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Late Glacial and Holocene Glacier and Vegetation Fluctuations at Little Swift Lake, Southwestern Alaska, U.S.A.

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Abstract

Multiproxy data from Little Swift Lake, an alpine lake in southwestern Alaska, provide evidence for pronounced late glacial and Holocene environmental change. An alpine glacier upvalley of Little Swift Lake retreated following the Younger Dryas chronozone, as evidenced by sedimentological changes in the lake record. Glacier retreat was accompanied by local and regional vegetation changes, including the expansion of *Betula* and contraction of Cyperaceae, in response to climatic amelioration. Warm, moist conditions between ~9800 and 8000 cal yr B.P. supported abundant *Betula* shrubs and high lake and watershed productivity. *Alnus* rapidly expanded near Little Swift Lake while the region cooled between 8000 and 7500 cal yr B.P. Environmental changes at Little Swift Lake appear to have been roughly synchronous with similar changes elsewhere in southwestern Alaska, but late glacial and Holocene changes in other parts of Alaska were different in nature and timing. The complex spatial and temporal patterns of late glacial and Holocene environmental change throughout Alaska point to the importance of local- and regional-scale factors, especially controls on moisture availability, as modulators of site-specific responses to hemispheric- and global-scale climate forcing.

Introduction

Late glacial climate was remarkable for its rapid, high-magnitude fluctuations, and Holocene climate for its relative stability. Both extremes hold clues about how the climate system works: late glacial paleoclimate studies reveal how quickly climate can switch gears. Studies of Holocene paleoclimate demonstrate the range of natural climatic variability given near-modern configuration of ice sheets and oceanic and atmospheric circulation.

The late glacial and Holocene climate history of the North Pacific region is poorly known compared with the North Atlantic. Although changes in the North Atlantic region were probably greater in magnitude during this time, an increasing number of records indicate that the northeast Pacific region also experienced significant environmental changes. Such changes included climate reversals during the Younger Dryas chronozone (YD, ~12,900 to 11,600 cal yr B.P. as defined by Alley [2000]; cal yr B.P. indicates calendar years before 1950) (e.g., Engstrom et al., 1990; Mathewes et al., 1993; Peteet and Mann, 1994; Patterson et al., 1995; Hu et al., 1995, 1998, 2002; Brubaker et al., 2001; Mann et al., 2002; Briner et al., 2002), an early Holocene "thermal maximum" (e.g., Hu et al., 1996, 1998), and Neoglacial glacier advances (e.g., Calkin, 1988).

Here we reconstruct the major features of late glacial and Holocene paleoenvironmental change at an alpine site in southwestern Alaska, a region closely linked to the Bering Sea. We present lacustrine sedimentological evidence for glacier fluctuation, along with a continuous pollen record of vegetation changes, from Little Swift Lake in the Ahklun Mountains.

Study Area

The Ahklun Mountains (Fig. 1A) are a large glaciated massif lying beyond the late Wisconsin extent of the Cordilleran Ice Sheet complex. Past research in the Ahklun Mountains has focused on the extensive Pleistocene stratigraphic and geomorphic records of this repeatedly glaciated region (e.g., Kaufman et al., 1996, 2001; Briner and Kaufman, 2000; Briner et al., 2001; Manley et al., 2001). Yet the Ahklun Mountains, containing many peaks over 1500 m a.s.l. and more than 100 extant glaciers, also preserve largely unstudied records of late glacial and Holocene glacier fluctuations.

The northwestern Ahklun Mountains include some of the highest cirques in the range, extensive well-preserved moraine sequences, and numerous lakes. Little Swift Lake (60°12′52″N; 159°45′57″W; 572 m a.s.l.) is a small (~0.7 km^2 surface area), 24-m deep drift-impounded basin capturing the drainage of a deeply carved ~10-km-long glacial valley (Figs. 1, 2). The drainage basin has no glaciers today, but extant glaciers occupy cirques 15 km to the southeast.

Little Swift Lake is at the mouth of a valley tributary to the Crooked Creek valley (Figs. 1, 2). Bedrock within the lake's catchment includes granodiorite, rhyolite tuff, and mixed sedimentary and volcaniclastic rocks, with no carbonate or carbonaceous bedrock reported (Box et al., 1993). The lake is dammed by a lateral moraine that was deposited along the right margin of the trunk glacier that occupied the Crooked Creek valley during the late Wisconsin (Briner et al., 1999; Axford, 2000). The lake is downvalley from a sequence of alpine-glacier end moraines that were deposited by local valley glaciers. At least three late glacial and Holocene ice positions are represented (Axford, 2000; Figs. 1C, 2). Although glaciofluvial sediment produced by small glaciers in the high upper reaches of the valley may have been captured by several ponds upvalley of Little Swift Lake, the lake should contain sedimentary evidence of larger glaciers, such as the ca. 8.5-km-long glacier that deposited a prominent late glacial end moraine 1.3 km upvalley of the lake (Figs. 1C, 2).

