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Harold W. Borns Jr.
University of Maine

Lisa A. Doner
University of Colorado Boulder

Christopher C. Dorion
University of Maine

George L. Jacobson Jr.
University of Maine

Michael R. Kaplan
University of Wisconsin-Madison

See next page for additional authors

Authors
Harold W. Borns Jr., Lisa A. Doner, Christopher C. Dorion, George L. Jacobson Jr., Michael R. Kaplan, Karl J. Kreutz, Thomas V. Lowell, Woodrow B. Thompson, and Thomas K. Weddle

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Harold W. Borns, Jr. [1], Lisa A. Doner [3], Christopher C. Dorion [1], George L. Jacobson Jr. [1], Michael R. Kaplan [4], Karl J. Kreutz, [1], Thomas V. Lowell [5], Woodrow B. Thompson [2] and Thomas K. Weddle, [2]

[1] *Institute for Quaternary and Climate Studies, University of Maine, Orono, Maine 04469, U.S.A.*
[2] *Maine Geological Survey, Augusta, Maine 04333, U.S.A.*
[3] *Institute of Arctic and Alpine Research, University of Colorado, Boulder, Colorado 80309-0450, U.S.A.*
[4] *Department of Geology and Geophysics, University of Wisconsin, Madison, Wisconsin 53706, U.S.A.*
[5] *Department of Geology, University of Cincinnati, Cincinnati, Ohio 45221, U.S.A.*

Abstract

The glacial geology of Maine records the northward recession of the Late Wisconsinan Laurentide Ice Sheet, followed by development of a residual ice cap in the Maine-Québec border region due to marine transgression of the St. Lawrence Lowland in Canada. The pattern of deglaciation across southern Maine has been reconstructed from numerous end moraines, deltas and submarine fans deposited during marine transgression of the coastal lowland. Inland from the marine limit, a less-detailed sequence of deglaciation is recorded by striation patterns, meltwater channels, scattered moraines and waterlain deposits that constrain the trend of the ice margin. There is no evidence that the northern Maine ice cap extended as far south-west as the Boundary Mountains and New Hampshire border.

Newly-obtained radiocarbon ages from marine and terrestrial ice-proximal environments have improved the chronology of glacial recession in Maine. Many of these ages were obtained by coring late-glacial sediments beneath ponds and lakes. Data from this study show that the state was deglaciated between about 14.5 and 11.0 ka BP (^{14}C years). The coastal moraine belt in southern Maine was deposited by oscillatory ice-margin retreat during the cold pre-Bølling time. Rapid ice recession to northern Maine then occurred between 13 and 11 ka BP, during the warmer Bølling/Allerød chronozones. Radiocarbon-dated pond sediments in western and northern Maine show lithologic evidence of Younger Dryas climatic cooling and persistence of the northern ice cap into Younger Dryas time.

A large discrepancy still exists between radiocarbon ages of deglaciation in coastal south-western Maine and the timing of ice retreat indicated by New England varve records in areas to the west. Part of this problem may stem from the uncertainty of reservoir corrections applied to the radiocarbon ages of marine organics.

Introduction

The termination of the Wisconsinan glaciation has been revealed in many parts of the world as a time of highly variable climate and vast changes in ice masses, oceans and vegetation. Maine experienced a rare combination of important global phenomena, with the margin of the Laurentide Ice Sheet meeting the sea, with sea level changing dramatically as a result of both eustatic and isostatic processes and with vegetation invading newly-exposed landscapes. Davis & Jacobson (1985) mapped the changing environments of Maine and adjacent areas for each millennium during the period 14-9 ka BP (all ages in this paper are in ^{14}C years) and concluded that the changes were monotonic, with neither the shrinking ice mass south of the St. Lawrence River nor the vegetation inferred from pollen analysis showing signs of reversals.

Since 1985, dramatic new insights about high-frequency climate variability during the late Quaternary have been revealed by detailed stratigraphical studies of terrestrial and marine deposits. For example, annually resolved ice-core records from Greenland (Dansgaard *et al*., 1989, 1993) show millennial-scale oscillations up to and including the Younger Dryas. Thanks to the development of accelerator mass spectrometer (AMS) dating, detailed studies of marine sediments from the North Atlantic could link those same oscillations with millennial-scale variations in ice-rafted detritus and oxygen isotopes, revealing Heinrich events and Bond cycles (Heinrich, 1988; Bond *et al*., 1992).

The Younger Dryas cold event that preceded Holocene warming has received considerable attention as an example of a large and abrupt excursion in climate. Evidence that the event was contemporaneous to within a few decades in areas as remote from one another as southern Sweden (Björck *et al*., 1996) and New Zealand (Denton & Hendy, 1994) showed the phenomenon to be global. In the Maritimes region of eastern Canada, recognition of the event came after buried peat deposits in New Brunswick and Nova Scotia were determined to be roughly contemporaneous with the Younger Dryas in north-west Europe (Mott *et al*., 1986). Subsequently, lacustrine evidence of the Younger Dryas was documented by Mayle & Cwynar (1995), Doner (1996), Stea & Mott (1998) and Dorion (1997a, b).

The recognition of late-glacial climate variability in Maine and the Maritimes stimulated the authors to collect additional field data about the deglaciation of Maine and the extent and behavior of the ice mass that remained in this region after marine transgression of the St. Lawrence lowland to the north. With a grant from the National Science Foundation, the University of Maine and Maine Geological Survey collaborated to address these goals. In this paper an overview is presented of the authors' research on late-glacial environments in Maine using AMS dating and stratigraphical analysis. To test the ice-retreat model of Davis & Jacobson (1985; also in Thompson & Borns, 1985), the authors present new information concerning the chronology of deglaciation and marine transgression in Maine.

This study also provides further means for comparing the timing of deglaciation in the northern and southern hemispheres. Recent work in mid-latitude regions of New Zealand and Chile has shown that Late Wisconsinan (and equivalents) glacial events in these areas were synchronous with glacial advances and retreats in the North Atlantic region (Denton & Hendy, 1994; Lowell *et al*., 1995). The radiocarbon-dated sites in Maine are favorably located for comparison with the southern hemisphere findings.

Research Methods

Ice-marginal positions during deglaciation

Superficial geological mapping by the Maine Geological Survey, during the last 30 years, has generated much data on the types and distribution of glacial deposits in the state. This mapping program, supplemented by topical investigations and academic thesis studies, has helped clarify the pattern and style of ice retreat. The sequence of deglaciation is most clearly-defined in coastal areas of southern Maine. The time-lag between ice retreat and isostatic crustal uplift resulted in a marine transgression of this region (Stuiver & Borns, 1975; Belknap *et al*., 1987; Barnhardt *et al*., 1997). End moraines were deposited along the grounded tidewater-glacial margin during brief stillstands or minor readvances. These moraines are very abundant and indicate the trend of the ice front and incremental retreat pattern over large areas of coastal Maine (e.g. Smith, 1980, 1982; Kaplan, 1999). Ice-contact submarine fans are common both within the moraines and elsewhere, providing further evidence of glacial recession. Some of the fans built up to the ocean surface, especially in shallow waters near the marine limit and evolved into deltas that record both the ice-margin position and elevation of sea-level where each delta was deposited (Thompson *et al*., 1989).

Fewer end moraines exist at elevations above the limit of marine submergence, so it is necessary to rely on other evidence to reconstruct the deglaciation history of inland areas. The morphosequence mapping technique, applied successfully in southern New England (Koteff & Pessl, 1981), has been used to discriminate ice-contact heads of meltwater deposits across interior south-western Maine. Each morphosequence consists of a contemporaneous group of waterlain deposits that were graded to a particular base level, such as a delta and associated outwash graded to the surface of a temporary glacial lake.

Additional evidence of ice-margin positions is provided by meltwater channels eroded by streams issuing from the fronts and sides of valley ice tongues. Multiple sets of glacial striations indicate shifting late-glacial ice flow directions that can be related to the other evidence of deglaciation patterns. In rare instances, pit exposures have revealed glaciotectonic structures that probably resulted from the latest, topographically controlled ice-flow events.
The deglaciation of northern Maine was distinguished from other parts of New England by the development of a residual ice mass resulting from glacial drawdown into the St. Lawrence Lowland to the north. This process caused reversal of ice-flow direction over a large region that extended into adjacent Québec (Lowell, 1985; Lowell & Calkin, 1987). Mapping of meltwater deposits formed by the local ice mass has been augmented by detailed analysis of striations to develop a sequence of late-glacial flow events in northern Maine (Lowell *et al*., 1990).

