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Lake-Level Reconstructions and Paleohydrology of Birch Lake, Central Alaska, Based on Seismic Reflection Profiles and Core Transects

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Lake-level history for Birch Lake, Alaska, was reconstructed using seismic profiles and multiproxy sedimentary analyses including sedimentology, geochemistry, magnetic susceptibility, and palynology. Twenty-two seismic profiles (18 km total) and eight sediment cores taken from the lake margin to its depocenter at 13.5 m provide evidence for low lake stands during the late Pleistocene and Holocene. Thirty-one AMS radiocarbon dates of macrofossils and pollen provide a century-scale chronology. Prior to 12,700 ^{14}C yr B.P., the lake, which now overflows, was either seasonally dry or desiccated for prolonged periods, indicating a severe period of aridity. Lake level rose more than 18 m between 12,700 and 12,200 ^{14}C yr B.P. before falling to 17 m below the level of overflow. Between 11,600 and 10,600 ^{14}C yr B.P. the water remained between 14 and 17 m below the overflow level. Onlap sedimentary sequences were formed during a transgression phase between 10,600 and 10,000 ^{14}C yr B.P. Between 10,000 and about 8800 ^{14}C yr B.P. the lake was between 6 and 9 m below the overflow level. Lake level again rose, approaching the overflow level, between 8800 and 8000 ^{14}C yr B.P. Seismic and core evidence of minor erosional events suggest lowstands of 2–6 m until 4800 ^{14}C yr B.P. There have been no prolonged periods of lake-level depression since that time. The major restructuring of the climate system during deglaciation evidently generated a complex set of fluctuations in effective moisture in interior Alaska, which likely affected eolian processes and vegetation development, as well as lake levels.

Key Words: **lakes; Alaska; paleolimnology; Holocene; radiocarbon; insolation.**

INTRODUCTION

Forecasts of future climate change are limited by our ability to model fluctuations in the climate system and to validate these models with knowledge of past changes. The emphasis of most current paleoclimate research is reconstruction of past temperature changes, but fluctuations in the precipitation–evaporation balance (P–E) are equally important and have recently been shown to occur over human time scales (Hodell *et al.*, 1995; Curtis and Hodell, 1996; Binford *et al.*, 1997; Abbott *et al.*, 1997a,b). The Arctic and sub-Arctic are particularly sensitive climatically, as shown by recent work indicating abrupt climatic shifts during the 19th and 20th centuries (IPCC, 1990; Douglas *et al.*, 1994; Bradley *et al.*, 1996). This paper reports the timing, magnitude, and duration of lake-level changes identified from seismic profiles (Moore *et al.*, 1994) and core transects (Digerfeldt, 1986) at Birch Lake, near Fairbanks, interior Alaska (Fig. 1).

This is the first detailed, quantitative lake-level reconstruction from Alaska based on stratigraphic analyses of sediment cores and seismic data. The inferred shifts in P–E balance of the region provide important new paleoclimatic data on the sub-Arctic (PALE initiative; Andrews and Brubaker, 1994), which can be applied in several ways. Data may be compared with those of other regions to assess long-distance and interhemispheric climatic teleconnections (PANASH initiative; Bradley *et al.*, 1995). When combined with hydrologic modeling, quantitative lake-level changes can provide estimates of paleoprecipitation (Barber and Finney, in press); these can be

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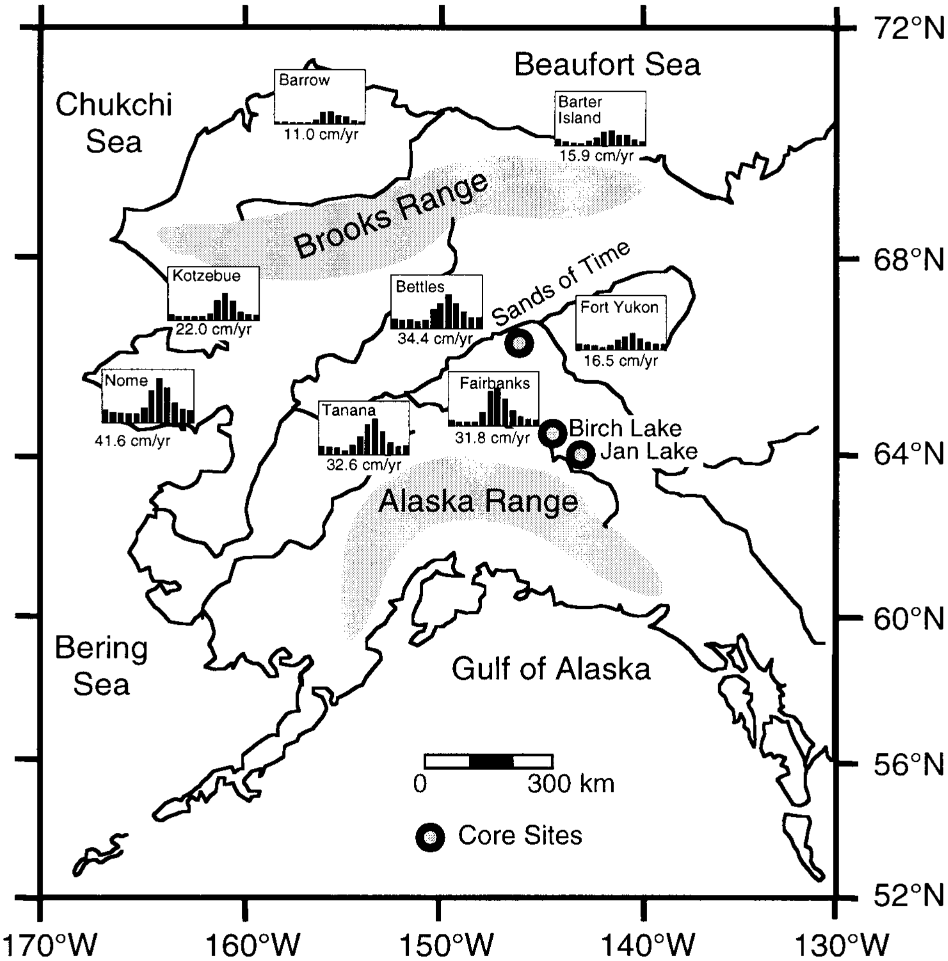

FIG. 1. Map showing location in central Alaska of the Birch Lake study site, as well as Sands of Time and Jan Lake. Graphs of monthly precipitation, with annual totals below the chart, are plotted from January on the left to December on the right with a vertical scale from 0 to 10 cm. Two points are illustrated by this figure: (1) the seasonal distribution of precipitation and (2) the south–north and west–east moisture gradients.

compared with data from adjacent regions and with modern synoptic patterns to gain a better understanding of the causes of P–E changes (Edwards *et al.,* in press). Paleoclimate data may also be compared with climate model output (e.g., Bartlein *et al.,* 1998) to assess model simulations of past effective moisture (Mock *et al.,* 1998).