Little Swift Lake is located within the transition between maritime and continental climate, in an area of discontinuous permafrost. The lake is surrounded by *Betula* shrub tundra, graminoid herb tundra, and steep, unvegetated rocky slopes. Dwarf birch (*Betula nana*) and diverse willows (*Salix* spp.), ferns (Polypodiaceae), sedges (Cyperaceae), grasses (Poaceae), club mosses (*Lycopodium*), and *Sphagnum* mosses dominate moist low-lying vegetation communities. Moraine ridge

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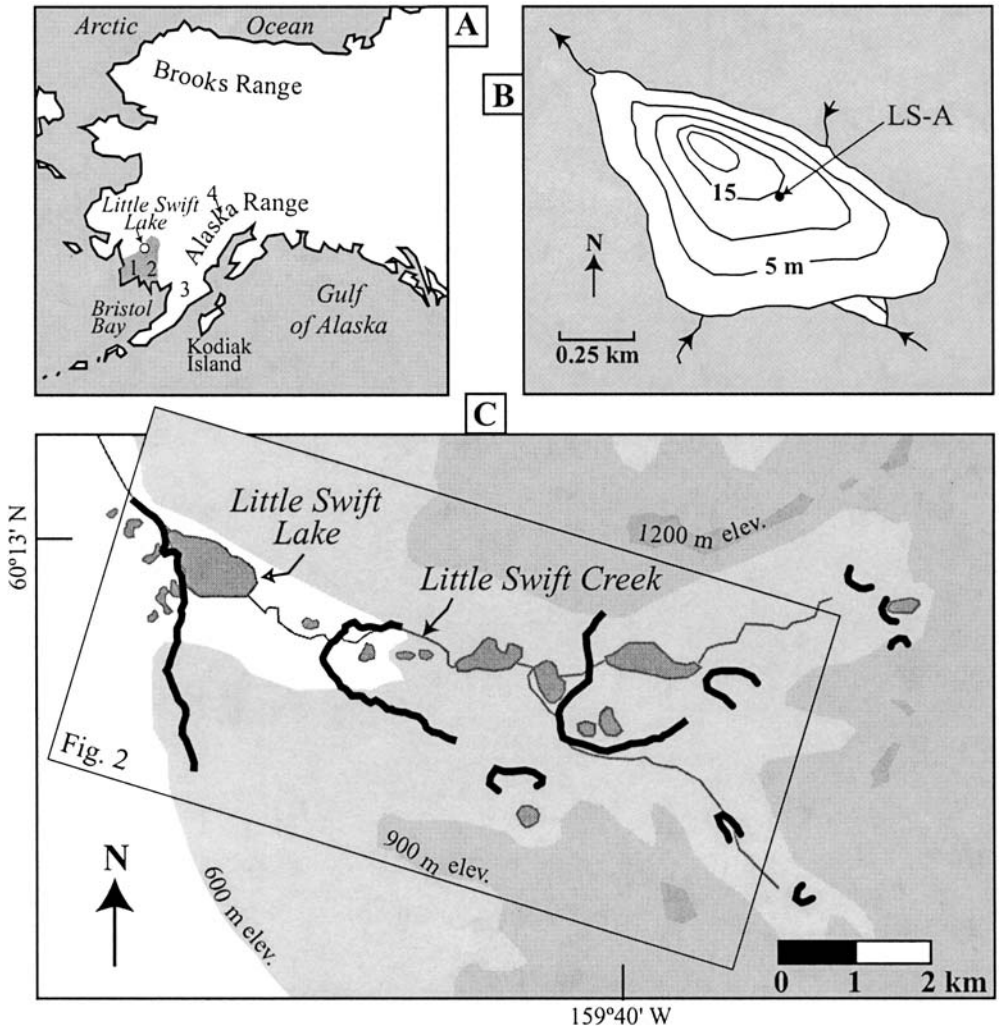


FIGURE 1. (A) Location of the Ahklun Mountains (shaded dark gray) and Little Swift Lake (white dot) relative to major geographic features of Alaska and other sites mentioned in the text, namely Nimgun Lake (1), Ongivinuk Lake (2), Idavain Lake (3), and Farewell Lake (4). (B) Bathymetry of Little Swift Lake, with location of coring site LS-A. (C) Little Swift Lake drainage area, with distal edges of prominent late glacial and Holocene moraines shown as bold lines. Box indicates area of Figure 2.

crests and highlands support sparse alpine tundra, with lichens, heaths (Ericaceae, e.g., labrador tea [*Ledum decumbens*], crowberry [*Empetrum nigrum*]), and herbs. Alder (*Alnus*) is not present within the lake's catchment, but *Alnus* thickets occupy creek beds ~5 km away. Spruce (*Picea*) grows 20 km to the northwest (where small *Picea glauca* stands inhabit river margins), and in boreal forest lowlands 50 km to the east.

Methods

Core LS-A, a 575-cm-long sediment core, was recovered from Little Swift Lake using a modified Nesje corer (Nesje, 1992) in 14.7 m of water. Magnetic susceptibility (MS) was measured on unsplit core segments at 2-cm increments using a Sapphire Instruments loop detector. Loss-on-ignition (LOI) at 550°C was used to estimate the organic matter content of 1-cc samples taken at 5-cm increments (Dean, 1974; Bengtsson and Enell, 1986). Sediment grain-size distribution was analyzed for 1-cm-thick samples taken at 5-cm increments using a Coulter LS230 laser detection particle-size analyzer. Grain-size samples were pretreated with H_2O_2, NaOH, and sodium hexametaphosphate to remove organic material, remove biogenic silica, and disaggregate sediments, respectively. Laboratory tests revealed a problem measuring the abundance of sand-sized grains in samples, due to the evolution of sand-sized air bubbles in the cold water used to circulate sediment through the analyzer. The problem was later corrected by installing a holding tank to store water at room temperature. Because measurements of sand percentages for this study are unreliable, we report only the unaffected clay:silt ratio.