Radiocarbon dating of deglaciation

One of the principal objectives has been to obtain new radiocarbon ages that would help refine the deglaciation chronology of Maine. As the Late Wisconsinan ice sheet receded from southern Maine, the accompanying marine submergence of lowland areas extended up to 175 km inland in the major river valleys. Deposits of silt, clay and fine sand, named the Presumpscot Formation by Bloom (1960), accumulated in quiescent areas on the sea floor. This glaciomarine environment was colonised by an arctic to subarctic flora and fauna, the fossil remains of which have been used by many investigators for radiocarbon age determinations. Detailed work by Kreutz (1994) using stable isotopic analyses of molluscs and forams provided direct evidence for palaeotemperatures and salinities in the transgressing sea.

The authors conducted systematic searches for deposits of fossiliferous Presumpscot clay in ice-marginal environments. Only a few gravel pits and coastal bluff exposures have yielded *in-situ* ice-proximal fauna suitable for radiocarbon dating. Fossils in surface exposures of the Presumpscot Formation often represent a regressive marine fauna that is somewhat younger than the time of deglaciation. In rare cases radiocarbon ages could be obtained for end moraines in which organic materials are included in till or glaciomarine sediment comprising the moraines. Where direct dating was not possible, dating of the first organic material to accumulate above these landforms provided minimum ages for their deposition and the time of deglaciation.

Organic materials have not been found in exposures of ice-contact deposits above the marine limit. In order to define the timing of deglaciation in this part of Maine, sediments beneath ponds and lakes were cored to obtain basal organic material that would provide minimum-limiting radiocarbon ages for ice retreat. This technique has been used for many years (e.g. Davis & Jacobson, 1985), but the advent of accelerator mass spectrometer (AMS) dating now enables the basal segments of piston cores from ponds to be dated with greater precision and smaller sample requirements than with earlier methods.

Our lake coring encompassed sites extending from the coastal end-moraine belt across the marine limit and north to the Canadian (Québec) border. Seaward from the marine limit, the cores document the transition to freshwater conditions as sea level regressed. At least one radiocarbon age was obtained from each of 51 ponds. Transects parallel to the north/north-westward direction of glacial recession provided radiocarbon ages that are believed to approximate the chronology of deglaciation.

The authors obtained pond sediment cores with a 5-cm and, more recently, a 7.5-cm diameter square-rod piston corer modeled after the design of Wright (1967). Each core was driven to the point where a mass of 250 kg could not make the core barrel penetrate deeper. The cores were extruded in the field, cleaned, logged, photographed, wrapped and stored at the University of Maine's palynology laboratory for later analysis. The AMS radiocarbon ages in Table 1 are reported without reservoir corrections or calibration to calendar years and follow the reporting guidelines of Stuiver & Polach (1977) and Stuiver (1980). A half-life of 5568 years was used and all sample ages were normalised to a d^{13}C PDB value of -25 $^{o}/_{oo}$.

Loss-on-ignition (LOI) determined the following: (1) the variability in organic percentage of sediments that accumulated in the pond basins during late-glacial time; and (2) the percentage of carbonate, or marl, precipitated authigenically in each lake. Samples of 6 cm^3 for LOI analyses were removed from the cores, weighed, dried at 100^{o}C for 48 hours, weighed again and then burned for 120 minutes at 550 ^{o}C to determine organic percent. A final burn at 925 ^{o}C combusted any carbonate in the sediment. The Younger Dryas study used 1 cm^3 samples for LOI analyses (Doner, 1995). All procedures and calculations followed the conventions of Bengtsson & Enell (1986).

Map of Deglaciation Chronology

The map accompanying this paper (Fig. 1) shows time lines for retreat of the Late Wisconsinan ice sheet in Maine. These are based on the radiocarbon ages in Table 1, which includes data from Maine and adjacent New Hampshire, Québec and New Brunswick. All ages shown on the map are in radiocarbon years. Figure 1 essentially shows the chronology of ice-cover in Maine, i.e. the interpreted extent of glacial ice remaining at the indicated time. Each time line designates a generalised position of the ice margin. The actual configuration of the margin would have been much more irregular in detail, given the uneven terrain, local embayments of the ice front, and more-or-less detached and stagnating ice masses fringing the main ice sheet.

A few of the ages shown on Figure 1 are too old in relation to the isochrons (e.g Ledge Pond, Site 51). Some of these sites are high enough that they may have been deglaciated earlier than surrounding lowlands; alternatively there may be problems with the ages. The isochrons are drawn so as to obtain a best-fit configuration based on the majority of ages and field evidence for the pattern of deglaciation. The authors considered major topographic elements (large mountain ranges and basins) along with end moraines, other ice-contact deposits and striation data shown on the Superficial geological Map of Maine (Thompson & Borns, 1985) and detailed local maps.

In many cases sample ages were adjusted to obtain a closer approximation of the time of deglaciation. For terrestrial pond-bottom ages, the inferred sedimentation rates were used in the manner of Davis & Jacobson (1985) to extrapolate from the earliest dated horizon in each core back to the onset of sedimentation, when glacial ice left the pond basin. In spite of these adjustments, some of the deglaciation ages are still minimum estimates. Other estimates probably are too old, either because the dated materials were contaminated or not in atmospheric equilibrium, or due to uncertainties in late-glacial sedimentation rates. For example, the anomalously old deglaciation estimates for Sites 7 and 8 in northern Maine are improbable given the other radiocarbon ages in that region. At these and a few other such localities, the original unadjusted ^{14}C ages are more credible and have been used to compile the deglaciation isochrons.

There is uncertainty regarding the magnitude of the reservoir correction that should be applied to marine radiocarbon ages in Maine. Corrections of approximately 400 years have been applied by various workers in the North Atlantic region (Donner, 1995), but the reservoir effect varies with time and location. For example, it was as much as 800 years in the North Atlantic during the Younger Dryas Chronozone, but close to today's value (379 ± 20 yr) during Allerød/Bølling time (Bondevik *et al.*, 1999).

AMS radiocarbon ages of modern mollusk shells from coastal sites in the Gulf of Maine average about 600 years (M. Kashgarian, personal communication, 1994). Dorion *et al.* (2001) have compiled recent work on the timing of deglaciation in eastern Maine and analyses of the late-glacial marine environment. Their comparison of the ages of basal lacustrine and marine sediments in this part of the state indicates a marine reservoir correction factor between -600 and -800 years. Radiocarbon dating of the remains of a mammoth found in shallow-marine deposits at Scarborough, in south-western Maine, support a correction of similar magnitude. The age of the mammoth is about 12,200 ^{14}C yr BP (B. Bourque, Maine State Museum, personal communication, 1999). Age estimates reported by Retelle & Weddle (2001) from marine shells at similar elevations elsewhere in this area yield ages of about 12,800 ^{14}C yr BP, suggesting a reservoir correction of as much as -600 years for the Late Pleistocene Gulf of Maine.

At Ross Pond in coastal Maine (Site 77 on Fig. 1), AMS bulk-sediment ages from cores in the pond yielded a

