QUATERNARY RESEARCH **49,** 157–170 (1998)
ARTICLE NO. QR981962

Glacioisostasy and Lake-Level Change at Moosehead Lake, Maine

Greg Balco and Daniel F. Belknap

Department of Geological Sciences, University of Maine, Orono, Maine 04469-5790

and

Joseph T. Kelley

Maine Geological Survey, State House Station 22, Augusta, Maine 04333

Received June 13, 1997

Reconstructions of glacioisostatic rebound based on relative sea level in Maine and adjacent Canada do not agree well with existing geophysical models. In order to understand these discrepancies better, we investigated the lake-level history of 40-km-long Moosehead Lake in northwestern Maine. Glacioisostasy has affected the level of Moosehead Lake since deglaciation ca. 12,500 ^{14}C yr B.P. Lowstand features at the southeastern end and an abandoned outlet at the northwestern end of the lake indicate that the lake basin was tilted down to the northwest, toward the retreating ice sheet, by 0.7 m/km at 10,000 ^{14}C yr B.P. Water level then rose rapidly in the southeastern end of the lake, and the northwestern outlet was abandoned, indicating rapid relaxation of landscape tilt. Lowstand features at the northwestern end of the lake suggest that the lake basin was tilted to the southeast at ca. 8750 ^{14}C yr B.P., possibly as the result of a migrating isostatic forebulge. After 8000 ^{14}C yr B.P., water level at the southeastern end was again below present lake level and rose gradually thereafter. We found no evidence suggesting that postglacial climate change significantly affected lake level. The rebound history inferred from lake-level data is consistent with previous interpretations of nearby relative sea-level data, which indicate a significantly steeper and faster-moving ice-proximal depression and ice-distal forebulge than geophysical models predict.

INTRODUCTION

The advance and retreat of large ice sheets causes crustal depression and rebound in areas beneath and marginal to the ice, as the crust and upper mantle equilibrate with the changing ice load. The record of these crustal motions can be used to reconstruct ice dynamics and determine earth properties, typically by forward modeling in which the global deglaciation history and rheological parameters are adjusted until the model output matches field observations (e.g., Peltier, 1994; Clark *et al.*, 1994). Most such models are tuned using relative sea-level (RSL) records, which are well studied at many locations worldwide.

Because these numerical models are widely used to reconstruct large-scale global changes (Peltier, 1994) as well as local glacial events (Quinlan and Beaumont, 1981), it is important to evaluate them accurately using field data. Existing models, presumably due to their coarse spatial resolution, duplicate observed RSL records poorly in some ice-marginal areas having complicated deglaciation histories, such as the Irish Sea (McCabe, 1997) and the Gulf of Maine (Tushingham and Peltier, 1991; Barnhardt *et al.*, 1995). Models and field evidence for the Gulf of Maine and Atlantic Canada indicate that the region experienced a time-transgressive RSL lowstand as a marginal isostatic forebulge followed the retreating ice margin northward (Barnhardt *et al.*, 1995; Liverman, 1994; Quinlan and Beaumont, 1981, 1982). However, RSL data from the Maine coast (Barnhardt *et al.*, 1995) and Quebec City (Dionne, 1988; Dionne and Coll, 1995) indicate that this forebulge was significantly steeper, narrower, faster-moving, and more persistent than predicted by geophysical models. We initiated this study to reconstruct crustal deformation in interior Maine independently of RSL data and thus better understand the discrepancies between actual and modeled regional crustal deformation. We selected Moosehead Lake because it lies directly between the Maine coast and Quebec City (Fig. 1), an ideal location to test reconstructions of glacioisostatic rebound derived from RSL curves at these locations.

RSL data measure vertical motions of the crust at a specific point. Changes in lake level, on the other hand, measure changes in the tilt of the landscape. The relationship between late-glacial and Holocene lake-level changes and glacioisostasy is well known in the Great Lakes (Goldthwait, 1908; Clark *et al.*, 1994), but the thousands of other lakes in glaciated North America, which presumably contain an equally useful record of glacioisostasy, have not been well studied from this perspective. In Maine, the ice sheet retreated to the northwest. Thus, lake basins were tilted to the northwest immediately after deglaciation. As the marginal bulge passed underneath them, they must have

0033-5894/98 $25.00

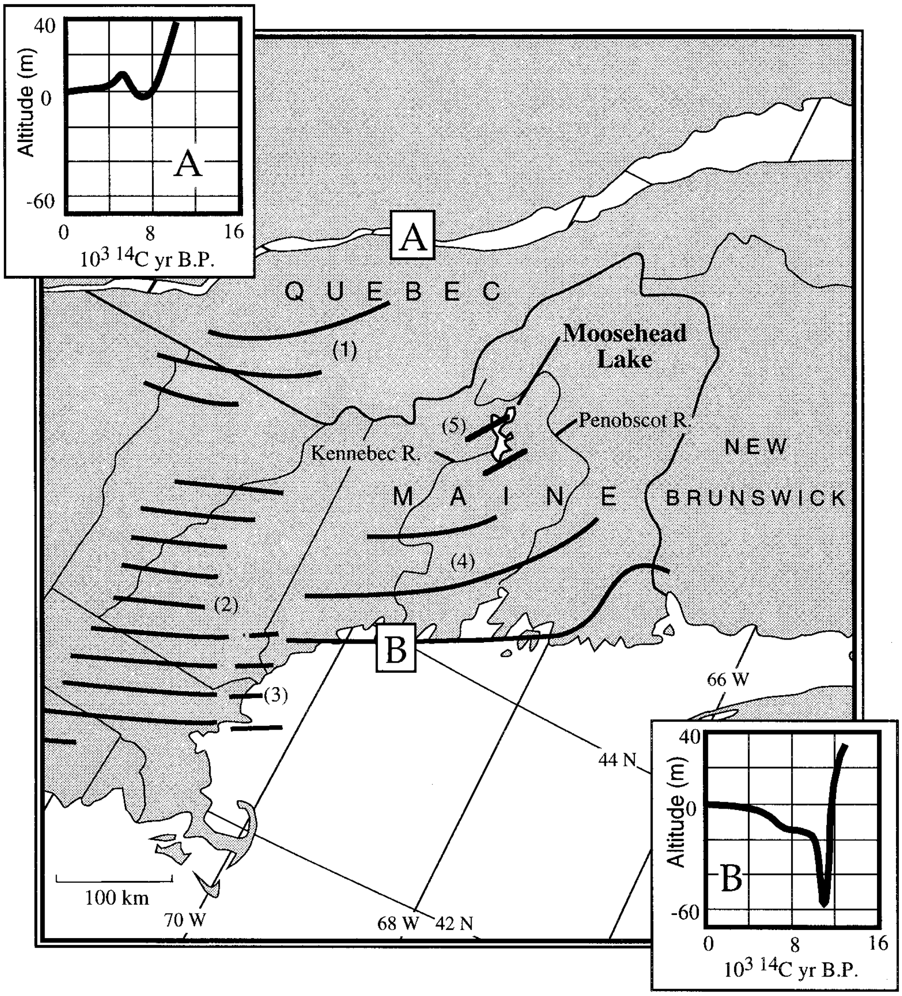

FIG. 1. Map of Maine and adjacent Canada, showing location of Moosehead Lake and evidence pertaining to regional glacioisostatic rebound. Dark lines are contours drawn on sets of late-glacial marine and lacustrine shoreline features. Contour interval, 25 m. Sources as follows: (1) Champlain II shorelines, 0.9 m/km tilt, ca. 11,700 ^{14}C yr B.P. (Parent and Occhietti, 1988); (2) Glacial Lake Hitchcock deltas, 0.9 m/km, 16,000–14,000 ^{14}C yr B.P., and (3) coastal Massachusetts/New Hampshire glaciomarine deltas, 0.9 m/km, 15,000–14,000 ^{14}C yr B.P. (Koteff *et al.*, 1993); (4) Maine glaciomarine deltas, 0.5 m/km, 14,000–13,000 ^{14}C yr B.P. (Thompson *et al.,* 1989); and (5) Moosehead Lake shoreline, 0.7 m/km, ca. 10,000 ^{14}C yr B.P. (this study). Insets show relative sea level curves for Montmagny, Quebec (A, Dionne, 1988) and the Maine coast (B, Barnhardt *et al.,* 1995).

been tilted in the opposite direction, to the southeast, for a short time. These events should be recorded by asynchronous lake-level changes: the early northwestward tilt by a highstand in the northwest end, and a lowstand in the southeast end, of each lake, then a later southeastward tilt by the opposite relationship (Fig. 2).

BACKGROUND AND RESEARCH STRATEGY

Physiography of Moosehead Lake

Moosehead Lake, the largest lake in Maine, forms the headwaters of the Kennebec River (Figs. 1 and 3). The lake surface is normally maintained at 313.6 m elevation (present lake level or PLL) by concrete dams located at East and West Outlets. Borings at West Outlet penetrate 15 m of till that overlies middle Wisconsinan lacustrine sediments (Balco, 1997). Bedrock crops out close to shore everywhere else in the lake. Thus, the lake occupies a bedrock basin which was enlarged slightly by a late Wisconsinan till dam, and further expanded by the modern dams. Borings through the dams intercept bedrock at 309.5 m (−4 m PLL) at East Outlet, and bouldery till at 312 m (−2 m PLL) at West Outlet (A. Straz, Central Maine Power Co., written communication, 1995). Historical accounts that predate dam construction indicate that West Outlet flowed only during spring high water (Calvert, 1983, p. 300), and it appears that original lake level (hereafter, OLL) was near or slightly above the East Outlet sill elevation at 309.5 m (−4 m PLL).