Current knowledge of past temperature and effective moisture changes in Alaska is based mainly on pollen studies. Environmental conditions for critical periods (e.g., 18,000, 12,000, and 6000 ^{14}C yr B.P.) have been mapped and primary vegetational features have been identified (Barnosky *et al.,* 1987; Anderson and Brubaker, 1993, 1994; Edwards and Barker, 1994). Data suggest a marked increase in temperature and moisture during the glacial–interglacial transition. Warmer growing conditions and a likely increase in the moisture balance are indicated by a change from herb tundra to birch-dominated tundra over all of northern Alaska north of the Alaska Range. This change has been dated in the Tanana valley (Fig. 1) to 14,000 ^{14}C yr B.P. (Ager, 1975), but at most sites in interior Alaska it occurs between 13,000 and 12,000 ^{14}C yr B.P. (Edwards and Barker, 1994). Subsequent increases in poplar and white spruce imply a further increase in temperatures at 11,000–9000 ^{14}C yr B.P., coinciding with the northern hemisphere summer insolation maximum. Multiproxy studies by Hu *et al.* (1996, 1998) indicated that the period between 11,000 and 8000 ^{14}C yr B.P. was drier than present at Farewell Lake in the northwestern Alaska Range. Widespread eolian activity during full-glacial time and intermittent reactivation of eolian systems in late-glacial and early Holocene times also suggest generally arid conditions prior to the middle Holocene (Péwé, 1975; Hopkins *et al.,* 1981; Carter, 1993). Expansion of black spruce and alder between 8000 and 6000 ^{14}C yr B.P. suggests wetter conditions.

Lake-level records provide an independent and more direct means to determine variations in the regional P–E balance (Street-Perrott and Harrison, 1985; COHMAP, 1988; Harrison and Digerfeldt, 1993). Lake systems tend to respond rapidly and be more sensitive to effective moisture changes than vegetation (annual-decadal vs century response) and usually have more highly resolved records than terrestrial eolian systems.

METHODS

Seismic Profiling

Twenty-two cross-lake seismic reflection profiles from Birch Lake were used to trace the acoustic stratigraphy and identify surfaces associated with water-level changes. The seismic profiles were obtained with an ORE-Geopulse seismic system (3–7 kHz) with an EPC 9800 digital graphic recorder. The acoustic source and hydrophone were towed ~4 m apart from the stern of a 7-m boat. Emergent vegetation in shallow water limited seismic tracks to water depths of greater than 4 m. The navigation was logged using the global positioning system (GPS) and by compass measurements. Seismic data were digitally recorded with a DAT recorder for further processing. Interpretations utilized both the unfiltered field records and processed data filtered between 3 and 7 kHz. We used 1425 and 1500 m s^{-1} for the speed of sound in water and sediment, respectively, to determine depth of reflectors and thickness of sediment units. The instrumental error for depth of reflectors on filtered profiles is about ±30 cm.

The seismic profiles from Birch Lake can be used to identify onlap sedimentary sequences and erosion surfaces associated with lake-level changes, despite two problems that affect the interpretation of acoustic records from shallow lakes. First, acoustic signals may be obscured by exsolved gas from decomposition of *in situ* organic matter. This effect is visible on records throughout the northern basin and in patches of the southern basin. A second problem occurs where sediment thickness exceeds the water depth, and ringing of the sonic signal within the water column masks the seismic stratigraphy by overprinting multiples. Despite these complications, the acoustic stratigraphy from Birch Lake is remarkably correlative with the physical and geochemical stratigraphy obtained from the sediment cores. One reason is the high impedance contrasts between different sediment facies encountered in lacustrine basins.

Sediment Core Transects

Although relative lake-level changes can be inferred from analysis of a single, deep-water core, these estimates remain qualitative. Water-level variations can be quantified using core transects to map and date the extent and timing of lake-level changes. Past lake-level changes were reconstructed using analysis of offshore changes in sediment facies determined by core data. We used the methodology of Digerfeldt (1986), supplemented by other lake-level indicators. The Digerfeldt method is based on determining changes with time in the depth of the sediment limit, which is the zone where the net accumulation of fine organic material is low due to removal by waves and other physical processes. We have been able to identify shallow-water environments and differentiate between lacustrine and subaerial sediments as additional methods of reconstructing lake-level changes.

Erosion surfaces were identified in cores by (1) abrupt transitions (<1 cm) characterized by coarser-grained (fine sand and greater) sediments with high bulk density (>1 g/cm^3) underlying fine-grained organic-rich muds (>10% organic matter), (2) highly fragmented shell material, (3) scour marks, and (4) mud cracks. We used detailed core descriptions, smear-slide mineralogy, and radiocarbon stratigraphy to delimit erosion surfaces formed during low-water stands and subaerial exposure. Shallow-water subfacies (<2 m water depth) were identified by comparison with modern shallow-water sediments, collected using a dredge, that are characterized by (1) the presence of high concentrations of emergent vegetation in a coarse-grained matrix (silt to fine sand), (2) large amounts of aquatic plant macrofossils (*Myriophyllum* and *Potamogeton*), and (3) sediments containing >60% $CaCO_3$ composed of calcified macrophyte coatings and fragmented gastropod and bivalve shells.

Sediment cores were taken on a transect from 2.5 to 13.5 m water depth in the southern basin (Fig. 2) with a square-rod piston corer (Wright *et al.,* 1984). A single core was collected from the deepest part of the northern basin. Magnetic susceptibility was measured on whole cores at 2-cm intervals with a Bartington MS-2 Susceptibility Meter. Core lithology was determined from smear-slide mineralogy and detailed core logging, including descriptions of Munsell color, sedimentary structures, and biogenic features. Other laboratory analyses included water content, bulk density, and both organic matter and calcium carbonate content by the loss-on-ignition (LOI) method (Bengtsson and Enell, 1986). The laboratory measurements and detailed core descriptions were used to characterize the sediment units and transitions.

RESULTS AND DISCUSSION

Seismic Stratigraphy

The two-dimensional perspective that seismic data provide allows detection and mapping of erosion surfaces and onlap/offlap sedimentary sequences that might otherwise go undetected in single-core lake studies. For example, three erosion surfaces and two onlap sequences are illustrated in Fig. 3. The onlap transgression sequences resulted in organic-rich fine-grained sediments being deposited at increasingly higher elevations in the lake basin as water level rose. During periods of transgression these newly deposited sediments formed a high-density contrast with the underlying shoreline sediments comprised of coarse-grained mineral matter. This high impedance contrast is detectable in the acoustic records.

Detailed documentation of acoustic reflectors and sediment layers is provided in Table 1, as well as an interpretation of the observations regarding changes in lake level. Comparisons of

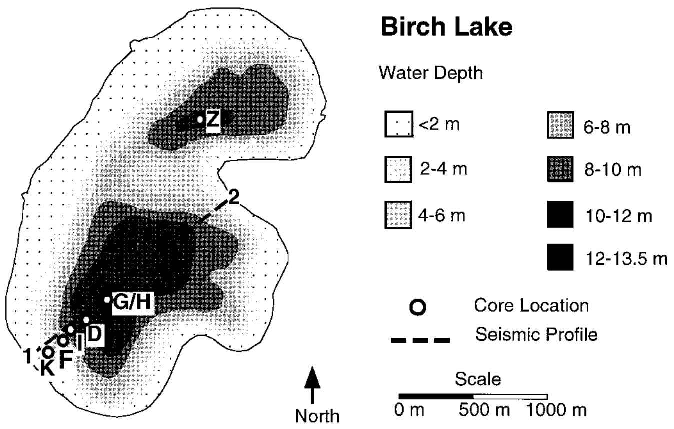

FIG. 2. Map of Birch Lake showing locations of seismic profile 1-2 and cores K, F, I, D, G/H, and Z in the two lake basins.

the depths associated with acoustic layers (defined from seismic profiles) with lithological units (defined by sedimentary variations in the cores) show relatively consistent boundaries (Table 2). The estimated match for depth and thickness of the acoustic and sediment units is generally within the 30-cm error range calculated for the filtered signal of 3–7 kHz used to process the seismic records. The match between the lake-bottom reflector (Bt) and the sediment–water interface in the cores from deep water ($>$10 m) is poor, probably because the water content of these fine-grained, organic-rich sediments is high prior to compaction during the early stages of burial. The relationship between the seismic reflectors and sediment units improves to the $\pm$30-cm range below the Bt reflector.