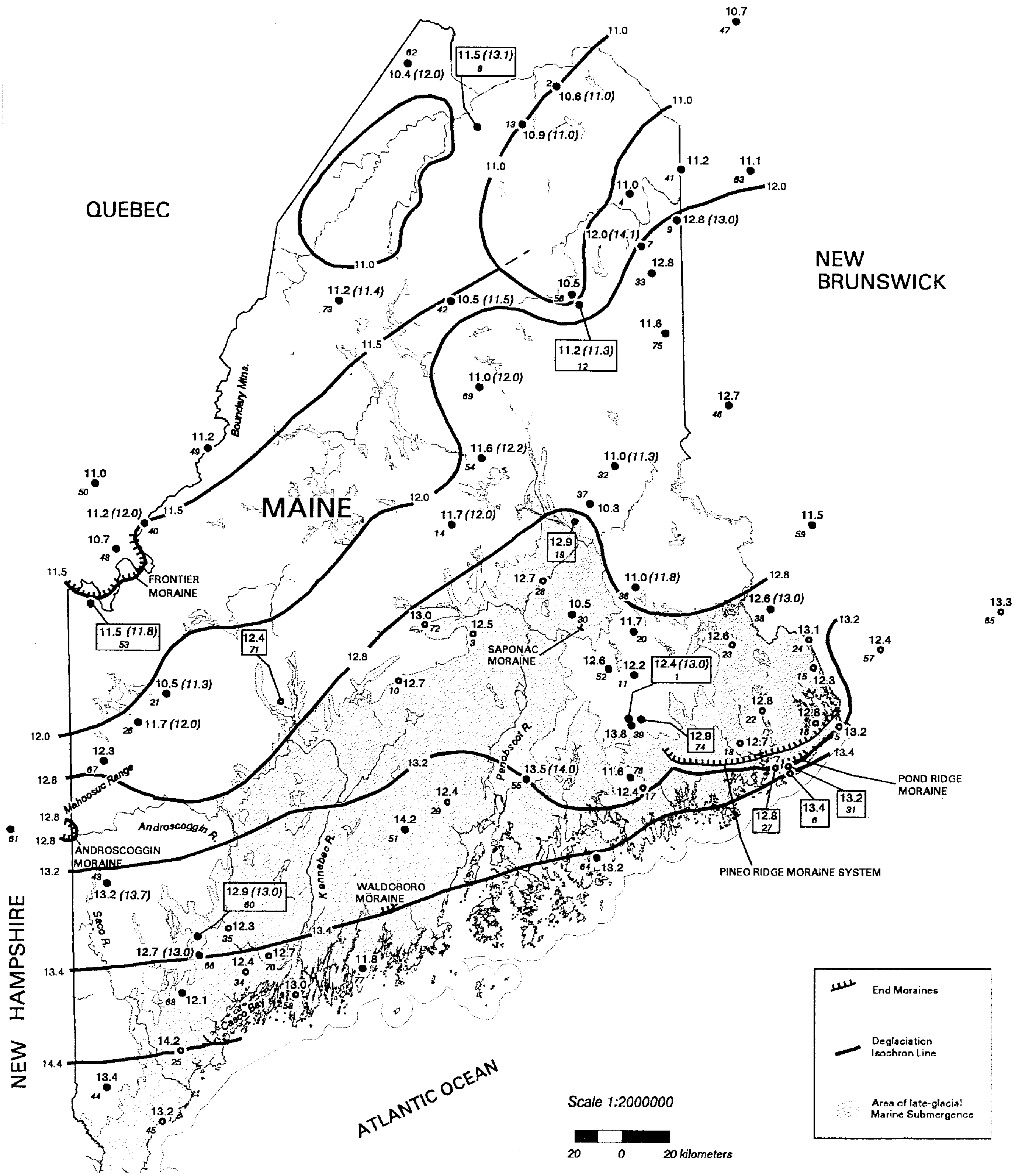

Figure 1. Map showing sites that yielded radiocarbon ages constraining the time of deglaciation in Maine and surrounding areas. Site numbers correspond to Table 1. Radiocarbon ages on the map are rounded to nearest 100 years. Reservoir-corrected ages are shown for marine samples (open circles); labels for terrestrial sites (black dots) include both the reported laboratory ages and (in brackets) the estimated time of deglaciation based on pond sediment cores.

Table 1. Radiocarbon ages used to establish the d[illegible]aciation chronology of Maine. Site numbers [illegible]rrespond to locations shown on Figure 1.

Site	Name (1)	Lat. (° N.)	Long. (° W.)	Setting (2)	Material	Laboratory Number (3)	^{14}C age (yr B.P.) A = AMS age	$\delta^{13}C$ (o/oo)	Deglaciation ^{14}C Age (4)	Reference
1	Bear Pond	44.853	68.094	Terr.	Terr. vegetation	OS-2093*	12,350 ± 55 (A)	-27.21	13.0	This paper
2	Black Lake	47.213	68.468	Terr.	Terr. vegetation	OS-4383*	10,600 ± 30 (A)	-28.82	11.0	This paper
3	Boyd Lake	45.170	68.924	Marine	*Portlandia arctica*	AA-9293*	13,075 ± 90 (A)	0.4	12.5	Dorion *et al.*, in press
4	Caribou Lake	46.811	68.066	Terr.	Terr. vegetation	OS-5993*	11,000 ± 160 (A)	-28.73	11.0	This paper
5	Carrying Place Bluff	44.813	66.979	Marine	*Nucula tenuis*	OS-2075*	13,800 ± 80 (A)	1.36	13.2	Dorion *et al.*, in press
6	Dennison Point	44.642	67.242	Marine	*Macoma calcarea*	OS-2154*	14,000 ± 85 (A)	-0.58	13.4	Kaplan, 1999
7	Echo Lake	46.615	68.005	Terr.	Terr. vegetation	OS-3002*	11,950 ± 190 (A)	-26.01	14.1	This paper
8	1st Pelletier Brook L.	47.061	68.906	Terr.	Terr. vegetation	OS-5305*	11,500 ± 60 (A)	-22.17	13.1	This paper
9	Fischer Lake	46.711	67.811	Terr.	Bulk sediment	OS-4839*	12,750 ± 40 (A)	-26.95	13.0	This paper
10	Gould Pond	44.993	69.319	Marine	*Portlandia arctica*	AA-7463*	13,290 ± 85 (A)	-0.51	12.7	This paper
11	Green Lake	45.016	68.063	Terr.	Terr. vegetation	OS-4843*	12,200 ± 60 (A)	-20.1	12.2	This paper
12	Hall Pond	46.398	68.348	Terr.	Terr. vegetation	OS-6148*	11,200 ± 75 (A)	-25§	11.3	This paper
13	Isie Lake	47.071	68.656	Terr.	Moss	OS-6683*	10,900 ± 90 (A)	-20.67	11.0	This paper
14	Jo Mary Pond	45.577	69.039	Terr.	Terr. vegetation	OS-6440*	11,650 ± 100 (A)	-23.41	12.0	This paper
15	Lewis Cove	45.034	67.108	Marine	*Nucula tenuis*	OS-2659*	12,900 ± 50 (A)	2.34	12.3	Dorion *et al.*, in press
16	Lily Lake	44.828	67.102	Marine	*Hiatella arctica*	OS-2151*	13,350 ± 50 (A)	0.15	12.8	Kaplan, 1999
17	Long Pond	44.595	68.023	Marine	*N. tenuis / P. arctica*	OS-3466*	12,950 ± 120 (A)	-17.9†	12.4	Dorion *et al.*, in press
18	Mark's Lake	44.758	67.505	Marine	*Nucula* sp.	OS-3161*	13,300 ± 65 (A)	1.29	12.7	Dorion *et al.*, in press
19	Mattaseunk Lake	45.590	68.378	Marine	*Portlandia arctica*	OS-1322*	13,450 ± 75 (A)	-16.8†	12.9	Dorion *et al.*, in press
20	Middle Unknown L.	45.178	68.064	Terr.	Terr. vegetation	OS-4844*	11,700 ± 50 (A)	-25.36	11.7	This paper
21	Midway Pond	44.931	70.542	Terr.	Terr. vegetation	AA-9508*	10,490 ± 75 (A)	-25.9	11.3	This paper
22	Patrick Lake	44.878	67.385	Marine	Seaweed	OS-3465*	13,400 ± 95 (A)	-19.33	12.8	Dorion *et al.*, in press
23	Pocomoonshine Lake	45.125	67.539	Marine	*Portlandia arctica*	OS-2661*	13,200 ± 60 (A)	-0.07	12.6	Dorion *et al.*, in press
24	Sand Point, NB	45.139	67.129	Marine	*Nucula* sp.	OS-2663*	13,700 ± 70 (A)	1.92	13.1	Dorion *et al.*, in press
25	Scarborough	43.606	70.430	Marine	*Portlandia arctica*	AA-10166*	14,820 ± 105 (A)	-3.5	14.2	This paper
26	Spencer Pond	44.822	70.688	Terr.	Terr. vegetation	AA-9506*	11,665 ± 85 (A)	-26.7	12.0	This paper
27	Sprague Neck	44.664	67.319	Marine	*Portlandia arctica*	AA-7462*	13,370 ± 90 (A)	-1.47	12.8	Dorion *et al.*, in press
28	T2 R8 NWP	45.368	68.549	Marine	*Nucula* sp.	OS-3160*	13,300 ± 65 (A)	0.31	12.7	Dorion *et al.*, in press
29	Toddy Pond	44.543	69.057	Marine	*Elphidium excav.*	OS-2662*	13,000 ± 60 (A)	-1.24	12.4	Dorion *et al.*, in press
30	Trout Pond	45.242	68.394	Terr.	Terr. vegetation	OS-2682*	10,450 ± 50 (A)	-28.47	10.5	This paper
31	Turner Brook	44.669	67.250	Marine	*Nucula* sp.	AA-7461*	13,810 ± 90 (A)	-1.77	13.2	Dorion *et al.*, in press
32	Wytopitlock	45.796	68.160	Terr.	Terr. vegetation	OS-6582*	11,000 ± 60 (A)	-26.98	11.3	This paper
33	Youngs Lake	46.514	67.950	Terr.	Terr. vegetation	OS-5305*	12,800 ± 100 (A)	-20.96	12.8	This paper
34	Seymour Pit	43.903	70.099	Marine	*Portlandia arctica*	OS-18899*	13,000 ± 55 (A)	-1.15	12.4	Retelle and Weddle, in press
35	Dube Pit	44.064	70.193	Marine	*Portlandia arctica*	AA-10165*	12,890 ± 85 (A)	-1.2	12.3	Retelle and Weddle, in press
36	Duck Lake	45.343	68.052	Terr.	Insect parts	OS-4842*	11,000 ± 40 (A)	-25.45	11.8	This paper

(1) L=Lake NB=New Brunswick Pd=Pond QC=Quebec R=Road (2) Terr. = terrestrial, *i.e.* above the limit of late-glacial marine submergence.