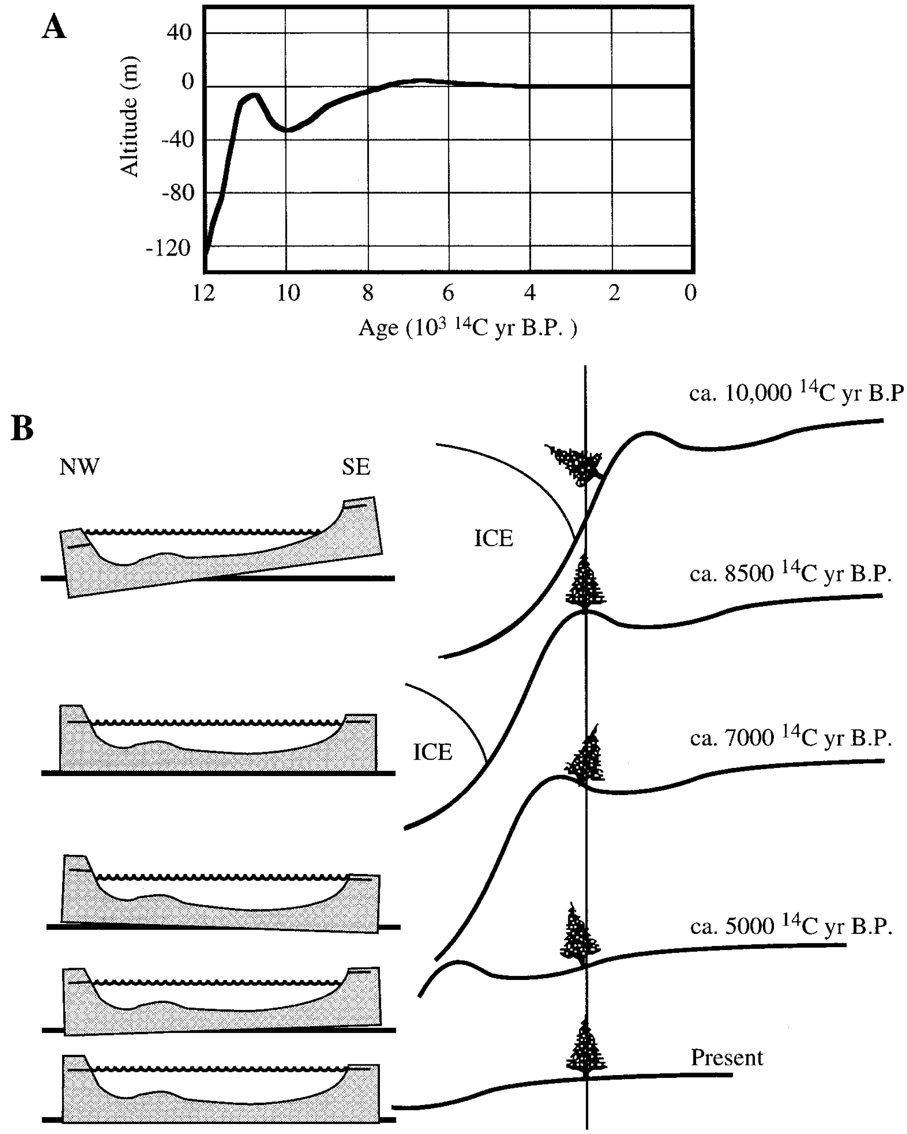

FIG. 2. (A) Glacioisostatic rebound curve for the Maine coast (Barnhardt *et al.*, 1995). (B) Diagrams showing the effect of a migrating ice-proximal depression and ice-distal forebulge on water level in a lake.

Suitability of Moosehead Lake for Reconstructing Isostatic Rebound

The ideal lake for reconstructing postglacial isostatic tilting should (1) be long enough in a direction normal to the ice margin that a subtle tilt of the landscape can produce measurable lake-level changes, (2) have outlets that have not significantly eroded since deglaciation, (3) have deglaciated quickly and not contained long-lasting glacial lakes, and (4) be insensitive to climatic and hydrologic changes that might affect lake level. Moosehead Lake satisfies these criteria. It is 30 km long in a direction normal to the ice margin, which is adequate to amplify a landscape tilt of 0.5–1 m/km into a relative lake-level change of 15–30 m. The outlets have not eroded significantly since deglaciation: the sill at East Outlet is metamorphic bedrock, while that at West Outlet is boulder-armored stony till, and a third, abandoned outlet (Figs. 3 and 4) is lined with boulders and exposed bedrock. None of these channels is incised more than 1–2 m below the surrounding landscape.

The Moosehead Lake basin contained a series of short-lived proglacial lakes during deglaciation (Lowell and Crossen, 1983; Balco, 1997). They have not been directly dated, but correlation of dated sites across Maine and adjacent Canada (Thompson and Dorion, 1996; Dorion, 1995; Thompson *et al.*, 1996; C. C. Dorion, unpublished data) indicates that the Moosehead area deglaciated fully between 13,000 and 12,500 ^{14}C yr B.P. After this time,

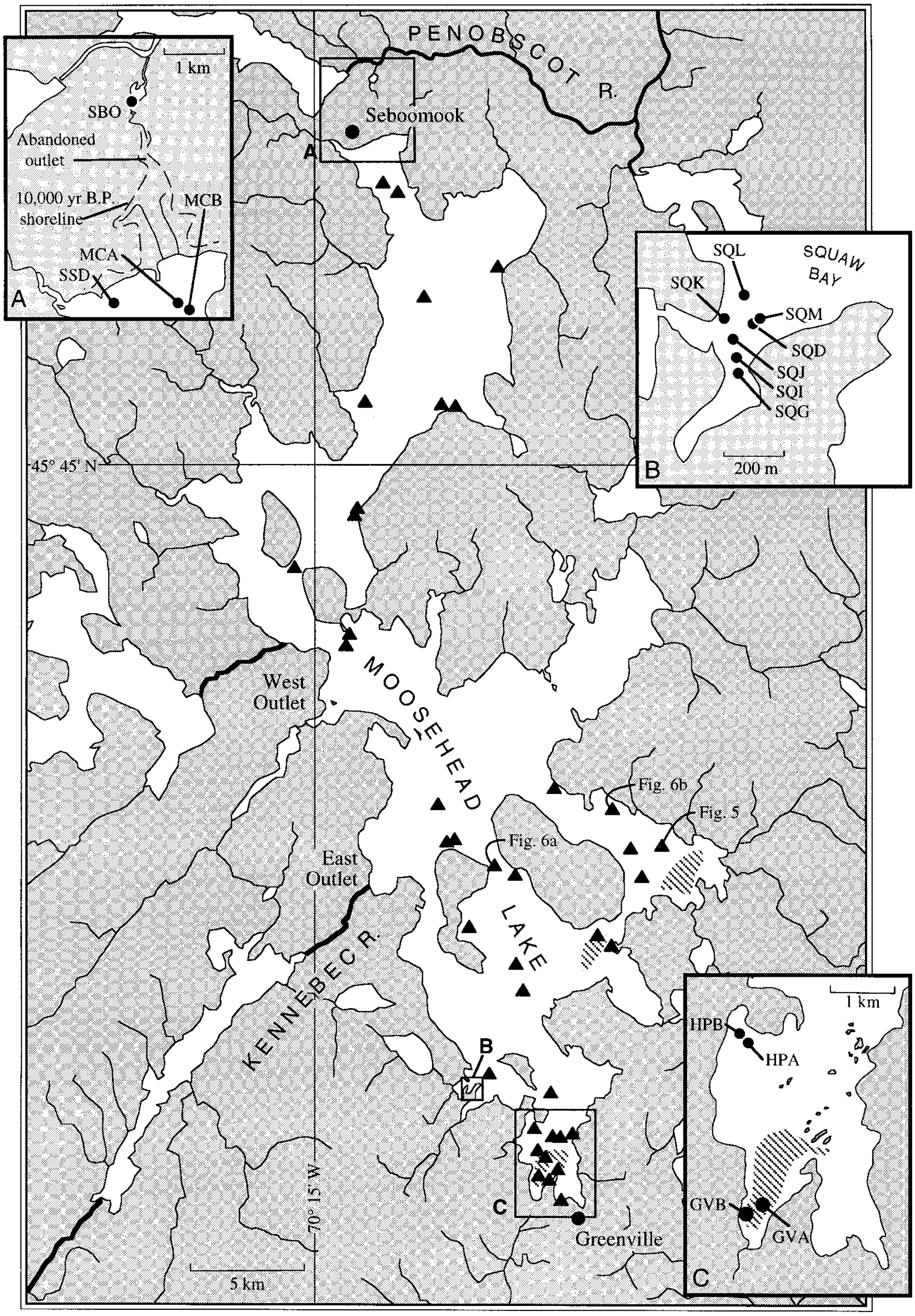

FIG. 3. Map of Moosehead Lake. Labeled circles are locations of sediment cores. Triangles are locations of erosional lowstand features observed on SPS and SSS records. Dark shaded regions are areas near the southeastern end of the lake where transgressive stratigraphy is observed on SPS records (see text).

lake level was presumably controlled only by outlet sill elevation.

