110 cal. yr BP calculated according to the age model for WL-1.

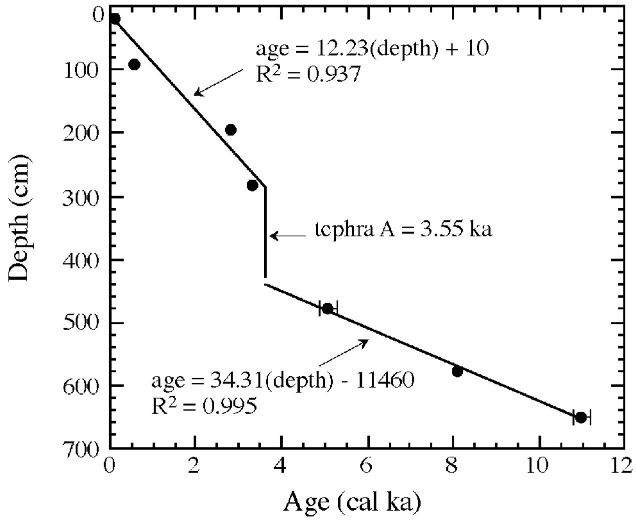

Figure 5 Age model for Waskey Lake core 1 (WL-1). Equations are least-squares linear regressions of three line segments discussed in the text. Error bars indicate the one sigma range of calibrated ages; symbols without bars are wider than age ranges (data listed in Table 4).

Table 5 Major-element geochemistry of tephras from Waskey Lake core 1 (WL-1)

Sample	Na_2O	MgO	Al_2O_3	SiO_2	P_2O_5	Cl	K_2O	CaO	TiO_2	MnO	FeO
Tephra A_1	3.68	2.94	16.69	59.86	0.64	0.12	1.61	6.16	1.33	0.19	6.78
n = 7	(0.40)	(0.37)	(0.79)	(1.25)	(0.05)	(0.02)	(0.09)	(0.53)	(0.11)	(0.04)	(0.85)
Tephra A_2	3.48	0.56	15.02	72.70	0.09	0.14	3.09	1.93	0.55	0.11	2.34
n = 6	(0.58)	(0.31)	(0.46)	(1.23)	(0.04)	(0.04)	(0.51)	(0.51)	(0.09)	(0.06)	(0.54)
$Aniakchak_1$	4.05	2.76	16.08	59.93	0.58	0.12	1.66	6.16	1.40	0.22	7.06
n = 15	(0.31)	(0.37)	(0.77)	(0.80)	(0.08)	(0.03)	(0.21)	(0.41)	(0.08)	(0.04)	(0.75)
$Aniakchak_2$	4.32	0.31	14.33	72.50	0.12	0.13	3.63	1.65	0.61	0.09	2.31
n = 7	(0.77)	(0.26)	(1.44)	(1.20)	(0.15)	(0.05)	(0.57)	(0.35)	(0.34)	(0.06)	(0.83)
Tephra B	3.46	1.97	15.32	62.88	0.22	0.13	1.74	4.95	1.24	0.21	7.88
n = 10	(0.42)	(0.13)	(0.49)	(0.03)	(0.01)	(0.01)	(0.13)	(0.19)	(0.06)	(0.03)	(0.48)

Two populations indicated by subscripts; n = number of glass shards analysed; mean values reported with standard deviation in parentheses. Complete data are in Levy (2002).

Indicators of glacier activity

Magnetic susceptibility

Whole-core (volume) magnetic susceptibility (MS) was measured prior to splitting the cores. Measurements were taken every 3 cm using a Sapphire SI-2 loop detector, and air measurements were taken every ~30 cm. The MS record of WL-1 varies in relation to lithological changes and exhibits distinctive maxima over the lower 0.5 m and the top 2.0 m of the core (Figure 4). MS is relatively high (~4 $\times$ 10^{-4} cgs) at the base of WL-1, corresponding to the basal unit of dense, sandy, mica-rich mud. The presence of strongly magnetic minerals, probably magnetite, appears to be responsible for the high MS in this lower unit. At 6.0 m, the top of the mica-rich unit, the MS sharply decreases. In the overlying silty mud, the MS decreases to ~1 $\times$ 10^{-4} cgs, but is interrupted by a sharp peak at 512 cm corresponding to tephra B. MS increases slightly in the overlying reworked mud with tephra A (unlike tephra that we ascribed to Aniakchak in other lakes from the area; Kaufman *et al.*, 2003). MS remains low from 290 to 230 cm, before increasing in a stepwise fashion to 4 $\times$ 10^{-4} cgs at 60 cm. MS remains high up to ~20 cm below the top of the core before decreasing. The stepwise increase is reproduced in our second core (WL-2; Levy, 2002), lending confidence to these results.