Three low lake stands at $>$18, 14–17, and 6–9 m below the overflow level (BOL) were identified by mapping the distribution of irregular hummocky reflectors interpreted to be erosion surfaces (Table 1). The basal reflector B-5 is characterized by irregular meter-scale hummocks that truncate underlying strata. This reflector can be traced throughout all depths in Birch Lake and is interpreted to be an erosion surface (ES-3), indicating a period when the lake system was at least seasonally dry or completely desiccated. Two other sets of irregular hummocky reflectors, ES-1 and ES-2, were identified in water depths above 9 and 17 m, respectively. The maximum depth of the ES-2 unconformity corresponds with an onlap sedimentary sequence formed from 14–17 m depth. The maximum depths of ES-1 and ES-2 are consistent at 9 and 17 m, respectively.

Radiocarbon Dates

Terrestrial macrofossils were not present in sufficient quantity for AMS radiocarbon dates at most stratigraphic levels, so aquatic macrofossils were used when necessary to constrain the timing of important events. We estimate ^{14}C reservoir-age associated with aquatic plant-macrofossils to be small for the Birch Lake system (Table 3), in part because limestone is not found in the watershed. Two sets of paired aquatic–terrestrial radiocarbon dates from the same stratigraphic level are not significantly different (compare CAMS-13587 with 13588 and CAMS-17026 with 18718). Reworking of organic matter is possible during lake-level changes; therefore, only well-preserved macrofossils that were $>$500 μm were used for ^{14}C measurements. Accumulation rates were calculated using calibrated radiocarbon ages (Stuiver and Reimer, 1993) to estimate the amount of time between dated strata.

Ager (1975) reported a series of conventional radiocarbon dates from bulk-sediment samples at pollen zone boundaries in his Birch Lake record. The interpolated ages for the same boundaries in this AMS ^{14}C study are significantly younger,

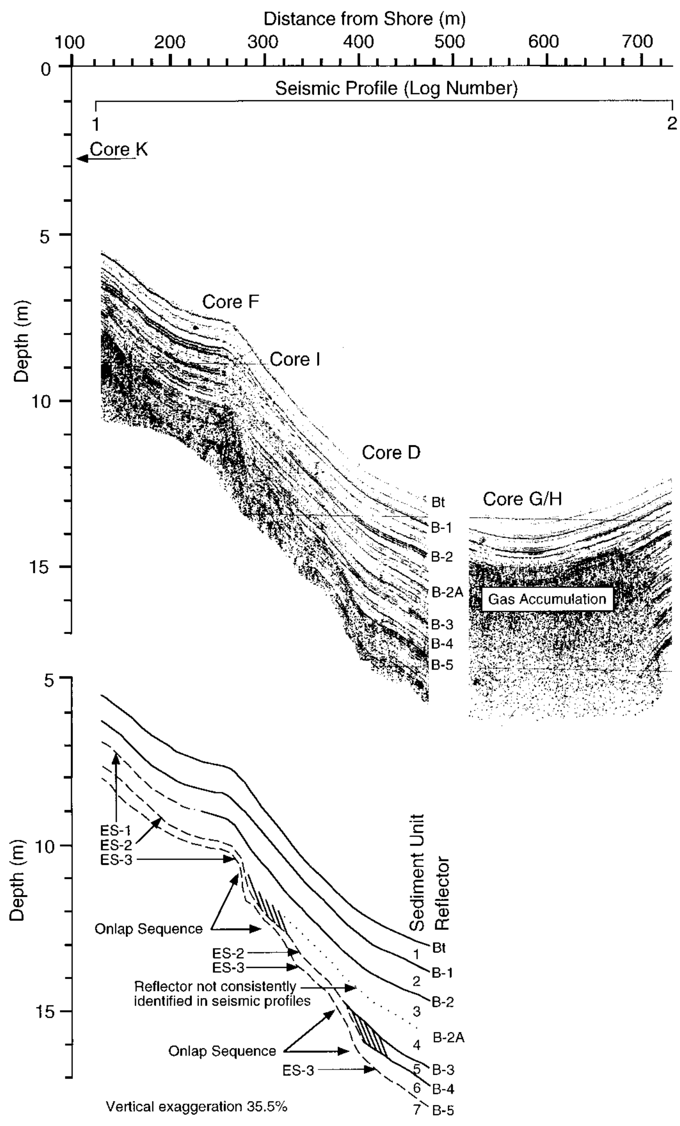

FIG. 3. Seismic profile 1-2 from the southern basin of Birch Lake, filtered between 3 and 7 kHz, illustrates the acoustic stratigraphy along the transect of cores K-G/H. Note the two sets of onlap sedimentary sequences formed during lake-level transgressions and erosion surfaces 1–3 highlighted by the illustration in the lower part of the figure. The zones of gas accumulation are limited to the center of the lake and to the sill separating the two basins.

TABLE 1
Description of Acoustic Reflectors and Layers

Reflector	Layer	Acoustic description	Interpretation
Bt		Continuous and traceable throughout the 2 basins, weak contrast in water in depths deeper than 10 m BOL	Reflector Bt represents the lake bottom; water and organic matter content of sediments increase with water depth corresponding with a decrease in grain size; the sediment–water interface becomes difficult to identify accurately in deeper water
	Bt to B-1	Faint parallel reflectors, draped, continuous, and traceable throughout the two basins	High lake-stand, water level at overflow stage, and stable as indicated by low variability in the acoustic and sediment properties
B-1		Widespread, continuous, parallel character, stronger contrast and small amplitude hummocks in shallow water (above 6 m BOL)	Reflector B-1 could represent a small-scale low stand of 6 m or less; however, possibly these features were formed by wave action or some combination of the two
	B-1 to B-2	Draped, continuous, parallel, and traceable throughout the two basins, small amplitude hummocks in shallow water (above 4 m BOL)	High lake-stand, water near or at present overflow level; short periods of lower lake level as much as 4 m BOL are suggested by hints of erosion surfaces in acoustic profiles from shallow water and by variations in sediment properties; however, these characteristics could be the result of wave action
B-2		In water depths below 9 m BOL this reflector is continuous and parallel, but in water depths above 9 m BOL this reflector is irregular, hummocky, and has very strong contrast	Reflector B-2 was formed during a lake-level rise from a low point 6 to 9 m lower to a level that approached or breached the overflow covering an erosion surface (ES-1) that had formed above the 9-m-BOL level
	B-2 to B-2A	Parallel and traceable in depths below 9 m BOL	Low lake-stand, water level between 6 and 9 m BOL
B-2A		This reflector was not identified consistently in the 18 km of seismic profiles, but where present it is parallel and traceable in depths below 9 m BOL	Reflector B-2A formed when water level decreased from approximately 5 m BOL to a level between 6 and 9 m BOL
	B-2A to B-3	Limited to depths below 9 m BOL, parallel and traceable reflectors, gas obscures sediments in basin depocenter	High lake-stand, water level rose from 14 m to within 5 m BOL as indicated by shallow water sediments in Core K between 5.3 and 4.8 m BOL
B-3		Strong contrast and parallel	Rising water level from 14 m BOL to within 5 m BOL as indicated by shallow water sediments formed in Core K between 5.3 and 4.8 m BOL
	B-3 to B-4	Parallel and traceable in water depths below 14–17 m BOL, onlap sequence between 14 and 17 m BOL, gas obscures sediments in basin depocenter	Low lake stand; water level dropped at least to a level 17 m BOL resulting in erosion of sediments above the 17 m BOL level; an onlap sequence formed between 14 and 17 m BOL indicates water level rose slightly during this time
B-4		Strong contrast, parallel, and limited to water depths below 14 m BOL	Small shallow lake with water level increasing from a low point of 17 to 14 m BOL; an erosion surface (ES-2) formed in depths above 14 m BOL
	B-4 to B-5	Continuous and traceable in water depths below 14 m BOL; irregular and hummocky in water depths above 14 m BOL; gas obscures sediments in basin depocenter	High lake stand; water level increased from complete desiccation at 18.5 m BOL to at least 5.3 m BOL as indicated from sediments in Core K
B-5		Very strong contrast, irregular, hummocky, and continuous throughout both basins; sonic ringing indicative of a strong density contrast	Reflector B-5 represents an erosion surface (ES-3) that formed when the lake was desiccated, at least on a seasonal basis if not for prolonged periods