Pollen samples were processed at Northern Arizona University's Laboratory of Paleoecology according to conventional procedures (e.g., Faegri and Iversen, 1975). An extended acetolysis time of 4 min improved the efficacy of saffranin staining. *Lycopodium* tracer tablets were added to samples to allow for calculation of pollen concentrations. All samples were sieved through a 9-μm screen following acetolysis to remove clays. Processed samples were mounted in silicon oil. A sum of 200 to 350 terrestrial pollen grains was identified per sample. Pollen abundances were calculated as percentages of the terrestrial pollen sum, and spore abundances as percentages of pollen plus spores. Pollen zones were determined using a stratigraphically constrained incremental sum of squares cluster analysis (the CONISS program packaged with TILIA software; Grimm, 1987) based on pollen and spore taxa with abundances >2%.

Nine AMS ^{14}C ages were obtained on mixed samples of hand-picked terrestrial and aquatic chitinous and floral macrofossil

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fragments >150 μm diameter. A tenth age—paired with a mixed sample—was obtained on terrestrial plant parts. $\delta^{13}C$ was accounted for in age calculations. Radiocarbon ages were calibrated to calendar years using CALIB 4.3 (Stuiver and Reimer, 1993).

Results

RADIOCARBON CHRONOLOGY

The 10 calibrated ages from core LS-A range from 1480 ± 65 cal yr B.P. on a sample from 20 to 25 cm depth to 13,110 ± 85 cal yr B.P. from the interval 545 to 550 cm depth (Table 1; Fig. 3). As described herein, correlations between the Little Swift Lake tephra stratigraphy and the regional tephra stratigraphy suggest that the radiocarbon ages are erroneously old, even though we reject the ages obtained from 508-513 cm and 545-550 cm depth, which contradict the basal age and apparently greatly overestimate the age of the core. A thick pumiceous tephra at 208 to 232 cm depth is most likely the Aniakchak tephra, which is prominent in lake records throughout the Ahklun Mountains (Kaufman et al., 2003) and has been described elsewhere in western Alaska (Riehle, 1985; Beget et al., 1992). Its age is well constrained at ~3600 ± 100 cal yr B.P., based upon over a dozen ^{14}C ages (Beget et al., 1992; Waythomas and Neal, 1998). A minimum limiting age of 4250 ± 100 cal yr B.P. from 202 to 207 cm depth in LS-A, immediately above the tephra, is apparently ~650 yr too old. A more mafic tephra at 318 cm depth in LS-A dates to ~6680 cal yr B.P., based upon a second-order polynomial interpolation between 8 of the 10 radiocarbon ages (Fig. 3). This tephra correlates with a ~6100 cal yr B.P. tephra found in nearby Arolik and Waskey lakes (Levy et al., 2003; Kaufman et al., 2003). The uppermost age of 1480 cal yr B.P. from 20 to 25 cm depth is probably also too old. Although our coring procedure did not capture the sediment-water interface, based on previous experiences with the same equipment, and assuming sedimentation rates throughout the core are a fair approximation of sedimentation rates over the last millennium, it is very unlikely that the core top is really 1000 yr old. We conclude that ^{14}C ages from the upper half of LS-A are ~600 yr too old. The polynomial trend through our original radiocarbon ages would estimate the age of the base of LS-A at 13,200 cal yr B.P. (Fig. 3), but a ~12,200 cal yr B.P. tephra found in cores from nearby Arolik and Nimgun lakes (Kaufman et al., 2003) is absent in LS-A; thus ages from the base of LS-A may be ~1000 yr too old.

In order to assign reasonable ages to events recorded in the core stratigraphy, we use an age model that corrects for offset in the radiocarbon chronology. We assume an offset of 600 yr for ages less than 10,000 cal yr B.P., and an offset of 1000 yr for older ages. 10,000 cal yr B.P. is an arbitrary cut-off, but is consistent with evidence that ages near the base of the core are 1000 yr too old. The ages immediately above and below this cut-off are ~2000 yr apart, so there is no sudden jump in the chronology.

A second-order polynomial trend line ($y = 0.012x^2 + 13.574x + 518.45$) through the eight adjusted ages has an R^2 value of 0.995, and agrees well with the tephrochronology (Fig. 3). Pollen correlation provides another test: Striking regional palynological changes occurred at Ongivinuk (Hu et al., 1995), Idavain (Brubaker et al., 2001), and Nimgun (Hu et al., 2002) lakes (Fig. 1) at the end of the YD chronozone; the same changes (including a sharp decline in Cyperaceae and increase in Poaceae and *Betula*) date to 11,430 cal yr B.P. in Little Swift Lake using our adjusted age model. Based on comparisons with the regional tephra and pollen stratigraphies, we estimate the uncertainty of our model to be ± 200 yr.