Studies of lakes and ponds in Maine (Thompson *et al.*, 1995; Talbot, 1996; H. Almquist-Jacobson, pers. commun., 1996) and climate reconstructions for northeastern North America (COHMAP, 1988; Webb *et al.*, 1993; Bierman *et al.*, 1997) suggest that a warm, dry period in the middle Holocene may have lowered lake levels throughout the region. Talbot (1996) dated a lowstand unconformity at −13 to −15 m in Lake Auburn, Maine to between 8435 ± 95 and 6930 ± 105 ^{14}C yr B.P., and Petersen *et al.* (1994) obtained a minimum age of 6100 ± 120 ^{14}C yr B.P. for a

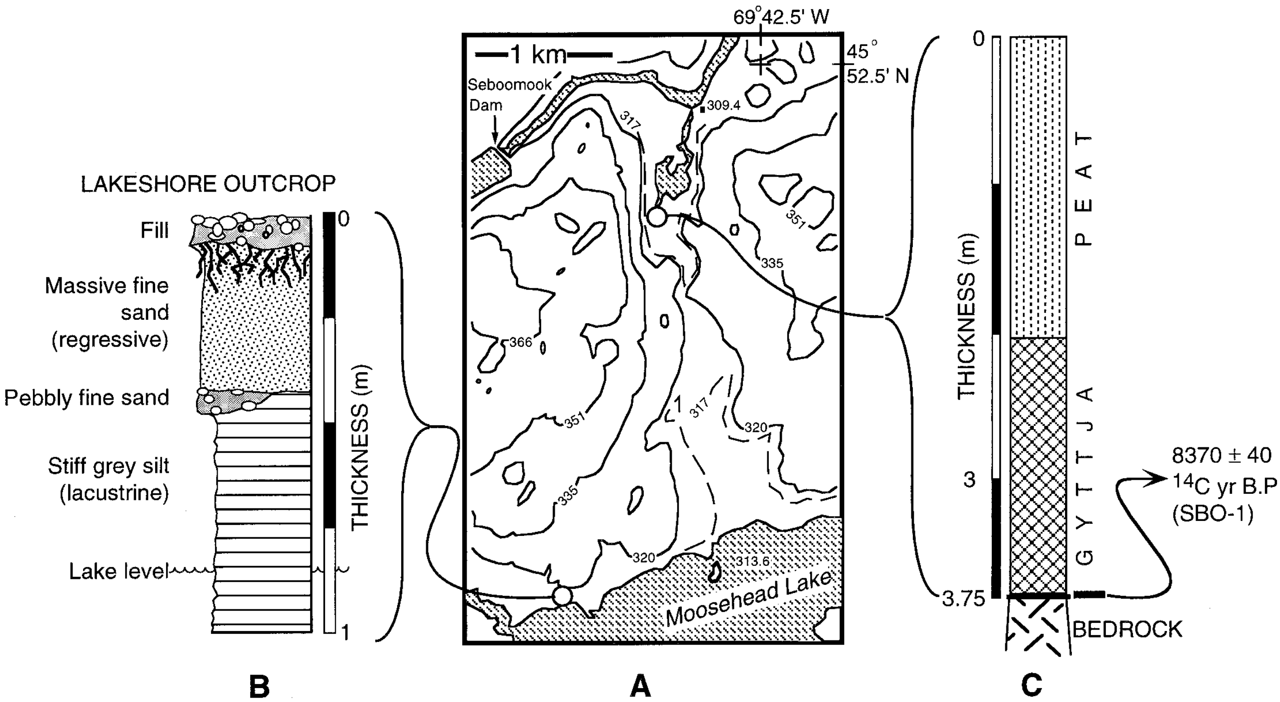

FIG. 4. (A) Map of abandoned outlet channel at the northwestern end of Moosehead Lake, after USGS 7.5′ Seboomook, ME Quadrangle. Elevations in meters. Area shown is same as inset A of Figure 2. (B) Regressive stratigraphy observed on the Moosehead side of the abandoned outlet. (C) Stratigraphy of marsh filling pools in abandoned channel and context of radiocarbon date limiting the age of channel abandonment.

similar unconformity at −8 m at Sebasticook Lake. Both studies invoked a drier climate to explain these lowstands. In general, large through-flowing lakes like Moosehead are insensitive to climate change (e.g., Street-Perrott and Harrison, 1985). However, since the historical lake-level record at Moosehead begins after the construction of dams, the actual sensitivity of the lake to climate can be inferred only from water-balance calculations and climate norms.

In a large bedrock basin like that of Moosehead Lake, the surface water and groundwater basins are presumably the same. Thus, the water balance for the lake is

$$EA_{\mathrm{L}} + Q = PA_{\mathrm{L}} + R\,A_{\mathrm{DB}}, \qquad (1)$$

where E is evaporation over water, R runoff, Q discharge at the outlet, P precipitation, and A_{L} and A_{DB} the area of lakes within the drainage basin and the land area within the drainage basin, respectively (Benson and Thompson, 1987). A_{L} in the Moosehead basin is 370 km^2, and A_{DB} is 2870 km^2. Solving Eq. (1) for $Q = 0$ gives the condition for evaporative drawdown:

$$370E = 370P + 2870R. \qquad (2)$$

At present, precipitation (P) in the Moosehead basin is approximately 1 m/yr (National Climatic Data Center, 1994), runoff is approximately 50 cm/yr (Lewis, 1997), and overwater evaporation (E) is approximately 60 cm/yr (Peixoto and Oort, 1992). Webb *et al.* (1993) suggested that annual precipitation in northern New England may have dropped to 75 cm/yr at the peak of mid-Holocene warmth at approximately 9000 yr B.P. Empirical precipitation–runoff relations for the United States (Langbein *et al.*, 1949) indicate that runoff would have been 18–28 cm/yr at this time. According to Eq. (2), evaporation rates of 215–293 cm/yr would be required to cut off outflow from Moosehead Lake under these conditions. Such evaporation rates are measured only in the most arid environments on Earth. The use of empirical precipitation/runoff relations for an ungauged drainage basin does introduce significant inaccuracies (Langbein *et al.*, 1949; Benson and Thompson, 1987). However, the unreasonable evaporation rates required by this calculation make it clear that estimated Holocene climate change is not compatible with evaporative drawdown of Moosehead Lake.

Stage–discharge curves for the natural outlets would be necessary to determine how far lake level could have risen above the outlet sill. Since the outlets are now dammed, this information is lost. However, the dams cross wide, shallowly incised channels with gently sloping banks, which must have had steep stage-discharge curves. The existence of a second outlet slightly above the primary one further increases the available discharge capacity. It is unlikely that lake level could rise above the outlet for more than a brief period during the spring freshet.

Thus, we assume that lake-level fluctuations at Moosehead Lake during the Holocene did not exceed the depth of the outlet channel, i.e., approximately 1–2 m. We interpret past lake-level change exceeding this amount as evidence of glacioisostasy rather than climate.

Field and Laboratory Methods

This study summarizes seismic profiling system (SPS) and side-scan sonar (SSS) surveys, sediment cores, and reconnaissance-level mapping. For geophysical surveys, we used a Raytheon RTT1000A 3.5-kHz SPS, an ORE Geopulse ''boomer'' SPS filtered between 0.5 and 2 kHz, and an EG&G SMS 260 digitally correcting SSS system. We collected 205 km of SPS trackline, including 145 km with simultaneous SSS coverage. Our SSS images cover 33.2 km^2, 11% of the lake area. We used postprocessed differential GPS for navigation.

We collected sediment cores using a modified Wright–Livingston piston coring system (Wright, 1967). Unless otherwise stated, all core depth measurements in this paper are normalized to PLL (313.6 m). We described cores in the field and again after cleaning in the lab, X-rayed some segments to evaluate sedimentary structures, and separated plant macrofossils from some segments by disaggregation in KOH and washing through a 425-μm sieve (Wasylikowa, 1986; Birks, 1980). We identified fossils by comparison to reference works and the plant macrofossil collection at the University of Maine. We obtained AMS radiocarbon dates on individually selected terrestrial plant remains and conventional radiocarbon dates on samples of bulk sediment (Table 1).