suggesting that the bulk-sediment dates were contaminated with “old” carbon from the watershed (Table 4). This is a common problem in Arctic lakes where carbon is stored on the landscape for prolonged periods in soils (Abbott and Stafford, 1996).

In the Ager (1975) study, the increase in birch pollen occurred at 14,730 ± 830 ^{14}C yr B.P. (I-8068), preceding the initial lake-level rise identified in this study, which dates to approximately 12,300 ^{14}C yr B.P. The AMS dates for the birch rise of 11,500 ^{14}C yr B.P. from aquatic macrofossils are interpreted to represent reasonable approximations, whereas the bulk-sediment date of 14,730 ± 830 ^{14}C yr B.P. appears to be older than the true age of deposition by >3000 ^{14}C yr. The difference between the bulk sediment and macrofossil radiocarbon ages is not as great after the initial transgression; however, they differ by 300 to >1000 yr, respectively.

Sediment Core Transects

The suite of eight cores characterizes the depth–distance distribution of modern sediment facies along the transect and shows stratigraphic changes indicative of water-level fluctuations. The results of sediment analyses from five representative cores (K, F, I, D, and G/H) are presented in Fig. 4. The three unconformities identified by seismic profiles are also noted in the cores. The ES-1

TABLE 2
Comparison of Acoustic and Sediment Boundary Depths

Reflector	Sediment unit transition	Core F		Core I depth (m)		Core D		Core G/H	
		Acoustic depth (m)	Sediment depth (m)	Acoustic depth (m)	Sediment depth (m)	Acoustic depth (m)	Sediment depth (m)	Acoustic depth (m)	Sediment depth (m)
B-1	1 to 2	0.90	0.80	0.90	0.95	0.70	1.30	1.15	1.80
B-2	2 to 3	1.70	190	1.90	1.90	1.70	2.00	2.20	2.60
B-3	3/4 to 5 or 5A					3.20	3.00	g	3.90
B-3 (ES-2)	3/4 to 6	2.60	260	3.20	3.00				
B-4	5 or 5A to 6	not present	not present	not present	not present	3.70	3.70	g	4.50
B-5 (ES-3)	6 to 7	2.80	2.85	3.40	3.30	4.20	4.10	g	5.00

Note. A "g" indicates not visible because of gas.

and ES-2 erosion surfaces are limited to water depths above 9 and 17 m BOL, respectively, whereas ES-3 was identified in all cores from both the northern and southern basins. A summary of acoustic and sedimentary data and their interpretation in terms of lake level is presented in Table 1.

Sediment unit 7 forms the basal section underlying the ES-3 surface in all water depths (Fig. 4). This section is characterized by dry, compacted sediments with high magnetic susceptibility (100 to >500 SI), coarse grain size, high wet and dry bulk densities, low organic matter (<2%), and low calcium

TABLE 3
AMS Radiocarbon Dates from Birch Lake Cores

Laboratory number (CAMS)	Core	Depth (cm)	Material measured	Measured age (^{14}C yr B.P.)	Median calibrated age (yr B.P.)
17032	K	370.5	>500 μm aquatic macrofossil	4750 ± 60	5540
17033	K	464.5	>500 μm aquatic macrofossil	8780 ± 60	9710
17034	K	477.5	>500 μm aquatic macrofossil	10,030 ± 70	11,235
17035	K	519.5	>500 μm aquatic macrofossil	10,830 ± 50	12,730
13592	K	548	>500 μm aquatic macrofossil	12,010 ± 70	14,000
13594	K	571.5	>500 μm aquatic macrofossil	12,180 ± 70	14,220
13586	F	843	>500 μm aquatic macrofossil	4270 ± 70	4840
13587	F	907	>500 μm aquatic macrofossil	6270 ± 70	7180
13588	F	907	charcoal	6230 ± 70	7110
13589	F	943	>500 μm aquatic macrofossil	8060 ± 130	8980
17029	F	974	>500 μm aquatic macrofossil	10,490 ± 60	12,410
17030	F	1006	>500 μm aquatic macrofossil	10,630 ± 60	12,560
13590	F	1030	wood	12,310 ± 90	14,390
13591	F	1044	wood	11,930 ± 70	13,910
17031	I	1227	>500 μm aquatic macrofossil	9770 ± 60	10,980
13583	D	1271	>500 μm aquatic macrofossil	2320 ± 70	2380
17026	D	1360	>500 μm aquatic macrofossil	4790 ± 90	5510
18718	D	1361.5	pollen	5100 ± 120	5900
17027	D	1417	>500 μm aquatic macrofossil	8450 ± 60	9440
17028	D	1456	>500 μm aquatic macrofossil	10,020 ± 60	11,190
13585	D	1589	>500 μm aquatic macrofossil	11,640 ± 160	13,570
9586	D	1596.5	wood	12,300 ± 90	14,370
13584	D	1611	>500 μm aquatic macrofossil	12,020 ± 90	14,020
25425	G/H	1530	pollen	4810 ± 60	5580
25426	G/H	1604	pollen	6590 ± 60	7400
25422	G/H	1604	>500 μm aquatic macrofossil	6630 ± 90	7500
25427	G/H	1643	pollen	8480 ± 60	9450
25423	G/H	1691	pollen	9210 ± 340	10,260
25424	G/H	1758.5	pollen	11,420 ± 120	13,330
25420	G/H	1842	pollen	11,840 ± 100	13,800
25421	G/H	1864	pollen	12,780 ± 60	15,080
22000	Z	1690	>500 μm aquatic macrofossil	12,440 ± 200	14,560

TABLE 4

Comparison of the Age of Pollen Zones Dated by Conventional Radiocarbon Measurements on Bulk Sediment (Ager, 1975) with AMS Radiocarbon Dates on Macrofossils from This Study