Loss-on-ignition

Loss-on-ignition (a proxy for total organic matter (OM) and total inorganic carbon (TIC)) was measured at regular intervals (5 cm for long cores and 1 cm for surface cores) using the procedure described by Dean (1974) and recently re-evaluated by Heiri *et al.* (2001). OM content at the base of WL-1 (650–590 cm) is relatively low, ranging between 1.0 and 2.5% (Figure 4). The values increase to ~6% at ~590 cm and remain relatively high to the base of the reworked tephra unit at 430 cm. OM content is nearly constant at ~2.5% throughout the tephra-bearing unit, then sharply increases to values similar to those immediately below the tephra-bearing unit. At ~260 cm, OM content decreases approximately inversely to MS, reaching values similar to the base of the core at ~100 cm. The surface core (WL-1A) shows trends similar to the top of WL-1, and exhibits an increase of ~1% at 25 cm depth, in concert with the MS decrease in WL-1. The TIC in WL-1 is <1% throughout (Levy, 2002) and is considered negligible.

Grain-size distribution

Grain size was measured using a Coulter LS230 laser diffraction grain-size analyser. Approximately 0.1 cc of wet sediment was sampled at 5 cm intervals and pretreated with 30% H_2O_2 to remove organic detritus, and with 1M NaOH to remove biogenic silica prior to analysis. Samples were stored in dispersant (sodium hexametaphosphate) until analysis to prevent grains from flocculating. Each sample was analysed six times, with each run lasting ~80 sec. The bottom 0.5 m of WL-1 contains a relatively high percentage (~35%) of fine sand (Figure 4). Above 600 cm, where MS decreases and OM increases, fine silt and clay increase at the expense of coarse particles. The grain size of the reworked tephra unit is distinct, with 40% medium silt, 35% coarse silt and 15% fine sand. Clay and fine silt are nearly absent (~10% total) throughout this unit. Above the tephra-bearing unit, the grain-size distribution returns to that of the sediment below the tephra-bearing unit. The top 2 m of WL-1 shows a coarsening-upward trend with medium silt increasing from 30 to 50%, and clay and fine silt decreasing from 30 to ~18%.

Bulk-sediment mineralogy

Nine samples from WL-1 were analysed using X-ray diffraction to detect changes in bulk-sediment mineralogy that might indicate glacier fluctuations upvalley. Sediment for XRD analysis was sampled from each of the major lithostratigraphic units. Samples were dried at 40°C overnight and then lightly crushed. Mineralogy was analysed using a Siemens Defraktometer 550. The proportion of kaolinite (4.5Å) to quartz (4.3Å) was estimated by comparing their peak areas. Kaolinite is a product of chemical weathering and, therefore, an increase in kaolinite relative to quartz indicates an increase in the input of sediment affected by chemical versus mechanical weathering (e.g., Menking, 1997). The kaolinite:quartz ratio doubles between 6 and ~5 m (from 0.2 to 0.4) and it remains high through 2 m, where it returns to low values (~0.1) for the remainder of the core (Figure 4).