The shell of a living snail collected from the lake in 1998 has a radiocarbon activity of 100.1% Modern, less than atmospheric values of the 1990s. This indicates that there is a source of ^{14}C-depleted dissolved inorganic carbon (DIC) to the lake. Possible sources of ^{14}C-depleted DIC include groundwater, dissolution of rocks within the lake's watershed (although no carbonate or carbonaceous bedrock has been reported from Little Swift Lake's drainage), and decomposition of

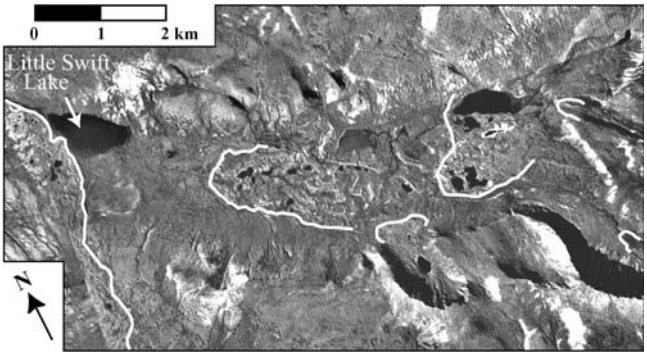


FIGURE 2. Mosaic of vertical aerial photographs depicting the lower portion of the Little Swift Lake drainage. The distal edges of prominent moraines are marked in white, including the right lateral moraine of the Crooked Creek valley, which dams Little Swift Lake along the west edge of the image. Location is shown in Figure 1.

TABLE 1

Radiocarbon Ages from Core LS-A, Little Swift Lake, Alaska

Lab number	Depth (cm)	Material[a]	$\delta^{13}C$	Fraction modern	^{14}C age (^{14}C yr B.P.)	Calibrated age (cal yr (B.P.)[c]	Adjusted age (cal yr B.P.)[d]
NSRL-10559	live aquatic snail	*Stagnicola* shell	−8.0	1.0011 ± 0.042	"modern"	"modern"	N/A
NSRL-11065	20–25	wood	−25.0[b]	0.8190 ± 0.0042	1600 ± 40	1480 ± 65	880
NSRL-11066	20–25	mixed	−26.3	0.8136 ± 0.0043	1660 ± 40	1520 ± 90	920
NSRL-11067	123–126	mixed	−26.7	0.7046 ± 0.0034	2810 ± 40	2870 ± 80	2270
NSRL-11068	202–207	mixed	−25.1	0.6198 ± 0.0032	3840 ± 40	4250 ± 100	3650
NSRL-11069	338–343	mixed	−25.6	0.4604 ± 0.0026	6230 ± 45	7130 ± 120	6530
NSRL-11070	403–407	mixed	−25.9	0.3908 ± 0.0036	7550 ± 75	8310 ± 100	7710
NSRL-11071	447–452	mixed	−25.8	0.3107 ± 0.0021	9390 ± 55	10,560 ± 125	9560
CAMS-76366	508–513	mixed	−25.0[b]	0.2659 ± 0.0053	10640 ± 170	12,718 ± 281	rejected
NSRL-11059	545–550	mixed	−25.2	0.2480 ± 0.0022	11,200 ± 75	13,110 ± 85	rejected
NSRL-11060	568–571	mixed	−25.3	0.2621 ± 0.0025	10,760 ± 75	12,800 ± 150	11,800

[a] "Mixed" samples contain floral and faunal aquatic and terrestrial macrofossils.

[b] $\delta^{13}C$ is estimated (not measured).

[c] Each calibrated age is the midpoint ± 1/2 the 1-sigma range calculated using CALIB 4.3. (Stuiver and Reimer, 1993).

[d] See text for an explanation of adjusted ages.

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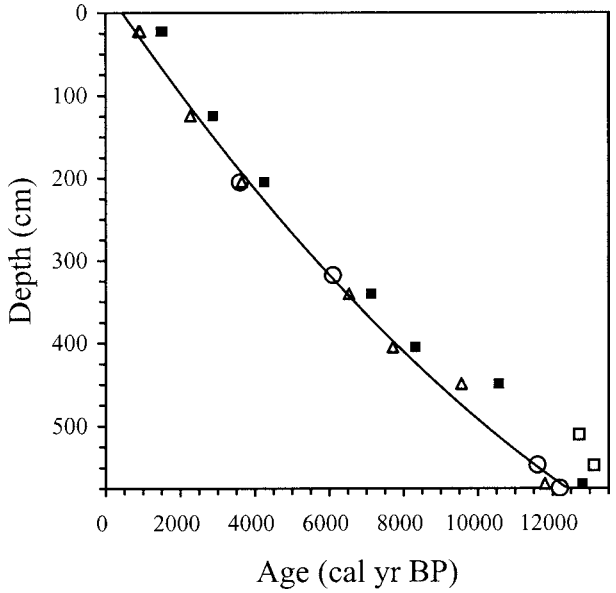


FIGURE 3. Chronology for core LS-A. Calibrated radiocarbon ages are shown as solid black squares; open squares are rejected ages; triangles represent adjusted radiocarbon ages (described in the text). The black line is our age model, a second-order polynomial trend through the adjusted ages. Note the agreement between the age model and independent chronological evidence (shown as open circles representing known tephra ages and one pollen correlation at 547 cm depth). Radiocarbon data are provided in Table 1.

organic soil horizons within the lake's watershed or old organic matter within the lake (e.g., Abbott and Stafford, 1996).