(3) * Indicates samples collected and dated for this study. § Assumed value of $\delta^{13}C$. † Age and $\delta^{13}C$ obtained from shell periostraca.

(4) Time of deglaciation (x 10^3 ka BP) estimated from inferred pond sedimentation rates (terrestrial ages) or by applying reservoir correction of 600 yr (marine ages).

Table 1. Radiocarbon ages used to establish the glaciation chronology of Maine. Site numbe correspond to locations shown on Figure 1.

37	Molunkus Lake	45.654	68.301	Terr.	Insect parts	OS-5992*	10,300 ± 55 (A)	-26.9	10.3	This paper
38	Basswood R. L., NB	45.255	67.330	Terr.	Gyttja	GSC-1067	12,600 ± 270		13.0	Mott, 1975
39	Beddington	44.828	68.080	Terr.	Basal organic sed.	Y-2220	13,760 ± 100		13.8	Stuiver and Borns, 1975
40	Boundary Pond	45.565	70.679	Terr.	Basal organic sed.	GSC-1248	11,200 ± 200		12.0	Shilts, 1981
41	California, NB	46.900	67.783	Terr.	Basal gyttja	GSC-662	11,200 ± 200		11.2	Rampton and Paradis, 1981
42	Chase Lake	46.411	69.049	Terr.	Wood fragments	AA-1935	10,480 ± 130 (A)		11.5	R. B. Davis, pers. comm.
43	Cushman Pond	44.220	70.832	Terr.	Insect parts	OS-7122*	13,150 ± 50 (A)	-23.46	13.7	Thompson *et al.*, 1999
44	Deering Pond	43.463	70.807	Terr.	Bulk sediment	OS-12839*	13,350 ± 75 (A)	-24.33	13.4	C. D. Neil, pers.comm.
45	Great Hill	43.342	70.517	Marine	Marine shells	QL-192	13,830 ± 100		13.2	Smith, 1985
46	Hillman, NB	46.018	67.540	Terr.	Sub-diamict organics	BGS-684	12,700 ± 200		12.7	Rampton and Paradis, 1981
47	Jardine Brook, NB	47.450	67.467	Terr.	Peat	BGS-73	10,664 ± 77		10.7	Korpijaakko, 1976
48	L.aux Araignees, QC	45.467	70.828	Terr.	Basal organic sed.	GSC-1353	10,700 ± 310		10.7	Shilts, 1981
49	Lac Dufresne, QC	45.850	70.350	Terr.	Basal organic sed.	GSC-1294	11,200 ± 160		11.2	Shilts, 1981
50	Lac a la Truite, QC	45.708	70.950	Terr.	Basal organic sed.	GSC-1289	11,000 ± 240		11.0	Shilts, 1981
51	Ledge Pond	44.439	69.279	Terr.	Basal organics	OS-4130	14,150 ± 95 (A)	-28.31	14.2	Donner, 1995
52	Loon Pond	45.039	68.200	Terr.	Plant macrofossils	SI-4953	12,615 ± 115		12.6	Davis and Jacobson, 1985
53	Lower Black Pond	45.260	70.957	Terr.	Insect parts	OS-7123*	11,500 ± 50 (A)	-28.1	11.8	Thompson *et al.*, 1999
54	Lower Togue Pond	45.825	68.881	Terr.	Organic sediment	SI-2992	11,630 ± 260		12.2	Davis and Davis, 1980
55	Moulton Pond	44.628	68.639	Terr.	Organic sediment	I-5639	13,510 ± 300		14.0	Davis *et al.*, 1975
56	Oxbow	46.436	68.388	Terr.	Peat	Beta-90768*	10,530 ± 90	-27.5	10.5	This paper
57	Pennfield Ridge, NB	45.097	66.755	Marine	*Portlandia* sp.	GSC-882	13,000 ± 240		12.4	Gadd, 1973
58	Stonybrook R. Pit	43.821	69.840	Marine	*Mytilus edulis*	GX-21939	13,600 ± 380	-0.3	13.0	Retelle and Weddle, in press
59	Pine Ridge Pond, NB	45.567	67.100	Terr.	Terr. vegetation	Beta-55257	11,490 ± 80 (A)		11.5	Levesque *et al.*, 1994
60	Poland Spring Pond	44.031	70.353	Terr.	Plant macrofossils	SI-4656	12,860 ± 325		13.0	Davis and Jacobson, 1985
61	Pond of Safety, NH	44.410	71.342	Terr.	Insect parts	OS-7125*	12,450 ± 60 (A)	-18.2	12.5	Thompson *et al.*, 1999
62	Rideout Pond	47.296	69.292	Terr.	Plant macrofossils	SI-4658	10,385 ± 140		12.0	Davis and Jacobson, 1985
63	Roulston Lake, NB	46.892	67.400	Terr.	Basal gyttja	GSC-2804	11,100 ± 90	-23.2	11.1	Rampton and Paradis, 1981
64	Sargent Mtn. Pond	44.336	68.269	Terr.	Organic sediment	SI-4042	13,230 ± 360		13.2	Lowell, 1980
65	Sheldon Point, NB	45.225	66.108	Marine	*Hiatella arctica*	GSC-3354	13,900 ± 620		13.3	Rampton *et al.*, 1984
66	Sinkhole Pond	43.961	70.342	Terr.	Plant macrofossils	SI-4657	12,710 ± 125		13.0	Davis and Jacobson, 1985
67	Surplus Pond	44.675	70.865	Terr.	*Salix herbacea*	OS-7119*	12,250 ± 55 (A)	-28.1	12.3	Thompson *et al.*, 1999
68	Tandberg Pit	43.819	70.427	Terr.	*Populus balsamifera*	OS-4416*	12,100 ± 110 (A)	-25.58	12.1	Thompson *et al.*, 1995
69	Upper S. Branch Pd.	46.090	68.893	Terr.	Plant macrofossils	SI-4463	10,965 ± 230		12.0	Anderson *et al.*, 1986
70	Webber Pit	43.964	69.982	Marine	*Portlandia arctica*	AA-10162*	13,315 ± 90 (A)	-2.3	12.7	Retelle and Weddle, in press
71	Embden Pond	44.910	69.939	Marine	Marine shells	Y-1477	13,020 ± 240		12.4	Stuiver and Borns, 1975
72	Dover-Foxcroft	45.204	69.181	Marine	*Macoma balthica*	OS-11022*	13,550 ± 60 (A)	-1.09	13.0	Dorion *et al.*, in press
73	Allagash Pond	46.423	69.668	Terr.	*Salix herbacea*	OS-14150*	11,150 ± 85 (A)	-28.1	11.4	Dorion, 1998
74	Mountain Pond	44.850	68.028	Terr.	Insect / plant parts	OS-15398	12,850 ± 65 (A)	-19.22	12.9	Lurvey, 1999
75	Conroy Lake	46.289	67.875	Terr.	Woody twig	Beta-65420	11,560 ± 60 (A)	-30.4	11.6	Doner, 1995
76	Mud Pond	44.851	68.029	Terr.	*Picea* needles	Beta-43725	11,620 ± 100 (A)		11.6	Doner, 1995
77	Ross Pond	43.923	69.143	Terr.	Cedar(?) twig	Beta-24945	11,770 ± 160 (A)		11.8	Kellogg, 1989

chronology of botanical and climatic events at least several hundred years older than at Mud Pond (Site 76), which is above the late-glacial marine limit. It has been concluded that the dated Ross Pond samples were contaminated by old carbon from the underlying marine mud (Doner, 1995; Kellogg, 1989). A reservoir correction of 500-700 years is suggested by the Ross Pond results (Doner, personal communication, 2001).

Based on the above considerations, we have applied an estimated reservoir correction of 600 years to the marine radiocarbon ages in Table 1, in order to compare them with deglaciation times based on radiocarbon ages from terrestrial sites.

Results

Stratigraphy of pond sediments

The ponds located below the marine limit were depositional basins in shallow marine waters approximately 14.0 to 12.0 ka BP. Water depths were generally less than 80 m. Certain species of invertebrate molluscs, foraminifera and ostracods flourished in the glaciomarine grounding-line environment. Their aragonite tests have been preserved in the anoxic and reduced sediments. In places, mollusc periostraca and marine algae were recovered. Most of the shells were articulated and show no evidence of abrasion or transport. Juveniles and adults were found together in what we interpret as a life assemblage without post-depositional reworking or transport. Isostatic uplift eventually raised the pond basins above sea level and lacustrine sediments were deposited on top of the marine sequence. The transition zone is marked by 10 to 30 cm of black FeS accumulation associated with the onset of meromixis. The isolation of each basin from the sea is marked by an abrupt transition from *in-situ* marine invertebrates preserved in the sediment to freshwater aquatic plant macrofossils.