Diatom assemblages

Two sediment samples from WL-1 were analysed for diatoms. The samples from the lower (640 cm) and upper (50 cm) stratigraphic units were analysed to determine if there were significant changes in the diatom assemblages that might reflect greater lake ice, turbidity or changes in inflow. Both samples contained low abundances of diatoms and a predominance of benthic species suggesting that the environment was marginal for algal growth. The floras of the two samples differ significantly. The bottom sample is dominated (~40%) by small benthic *Fragilaria* taxa (*F. brevistriata* v. *inflata*, *F. pseudoconstruens* and *F. pinnata*), which are indicative of alkaliphilic environments. The small benthic *Fragilaria* taxa are also pioneering taxa, successful in harsh conditions (e.g., under short growing season, in moats of ice-covered lakes; Smol, 1988). In the upper sample, small benthic *Fragilaria* spp. make up <5% of the assemblage, and instead the community is dominated by species in the genera *Achnanthes*, *Cymbella* and

Navicula, as well as indicators of stream input (*Hannaea arcus* v. *arcus* and *Meridion circulare*).

Interpretation of lake-core data

The lower unit (6.6 to 6.0 m) of WL-1 was deposited 11 200 to 9100 cal. yr BP, following the retreat of ice from its pronounced lateglacial readvance (Briner *et al.*, 2002). This basal unit exhibits high MS and low OM content. The high MS reflects the presence of magnetite, which is derived from the granodiorite that underlies the glaciated watershed (Hoare and Coonrad, 1978). We suggest that the flux of magnetite is proportional to the aerial extent of ice within the watershed and the rate that it erodes its bed (i.e., velocity). Apparently, these remained high as glacier ice lingered in this alpine setting until ~9100 cal. yr BP. Further indicators of ice proximity include: (1) low OM content suggesting low level of organic productivity within the lake and/or its catchment; (2) depauperate diatom assemblage suggesting cold, turbid water; (3) relatively poorly sorted grain size suggesting high-energy fluvial input; (4) low kaolinite:quartz ratio suggesting a relatively high proportion of mechanically weathered sediment (i.e., rock flour); and (5) strong affinity between the major-element geochemistry of this unit and that of outwash in the active glacier floodplain in Waskey Lake valley (discussed by Levy, 2002).

The sharp decrease in MS and the dramatic fining of grain size in the overlying unit (600–430 cm = 9100–3500 cal. yr BP) suggest a reduction in the extent of glaciated area and its erosive capacity (i.e., glacial activity), and perhaps an increase in the distance to the glacier termini. The proportion of kaolinite to quartz increases through this unit, indicating that the influx of mechanically weathered sediments decreased. OM content increased rapidly, attaining maximum Holocene levels ~7400 cal. yr BP. The linear sedimentation rate indicates that the increase in OM content is not an artifact of progressively lower sediment flux, but more likely a response to astronomically driven summer insolation maximum that lengthened the growing season and promoted productivity in the lake and the watershed. Other authors (e.g., Nesje *et al.*, 1991; 2000; 2001) have associated increasing OM content of proglacial lake sediment to retreating glaciers.

The 30 cm thick interval overlying the mud with reworked tephra exhibits MS, OM and grain-size characteristics identical to the sediment that underlies the tephra-bearing unit. This indicates that the limnological system recovered quickly following the deposition of the tephra and that the physical changes that occur in this 30 cm thick interval are not a response to the perturbation caused by the deposition of the tephra on the landscape.

At 260 cm (3200 cal. yr BP), MS increases, OM content decreases and grain size coarsens. This coarsening is registered by an increase in medium and coarse silt, typical grain size for rock flour derived from plutonic source rocks (Haldorsen, 1981). This, together with the increase in the flux of magnetic minerals, suggests that glaciers were reactivated and increased their meltwater discharge into Waskey Lake. The OM decreases upward through this unit; given the linear sedimention rate, this suggests a decrease in productivity rather than a progressive dilution by clastic material. Sedimention rate is higher overall in this interval (0.82 mm yr^{-1}), nearly three times the rate of the bottom half of the core, despite slightly higher bulk densities upwards (Levy, 2002). The proportion of kaolinite decreases, indicating an increase in mechanically weathered sediment. MS peaks and OM reaches its minimum at ~60 cm and remains relatively constant, indicating that the glaciers reached their maximum Neoglacial extent sometime following ~700 cal. yr BP The decrease in MS in the upper 20 cm of WL-1, and the increase in OM content over the upper 25 cm of WL-1A, signals the retreat of glaciers beginning about 250 cal. yr BP.