Stratigraphic level	Macrofossil (^{14}C yr B.P.)	Bulk sediment (^{14}C yr B.P.)	Laboratory number	Age difference (^{14}C yr B.P.)
Alder rise	7200	8450 ± 150	I-8066	1250
Spruce rise	8800	9185 ± 325	I-8070	400
Birch rise	11,500	14730 ± 830 (from Ager, 1975)	I-8068	3200

carbonate content. These characteristics combined with extremely low pollen concentrations (Bigelow, 1997) and the lack of aquatic macrofossils support the interpretation that this unit was deposited in a lake that desiccated at least on a seasonal basis if not for prolonged periods, resulting in the oxidation of organic matter. The transition from sediment unit 7 to unit 6 is distinguished by decreased magnetic susceptibility, decreased dry bulk density (from >1.4 to <1.2 g/cm^3), and increased organic matter content (from <2 to ~5%) (Table 1). The transition from unit 7 to unit 6 is abrupt (cm-scale) in cores

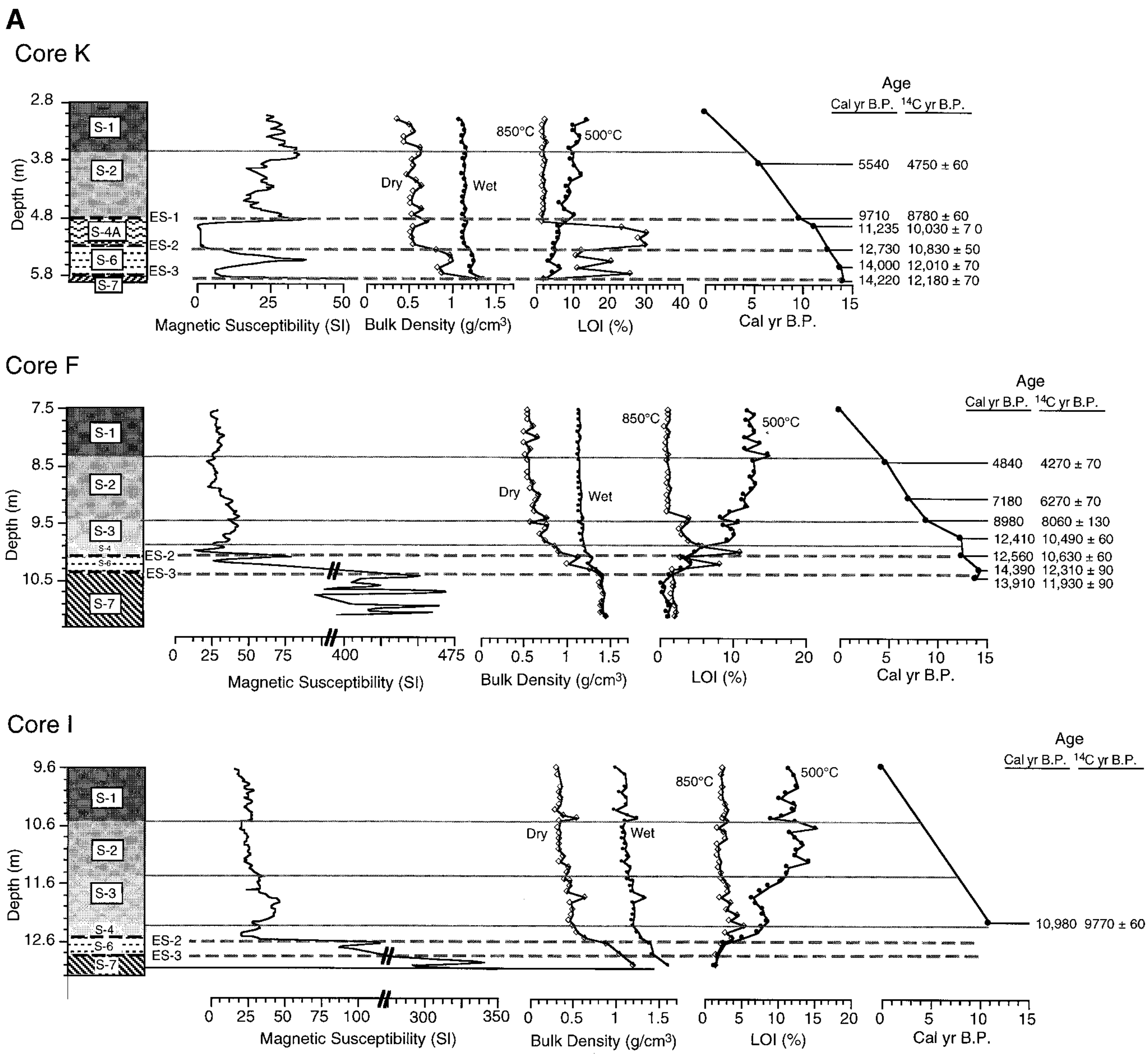

FIG. 4. Sediment analyses of cores K (2.8 m BOL), F (7.5 m BOL), I (9.6 m BOL), D (12.0 m BOL), and G/H (13.5 m BOL), including magnetic susceptibility, bulk density, and loss on ignition (LOI) at 500° and 850°C. Core lithology was determined from sediment descriptions and smear-slide mineralogy. Age–depth plots were made using calibrated radiocarbon ages.