EVIDENCE FOR LAKE-LEVEL CHANGE

The Abandoned Outlet at Seboomook

A low pass into the Penobscot River drainage at the north end of the lake near Seboomook contains an abandoned outlet channel at approximately 4.5 ± 1.5 m PLL, which was active during a past relative lake-level highstand (Fig. 4). A broad, flat area on the Moosehead side of the pass is underlain by stiff, weakly laminated gray sand and silt. This sediment resembles nearshore lacustrine sediment from a nearby core (core SSD, Fig. 3), suggesting that it was deposited in 1–5 m of water when relative lake level was high here and this outlet was active. This unit is overlain by 20–30 cm of massive fine gray sand, which is similar to present shoreface sediment. We interpret it as a regressive deposit formed during relative lake-level fall. On the Penobscot side of the pass, the channel is incised into thin till, exposing boulders and scoured bedrock. Former pools in the channel are filled by marsh peat overlying gyttja, indicating that they were isolated by channel abandonment due to relative lake-

TABLE 1
Radiocarbon Sample Information

Sample No.	Laboratory No.	Analysis type	Depth, cm (PLL) Top	Depth, cm (PLL) Bottom	Stratigraphic significance	Material dated	Paleo-water depth	$\delta^{13}C$ (‰)	Age (^{14}C yr B.P.)	Analytical error (yr)
GVA-1	OS-8337	AMS	1541	1552	Shallow-water sediment	Mixed terrestrial plant frags.	0–2 m	−28.88	9840	70
GVA-2	OS-8388	AMS	1526	1536	Basal gyttja	Mixed terrestrial plant frags.	2–4 m	−26.36	9730	55
GVB-1	OS-8339	AMS	1636	1641	Beach deposit	Leaves, twigs of tundra shrubs, e.g., *Dryas integrifolia, Salix herbacea, Ericaceae* sp.	± 0.5 m	−29.54	10,050	45
GVB-2	OS-8340	AMS	1581	1586	Basal gyttja	Mixed terrestrial plant frags.	2–4 m	−29.13	9640	80
SBO-1	OS-6794	AMS	373.5[a]	375[a]	Basal organics from outlet channel	Conifer needles, leaves of *Ericaceae* sp.	N/A	−28.95	8370	40
SQD-1	OS-8341	AMS	945	950	Basal marsh sediment	Mixed terrestrial plant frags.	0–1 m	−27.89	9950	45
SQD-2	OS-8342	AMS	820	825	Marsh sediment	Mixed terrestrial plant frags.	0–2 m	−28.29	6880	50
SQD-3	OS-8343	AMS	695	700	Marsh sediment	Mixed terrestrial plant frags.	0–2 m	−27.51	2190	35
SQH-1	GX-22120	bulk	606	616	Basal marsh sediment	Bulk sediment	0–1 m	−25.9	9400	130
SQI-1	GX-22121	bulk	702	712	Basal marsh sediment	Bulk sediment	0–1 m	−27.3	9470	130
SQJ-1	GX-22122	bulk	681	691	Basal marsh sediment	Bulk sediment	0–1 m	−23.1	9280	130
SQK-1	GX-22123	bulk	1135	1148	Basal marsh sediment	Bulk sediment	0–1 m	−27	9830	140
SQK-2	GX-22124	bulk	928	938	Marsh sediment	Bulk sediment	0.4–2 m	−27.2	8135	120
SQL-1	GX-22125	bulk	1319	1335	Basal gyttja	Bulk sediment	0–3 m	−27.4	10,475	275
SQM-1	GX-22126	bulk	949	959	Basal marsh sediment	Bulk sediment	0–1 m	−25.8	8940	240

[a] Measured from sediment surface, not present lake level.

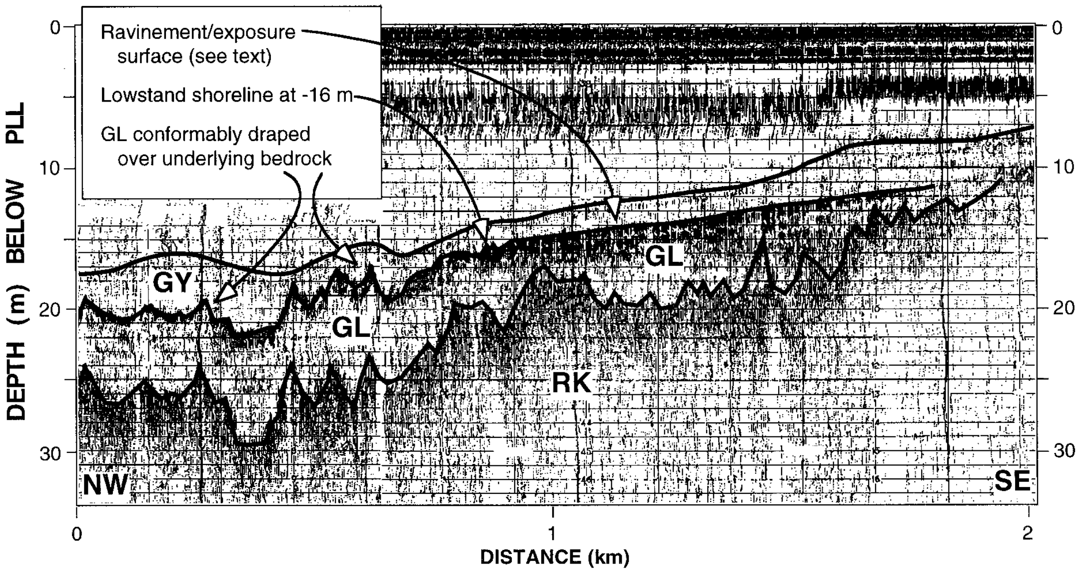

FIG. 5. Interpreted 3.5-kHz SPS record showing transgressive stratigraphy at southeastern end of Moosehead Lake. Glaciolacustrine sediment (GL) is conformably draped over underlying bedrock (RK) below 17 m water depth, but is truncated by a ravinement surface with enhanced reflectivity above this depth. The ravinement surface is covered by gyttja (GY), which records later deep-water deposition. Depth scale calculated using a seismic velocity of 1500 m/s.

level fall, colonized by marsh vegetation, and eventually filled with organic sediment. Terrestrial plant remains from the basal 1.5 cm of one of these pools (Fig. 4) have an AMS age of 8370 ± 40 ^{14}C yr B.P. (SBO-1), which provides a lower limiting age for outlet abandonment.

Stratigraphic Evidence for Relative Lake-Level Lowstands

SPS records from the southeast end of the lake show a transgressive sequence that records relative lake-level rise from a past lowstand. Well-laminated glaciolacustrine sediments, presumably deposited in a proglacial lake which was near or above OLL, are draped conformably over underlying topography in the deep basins, but are truncated in shallower areas (Fig. 5) by a strong flat reflector which represents a subaerial exposure surface, a transgressive ravinement surface, or both. The present maximum depth of significant wave erosion in this part of the lake does not exceed 2–3 m (Balco, 1997). Thus, local relative lake level must have been at least 10 m below OLL to form this erosion surface. The erosion surface is overlain by deep-water gyttja, which records later transgression during relative lake-level rise. We observed this stratigraphy at several places at the southeast end of the lake (Fig. 3).

Two cores from the southernmost basin in the lake (cores GVA, GVB, Fig. 3) penetrated lithologic discontinuities at the approximate depth of this erosion surface. In core GVB, glaciolacustrine mud is unconformably overlain by 65 cm of cross-bedded medium to coarse sand. The sand contains subhorizontal layers of coarse terrestrial plant debris, including abundant leaves, twigs, seeds of *Dryas integrifolia, Salix herbacea,* and *Ericaceae,* and needles of *Picea* sp. This unit is analogous to modern beaches in sheltered parts of the lake, where plant debris accumulates in the swash zone and is periodically buried by sand, and the species assemblage reflects the shrub–tundra vegetation that predominated in Maine during deglaciation (Davis and Jacobson, 1985). This unit is overlain by sandy gyttja and then by several meters of massive gyttja. This lithologic sequence records transgression following a relative lake-level lowstand, which agrees with the seismic stratigraphy. An AMS age for the plant debris of 10,050 ± 45 ^{14}C yr B.P. (GVB-1) dates a time when relative lake level stood near −12 m OLL.

SPS records throughout the lake show truncated reflectors in shallow areas, as well as terraces incised into sediments (Fig. 6). These erosional features are now preserved at −5 to −20 m PLL (−1 to −16 m OLL), and we interpret them as evidence of past lake-level lowstands. In addition, we recognize submerged shorelines on SSS records from exposed areas of the lake in water depths up to 16 m PLL (12 m OLL). Like the modem shoreline, these features comprise concentric, shore-parallel zones of wave-washed boulders, eroding glaciolacustrine sediment, and acoustically transparent gyttja. Often, shore-parallel boulder berms are present within the bouldery zone. These are identical to modern berms at Moosehead Lake and other seasonally frozen lakes and estuaries, which are formed as ice bulldozes rocky shorelines in winter (Dionne and Laverdiere, 1972; Hansom, 1983). By analogy with the present shoreline, we infer that