In contrast, at elevations above the upper marine limit, many ponds were proglacial lakes for a brief time as the ice margin withdrew from their basins. Continued recession of the terminus eventually would allow the meltwater to seek a lower alternate path. When meltwater ceased entering a pond's watershed, minerogenic deposition decreased abruptly and algal gyttja became the dominant sediment type in the basin bottom during post-glacial time.

When the pond sediment cores are compared, including those from both former marine basins and proglacial lakes, most include a similar stratigraphical sequence. The pond-bottom sediments generally overlie till, which locally is mantled by subaqueous outwash deposited as the glacier terminus stood in or near the basin. Above the latter unit is rhythmically-laminated mud and very fine sand. This unit was deposited in the ice-marginal environment by sediment-laden glacial meltwater and/or paraglacial sedimentation; the rhythmites are most probably varves. They are present in both lacustrine and marine environments. This rhythmically-laminated unit grades upward into minerogenic algal gyttja.

Ice recession to the marine limit

Recession of the Late Wisconsinan ice sheet was accompanied by marine transgression of Maine's coastal lowland. Kreutz (1994) and Kaplan (1999) concluded that the glaciomarine environment was analogous to that described by Powell (1990) in which meltwater discharging from the ice rises directly to the ocean surface as muddy overflow plumes, while coarser sand, gravel and flowtill are deposited along the ice margin. Much of the glaciomarine sand and gravel in southern Maine was incorporated into end moraines, submarine fans and deltas. Associated ice-tunnel deposits suggest that most of the glacial sediments that washed into the sea were deposited at the mouths of subglacial meltwater conduits (Ashley *et al*., 1991). Thick deposits of marine mud blanket much of the coastal zone and inland valleys below the marine limit. Water depths were shallow, typically less than 80 m and often shallower than 40 m. The edge of the ice sheet was a grounded tidewater glacier in this environment (Thompson *et al*., 1989).

Many investigations have shown that end moraines are common in the zone of marine submergence and that marine sediments constitute significant percentages of these moraines (e.g. Smith & Hunter, 1989; Dorion, 1997a,b; Kaplan, 1999). Glaciotectonic structures and the stratigraphy of the moraines indicate that the ice remained active during deglaciation of coastal Maine. In some places the associated marine deposits are fossiliferous, enabling the moraines to be dated. For example, the age of the major Pond Ridge Moraine in eastern coastal Maine is bracketed between 13.2 and 12.8 ka B.P. (Sites 27 and 31; Kaplan, 1999). This is close to the age of 13.3 ka (12.7 ka with reservoir correction) obtained by Stuiver & Borns (1975) on seaweed from the Pond Ridge Moraine.

Previous workers have suggested two significant glacial readvances in the coastal zone: the Kennebunk advance in south-western Maine (Bloom, 1960, 1963) and the Pineo Ridge readvance in eastern Maine (Borns, 1973). However, our current findings support the contention of Smith (1981) and Smith & Hunter (1989) that deglaciation of this region proceeded without major readvance of the ice margin. Large moraines such as Pond Ridge and Pineo Ridge locally cross-cut earlier moraines and indicate persistent stillstands or minor readvances of active ice (Kaplan, 1999), but no evidence of regionally extensive overriding of older deposits has been found.

Another noteworthy readvance occurred in the vicinity of South Pond in Warren, Maine, about 7 km south of the well known Waldoboro moraine (the largest moraine in coastal south-western Maine). G.W. Smith mentioned this readvance site in an unpublished field guide for the INQUA Commission on Genesis and Lithology of Quaternary Deposits in 1980. He noted that this locality was originally

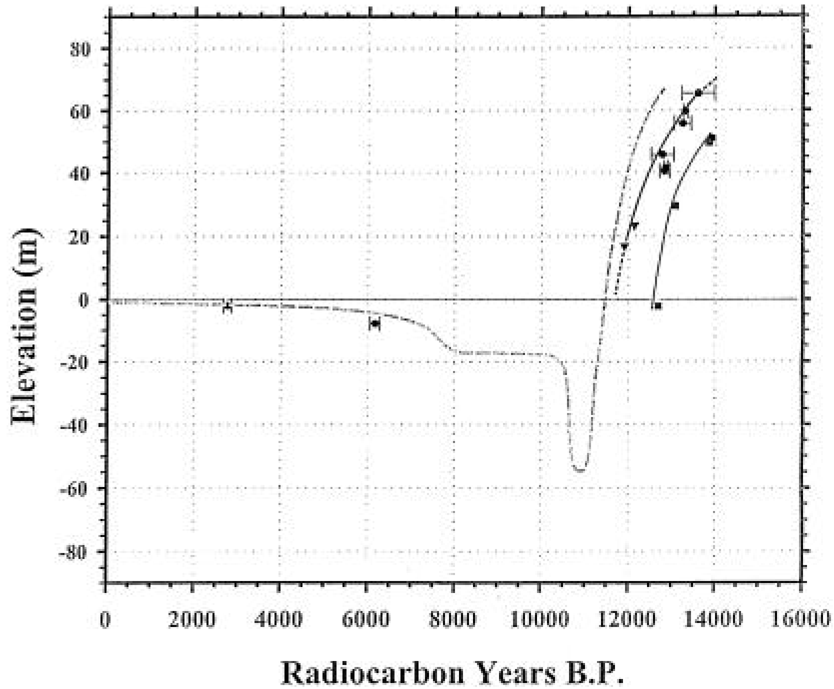

Figure 2. Relative sea-level curves from Maine. Circles are Casco Bay Lowland data (Retelle & Weddle, 2001. refer to Table 1); inverted triangles are data from islands in Casco Bay (Stuiver & Borns, 1975; Maine Geological Survey unpublished data from W.B. Thompson). Squares are eastern Maine data (Dorion, 1997a). Grey dashed line is regional curve for Maine (Barnhardt et al., 1995). Age-estimate elevations were projected to local common values of rebound (Thompson et al., 1989; 80 m for Casco Bay Lowland; 54 m for eastern Maine). Ages have not been corrected for the marine reservoir effect.

described by B.G. Andersen (University of Oslo, Norway) in unpublished reconnaissance work for the Maine Geological Survey. Cross-cutting moraines and striation evidence at South Pond record an apparent oscillation of the local ice margin over a distance of at least 2 km (Maine Geological Survey, unpublished data). Conventional radiocarbon analysis of a bulk-organic sediment sample from a core collected from Kalers Pond, south of the Waldoboro Moraine and west of South Pond, gave an age for marine emergence of 13,240 ± 190 ^{14}C yr BP (GX-23803, $d^{13}C$ value -28.0 $^{o}/_{oo}$; Voisin, 1998). Pending the validity of the age, it provides a minimum constraint on timing of the South Pond readvance and the formation of the Waldoboro Moraine.

Marine-limit deltas

Many of the ice-contact glaciomarine deltas in eastern and south-western Maine are located at or near the inland marine limit (Thompson *et al*., 1989). These Gilbert-type deltas commonly are associated with prominent esker systems. In some cases migration of the tunnel mouth during ice retreat was accompanied by deposition of multiple deltas along a single esker. Surface features on the proximal delta margins include steep ice-contact slopes, kettles and large boulders. Pit exposures in these parts of the deltas reveal coarse gravels with collapse structures and rare flowtills produced by debris flows off the ice margin. The ice-contact zones and their relation to other deposits consistently indicate glacial recession in directions ranging from north to north-west.

The elevation of the contact between topset and foreset beds in the glaciomarine deltas has been precisely measured at numerous locations (Thompson *et al*., 1989). This information was used to established the inland limit of marine submergence shown on the deglaciation map (Fig. 1).

Deglaciation inland from the marine limit

The mode of deglaciation varied across the interior of Maine, depending on ice dynamics and topography. These factors influenced the development and preservation of evidence concerning the history of ice retreat. End moraines are not common over most of this large region, but they do exist in places where conditions favored late-glacial ice flow and focused sediment accumulation at the glacier margin. Some river valleys and lake basins contain clusters of ridges composed of till or waterlain sediments whose origin and relationship to the ice margin are uncertain. The latter ridges are sometimes mapped as ‘ribbed moraine’ and they may at least indicate the direction of ice flow in the marginal zone of the ice sheet (Caldwell *et al*., 1985; Davis, in Thompson *et al*., 1995).