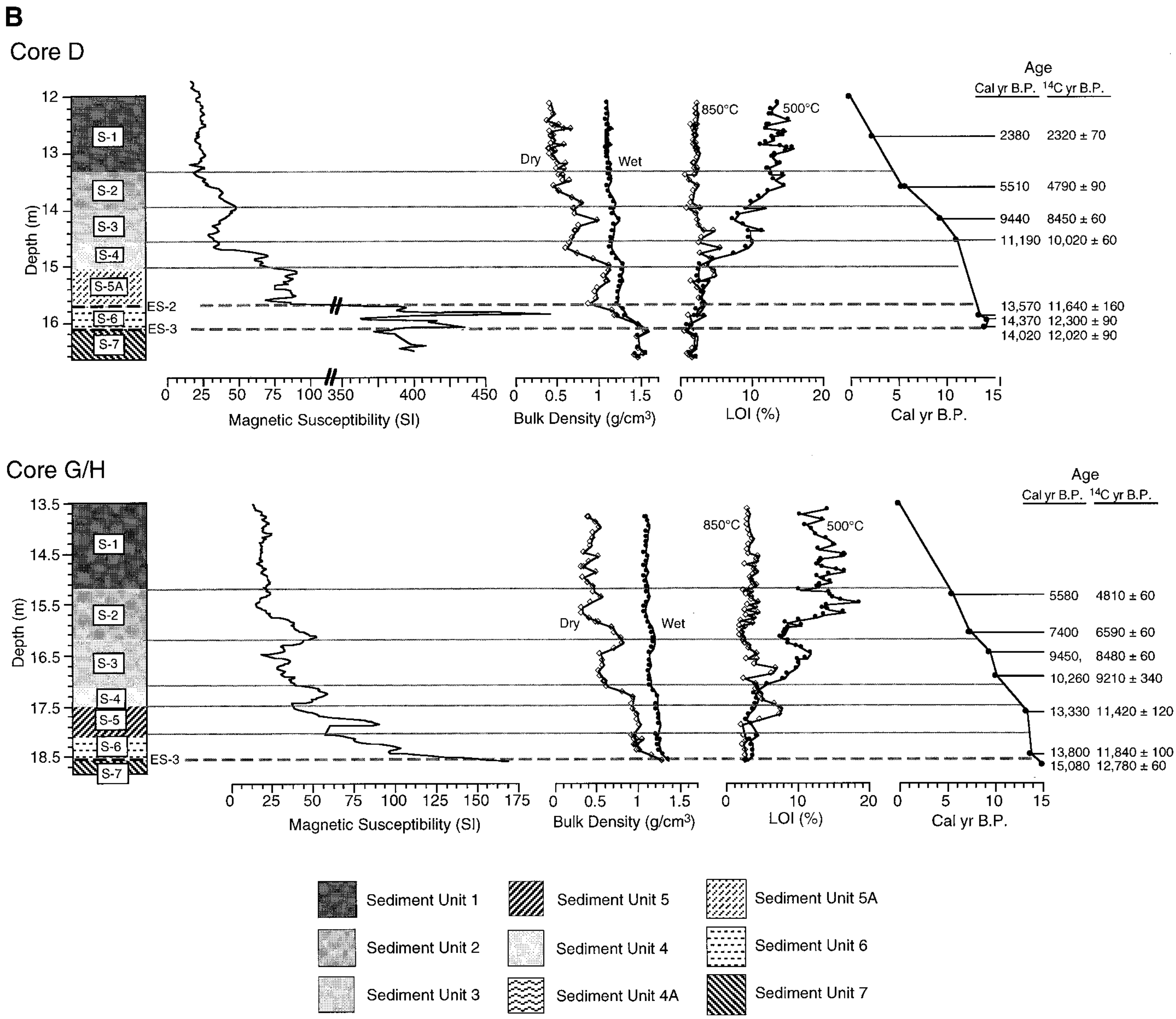

FIG. 4—*Continued*

from all water depths and is identified as the erosion surface ES-3.

Sediment unit 6 is characterized by an increase in calcium carbonate content in cores K and F, which corresponds to the presence of concentrated shallow-water shell debris. The age of the base of unit 6 is 12,780 ± 60 ^{14}C yr B.P. (CAMS-25421) in core G/H, 12,440 ± 200 ^{14}C yr B.P. (CAMS-22000) in core Z, and 12,180 ± 70 ^{14}C yr B.P. (CAMS-13594) in core K. These data indicate that the water level rose in both basins from the paleo base-level at ~19 m BOL to near the overflow stage within several centuries. Cores from deeper water contain only fragmented shells and are characterized by greater magnetic susceptibility, coarser grain size, and higher accumulation rates. The transition from sediment unit 6 to unit 5 is found only in cores from deeper than 17 m BOL, indicating it was formed during a regression phase and a low lake stand. Core D contains the transition from unit 6 to unit 5A, which is characterized dry compacted sediments with very low pollen concentrations (Bigelow, 1997) overlain by sediments with a marked decrease in magnetic susceptibility, bulk density, and grain size. The transition from sediment unit 6 to unit 5A is abrupt (cm-scale) and is identified as the erosion surface ES-2. In water depths above the limit of sediment unit 5A, the ES-2 boundary occurs between sediment units 6 and 4.

Sediment unit 5 is distinguished by an increase in shell material and calcium carbonate content, suggesting shallow-water conditions at core site G/H, which is currently more than 17 m BOL. Similarly, a decrease in magnetic susceptibility and an increase in shell material and aquatic macrofossils characterize sediment unit 5A.

Sediment unit 4 occurs in cores collected from more than 5 m BOL. Core K contains sediment unit 4A which is a shallow-water deposit formed during the same period, indicating that the lake was approximately 4–5 m below the overflow level. Decreased magnetic susceptibility, bulk density, and grain size, and increased organic matter content, characterize sediment unit 4. The radiocarbon age of the base of this unit is 10,630 ± 60 ^{14}C yr B.P. (CAMS-17030) in core F. Sediment unit 4A is nearly pure calcium carbonate, with high concentrations. The radiocarbon age from the base of sediment unit 4A is 10,830 ± 50 ^{14}C yr B.P. (CAMS-17035) in core K. The top of sediment unit 4A has a radiocarbon age of 10,030 ± 70 ^{14}C yr B.P. (CAMS-17034) for core K.

Sediment unit 3 is characterized in cores with sediments from more than 9 m BOL (cores F, I, D, and G/H) by an increase in organic matter content followed by a decrease. The increase in organic matter content is coupled with a decrease in magnetic susceptibility, grain size, and bulk density. The transition between units 3 and 2 is characterized by an interval of lower organic-matter content (3% decrease), increased magnetic susceptibility (>10 SI unit increase), and increased dry bulk density (0.2 g/cm^3 increase). The contact between units 4A and 2 in cores from above 5 m BOL is abrupt (1 cm) and marked by a shift from nearly pure calcium carbonate sediments containing high concentrations of shell debris (>500 μm) to an organic-rich silt containing <10% calcium carbonate. We interpret this sequence as a lake-level drop with erosion followed by a rise in water level forming erosion surface ES-1. Organic matter content increases upward in deep-water cores in sediment unit 2 from ~10 to >15%.

Sediment unit 2 is characterized by an increase in organic matter content at all water depths coupled with a decrease in calcium carbonate in shallow water cores. The rise in organic matter corresponds with a decrease in magnetic susceptibility and bulk density in all cores. Radiocarbon ages for the base of unit 2 range from 8780 ± 60 ^{14}C yr B.P. (CAMS-17033) in core K to 8060 ± 130 ^{14}C yr B.P. (CAMS-13589) in core F. The transition from unit 2 to unit 1 is characterized by slightly higher magnetic susceptibility (~5 SI increase) and higher organic matter (~3% increase).

Sediment unit 1 is characterized by high organic carbon content in all cores coupled with low magnetic susceptibility and bulk density. Radiocarbon ages from the base of sediment unit 1 are 4750 ± 60 ^{14}C yr B.P. (CAMS-17032) from core K, 4270 ± 70 ^{14}C yr B.P. (CAMS-13586) from core F, and 4790 ± 90 ^{14}C yr B.P. (CAMS-17026) from core D.

Lake-Level History

Water levels in both basins of Birch Lake rose between 12,700 and 12,200 ^{14}C yr B.P. after the lake was either seasonally dry or desiccated for a prolonged period, indicating an arid phase prior to 12,700 ^{14}C yr B.P. Although initially the northern and southern basins did not have a surfice connection, water levels rose at approximately the same time in both basins, as indicated by the radiocarbon age of the basal lacustrine sediments. The transition into lacustrine sediments occurred at 12,780 ± 60 ^{14}C yr B.P. at 18.6 m BOL in the southern basin and 12,440 ± 200 ^{14}C yr B.P. at 16.9 m BOL in the northern basin. The younger age in the northern basin can be attributed to the higher stratigraphic level of the basin. Likewise, the radiocarbon age of the basal lacustrine sediments is generally younger at higher stratigraphic levels in the cores from the southern basin: (1) 12,780 ^{14}C yr B.P. at 18.6 m BOL, (2) 12,300 ± 90 ^{14}C yr B.P. at 15.9 m BOL, (3) 11,930 ± 70 ^{14}C yr B.P. (CAMS-13591) at 10.4 m BOL, and (4) 12,180 ± 70 ^{14}C yr B.P. at 5.7 m BOL. The age distribution among core sites suggests that the transgression may have spanned ~800 ^{14}C yr and may have had a complex history of small-scale water-level fluctuations. Water levels reached ~5 m below the overflow level, as indicated by a shallow water deposit in core K, before falling again by 17 m between 11,600 and 11,400 ^{14}C yr B.P. (Fig. 5).