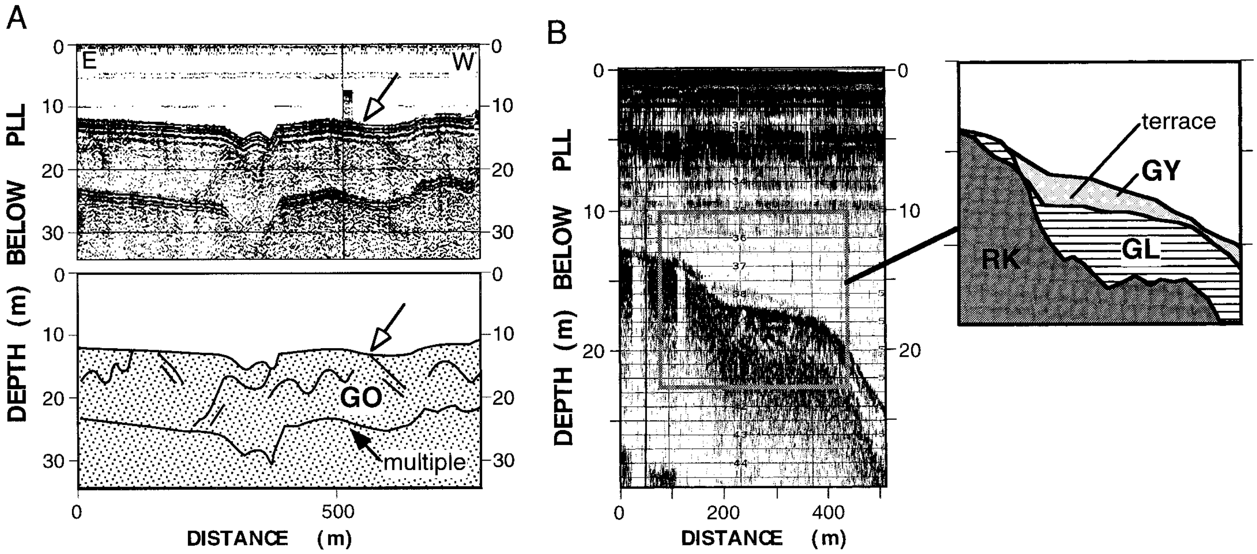

FIG. 6. Examples of erosional lowstand features observed on SPS records. (A) ORE Geopulse record and interpretation showing glacial outwash (GO) containing internal reflectors that are truncated by an erosion surface. (B) 3.5-kHz record and interpretation showing an erosional terrace developed in glaciolacustrine sediment (GL) overlying bedrock (RK) and subsequently covered by gyttja (GY). Enhancement of reflector at top of GL above 21 m depth may indicate reworking or exposure. Depth scale calculated using a seismic velocity of 1500 m/s.

relative lake level must have dropped to 1 m above the bottom of the bouldery zone to create these features.

Several cores in a protected inlet at Squaw Bay (Fig. 3) penetrated organic-rich, shallow-water marsh sediments, which we interpret as lowstand deposits, lying at −5 to −12 m PLL (−1 to −8 m OLL). The plant macrofossil assemblage in these sediments indicates that deposition mostly took place in very shallow water. In core SQD, for example, the basal 20 cm contains *Characeae* oospores and very abundant (>30/200 cm^3) *Ranunculus trichophyllus* seeds (Fig. 7). *Characeae* have a wide depth range, but *R. trichophyllus* is an emergent aquatic species which typically grows less than 50 cm high (Fassett, 1957). The presence of its seeds in quantity suggests that this sediment was deposited very close to water level.

Between 900 and 930 cm in this core, the submerged aquatic plants *Characeae, Najas flexilis,* and *Potamogeton* spp. and the emergent species *Hippuris vulgaris, Typha* sp., and *Cyperaceae* predominate. We interpret this as a period of deeper-water sedimentation. Above 900 cm, however, submerged species are nearly absent, and emergent species along with wet shore plants such as *Myrica gale* and *Triadenum virginicum* are dominant (Fig. 7). This species assemblage is characteristic of the shoreline fringe of lakes (Fassett, 1957; Birks, 1973; Harrison and Digerfeldt, 1993). In addition, the seeds of many of these plants are buoyant and usually accumulate at the water's edge (Birks, 1980). Thus, with the exception of a period of deeper water shortly after the onset of deposition, the bulk of the sediments at Squaw Bay accumulated in very shallow water near the shore as sedimentation kept up with a slow rise in relative lake level.

We used radiocarbon dates of shallow-water sediment from various elevations to construct a lake-level curve (Table 1, Figs. 8, 9). This is most accurately done using basal marsh sediment overlying an incompressible substrate, a technique used in numerous studies of tidal marshes (Gehrels *et al.*, 1996). We obtained seven radiocarbon dates of basal sediment. Three other samples may have been displaced by compaction. All samples from Squaw Bay except SQL-1 contain abundant macrofossils of emergent and wet shore plants, and we infer that they were deposited in water <1 m deep. Sample SQL-1 dates the onset of gyttja deposition above a lowstand unconformity. By analogy with present conditions, this probably occurred when the water depth reached 2–3 m. The inferred vertical errors given in Table 1 incorporate these factors.

Bulk sediment sample SQM-1 and AMS sample SQD-1 were taken at the same location and elevation, but differ in age by 1000 yr. Since bulk sediment ages are considered to be less accurate than AMS ages on picked terrestrial macrofossils (e.g., Aravena *et al.*, 1992), and SQM-1 has the highest analytical error of the dates in this study, we believe that it is necessary to disregard SQM-1. Two more cores from the southeastern end of the lake limit the depth of any past lowstand. Cores HPA and HPB contain a distinctive band of minerogenic gyttja that is found in lakes throughout Maine and Atlantic Canada. It has been described and well dated in many studies and reflects lacustrine sedimentation during the Younger Dryas climate oscillation which occurred between ca. 11,000 and 10,000 ^{14}C yr B.P. (Thompson *et al.*, 1996; Mayle *et al.*, 1993; Levesque *et al.*, 1993). Thus, local

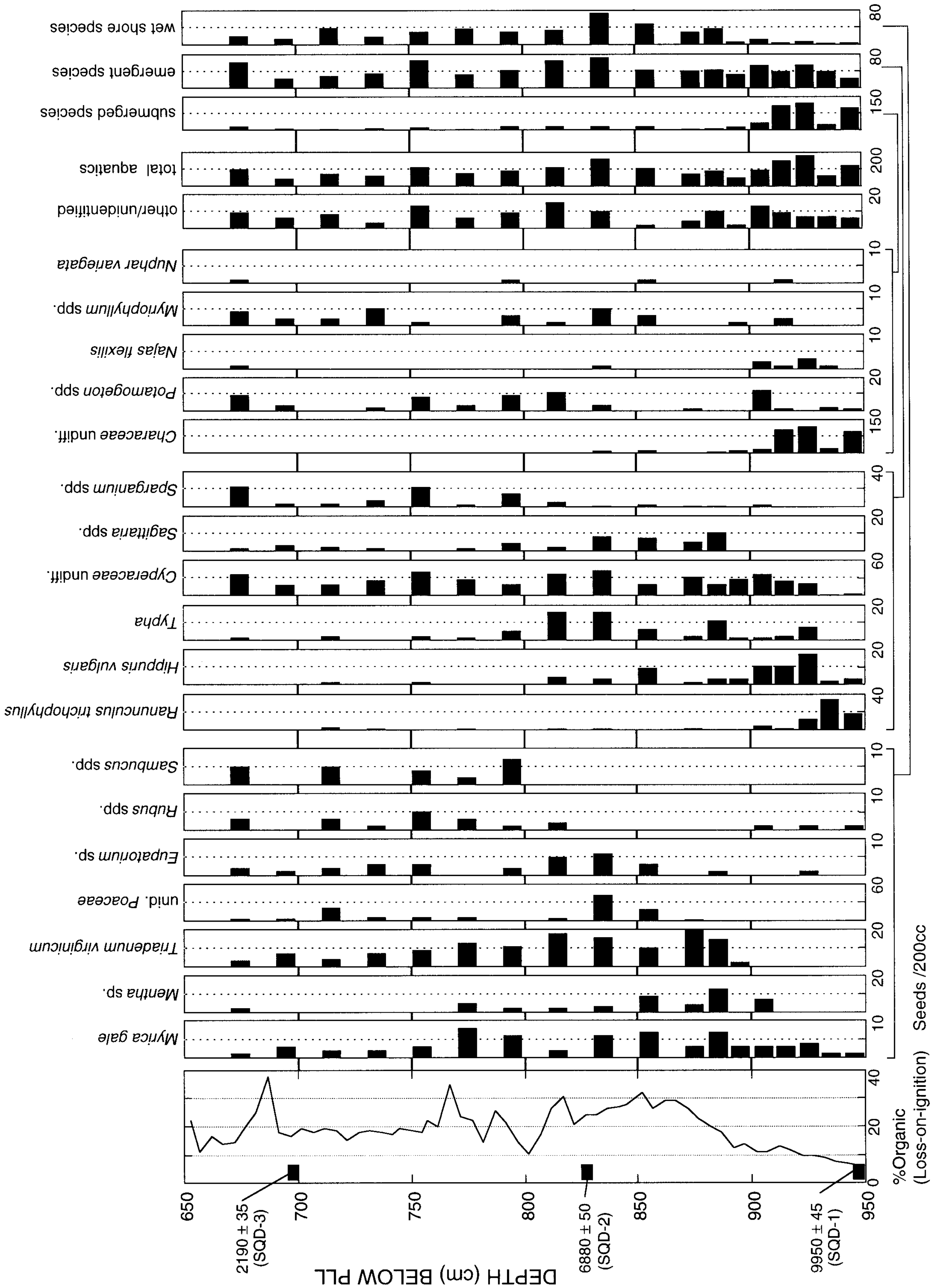

FIG. 7. Aquatic plant macrofossil abundance, percentage organic content by loss-on-ignition, and radiocarbon dates for core SQD. Identifiable fossils of trees and other upland species are not shown. Grouping of species into submerged, emergent, and wet-shore categories is indicated by brackets at bottom of figure.