Other studies on lake cores from glaciated basins used similar data to derive similar conclusions. For example, Leonard's (1986) study of lake cores from Banff National Park, Canada, found that low OM content combined with high sedimentation rates were indicative of a high glacial sediment contribution. Similarly, Benson *et al.* (1998) attributed high MS and low OM content to glacial intervals in the Sierra Nevada of California and Nevada.

Conclusions and integration

This study combines morainal evidence of upvalley glacier fluctuations with the continuous sediment record of a downvalley lake to determine the timing and extent of Neoglacial glacier fluctuations. The sediments of Waskey Lake appear to reliably record glacier fluctuations in its drainage basin. In part, this might be attributed to the relatively large (but not too large) glaciated area proximal to the lake. At the peak of Neoglaciation in the Waskey Lake valley, the glaciers covered 40% of the drainage basin. Since ~250 yr BP ice cover has shrunk by 50%. A similar study at Sunday Lake, a larger lake with significantly less glacial ice located ~25 km to the southwest, apparently did not register Neoglacial ice fluctuations (Levy, 2002), contrary to earlier interpretations (Feinberg, 2000).

Glaciers, larger than any that formed during the late Holocene, persisted near Waskey Lake until 9100 cal. yr BP, although no moraines dating to the early Holocene were found. Around this time, palaeoecological and geochemical evidence from elsewhere in southwestern Alaska indicates a thermal maximum (e.g., Hu *et al.*, 1998; Axford, 2000). The juxtaposition of significant glacier ice extent concurrent with peak warmth suggests that glaciers survived under conditions of abundant winter precipitation, more than any time since. Higher temperatures may have reduced the duration of ice cover over the large lakes proximal to Waskey Lake, resulting in an increase of lake-effect precipitation. On the other hand, mountain glaciers do not appear to have persisted long after the Younger Dryas interval in Little Swift Lake, just 55 km north of Waskey Lake (Axford, 2000). Peak OM content in Waskey Lake sediment was attained by ~7400 cal. yr BP, within 2000 yr following deglaciation of the drainage basin. We do not know how far the glaciers retreated within the basin, but they were less extensive than the present, and might have been melted entirely. Widespread evidence from the region (Hu *et al.*, 1998), and elsewhere in Alaska (Edwards *et al.*, 2001), indicates an increase in effective moisture after about 9100 cal. yr BP. The reduction in glacier extent after 9000 cal. yr BP suggests a shift to summer precipitation in the Waskey Lake area (cf. Hu *et al.*, 1995).

The moraine evidence does not attest to the onset of Neoglaciation because advance-phase deposits were probably obliterated. In contrast, the onset of Neoglaciation is clearly recognized in the lake sediment at 3100 cal. yr BP. This age is somewhat younger than suggested for the culmination of major Neoglacial advances elsewhere in Alaska, including at 4000–3500 cal. yr BP in the Brooks Range (Ellis and Calkin, 1984), at 4000–3000 cal. yr BP on the Seward Peninsula (Calkin *et al.*, 1998) and at 3600 cal. yr BP in the Kenai Mountains (Wiles and Calkin, 1994). However, we cannot rule out the possibility that some glaciological or sedimentological threshold had to be crossed before lake sediments in Waskey Lake began registering the glacial influence.

Multiple late-Holocene terminal end-moraines in the forefield of modern glaciers are in accord with the sedimentological indicators from core WL-1. Similar-age moraines (based on lichen diameters) in Waskey Lake valley and in adjacent valleys indicate climatic, rather than glaciological, controls. On the basis of tentative lichenometric age estimates, most moraines appear to record the peak of Neoglaciation sometime between 650 and 200 yr BP. This age agrees with the lake-core record, which suggests peak Neoglacial activity from ~700 to 250 yr BP, and is consistent with studies in other parts of Alaska (e.g., Calkin, 1988; Wiles and Calkin, 1994) and elsewhere (e.g., Grove, 1988).