Supporting evidence within the region for an increase in the P–E balance over interior Alaska is found at several other sites. During this period water levels also rose in Jan Lake in the Tanana Valley and Sands of Time on the Yukon Flats (Finney *et al.*, 1994, Fig. 1). Hamilton (1986) dated the Itkillik II glacier advance in the Brooks Range between 13,000 and 11,500 ^{14}C yr B.P., and he correlated it with the McKinley Park III advance in the Alaska Range (Hamilton, 1994), which occurred between 12,500 and 11,500 ^{14}C yr B.P. (Ten Brink and Waythomas, 1985; Fig. 5). This evidence is consistent with wetter conditions between ca. 12,500 and ca. 11,500 ^{14}C yr B.P. in the whole intermontane interior and in the Alaska and Brooks ranges.

After the initial lake-level rise in Birch Lake, water levels decreased to ~17 m BOL after 11,600 ^{14}C yr B.P. and remained low until 10,600 ^{14}C yr B.P. Shallow-water deposits below 17 m BOL and an onlap sedimentary sequence that formed between 14 and 17 m BOL indicate a low lake stand during this period. Low water levels are supported by an erosion surface formed above the 14 m BOL level.

Bigelow *et al.* (1990) identified an increase in wind intensity between 11,100 and 10,700 ^{14}C yr B.P. from grain-size shifts at eolian sections along the Nenana River (Fig. 5). Although the climatic mechanism and link to the Younger Dryas remain controversial, these findings are consistent with more-arid conditions during this period. Further evidence of dune reactivation during this time is found in the Teshepuk dune field in northern Alaska at ca. 11,000 ^{14}C yr B.P. (Carter, 1993). Evidence from mountain glaciers indicates a general retreating trend during this period in both the Brooks and Alaska ranges, consistent with drier conditions.

After ca. 10,600 ^{14}C yr B.P., water in Birch Lake rose to within 5 m of its overflow stage, as indicated by shallow-water deposits in core K. Onlap sedimentary sequences between 9

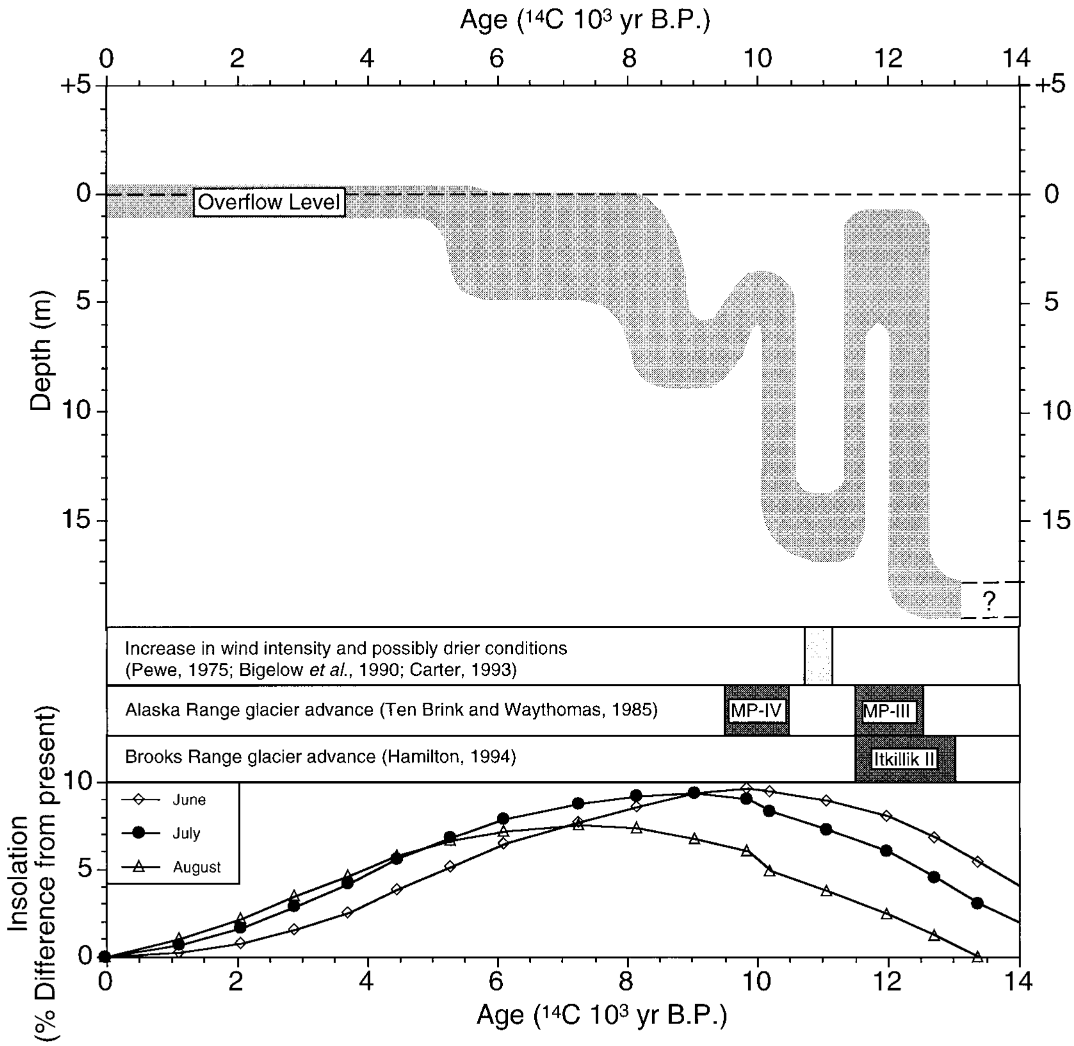

FIG. 5. Lake-level history of Birch Lake determined from seismic profiles and core transects. The thickness of the curve represents uncertainty in water levels. Correlation of transgression and regression phases with other studies is noted below the lake-level curve. Insolation changes for the last 14,000 ^{14}C yr B.P. relative to present are calculated from values given by Berger (1978a), Berger (1978b), and Berger and Loutre (1991). The insolation curve is plotted in radiocarbon years for comparison with the Birch Lake data. Monthly insolation for the summer wet season (June–August) and transition months (May and September) peaks between 12,000 and 9000 ^{14}C yr B.P.

and 12 m BOL in seismic profiles clearly show the lake transgression. Ten Brink and Waythomas (1985) dated the McKinley Park II advance between 10,500 and 9500 ^{14}C yr B.P., roughly consistent with colder and/or moister conditions at Birch Lake. Sand deposition waned on the Nenana River and soils began to form during this period (Bigelow *et al.*, 1990).

Lower lake levels occurred during the early Holocene between 10,000 and 8800 ^{14}C yr B.P. as indicated by an erosion surface in core K and a shallow-water deposit in core F. This period is marked by continued eolian activity, with considerable loess accumulation in the Tanana valley (Péwé, 1975). The eolian activity may reflect a drier climate directly or increased alluviation of glacially derived material as mountain glaciers retreated under warmer and drier conditions. In either case, this is compatible with lowered lake levels.

The lake rose at 8800 ^{14}C yr B.P. and likely overflowed at least on a seasonal basis after this time. The Birch Lake record indicates that early to middle Holocene climate became increasingly wet, but was still drier than today. Neither acoustic nor sediment records show strong evidence of lower lake stands during the middle or late Holocene, implying that the hydrologic balance of the Birch Lake system has remained positive since this time. Hints of low water stands identified from seismic profiles suggest possible low lake stands throughout sediment unit 2 until approximately 4800 ^{14}C yr B.P.