The statistically indistinguishable paired ages obtained on terrestrial plant material and mixed aquatic and terrestrial material at 20 to 25 cm depth (Table 1) indicate that there must also be an influx of ^{14}C-depleted terrestrial macrofossils to the lake. Reworked organic carbon from old soils can contribute depleted carbon to arctic lakes (Abbott and Stafford, 1996), but old soils should have been removed from this watershed during the extensive late Wisconsin glaciation, and we found no buried soils in deep river bluff exposures. Thus the source of old terrestrial macrofossils to the lake is unknown. We note that we would have been unaware of problems with this chronology, had it not been for comparison with tephras of known age. Similar unexpected chronological problems at other arctic lakes may go unrecognized, emphasizing the importance of reliable chronostratigraphic markers.

LITHOSTRATIGRAPHY

Most of core LS-A (0 to 547 cm depth) is massive to laminated medium-brown to gray-brown gyttja, with only subtle changes in appearance (Fig. 4). The gyttja is banded throughout, with finer (millimeter-scale) laminations at the base of this interval. Magnetic susceptibility (MS) within the gyttja is low (1.9×10^{-5} to 3.6×10^{-5} cgs), with little variation except for a subtle upward-increasing trend. The gyttja exhibits only minor changes in organic content (7 to 11% LOI), with a zone of highest LOI between 380 and 455 cm depth. Clay:silt decreases (i.e., the sediment becomes coarser grained) from the bottom of the gyttja (maximum of 0.76 clay:silt at 545 cm depth) to the top (minimum of 0.26 clay:silt at 50 cm depth).

A 24-cm-thick, light brownish-gray tephra at 208 to 232 cm depth is composed of sand- and silt-sized pumice grains, bubble-wall glass shards, and fine mafic mineral grains. There is also a dark brown fine sandy tephra, ca. 0.2 cm thick, at 318 cm depth, and a light gray fine sandy tephra of similar thickness at 398 cm depth. Each tephra layer is coincident with a prominent peak in MS and a zone of low LOI.

The core terminates in medium gray silty clay, extending from an abrupt transition at 547 cm depth (~11,400 cal yr B.P.) to the base of the core at 575 cm (~12,200 cal yr B.P.). (Note that we did not meet resistance in coring, so the base of LS-A does not represent the beginning of the sedimentary record in Little Swift Lake.) The basal gray mud has a lower clay:silt ratio (0.50 to 0.57) than the brown gyttja immediately above (0.76 at 545 cm depth) (Fig. 4), and low LOI (3 to 4%). The MS of the basal unit is an order of magnitude higher than that of the gyttja, reflecting its low organic content, coarse grain size, and probably distinct mineralogy. None of the sediment in core LS-A fizzes in contact with dilute HCl.

PALYNOLOGY

Modern pollen deposition was estimated by analyzing pollen from the top 0.5 cm of a short gravity core that captured the undisturbed sediment-water interface near core LS-A. The modern pollen rain is dominated by *Alnus* (44%), Cyperaceae (20%), and *Betula* (19%), with low percentages (<5%) of Poaceae, *Picea*, *Artemisia*, *Salix*, Ericales, and numerous herb taxa. Monoletes, *Lycopodium*, and *Sphagnum* are the most abundant spores.

Pollen assemblages from Zone LS1 (575 to 547 cm depth; ~12,200 to 11,400 cal yr B.P.) include abundant Cyperaceae (>45%), and the core's highest percentages of *Salix*, *Artemisia*, Tubuliflorae, and Saxifragaceae (Fig. 5), all of which decrease toward the top of this zone. *Betula* pollen increases slightly from its lowest value in the core (11%). Spores (including *Lycopodium*, *Sphagnum*, monoletes, and *Equisetum*) and *Pediastrum* cell nets are rare or absent (Fig. 5).

Zone LS2 (547 to 485 cm depth; ~11,400 to 9800 cal yr B.P.) is defined by a Poaceae peak reaching a maximum abundance of 34%. Cyperaceae decrease to about half their earlier abundance. Ericales and Brassicaceae are relatively abundant throughout much of LS2. *Alnus* pollen first appears, but remains rare. Monolete spores and *Pediastrum* rapidly increase, and *Pediastrum*, *Sphagnum*, and *Equisetum* reach their maximum abundances in this zone.

Within Zone LS3 (473 to 390 cm depth; ~9800 to 7600 cal yr B.P.), Poaceae pollen declines to its earlier low percentage (~10%). *Betula*, *Sanguisorba*, and monolete spores reach peak abundance, with *Betula* reaching maximum abundance (44%) at 428 cm depth (~8500 cal yr B.P.). *Artemisia*, *Pediastrum*, and *Equisetum* decline. *Picea* pollen first appears at 399 cm depth (~7800 cal yr B.P.).

At the base of Zone LS4 (390 to 0 cm depth; ~7600 cal yr B.P. to present), *Alnus* pollen rapidly increases from 2 to 25%. Above the increase in *Alnus*, Zone LS4 is relatively stable, with abundant *Alnus*, *Betula*, and Cyperaceae pollen, and lower but significant percentages of *Salix*, Ericales, *Artemisia*, and Poaceae. Ericales are relatively abundant and increase slightly throughout LS4. *Picea* pollen reaches a maximum abundance of 5% at 18 cm depth (~800 cal yr B.P.). Monolete spore and *Pediastrum* percentages decline throughout LS4, whereas *Lycopodium* increases slightly.