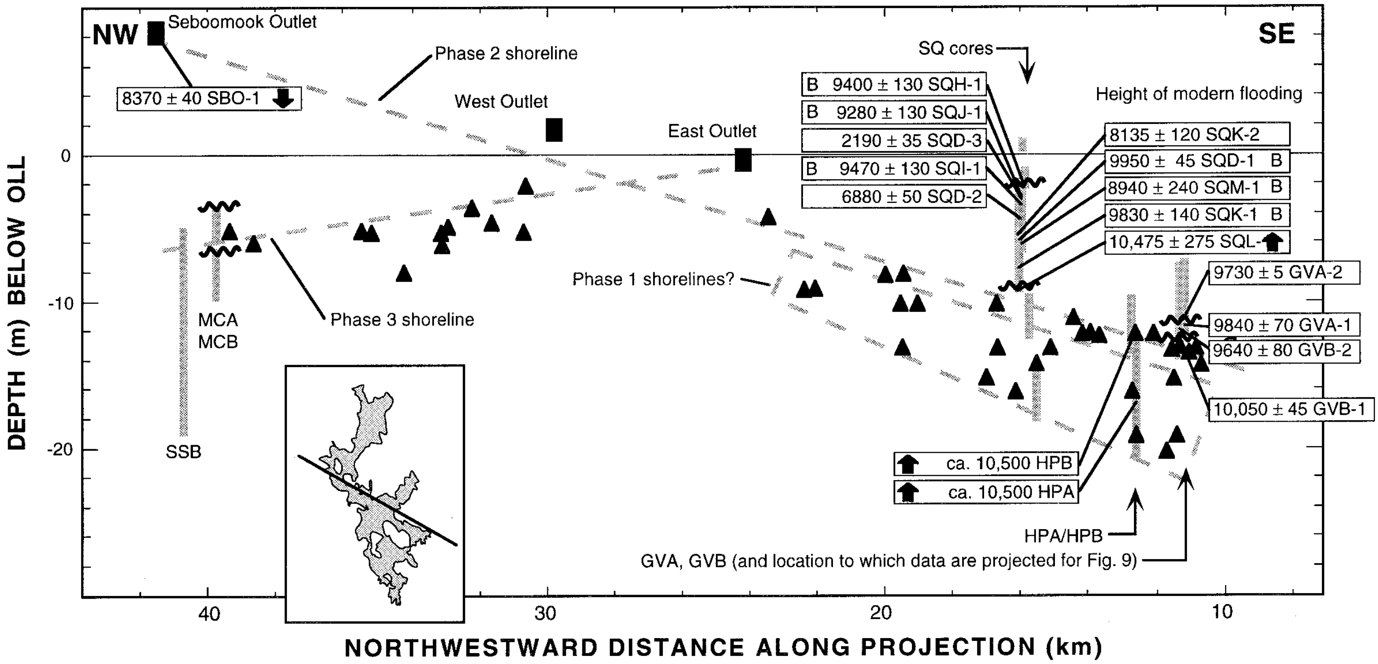

FIG. 8. Evidence for lake-level change projected onto a line with azimuth 120°. (Inset) Line of projection. Triangles are erosional lowstand features observed on SSS and SPS records. Gray bars show core locations. Dark wavy lines are inferred erosional unconformities in cores. Radiocarbon dates are placed at depth of sample, without consideration of sediment compaction or water-depth uncertainty. Dates with upward pointing arrows are lower limits on water level. Date with a downward pointing arrow is an upper limit on water level. ''B'' indicates a date for basal marsh sediment overlying an incompressible substrate. All depths are normalized to OLL (see text).

water level must have been above the depth of this unit during Younger Dryas time.

Correlation of Evidence of Lake-Level Change

Unconformities in cores, erosional features observed on SPS and SSS records, and active and abandoned outlets fall into two groups which we interpret as former water planes (Fig. 8). The majority of the features in the southern part of the lake fall near a southeastward-dipping plane that runs from −12 m OLL (−16 m PLL) in the extreme southeastern end of the lake to −5 m OLL (−9 m PLL) in the middle of the lake. This plane intersects the abandoned outlet channel at Seboomook, indicating that these features (Shoreline Phase 2) were formed synchronously at a time when lake level was controlled by the Seboomook outlet and the lake basin was tilted down to the northwest. In addition, a group of scattered features (Phase 1) lies between this plane and a plane which lies at −20 m OLL at the southeastern end of the lake and also appears to intersect the Seboomook outlet. Erosional features at the northern end of the lake (Phase 3) lie along a northwestward-dipping plane that approximately intersects East Outlet. These features were probably formed when the lake basin was tilted down to the southeast, and East Outlet controlled lake level. Trend surface analysis of these groups of features (Balco, 1997) indicates that the Phase 2 and Phase 3 water planes have dip directions of approximately 300° and 120°, respectively.

If Moosehead Lake had been drawn down below its outlet during a dry period, we would expect to see evidence for a shoreline lying below all of the outlets. We do not observe this, which supports our earlier assumption that lake-level changes at Moosehead are unrelated to climate.

Lake-Level Chronology

Radiocarbon dates from cores GVB and SQL date the Phase 1 shoreline to approximately 10,000 ^{14}C yr B.P. At this time, the lake basin was tilted down to the northwest at 0.7 m/km, relative lake level at Greenville stood near −13 m OLL, and the lake drained through the Seboomook outlet into the Penobscot River (Figs. 8 and 9).

Phase 1 erosional features suggest that relative lake level at Greenville was as low as −20 m OLL at some time. The radiocarbon dates younger than 10,000 ^{14}C yr B.P. from above Phase 2 in the south end of the lake (Figs. 8 and 9) indicate that this could have happened only before 10,000 ^{14}C yr B.P. However, the Younger Dryas marker horizon in core HPA constrains lake level at Greenville to above −13 m OLL at ca. 11,000 ^{14}C yr B.P., and probably earlier, since there is no unconformity in the core. Also, the seismic stratigraphy of some parts of the southeastern end of the lake

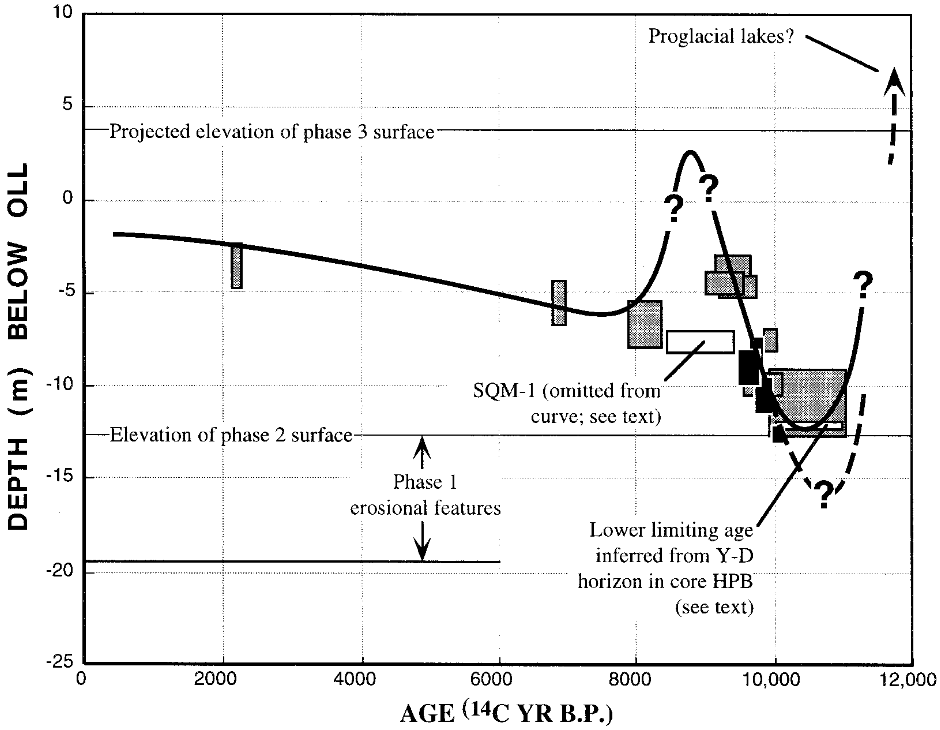

FIG. 9. Proposed relative lake-level curve for the south end of Moosehead Lake. Depths of radiocarbon samples have been corrected for tilting of the lake basin and normalized to core site GVB (Fig. 3) assuming that the Seboomook outlet was active before 9000 ^{14}C yr B.P. and East Outlet was in use thereafter. Width of boxes represents 2σ analytical error for radiocarbon dates. Height of boxes represents inferred water-level range reflecting possible displacement by compaction and uncertainty in depositional environment (Table 1). Black boxes are radiocarbon dates from cores GVA and GVB. Gray boxes are radiocarbon dates from Squaw Bay cores.