Western Maine. Thompson (2001) has described the sequence of deglaciation in this part of the state. Bouldery cross-valley till ridges which are believed to be end moraines have been found at widely scattered locations across western Maine. Along with meltwater channels and waterlain glacial deposits, the moraines indicate a progressive retreat of the Late Wisconsinan ice margin in a generally northwards direction to the Canadian border. The largest and best defined moraines in this part of the state are on the borders with New Hampshire and Québec (Fig. 1). The Androscoggin Moraine system formed where an ice stream in the upper Androscoggin River valley built a large arcuate cluster of moraine ridges that overlap the Maine-New Hampshire border (Thompson & Fowler, 1989). The Frontier Moraine system is banked against the north-west slopes of the Boundary Mountains on the border with Québec (Shilts, 1981).

Recent detailed mapping in the White Mountain foothills of south-western Maine has shown that many sand and gravel deposits in this region can be resolved into ice-contact morphosequences, as defined by Koteff & Pessl (1981). Most commonly these are glaciolacustrine delta complexes, but they may also include associated outwash in valleys that fed the deltas. The proximal margins (heads of outwash) of the ice-contact assemblages are helpful deglaciation indicators since they approximate the ice-margin positions from which the sequences were built.

The glaciolacustrine morphosequences in south-western Maine often formed in valleys that slope in directions between north and west and thus were temporarily blocked by the receding glacier margin. Ice-dammed lakes in these valleys drained through the lowest gaps in the surrounding hills, the elevation of which determined the water level to which deltas were graded during each lake stage. Depending on local circumstances, ice-margin retreat either opened lower spillways for the lakes (resulting in new delta sequences) or allowed them to drain completely. Other lakes formed behind temporary sediment blockages in south-draining valleys such as the Saco River (Thompson *et al*., 1995). Ice-contact deltas indicate the pattern of ice retreat in both types of lakes, but are essentially limited to valleys. The ability to correlate ice margins between neighbouring valleys depends on the distribution of ice-contact features in the intervening uplands. Proglacial and lateral meltwater channels are useful for this purpose in some areas.

The distribution of morphosequences and age relations of multiple striation sets demonstrate a late-glacial shift in ice-flow direction in south-western Maine, south of the Mahoosuc Range (Fig. 1). Earlier flow during the maximum phase of the Laurentide Ice Sheet was generally south-east to south-south-east, while the late-glacial flow direction varied between south and south-south-west. The shift apparently did not occur north of the Mahoosucs, so it probably resulted from thinning of the ice over the mountains of western Maine and local reorganization of ice flow (Thompson, 2001).

Figure 1 shows a broad ice lobe reaching south to the Androscoggin River at 12.8 ka, at the same time that the Androscoggin Moraine is interpreted as having formed on the Maine-New Hampshire border to the west. This ice-margin configuration is proposed as a possible control for a glacial lake that existed in the Androscoggin valley when the moraine was deposited. Glaciolacustrine sediments extend eastward down the valley from the moraine, but the valley becomes very narrow just west of Rumford and the lake sediments are interrupted in this area. It is suggested that when the Androscoggin Moraine was deposited, the ice margin wrapped around the north side of the Mahoosuc Range and dammed the Androscoggin valley drainage at Rumford.

The chronology of ice retreat in western Maine is based on minimum-limiting radiocarbon ages from pond sediment cores. Recently obtained ages range from approximately 13.5 ka at Site 44 in Sanford to 11.5 ka at Lower Black Pond near the Québec border (Site 53). The latter site is immediately in front of a previously unmapped segment of the Frontier Moraine system (Thompson *et al*., 1999). Ice recession continued from the New England border into the Magog area of south-eastern Québec, until the marine embayment eventually opened in the St. Lawrence Lowland to the north at around 12 ka (Parent & Occhietti, 1988).

The major ice-flow reversal that occurred further north-east in the Maine-Québec border region is not recorded in western Maine or adjacent New Hampshire and Québec. Thus the authors conclude that the glacial ice in this area remained part of the Laurentide Ice Sheet during deglaciation. However, there is a discrepancy between marine shell dates indicating Champlain Sea incursion by 12 ka and basal-organic dates from ponds and lakes suggesting Laurentide ice-margin retreat from the Québec border as recently as 11.5-11.0 ka (Thompson *et al*., 1999). Rodrigues (1992) suggested that marine-limit shell dates for the Champlain Sea transgression may be too old because of a greater concentration of old carbon in the upper part of the marine water column.

Central and northern Maine. In contrast to south-western Maine, glacial meltwater deposits are less abundant in the central and northern portions of the state. In the latter areas, the mapped deposits of stratified drift are limited to esker systems and local concentrations of outwash and glaciolacustrine sediments (Thompson & Borns, 1985). The deglaciation products in this region are also distinctive in other respects. The broad lowland extending from the marine limit north to Mount Katahdin, as well as several smaller basins, contain extensive areas of ribbed moraine. The ridges comprising these deposits are largely composed of till and generally conform to the trend of the ice margin, but it has been debated whether they formed subglacially or along the receding glacial terminus (Caldwell *et al*., 1985).

In the extreme northern part of the state, there are large areas of stagnation moraine, a hummocky mixture of till and meltwater deposits resulting from dissipation of the residual ice cap over northern Maine and adjacent Canada (Lowell, 1985, 1986). Lowell's work showed that the ice cap initially remained active as it diminished over northern Maine and receded southward from the Québec border. The north-flowing ice striated the bedrock, sculpted stoss-and-lee forms and accumulated end moraines (Lowell, 1985).

In north-western Maine, the remnant ice mass deposited varved sediments into Allagash Pond (Site 73) from about 11.4 to 10.9 ka. This side of the ice cap receded to the north or north-east because it dammed the northwards-flowing Allagash drainage in a series of glacial lakes. In north-eastern Maine, the ice margin retreated northward, leaving a few clusters of short end moraines and associated outwash in the region south of Caribou. The Mars Hill Moraine complex is one of the most prominent of these (Newman *et al*., 1985). Most of this area lay to the south of the northern Maine ice divide, with the result that southward ice flow was maintained during deglaciation. A series of glacial