Discussion

LATE GLACIAL ENVIRONMENTAL CHANGE

The record from core LS-A begins during a period of climatic amelioration, characterized by increasing *Betula* shrub tundra and decreasing *Salix*, *Artemisia*, and Tubuliflorae. These vegetation changes correlate with similar changes that occurred during a period of warming and increased moisture at the end of a YD-age cold reversal at nearby Grandfather and Nimgun lakes (Hu et al., 1995, 2002), and hint that Little Swift Lake may have experienced a similar reversal to colder conditions and herb-dominated tundra.

There is evidence that glacier retreat accompanied vegetation change at the end of the YD chronozone at Little Swift Lake.

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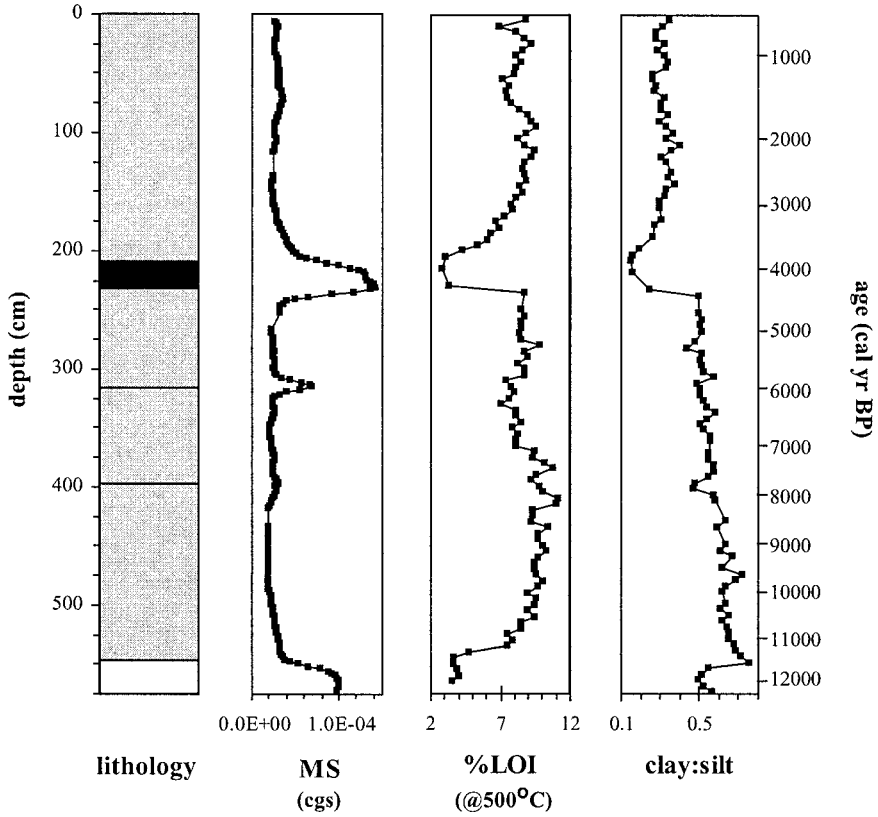


FIGURE 4. Lithology, magnetic susceptibility (MS), loss on ignition (LOI), and grain size data from core LS-A. Tephra layers are shown in black, light gray shading is organic brown gyttja, and white is inorganic silty clay. Ages are modeled calibrated ages.

Considering the proximity of the lake to upvalley late glacial moraines (Briner et al., 1999; Axford, 2000), we interpret the basal, inorganic gray silt in LS-A as ice-proximal sediment transported to the lake by glacial meltwater streams. An upvalley glacier retreated (probably to a position behind an upvalley pond) ~11,400 cal yr B.P., as recorded by the transition to organic clay-rich mud.

Alternatively, these sedimentological changes may record an abrupt increase in lake and watershed productivity without the influence of an upvalley glacier. However, diverse additional data support our original interpretation: The distinct bulk major-element geochemistry of the core's basal unit, exceptionally rich in Mg, Na, K, Al, and Ti, suggests an abundance of clastic minerals and a unique provenance. Geochemical analyses of likely sediment sources within the lake's watershed indicate that a granodiorite, which outcrops only in upvalley cirques, is the most likely source (Carey et al., 2000). The low LOI and lack of *Pediastrum* cell nets within the basal unit suggest low aquatic production, which could result from a variety of conditions, but is consistent with high turbidity and low water temperatures due to meltwater influx. The near-disappearance of the diatoms *Hannaea arcus* and *Nitzschia* spp. from the lake at ~11,400 cal yr B.P. further supports decreased meltwater flux to the lake (I. Gregory-Eaves, pers. comm., 2002); *Hannaea arcus* is found in cold, flowing water and *Nitzschia* species thrive in turbid, unproductive, river-influenced arctic lakes (e.g., Ludlam et al., 1996; Hay et al., 1997).