(Fig. 5) does not indicate any erosion below Phase 2. It is difficult to resolve this question without any age control on Phase 1 features. Since comparison of RSL curves from Maine and Quebec (Fig. 1) indicates that average regional landscape tilt increased between 12,000 and 10,500 ^{14}C yr B.P., it is most likely that relative lake level at Greenville dropped during this time as the northwestward tilt of the lake basin increased, reaching a lowstand at 10,000 ^{14}C yr B.P. (Phase 2) before beginning to rise. However, this interpretation does not explain Phase 1 erosional features. In the absence of further evidence, the lake-level history before 10,000 ^{14}C yr B.P. is uncertain.

Radiocarbon dates at Squaw Bay and Greenville show that relative lake level in the southeastern end of the lake rose rapidly between 10,000 and 9400 ^{14}C yr B.P. (Fig. 9). The Seboomook outlet was abandoned in favor of East Outlet when relative lake level reached -7 m OLL at Greenville at approximately 9750 ^{14}C yr B.P., which is consistent with the limiting age of 8370 ^{14}C yr B.P. from core SBO. Thus, the northwestward tilt of the basin decreased rapidly during this period.

The plant macrofossil record from Squaw Bay (Fig. 7) indicates deepening after the onset of lacustrine conditions at 9950 $\pm$ 45 ^{14}C yr B.P., followed by a return to shallow-water sedimentation by 8135 $\pm$ 120 ^{14}C yr B.P. We correlate the Phase 3 shoreline in the northwestern end of the lake with this period of deep-water sedimentation at the southeastern end. Thus, Phase 3 features were most probably formed at approximately 8750 ^{14}C yr B.P., indicating that the lake basin was tilted to the southeast at this time.

Other radiocarbon dates indicate that relative lake level at the southeastern end of the lake dropped again to -6 m OLL at 8135 $\pm$ 120 ^{14}C yr B.P., and then gradually rose to its modern predam elevation (Fig. 9). Again, we interpret this as evidence that the lake basin was tilted slightly to the northwest at this time. However, samples SQD-2 and SQD-3 (Table 1) indicate that relative lake level at Squaw Bay was near -4 m OLL at 6880 ^{14}C yr B.P. and -3 m OLL at 2190 ^{14}C yr B.P., suggesting that the lake basin was slightly tilted well into the late Holocene. This interpretation conflicts with RSL curves from coastal Maine and Quebec (Fig. 1), which converge after 6000 ^{14}C yr B.P. After this time, regional landscape tilt was less than 0.05 m/km, which can only account for a relative lake level 0.4 m below present at Squaw Bay. Thus, these samples may have been displaced by compaction or deposited below water level, the OLL datum may not accurately represent predam lake level, or seasonal water-level changes may have affected the distribution of sediment. Further study of the relationship between nearshore lacustrine sediments and lake level would significantly improve our interpretation of these data.

DISCUSSION AND CONCLUSIONS

The radiocarbon-dated Phase 2 shoreline indicates that the Moosehead basin was tilted to the northwest at 0.7 m/km at 10,000 ^{14}C yr B.P. Additional radiocarbon dates for transgressive deposits at Greenville and Squaw Bay show that this tilt decreased rapidly between 10,000 and 9400 ^{14}C yr B.P. Based on correlation of the Phase 3 shoreline at the northwestern end of the lake with evidence of deep-water sedimentation at Squaw Bay, we believe that the basin was tilted to the southeast, at a maximum of 0.3 m/km, between 9000 and 8500 ^{14}C yr B.P. By 8200 ^{14}C yr B.P., it was tilted slightly to the northwest again, and this tilt gradually decreased until the present time. This history is consistent with the movement of a steep marginal forebulge across Maine. Our relative lake-level curve (Fig. 9) for the southern end of Moosehead Lake resembles that proposed by Astley (1997) for Lake Champlain south of the controlling outlet. Her data show a rapid rise immediately after deglaciation, a plateau or slight fall at approximately 7500 ^{14}C yr B.P., and then a gradual rise until the present. Although Astley did not interpret this as unequivocal evidence for a migrating forebulge, the fact that at least two large lakes in New England show a time-transgressive pattern of similar lake-level changes supports a glacioisostatic rather than climatic origin for these events.

Our evidence is weak or contradictory during several time periods. First, the lake-level history between deglaciation and 10,000 ^{14}C yr B.P. is uncertain. Age control on Phase 1 erosional features is required to resolve this problem. Second, our evidence pertaining to the lake-level history since 8000 ^{14}C yr B.P. is very sparse. Third, since we did not obtain datable lowstand deposits associated with Phase 3, the timing and magnitude of southeastward tilting of the basin are uncertain. Since the existence, duration, and magnitude of this event are critical to reconstructing the shape of the marginal bulge, our highest priority in future studies at Moosehead Lake is to obtain and date unequivocal lowstand deposits from the north end of the lake.

The direction of tilt of the Phase 2 shoreline is consistent with other evidence throughout the region (Fig. 1). However, the evidence concerning the magnitude of tilt is contradictory. Champlain Sea shorelines (Parent and Occhietti, 1988) and Lake Hitchcock glaciolacustrine deltas (Koteff *et al.*, 1993) are older, closer to the center of the ice sheet, and lie on a plane which is tilted more steeply than the Phase 2 shoreline at Moosehead. In contrast, glaciomarine deltas in coastal Maine (Thompson *et al.*, 1989) are older, but lie on a less steeply tilted plane. Also, comparison of RSL curves from Maine and Quebec indicates that the landscape tilt integrated across the entire region increased from 0.19 m/km at 13,000 ^{14}C yr B.P. to 0.35 m/km at 11,000 ^{14}C yr B.P. and then decreased to 0.16 m/km by 10,000 ^{14}C yr B.P. The fact that the local magnitude of tilt at Moosehead exceeds the regional tilt shows that the crust was bent on a wavelength significantly shorter than the 300-km distance across Maine. In addition, these discrepancies suggest that measurements of crustal deformation made on landforms such as deltas, which were probably deposited time-transgressively and graded to a changing water level, may not accurately represent the tilt of the late-glacial landscape. Radiocarbon dates of nearshore lake sediment, when the controlling outlet of the lake is known, are an instantaneous measurement of tilt. Performing similar studies at other lakes in the region would help to resolve the shape of crustal deformation. Barnhardt *et al.* (1995) report that the crest of the marginal forebulge passed the Maine coast at 10,500 ^{14}C yr B.P. and Quebec City at ca. 6500 ^{14}C yr B.P. This event can be recognized in a lake-level record as a time when the lake basin was momentarily not tilted, and relative lake-level was similar to that at present. In our reconstruction, it took place 9000 ^{14}C yr B.P. The isobase-normal distance from Moosehead to Quebec is 200 km, and that from Moosehead to the Maine coast is 180 km. Thus, the forebulge crest moved at 0.11–0.12 km/yr from the Maine coast to Moosehead and at 0.07–0.10 km/yr from Moosehead to Quebec. This agrees with the observed deceleration from 0.11 to 0.045 km/yr for forebulge migration across Newfoundland (Liverman, 1994). Since lake-level changes respond to tilting rather than vertical motions, it is not possible to estimate the forebulge height from this information.

In summary, our interpretation of the lake-level record at Moosehead Lake as a record of a migrating forebulge is consistent with existing RSL data. This confirms that studies of lakes can be useful for reconstructing crustal deformation in inland areas. Also, our results support the conclusions of Barnhardt *et al.* (1995) that a steep marginal forebulge, which is not accurately reproduced by existing geophysical models, traveled rapidly across Maine. We anticipate that future efforts to develop lake-level reconstructions with accuracy similar to existing RSL curves will greatly improve our understanding of this feature. As noted by Barnhardt *et al.* (1995) and Tushingham and Peltier (1991), this mismatch is probably the result of local lithospheric variability and a complicated history of ice retreat that cannot be duplicated by a coarse-resolution model.

ACKNOWLEDGMENTS

This study was funded by an NSF Graduate Research Fellowship awarded to Greg Balco. Geochron Laboratories provided bulk radiocarbon analyses through their Research Awards Program. The National Ocean Sciences AMS Laboratory at Woods Hole, Massachusetts, performed AMS radiocarbon analyses under NSF Cooperative Agreement OCE-801015. We greatly appreciate the assistance of Chris Dorion, Heather Almquist-Jacobson, George Jacobson, Ann Dieffenbacher-Krall, Andy McIntosh, Greg Miller, Doug Reusch, Tom Henze, and Mike Retelle. Reviewers J. C. Dionne and S. Harrison made numerous helpful corrections to the first draft of this paper. Captain Roger Lane of Jolly Roger's Moosehead Tours, Rockwood, Maine, graciously sacrificed one of the M/V Socatean's propellers to the cause of science.

REFERENCES

Aravena, R., Warner, B. G., MacDonald, G. M., and Hanf, K. I. (1992). Carbon isotope composition of lake sediments in relation to lake productivity and radiocarbon dating. *Quaternary Research* **37,** 333–345.

Astley, B. N. (1997). ''Holocene Lake-level Change in Lake Champlain.'' Unpublished MS thesis, Univ. of Vermont, Burlington, VT.

Balco, G. (1997). ''Postglacial Land and Lake Levels, Moosehead Lake, Maine.'' Unpublished MS thesis, Univ. of Maine, Orono, ME.

Barnhardt, W. A., Gehrels, W. R., Belknap, D. F., and Kelley, J. T. (1995). Late Quaternary relative sea level change in the western Gulf of Maine: Evidence for a migrating glacial forebulge. *Geology* **23,** 317–320.