The ELA in the Waskey Lake region was lowered by only ~35 ± 22 m (n = 6) during the maximum Neoglaciation, which is three to five times less than for other regions of Alaska (e.g., Calkin, 1988) and elsewhere in the Northern Hemisphere (e.g., Dahl and Nesje, 1996; Nesje *et al.*, 2001). Assuming a lapse rate of 6°C/km (Porter, 2000) and no change in accumulation rate, this minor ELA depression indicates that temperature was only 0.2 ± 0.2°C lower than present during the Neoglaciation maximum. If accumulation rates were lower, which is likely, then summer temperatures would also have been lower. Regardless, either Neoglacial cooling in the Ahklun Mountains was less pronounced than elsewhere in Alaska, or twentieth-century warming has been more extreme in other mountainous regions of the state. This apparent discrepancy adds to evidence of the spatial variability of climatic changes across the Arctic during the past several centuries (Overpeck *et al.*, 1997).

Acknowledgements

Yarrow Axford, Jason Briner, Ali Feinberg and Ben Laabs assisted in the field. Jocelyn Turnbull prepared the targets for AMS analysis; James Wittke helped with the tephra geochemistry; Roderic Parnell assisted with the sediment mineralogy; Irene Gregory-Eaves analysed the diatoms; Kristi Wallace and Christian deFontaine helped examine the tephras; Bill Manley contributed to the spatial analyses; Benjamin Borowski provided computer support; Atle Nesje and Parker Calkin provided helpful comments on an early draft; Togiak National Wildlife Refuge provided logistical support. This study was funded by NSF-ESH grants (ATM-9808593 and EAR-9809330) and a GSA graduate student research grant, and is PARCS contribution no. 191.

References

Axford, Y. 2000: Late Quaternary glacier fluctuations and vegetation change in the northwestern Ahklun Mountains, southwestern Alaska. Masters thesis, Utah State University, Logan, UT.

Beget, J.E., Owen, M. and **Anderson, P.** 1992: Age, extent and climatic significance of the *c*. 3400 BP Aniakchak tephra, western Alaska, USA. *The Holocene* 2, 51–56.

Benson, L.V., May, H.M., Antweiler, R.C., Brinton, T.I., Kashgarian, M., Smoot, J.P. and **Lund, S.P.** 1998: Continuous lake-sediment records of glaciation in the Sierra Nevada between 52,600 and 12,500 ^{14}C yr BP. *Quaternary Research* 50, 113–27.

Bradley, R.S. and **Jones, P.D.** 1993: 'Little Ice Age' summer temperature variations: their nature and relevance to recent global warming trends. *The Holocene* 3, 367–76.

Briner, J.P., Kaufman, D.S., Werner, A., Caffee, M., Levy, L.B., Kaplan, M.R. and **Finkel, R.C.** 2002: Glacier readvance during the late glacial (Younger Dryas?) in the Ahklun Mountains, southwestern Alaska. *Geology* 30, 679–82.

Calkin, P.E. 1988: Holocene glaciation of Alaska (and adjoining Yukon Territory, Canada). *Quaternary Science Reviews* 7, 159–84.

Calkin, P.E. and **Ellis, J.M.** 1980: A lichenometric dating curve and its application to Holocene glacial studies in the central Brooks Range, Alaska. *Arctic and Alpine Research* 12, 245–64.

Calkin, P.E., Ellis, J.M., Haworth, L.A. and **Burns, P.E.** 1985: Cirque glacier regime and neoglaciation, Brooks Range, Alaska. *Zeitschrift für Gletscherkunde und Glazialgeologie* 21, 371–78.

Calkin, P.E., Kaufman, D.S., Przybyl, B.J., Whitford, W.B. and **Peck, B.J.** 1998: Glacier regimes, periglacial landforms, and Holocene climate change in the Kigluaik Mountains, Seward Peninsula, Alaska, USA. *Arctic and Alpine Research* 30, 154–65.

Calkin, P.E., Wiles, G.C. and **Barclay, D.J.** 2001: Holocene coastal glaciation of Alaska. *Quaternary Science Reviews* 20, 449–62.

Dahl, S.O. and **Nesje, A.** 1994: Holocene glacier fluctuations at Hardangerjøkulen, central southern Norway: a high-resolution composite chronology from lacustrine and terrestrial deposits. *The Holocene* 4, 269–77.