There is widespread evidence from southwestern Alaska indicating a reversal to cold, dry conditions during the YD, followed by warming (Peteet and Mann, 1994; Hu et al., 1995, 2002; Brubaker et al., 2001; Hu and Shemesh, 2003). Briner et al. (2002) reported evidence for a probable YD-age glacier readvance at Waskey Lake, 50 km south of Little Swift Lake. It is possible that there was a similar readvance above Little Swift Lake, but we cannot evaluate this possibility with available data. It does appear that Little Swift Lake experienced warming at the end of the YD comparable to that at lower-elevation sites in southwestern Alaska.

The dramatic expansion of Poaceae at this site and others in southwestern Alaska after the YD (Hu et al., 1995, 2002; Brubaker et al., 2001) is difficult to interpret because Poaceae are ecologically diverse. We note that post-YD Poaceae expansion was most dramatic at Little Swift, Ongivinuk, Nimgun, and Idavain lakes, which are distinguished from other sites in that they receive most of their moisture from Bristol Bay rather than the Gulf of Alaska. One explanation is that Poaceae expansion resulted from aridity in the Bristol Bay region during post-YD warming. The postglacial transgression of the shallow Bering Shelf was not yet complete at the end of the YD (Fairbanks, 1989; Elias et al., 1996), and all of these sites were significantly more continental than at present. The Bristol Bay shoreline probably transgressed to a near-modern location between 10,000 and 9000 cal yr B.P. (Manley, 2002), coincident with the decline of Poaceae at these sites.

POSTGLACIAL TREE AND SHRUB EXPANSION

Pollen records from southwestern Alaska record the gradual postglacial expansion of *Betula*, *Alnus*, and *Picea* to their modern ranges (e.g., Ager, 1983; Hu et al., 1995, 1996, 2002; Brubaker et al., 2001). *Betula* was already abundant at the beginning of the Little Swift Lake record ~12,200 cal yr B.P., consistent with the establishment of *Betula* shrub tundra by ~13,600 cal yr B.P. at Nimgun Lake (Hu et al., 2002) and elsewhere in the Ahklun Mountains (Brubaker et al., 2001; Hu et al., 2002). *Alnus* rapidly increased from less than 5% of the pollen rain at Little Swift Lake to more than 25% ~7600 cal yr B.P., roughly coincident with the rapid regional expansion of *Alnus* at many sites throughout southwestern (Hu et al., 1996; Brubaker et al., 2001),

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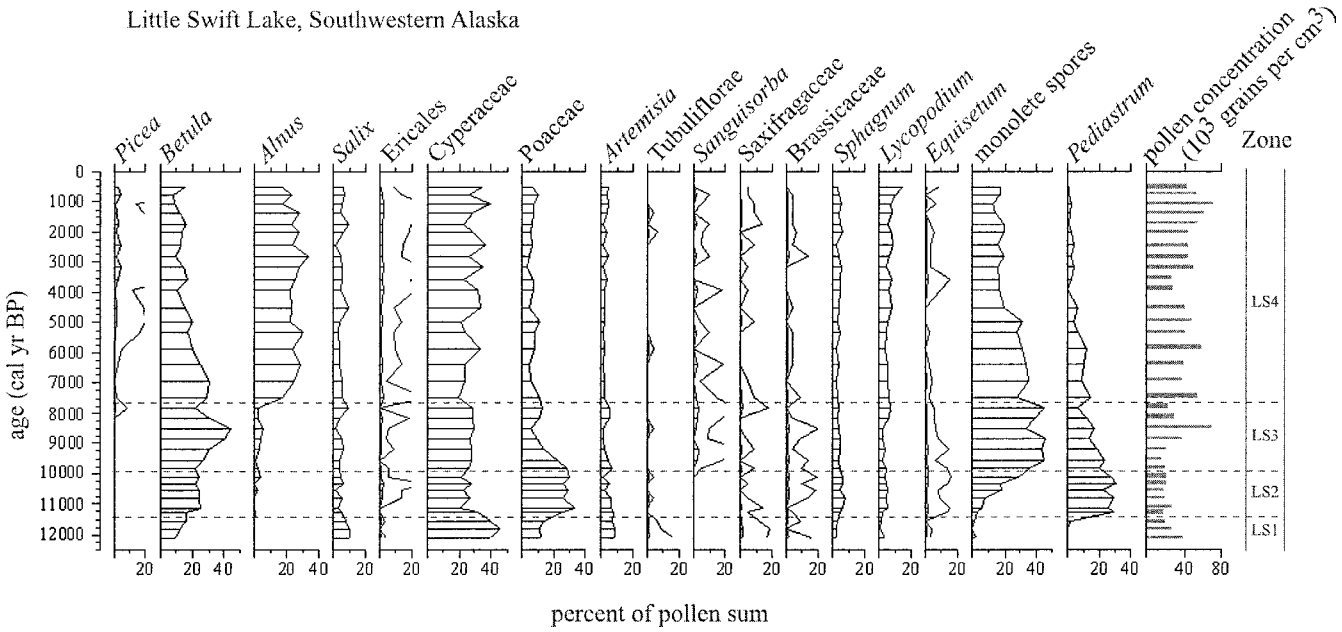


FIGURE 5. Percentages of pollen, spores, and aquatics from core LS-A. Unfilled curves (e.g., for Picea *and Ericales) represent 10 × exaggeration.*

central (Bigelow and Edwards, 2001), and northern Alaska (Anderson and Brubaker, 1994). Middle and late Holocene *Alnus* percentages were 10 to 25% lower than the modern value of 44%.