Benson, L., and Thompson, R. S. (1987). The physical record of lakes in the Great Basin. *In* ''The Geology of North America: North America and Adjacent Oceans During the Last Deglaciation'' (W. F. Ruddiman and H. E. Wright, Jr., Eds.), Vol. K-3, pp. 241–260. Geological Society of America, Boulder, CO.

Bierman, P., Lini, A., Zehfuss, P., Church, A., Davis, P. T., Southon, J., and Baldwin, L. (1997). Postglacial ponds and alluvial fans: Recorders of Holocene landscape history. *GSA Today* **7** (October), 1–8.

Birks, H. H. (1973). Modern macrofossil assemblages in lake sediments in Minnesota. *In* ''Quaternary Plant Ecology: The 14th Symposium of the British Ecological Society'' (H. J. B. Birks and R. G. West, Eds.), pp. 173–189. Wiley, New York.

Birks, H. H. (1980). Plant macrofossils in Quaternary lake sediments. *Archiv fur Hydrobiologie Beihefte Ergebnisse der Limnologie* **15,** 1–60.

Calvert, M. (1983). ''Dawn Over the Kennebec.'' The Monmouth Press, Monmouth, Maine.

Clark, J. A., Hendricks, M., Timmermans, T. J., Struck, C., and Hilverda, K. J. (1994). Glacial isostatic deformation of the Great Lakes region. *Geological Society of America Bulletin* **106,** 19–31.

COHMAP members (1988). Climatic Changes of the last 18,000 years: Observations and model simulations. *Science* **241,** 1043–1052.

Davis, R. B., and Jacobson, G. L. (1985). Late Glacial and Early Holocene landscapes in northern New England and adjacent areas of Canada. *Quaternary Research* **23,** 341–368.

Dionne, J. C. (1988). Holocene relative sea-level fluctuations in the St. Lawrence estuary, Quebec. *Quaternary Research* **29,** 233–244.

Dionne, J. C., and Coll, D. (1995). Le niveau marin relatif dans la région de Matane (Québec), de la deglaciation à nos jours. *Géographie physique et Quaternaire* **49,** 363–380.

Dionne, J.-C., and Laverdiere, C. (1972). Ice formed beach features from Lake St. Jean, Quebec. *Canadian Journal of Earth Sciences* **9,** 979–990.

Dorion, C. C. (1995). Late Glacial chronology and paleoenvironment along a south to north transect in Maine and adjacent Canada. *In* ''Abstracts with Programs, Third Annual Research Symposium.'' Institute for Quaternary Studies, University of Maine, Orono, ME.

Fassett, N. C. (1957). ''A Manual of Aquatic Plants.'' Univ. of Wisconsin Press, Madison, WI.

Gehrels, W. R., Belknap, D. F., and Kelley, J. T. (1996). Integrated high-precision analyses of Holocene relative sea-level changes: Lessons from the coast of Maine. *Geological Society of America Bulletin* **108,** 1073–1088.

Goldthwait, J. W. (1908). A reconstruction of water planes of the extinct glacial lakes in the Lake Michigan basin. *Journal of Geology* **14,** 459–476.

Hansom, J. D. (1983). Ice-formed intertidal boulder pavements in the sub-antarctic. *Journal of Sedimentary Petrology* **53,** 135–145.

Harrison, S. P., and Digerfeldt, G. (1993). European lakes as paleohydrological and paleoclimatological indicators. *Quaternary Science Reviews* **12,** 233–248.

Koteff, C., Robinson, G. R., Goldsmith, R., and Thompson, W. (1993). Delayed postglacial uplift and synglacial sea levels in coastal New England. *Quaternary Research* **40,** 46–54.

Langbein, W. B., and others (1949). ''Annual Runoff in the United States.'' U.S. Geological Survey Circular 52.

Levesque, A. J., Mayle, F. E., Walker, I. R., and Cwynar, L. C. (1993). A previously unrecognized late-glacial cold event in eastern North America. *Nature* **361,** 623–626.

Lewis, E. B. (1997). ''Stratigraphy and Geomorphology of the Glacial Outwash and Modern Songo River Deltas, Sebago Lake State Park, Maine.'' Unpublished MS thesis, University of Maine, Orono, ME.

Liverman, D. G. E. (1994). Relative sea-level history and isostatic rebound in Newfoundland, Canada. *Boreas* **23,** 217–230.

Lowell, T. V., and Crossen, K. (1983). The Seboomook Lake area became ice-free, but how? *In* ''Guide to Field Trips In North-Central Maine: New England Intercollegiate Geologic Conference 75th Annual Meeting'' (D. Caldwell and L. Hanson, Eds.), pp. 147–174.

Mayle, F. E., Levesque, A. J., and Cwynar, L. C. (1993). Accelerator-mass-spectrometer ages for the Younger Dryas event in Atlantic Canada. *Quaternary Research* **39,** 355–360.

McCabe, A. M. (1997). Geological constraints on geophysical models of relative sea-level change during deglaciation of the western Irish Sea. *Journal of the Geological Society of London* **154,** 601–604.

National Climatic Data Center (1994). ''Climatological Data, 1994 Annual Summary: New England.'' National Oceanic and Atmospheric Administration, Environmental Data and Information Service, Asheville, NC.

Parent, M., and Occhietti, S. (1988). Late Wisconsinan deglaciation and Champlain Sea invasion in the St. Lawrence Valley, Quebec. *Géographie physique et Quaternaire* **42,** 215–246.

Peixoto, J. P., and Oort, A. H. (1992). ''Physics of Climate.'' American Institute of Physics, New York.

Peltier, W. R. (1994). Ice age paleotopography. *Science* **265,** 195–201.

Petersen, J. B., Robinson, B. S., Belknap, D. F., Stark, J., and Kaplan, L. K. (1994). An Archaic and Woodland Period fish weir complex in central Maine. *Archaeology of Eastern North America* **22,** 197–222.

Quinlan, G., and Beaumont, C. (1981). The deglaciation of Atlantic Canada as reconstructed from the postglacial relative sea level record. *Canadian Journal of Earth Science* **18,** 2232–2246.

Quinlan, G., and Beaumont, C. (1982). A comparison of observed and theoretical postglacial relative sea level in Atlantic Canada. *Canadian Journal of Earth Science* **19,** 1146–1163.

Street-Perrott, F. A., and Harrison, S. P. (1985). Lake levels and climate reconstruction. *In* ''Paleoclimate Analysis and Modeling'' (A. D. Hecht, Ed.), pp. 147–174. Wiley, New York.

Talbot, J. (1996). A reconstruction of an early to middle Holocene climatic optimum and the paleoenvironmental landscapes of Lake Auburn, Auburn, ME. *Geological Society of America Abstracts with Programs* **29,** 103.

Thompson, W. B., Crossen, K. J., Borns, H. W., and Andersen, B. G. (1989). Glaciomarine deltas of Maine and their relation to late Pleistocene–Holocene crustal movements. *In* ''Neotectonics of Maine'' (W. A. Anderson and H. W. Borns, Jr., Eds.), pp. 43–67. Maine Geological Survey, Augusta, ME.

Thompson, W. B., and Dorion, C. C. (1996). Deglaciation chronology and marine-limit refinements in southwestern Maine. *In* ''Abstracts with Programs, Fourth Annual Research Symposium,'' p. 203. Institute for Quaternary Studies, University of Maine, Orono, ME.

Thompson, W. B., Fowler, B. K., Flanagan, S. M., and Dorion, C. C. (1996). Recession of the late Wisconsinan ice sheet from the northwestern White Mountains, New Hampshire. *In* ''Guide to Field Trips in Northern New Hampshire and Adjacent Regions of Maine and Vermont: New England Intercollegiate Geologic Conference 88th Annual Meeting'' (M. R. Van Baalen, Ed.), pp. 203–234.

Thompson, W. B., Hildreth, C. T., Johnston, R. A., and Retelle, M. J. (1995). Glacial Geology of the Portland-Sebago Lake Area. *In* ''Guide to Field Trips in Southwestern Maine: New England Intercollegiate Geologic Conference 87th Annual Meeting'' (A. M. Hussey II and R. A. Johnston, Eds.), pp. 51–70, Bowdoin College, Brunswick, ME.

Tushingham, A. M., and Peltier, W. R. (1991). Ice-3G: A new global model of late Pleistocene deglaciation based upon geophysical predictions of post-glacial relative sea level change. *Journal of Geophysical Research* **96,** 4497–4523.

Wasylikowa, K. (1986). Analysis of Fossil Fruits and Seeds. *In* ''Handbook of Holocene Paleoecology and Paleohydrology'' (B. E. Berglund, Ed.), pp. 571–594. Wiley, New York.

Webb, T., Bartlein, P. J., Harrison, S. P., and Anderson, K. H. (1993). Vegetation, lake levels, and climate in eastern North America for the past 18,000 years. *In* ''Global Climates Since the Last Glacial Maximum'' (H. E. Wright, Jr., and 5 others, Eds.), pp. 415–467. University of Minnesota Press, Minneapolis, MN.

Wright, H. E., Jr. (1967). A square-rod piston sampler for lake sediments. *Journal of Sedimentary Petrology* **37,** 975–976.