—— 1996: A new approach to calculating Holocene winter precipitation by combining glacier equilibrium line altitudes and pine-tree limits: a case study from Hardangerjøkulen, central south Norway. *The Holocene* 6, 381–98.

Dean, W.E. 1974: Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: comparison with other methods. *Journal of Sedimentary Petrology* 44, 242–48.

Denton, G.H. and **Karlén, W.** 1973: Lichenometry; its application to Holocene moraine studies in southern Alaska and Swedish Lapland. *Arctic and Alpine Research* 5, 347–72.

Edwards, M.E., Mock, C.J., Finney, B.P., Barber, V.A. and **Bartlein, P.J.** 2001. Potential analogues for paleoclimatic variations in eastern interior Alaska during the past 14,000 yr: atmospheric-circulation controls of regional temperature and moisture responses. *Quaternary Science Reviews* 20, 189–202.

Ellis, J.M. and **Calkin, P.E.** 1984: Chronology of Holocene glaciation, central Brooks Range, Alaska. *Geological Society of America Bulletin* 95, 897–912.

Feinberg, A.E. 2000: Evidence of Holocene glacial activity from the Ahklun Mountains, southwestern Alaska. Bachelor of Arts thesis, Mount Holyoke College, South Hadley, MA.

Grove, J.M. 1988: *The Little Ice Age*. London: Methuen.

Haldorsen, S. 1981: Grain-size distribution of subglacial till and its relation to glacial crushing and abrasion. *Boreas* 10, 91–105.

Heiri, O., Lotter, A.F. and **Lemcke, G.** 2001: Loss on ignition as a method for estimating organic and carbonate content in sediments: reproducibility and comparability of results. *Journal of Paleolimnology* 25, 101–10.

Hoare, J.M. and **Coonrad, W.L.** 1978: Geologic map of the Goodnews and Hagemeister Island Quadrangles region, southwestern Alaska. *US Geological Survey Open File Report* 78–9.

Hu, F.S., Brubaker, L.B. and **Anderson, P.M.** 1995: Postglacial vegetation and climate change in the northern Bristol Bay region, southwestern Alaska. *Quaternary Research* 3, 382–92.

Hu, F.S., Ito, E., Brubaker, L.B. and **Anderson, P.M.** 1998: Ostracode geochemical record of Holocene climatic change and implications for vegetational response in the northwestern Alaska Range. *Quaternary Research* 49, 86–95.

Innes, J.L. 1985: Lichenometry. *Progress in Physical Geography* 9, 187–254.

Karlén, W. 1976: Lacustrine sediments and tree-limit variations as indicators of Holocene climatic fluctuations in Lapland, northern Sweden. *Geografiska Annaler* 58A, 1–34.

Kaufman, D.S., Hu, F.S., Briner, J.P., Werner, A., Finney, B.P. and **Gregory-Eave, I.** 2003: A ~33,000 year record of environmental change from Arolik Lake, Ahklun Mountains, Alaska. *Paleolimnology* 29, in press.

Lemmen, D.S., Gilbert, R., Smol, J.P. and **Hall, R.I.** 1988: Holocene sedimentation in glacial Tasikutaaq Lake, Baffin Island. *Canadian Journal of Earth Science* 25, 810–23.

Levy, L.B. 2002: A record of late Holocene glacier fluctuations, Ahklun Mountains, southwestern Alaska. Masters thesis, Northern Arizona University, Flagstaff, AZ.

Leonard, E.M. 1986: Use of lacustrine sedimentary sequences as indicators of Holocene glacial history, Banff National Park, Alberta, Canada. *Quaternary Research* 26, 218–31.

Locke, W.W. II, Andrews, J.T. and **Webber, P.J.** 1979: A manual for lichenometry. *British Geomorphological Research Group Technical Bulletin* 26, 1–47.

Manley, W.F. 2000: Demonstration of GIS applications in paleoglaciology, glaciology, and bathymetry, Alaska. *Abstracts, 30th Annual Arctic Workshop (Boulder, Colorado),* 14–19 March, 20–21.

Matthews, J.A. and **Karlén, W.** 1992: Asynchronous neoglaciation and Holocene climatic change reconstructed from Norwegian glaciolacustrine sedimentary sequences. *Geology* 20, 991–94.