Controls Over Effective Moisture Changes

The lake-level results from Birch Lake highlight four periods with different effective moisture characteristics: (1) a severe phase of aridity prior to 12,700 ^{14}C yr B.P., (2) a two-step increase in effective moisture between 12,700 and 10,000 ^{14}C yr B.P., (3) a prolonged period of drier than present conditions between 10,000 and 4800 ^{14}C yr B.P., and (4) the late Holocene, when P–E balance became more positive and Birch Lake overflowed (which made the watershed and therefore the lake record less sensitive to further increases in moisture balance).

Today in interior Alaska the main source of moisture is from air advected from the North Pacific in a westerly flow, predominantly during the summer months. The strongly continental climate means that during the short, hot summer, evaporation levels can be high and water deficits are relatively common. Edwards *et al.* (in press) discuss possible synoptic-scale variations that could give rise to different precipitation–evaporation patterns in the eastern interior of Alaska. Here we focus on the larger-scale controls that may underlie the major changes in the late-Quaternary record, assuming that these changes are representative of the Alaskan interior as a whole.

Late-Pleistocene arid phase (pre-12,700 ^{14}C yr B.P.). Our results are consistent with past work that suggests a prolonged dry period during the late Pleistocene in interior Alaska, which was considerably more arid than at anytime during the Holocene (Barnosky *et al.*, 1987; Bartlein *et al.*, 1991; Anderson and Brubaker 1993; Edwards and Barker, 1994; Hu *et al.*, 1996; Hu *et al.*, 1998). The presence of the Laurentide ice sheet, increasing summer insolation (greater than modern), cool sea-surface temperatures, and lower sea level (~100 m) are important factors that would have influenced climate at the end of the last glaciation. During the last glacial maximum, the Laurentide ice sheet generated a stable anticyclone that deflected westerlies southward of their modern track; this would have reduced moisture advection to interior Alaska and generated conditions drier than present (COHMAP, 1988). This pattern probably persisted until late-glacial time (Bartlein *et al.*, 1991). A cooler North Pacific probably had a moderating effect on summer temperatures (Bartlein *et al.*, 1991). Peteet *et al.* (1997) showed with modeling experiments that cooling north Pacific sea-surface temperatures resulted in cooler atmospheric temperatures and a strong decrease in precipitation in adjacent regions. Thus a cooler North Pacific would likely result in drier conditions in interior Alaska. Sea level was more than 100 m lower than present prior to 12,700 ^{14}C yr B.P. (Fairbanks, 1989), which left vast expanses of the Bering, Beaufort, and East Siberian shelves exposed. This would have increased the distance for moisture transport into the interior of the continent, in contrast to present conditions. The combined effects of these factors are likely to have led to a much more arid environment and to desiccated lakes or lakes at very low stands.

Two-step increase in effective moisture, 12,700–10,000 ^{14}C yr B.P. The data from Birch Lake suggest two wet periods punctuated by a dry phase between ca. 12,700 and 10,000 ^{14}C yr B.P. These wet phases appear synchronous among sites in the Alaskan interior (Sands of Time and Jan Lake; B. Finney, unpublished data). The dry phase at Birch Lake (ca. 11,600–10,600 ^{14}C yr B.P.) began prior to the onset and terminated in the middle of the Younger Dryas interval.

As the Laurentide ice sheet waned, its effect on circulation would have relaxed, and a westerly flow more typical of today would have been reestablished (Bartlein *et al.*, 1991). The first rapid rise in Birch Lake between ca. 12,700 and 12,200 ^{14}C yr B.P. may be related to this major shift in hemispheric circulation patterns.

The wet phases can also be roughly correlated with the two periods of accelerated sea-level rise identified by Fairbanks (1989), Edwards *et al.* (1993) and Bard *et al.* (1996; Fig. 5). Between 12,500 and 11,600 ^{14}C yr B.P., sea level rose from 100 to 70 m below present (Fairbanks, 1989). The rate of sea-level rise slowed dramatically between 11,600 and 10,600 ^{14}C yr B.P. Edwards *et al.* (1993) present New Guinea coral-reef data that indicate a second rise, coupled with vigorous ocean circulation, during the second half of the Younger Dryas, ca. 11,200 U/Th yr B.P. The radiocarbon plateau makes comparison of this second event and the record at Birch Lake difficult. However, the timing appears to be similar. Thus, there may be a possible hemispheric-scale link between more vigorous ocean circulation and increased advection of moisture into the Alaskan interior.

The ice-sheet-circulation effect was presumably not the cause of the dry phase between 11,600 and 10,600 ^{14}C yr B.P. However, summer insolation levels at 65°N were rising during deglaciation, and peaked between ca. 10 and 7% greater than present during the period 12,000 to 9000 yr B.P.; values approached modern over the last few thousand years (Fig. 5). Thus, high levels of summer insolation, cool sea-surface temperatures, and sea-level still greatly lower than present may have combined during this period to produce conditions of relatively low effective moisture, perhaps enhanced by a less-vigorous hemispheric circulation.

Drier-than-present early and middle Holocene (ca 10,000–4800 ^{14}C yr B.P.). The early-Holocene lake lowering probably reflects a suite of changes in the climate system. Levels did not fall as far as previously. Precipitation and evaporation were probably greater than before, as the climate system developed a truly interglacial character. The rapidly flooding continental shelves would have brought moisture sources closer to the interior. Increasing sea-surface temperatures would probably have generated generally warmer and moister air masses that advected eastward. On the other hand, high summer isolation and warmer air temperatures in the interior would have raised evaporation rates. The result appears to have been effective moisture levels lower than those of the present during the early and middle Holocene.

Birch Lake shows a clear rise at ca. 8800 ^{14}C yr B.P., although it did not apparently permanently overflow at that time. The mechanisms for this further rise are as yet unclear, but they may reflect the onset of positive feedbacks in the climate system that would further influence temperature and moisture levels, such as might be generated by decreasing sea-ice extent and a longer ice-free season in the Bering sea (Bartlein *et al.*, 1991) or the expansion of coniferous forest across the region (Anderson and Brubaker, 1994). Given the interest in the possible feedback effects of coniferous forest on the climate system (e.g., Foley *et al.*, 1994), the close correlation between this moisture increase and the expansion of *Picea* forest at Birch Lake and other lakes in the eastern interior of Alaska (Bigelow, 1997) should be further investigated.

Late Holocene moist phase. During the late Holocene the (past 4800 ^{14}C yr B.P.) values for most of the major controls over the Alaskan climate system (sea-surface temperatures, sea-ice extent, insolation) have been close to those of the present (Bartlein *et al.*, 1991). Birch Lake began to overflow by this time and apart from indicating that this is the wettest period in the past ca. 12,700 ^{14}C yr B.P., its record provides no information on subsequent changes in moisture.

The complexities of the climate record require a multiproxy approach to the reconstruction of past climate changes, and this study demonstrates that the study of lake levels can provide an important contribution to documenting Alaskan climate history. The data presented here are strong evidence for climate fluctuations in interior Alaska during deglaciation; this period was not characterized by unidirectional increases in temperature and moisture. On top of large-scale changes related to changing ice volumes and insolation, there may have been a response to the hemispheric-scale modification of circulation that occurred before and during Younger Dryas time. Other patterns may only be fully explained when we better understand the effect of positive feedbacks (e.g., sea-ice, forest cover) and the synoptic-scale patterns of climate variability that result from the major changes discussed here.

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