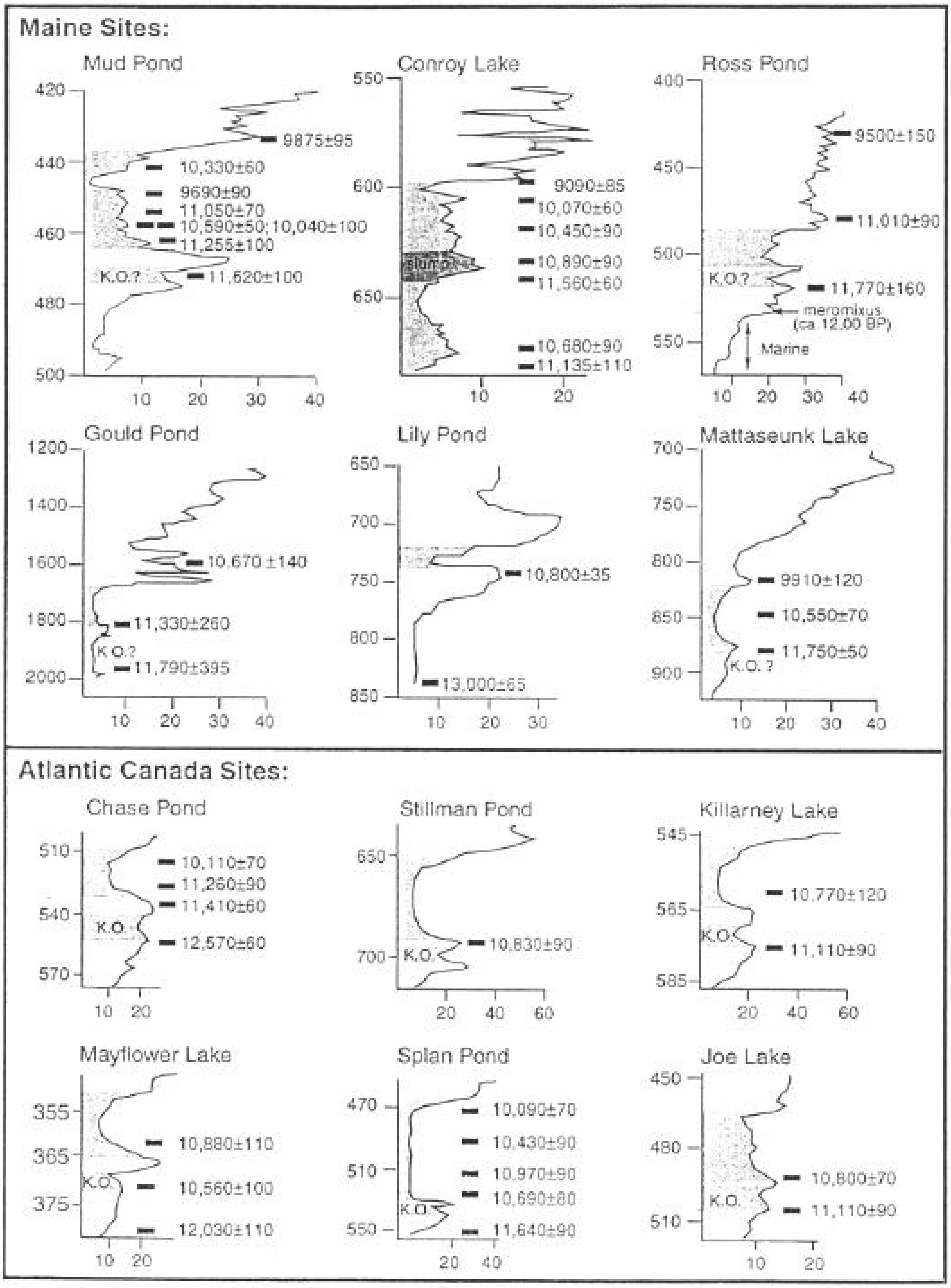

Figure 3. Loss-on-igition (LOI) curves from sites in Maine (top) and Atlantic Canada (bottom), with associated, uncorrected radiocarbon dates. The Younger Dryas interval is the uppermost shaded region in each plot. The Killarney Oscillation (K.O.) is the lower shaded region, where present. Modified from Doner (1995), Dorion (1997), Mayle & Cwynar (1995) and Levesque et al. (1993a). The evidence for marine sediments in Ross Pond (top right) comes from foraminifera (Doner, 1995; Kellogg, 1989).

lakes were impounded in north-draining tributaries of the Aroostook River.

With continued retreat of the northern Maine ice cap, the ice divide shifted to the south-east and the thinning glacier lost most of its erosional capacity and ultimately stagnated. The final ice remnants contained little debris and upon melting left a broad zone in which the glacial sediment cover is thin (Lowell & Calkin, 1987; Lowell *et al*., 1990). By about 11.0 ka, the ice may have separated into two large masses occupying lowland areas, as proposed in Figure 1. These ice remnants may have stabilised or readvanced during the Younger Dryas Chronozone, as

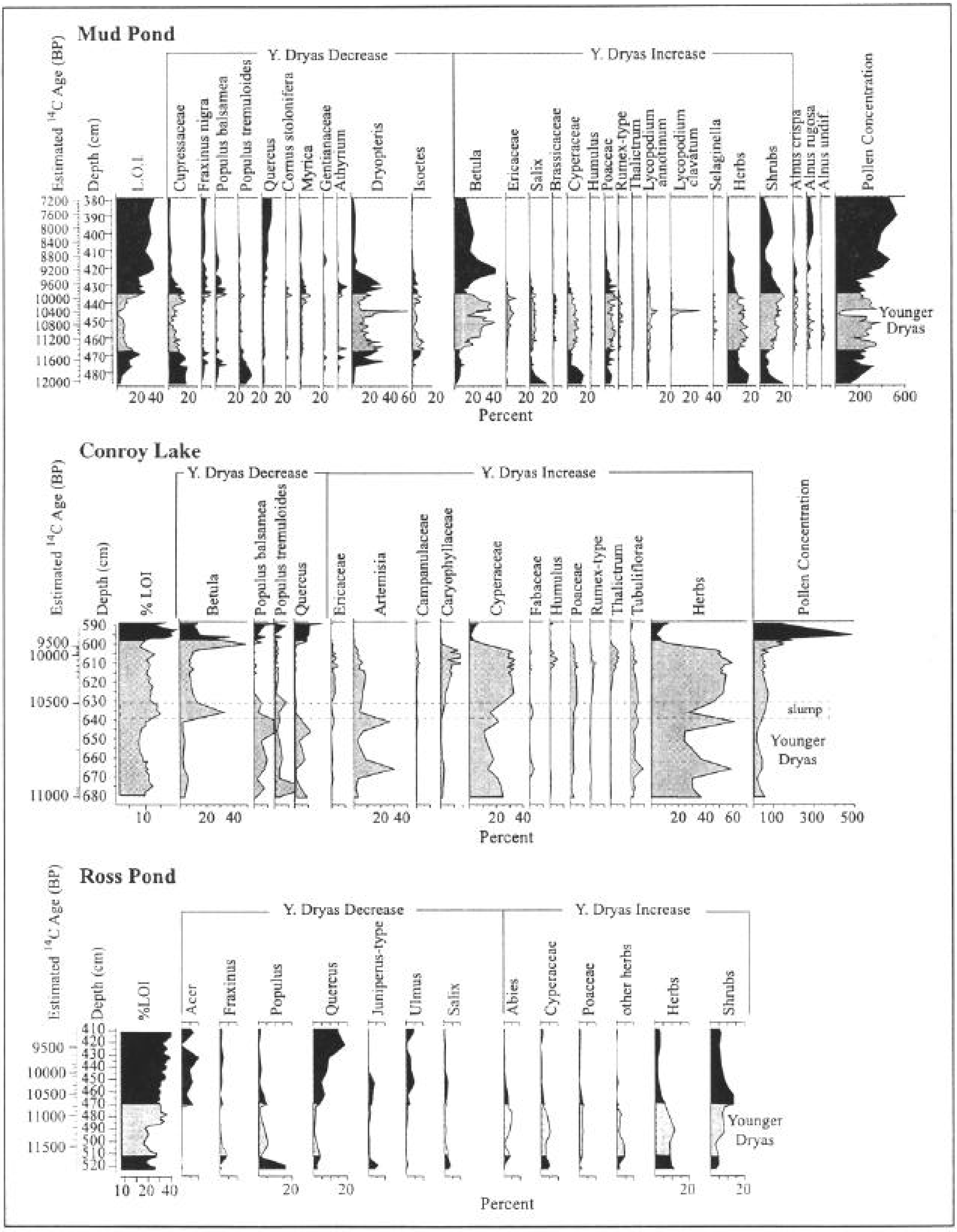

Figure 4. Pollen percentage diagrams from three lake sites in Maine, showing types that change during the Younger Dryas Chronozone. The estimated ages are based on linear interpolations of uncorrected radiocarbon ages. Pollen counts generally exceeded 500 grains. Percentages derive from totals of tree, shrub, herb and indeterminate types. Cyperaceae pollen are assumed to be non-aquatic tundra species and are included in the pollen sum. Spore percentages are based on the sum of spores alone, to highlight changes in spore types during this interval.

discussed below, but they finally dissipated at the onset of the Holocene.

The Younger Dryas chronozone

The Younger Dryas chronozone (10,000-11,000 radiocarbon years BP) is characterised by an 800 to 1000 year period of cold climate, resembling glacial conditions, in the northern North Atlantic and on nearby land masses. During the Younger Dryas, glaciers readvanced in parts of north-west Europe (Scandinavia, Great Britain) and some vegetation changed from forest to tundra **(Where?)** (Watts, 1980). Concurrently, in the North Atlantic Ocean, the polar water front shifted southward from about 58°N to 50°N latitude (Ruddiman *et al*., 1977). Greenland ice-core records show a 10-20 year onset for Younger Dryas cooling, revealed in stable isotopes of oxygen, deuterium, carbon dioxide and in records of dust and methane (Chappellaz *et al*., 1993; Johnsen, *et al*., 1992; Mayewski *et al*., 1993; Taylor *et al*., 1993). Evidence that the Younger Dryas cooling was part of a global phenomenon is often inconsistent (e.g. Heusser & Morley, 1990; Heusser & Rabassa, 1987; Kudrass *et al*., 1991 and Markgraf, 1993); but recent reports do suggest that it was a global event (Thompson *et al*., 1995; Denton & Hendy, 1994; Roberts *et al*., 1993; Almogi-Labin, *et al*., 1991 and Mathewes *et al*., 1993).

Within North America, the evidence for Younger Dryas cooling is supported by numerous studies in the Canadian Maritimes (Cwynar & Levesque, 1995; Levesque *et al*., 1993a; Mayle *et al*., 1993a,b; Mayle & Cwynar, 1995; Mott *et al*. (1986); Mott & Stea, 1993; Ogden, 1987; Rawlence, 1988; Stea & Mott, 1989; Walker *et al*., 1991 and Wolfe & Butler, 1994), in southern New England and New Jersey (Peteet *et al*., 1993), in the southern Great Lakes region (Shane & Anderson, 1993), and in western North America in Wyoming (Gosse *et al*., 1995) and western Canada (Reasoner *et al*., 1994).