Menking, K.M. 1997: Climatic signals in clay mineralogy and grain-size variations in Owns lake core OL-92, eastern California. In Bischoff, J., editor, *An 800,000-year paleoclimatic record from core OL-92, Owens Lake, southeast California,* Boulder, CO: Geological Society of America Special Paper 317, 25–36.

Nesje, A., Dahl, S.O., Andersson, C. and **Matthews, J.A.** 2000: The lacustrine sedimentary sequence in Sygneskardvatnet, western Norway: a continuous, high-resolution record of the Jostedalsbreen ice cap during the Holocene. *Quaternary Science Reviews* 19, 1047–65.

Nesje, A., Kvamme, M., Rye, N. and **Løvlie, R.** 1991: Holocene glacial and climate history of Jostedalsbreen region, western Norway; evidence from lake sediments and terrestrial deposits. *Quaternary Science Reviews* 10, 87–114.

Nesje, A., Matthews, J.A., Dahl, S.O., Berrisford, M. S. and **Andersson, C.** 2001: Holocene glacier fluctuations of Flatebreen and winter-precipitation changes in the Jostedalsbreen region, western Norway, based on glaciolacustrine sediment records. *The Holocene* 11, 267–80.

Overpeck, J., Hughen, K., Hardy, D., Bradley, R., Case, R., Douglas, M., Finney, B., Gajewski, K., Jacoby, G., Jennings, A., Lamoureux, S., Lasca, A., MacDonald, G., Moore, J., Retelle, M., Smith, S., Wolfe, A. and **Zielinski, G.** 1997: Arctic environmental changes of the last four centuries. *Science* 278, 1251–56.

Porter, S.C. 2000: Snowline depression in the tropics during the Last Glaciation. *Quaternary Science Reviews* 20, 1067–91.

Reihle, J.R., Meyer, C.E., Ager, T.A., Kaufman, D.S. and **Ackerman, R.E.** 1987: The Aniakchak tephra deposit – a late Holocene marker horizon in western Alaska. In Hamilton, T.D. and Galloway, J.P., *US Geological Survey (USGS) Circular* 998, 19–22.

Smol, J.P. 1988: Chrysophycean microfossils in paleolimnological studies. *Palaeogeography, Palaeoclimatology, Palaeoecology* 62, 287–97.

Snyder, J.A., Werner, A. and **Miller, G.H.** 2000: Holocene cirque glacier activity in western Spitsbergen, Svalbard, sediment records from proglacial Linnevatnet. *The Holocene* 10, 555–63.

Solomina, O. and **Calkin, P.E.** 2003: Lichenometry as applied to moraines in Alaska, U.S.A., and Kamchatka, Russia. *Arctic, Antarctica, Alpine Research* 35, 129–43.

Spatial Climate Analysis Service (SCAS) 2000: PRISM data set 1961–1990, mean monthly precipitation and temperature for Alaska and Yukon Territory. Oregon State University. Digital data set http://www.ocs.orst.edu/prism/ (last accessed June 2002).

Stuiver, M. and **Reimer, P.J.** 1993: Extended ^{14}C database and revised CALIB 3.0 ^{14}C calibration program. *Radiocarbon* 35, 215–30.

Torsnes, I., Rye, N. and **Nesje, A.** 1993: Modern and Little Ice Age equilibrium-line altitudes on outlet valley glaciers from Jostedalsbreen, western Norway; an evaluation of different approaches to their calculation. *Arctic and Alpine Research* 25, 106–16.

Waythomas, C.F. and **Neal, C.A.** 1998: Tsunami generation by pyroclastic flow during the 3500-year BP caldera-forming eruption of Aniakchak Volcano, Alaska. *Bulletin of Volcanology* 60, 110–24.

Werner, A. 1990: Lichen growth rates for the northwest coast of Spitsbergen, Svalbard. *Arctic and Alpine Research* 22, 129–40.

Wiles, G.C. and **Calkin, P.E.** 1994: Late Holocene, high-resolution glacial chronologies and climate, Kenai Mountains, Alaska. *Geological Society of America Bulletin* 106, 281–303.