Picea expanded southwestward across Alaska for several thousand years after deglaciation, apparently reaching southwestern Alaska after other parts of eastern Beringia. The Ahklun Mountains region may have been one of the last areas to be colonized. *Picea* did not become important at Grandfather Lake until after ~4500 cal yr B.P. (nearly 5000 yr after *Picea* colonized the area around Farewell Lake ~400 km away in the northwestern Alaska Range) and reached modern abundance at Grandfather Lake ~2000 cal yr B.P. (Hu et al., 1995; Brubaker et al., 2001). Windblown *Picea* pollen percentages at Little Swift Lake reached modern values ~3500 cal yr B.P. *Picea* groves were probably never much closer to the lake than at present, given that the highest Holocene percentages of *Picea* pollen are less than 1% higher than modern values.

HOLOCENE ENVIRONMENTAL CHANGE

After ~9800 cal yr B.P., Little Swift Lake experienced an early-Holocene period of warm, moist conditions. Abundant *Betula* shrubs, *Sanguisorba*, and Polypodiaceae thrived on land, and high terrestrial and aquatic production and/or stable soils resulted in the deposition of organic-rich lake sediments. *Betula* pollen percentages reached a Holocene maximum of 44% at ~8500 cal yr B.P. Early Holocene warmth at Little Swift Lake coincides with peak warmth ~9500 to 8900 cal yr B.P. at Farewell Lake ~400 km to the northeast, as inferred from ostracode trace-element geochemistry (Hu et al., 1998). Farewell Lake experienced a coincident peak in *Betula* abundance (Hu et al., 1996).

Warmth presumably resulted from the tail end of a peak in summer insolation (Berger and Loutre, 1991) at high northern latitudes combined with a weakening of the Laurentide glacial anticyclone (e.g., Heusser et al., 1985; Bartlein et al., 1991, 1998). As Bering Sea sea-surface temperatures warmed, sea-ice extent was reduced (Sancetta et al., 1984), decreasing regional albedo and increasing moisture availability. The primary moisture source was also more proximal to the Ahklun Mountains and Little Swift Lake because the Bering Sea had transgressed much of its shelf by this time.

The end of the thermal maximum at Little Swift Lake was punctuated by a brief low in *Betula*, accompanied by relative increases in *Artemisia*, *Salix*, *Sanguisorba*, and Polypodiaceae, and decreases in Ericales and *Pediastrum*. These changes, which occurred ~8000 cal yr B.P. and just before the local expansion of *Alnus*, may have resulted from cooling during a period of relatively moist climate. Previous researchers have suggested that the rapid regional expansion of *Alnus* resulted from increased effective moisture (possibly more winter snowfall) associated with climatic cooling (Anderson and Brubaker, 1994; Hu et al., 1995); and our pollen data, which suggest a cool, moist climate immediately preceding the expansion of *Alnus*, support this inference. Effective moisture continued to increase at other sites until ~7000 cal yr B.P. (Hu et al., 1998; Edwards et al., 2001). It is possible that a small but distinct upvalley readvance moraine with a mean cosmogenic surface exposure age of ~5500 cal yr B.P. (Briner et al., 1999; Axford, 2000) records a glacier advance that was driven by mid-Holocene cooling and increased effective moisture.

Widespread Neoglacial glacier advances were initiated after ~4000 cal yr B.P. at sites throughout Alaska (e.g., Calkin, 1988; Calkin et al., 1998). In the Ahklun Mountains, glaciers in the Waskey Lake valley were reactivated 3100 cal yr B.P. (Levy et al., 2003). Although proxy data from Little Swift Lake do not necessarily suggest Neoglacial advances, the subtle upcore decrease in organic content and coarsening of sediment grain size may record late Holocene climatic deterioration. Glaciofluvial sediment from limited Holocene glacier advances probably would have been trapped by the string of ponds upstream of Little Swift Lake, so we cannot rule out the possibility that glaciers expanded at Little Swift Lake during this time. Alaskan pollen records generally offer little evidence for late Holocene climate changes, partly because it is difficult to resolve subtle changes in tundra composition. CONISS analysis identifies a change in the Little Swift Lake pollen assemblage ~4800 cal yr B.P., with lower percentages of *Betula* and monolete spores, and more abundant Cyperaceae, possibly hinting at cooler temperatures in the late Holocene.

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Conclusions

Our results indicate that both glaciers and alpine vegetation in southwestern Alaska responded to climatic changes at the end of the YD chronozone, a conclusion consistent with previous work. Paleoclimate records reveal a complex pattern of climate changes across Alaska during late glacial time. Southeastern Alaska apparently became cooler and wetter during the YD (e.g., Engstrom et al., 1990; Mathewes et al., 1993; Patterson et al., 1995). In contrast, southwestern Alaska experienced a reversal to cold, dry conditions (e.g., Peteet and Mann, 1994; Hu et al., 1995, 2002). The spatial heterogeneity of responses to probable North Pacific cooling suggests that local- and regional-scale factors, such as proximity to the Bering Sea, must have modulated responses to larger-scale, late glacial climate forcing. Evidence for temporal and spatial variability of moisture availability throughout the late glacial and Holocene is a reminder that predictions of future conditions in the Arctic should take into account the effects of local- and regional-scale influences on moisture balance.

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