The Younger Dryas lithological zone is readily seen in pond sediment cores from parts of north-eastern Maine, supported by evidence from LOI analyses (Fig. 3) (Dorion, 1997; Doner, 1995). Deevey (1951) first recognised and described the Younger Dryas record in cores from this region. The authors found the Younger Dryas lithological zone in pond cores from areas of northern Maine outside our 11.0 ka BP isochrons (Fig. 1). Because the lithological zone in the cores is not overlain by diamicton or till, it is concluded that glacial ice did not readvance into those areas. In cores from sites near the eastern 11.0 ka isochron, basal ages from early Younger Dryas time have been obtained.

In neighbouring New Brunswick, Lamothe (1992) presented the most conclusive evidence yet for reactivation of ice during the Younger Dryas. He excavated a 25 m long trench in which floral and faunal remains belonging to a tundra environment and dating to 11.5 ka BP were overlain by glacial till. This site is approximately 90 km east of the Mars Hill Moraine in northern Maine. Thus, the pond sediment cores and glacial deposits from these parts of Maine and New Brunswick appear to indicate that glacial ice either remained active or readvanced during Younger Dryas time. Further evidence for persistence of glacial ice through the Younger Dryas was reported by Stea & Mott (1998) from Nova Scotia. Organic deposits are overlain by a variety of outwash, lacustrine and diamicton facies which they interpreted as indicating a return to periglacial conditions and reactivation of glacial ice in some areas.

Newman *et al*. (l985) reported a sheared and convoluted peat layer within a diamict at Oxbow in north-central Maine (Site 56). These authors dated the peat at l0,500 ^{14}C yr BP by averaging two radiocarbon ages. They concluded that the area was deglaciated by that time. Borns and Dorion (personal communication) re-excavated the Oxbow site and revealed a lower basal till overlain by about l.5 m of organic-rich sand layers. The latter unit in turn was overlain by a basal till containing ripped-up masses of the organic-rich sand. Radiocarbon ages from the lowest and highest organics in the sand unit are virtually identical: 10,530 ± 90 and 10,560 ± 100 yr BP respectively. These ages suggest that the Oxbow stratigraphy represents a readvance of the northern Maine ice cap during the Younger Dryas.

Other evidence about climate conditions during the Younger Dryas comes from high-resolution pollen studies on three Maine lakes covering the interval from about 11.6 to 9.0 ka BP (Doner, 1995). A summary diagram of these results shows a distinct vegetation response to climate changes beginning as early as 11.4 ka (Fig. 4). Mud Pond and Ross Pond in coastal Maine (Sites 76 and 77) have a distinct, organic-rich, tree-pollen-rich interval prior to 11.0 ka. This period is not well represented at Conroy Lake in northern Maine (Site 75), which has a bottom date of 11,135 ± 110 yr BP. Percentages of many tree types declined at Mud and Ross ponds from 11.6 to 11.0 ka, including *Fraxinus*, *Populus*, *Quercus* and *Juniperus*. These types also had lower values at Conroy Lake between about 10.7 and 9.5 ka. Between 11.4 and 9.8 ka, there were increases in Ericaceae, Cyperaceae, *Humulus*, *Rumex* and *Thalictrum* pollen at Mud Pond and Conroy Lake. Of this group at Ross Pond, only the Ericaceae and Cyperaceae pollen clearly increased during that interval (Fig. 4). Neither Mud Pond nor Conroy Lake became densely-populated by trees until after 9.4 ka.

These three pollen-analyzed sites also show low rates of organic sedimentation and high sand influxes between 11,400 and 9,800 BP. Around 9,200 BP, at Mud Pond and Conroy Lake, sand influx dropped to near zero and organic deposition increased. This period probably marked the establishment of a more stabilised soil because of forest development. Ross Pond, in south-central Maine, was forested much earlier, with tree percentages above 80% after 11,800 BP and consistent with the Davis & Jacobson (1985) summary maps of vegetation for this region.

The Younger Dryas event is generally associated with the period between 11,000 and 10,000 BP (Mangerud *et al*., 1974). Although some records suggest a shorter duration, from 10,600 to 10,000 BP (Cwynar & Watts, 1989), pollen and LOI records from Maine suggest a longer duration of cooler conditions. The late-glacial period was punctuated by repeated incursions and retreats of the relatively warm Atlantic Water into the northern North Atlantic (Lehman & Keigwin, 1992), with episodes of retreat bringing about concomitant cooling that agrees well with the timing of 'flickers' in the Greenland ice-core records (Dansgaard *et al*., 1982; Taylor *et al*., 1993). These multiple cooling events may be difficult to detect with decadal-scale sampling resolution. In fact, early publications on the Younger Dryas in Maine and New Hampshire used sampling resolutions for sediment, pollen and radiocarbon analyses that were too coarse to detect the Younger Dryas event (Anderson *et al*., 1992; Spear, 1989; Davis & Jacobson, 1985; Davis *et al*., 1975). It seems likely that the prolonged cooling shown in these new 'Younger Dryas' records from Maine represents a combined signal of multiple cooling events during the Allerød, Younger Dryas and Preboreal and that higher resolution sampling and dating might distinguish separate cooling events during this interval.

The LOI summary diagrams from Maine do show indications of multiple low-carbon events, including the Killarney Oscillation, at about 11.6 ka BP (Fig. 3). The Killarney Oscillation in New Brunswick and Nova Scotia (11.2-11.1 ka) is associated with the d^{18}O Gerzensee Oscillation in Switzerland, an Allerød-age cold event recorded in fossil beetle and pollen records from Europe, a 200-year decline in electrical conductivity in the GISPII ice core from Greenland and increases in the cold-water foraminiferan *N. pachyderma* in the North Atlantic prior to the Younger Dryas (Levesque *et al*., 1993a, 1993b). Some dates for the European events approximately equivalent to the Killarney Oscillation are closer to 11.5-11.3 ka; those sites include Blelham Bog, England (Pennington, 1975), Llanilid and Gransmoor, England (Walker & Harkness, 1990; Walker *et al*., 1993) and Notsel and Bosscherheide, The Netherlands (Bohncke, 1993). Marine evidence of an earlier date for the Killarney Oscillation comes from the North Atlantic deep-sea core V23-81, documenting an intra-Allerød cold period increase in *N. pachyderma* at about 11,450 BP (Bond *et al*., 1993; Lehman & Keigwin, 1992). Another North Atlantic core, Troll 3.1, has a similar increase in this foraminiferan beginning at 11,730 ± 100 BP and lasting about 200 years (dates corrected by -440 yr for ocean reservoir effect) (Lehman & Keigwin, 1992). In addition, a post-Younger Dryas cooling of the North Atlantic is interpreted from the Troll 3.1 record for 9,700 BP (c), suggesting a possible marine-source for the several hundred years of continued cool climates after 10.0 ka, recorded at Mud Pond and Conroy Lake.

Relative Sea Level During Deglaciation

The revised chronology and detailed mapping in the coastal lowland allows for a new look at deglaciation and post-glacial sea-level history in Maine. The beginning of ice retreat from the Gulf of Maine into south-western Maine falls within the span of Heinrich Event 1 (H-1), dated at 15.0 to 14.0 ka B.P. (Bond *et al*., 1992; Bond & Lotti,1993; Andrews, 1998). At present, the timing of initial response of the Late Wisconsinan tidewater glaciers in the Gulf of Maine to H-1 cannot be determined. The answer probably lies in the offshore record as the oldest ages on ice marginal deposits are 14.8 ka BP south of Portland (Site 25) and *c*. 14.0 ka north of Portland and in the eastern coastal region of the state (Kaplan, 1999; Weddle & Retelle, 2001).

The relative sea-level record in the Casco Bay Lowland of south-western Maine indicates that sea level fell, or conversely, isostatic rebound began concomitant with ice retreat, as demonstrated by both geomorphological and geochronological evidence (Crossen, 1991; Weddle & Retelle, 1995; Barnhardt *et al*., 1997). By contrast, Koteff *et al*. (1993) have determined that rapid uplift in southern New England and extreme south-western Maine was delayed until after 14.0 ka BP. If this were the case in the Casco Bay Lowland, the post-glacial rebound pattern (tilt) would be similar to that of coastal New Hampshire. However, if rapid uplift had begun by the time the ice margin was in the Casco Bay area, the post-glacial gradient of the water plane (as represented by delta topset/foreset contacts) in the study area should be less steep than in coastal New Hampshire and south-western Maine, where the tilt gradient was determined by Koteff *et al*.(1993) to be 0.85 m/km. An estimate of post-glacial tilt in the Casco Bay Lowland using elevations of delta tops from topographic maps and topset/foreset contact measurements by Thompson *et al*. (1989), is 0.63-0.71 m/km. Farther east in the Kennebec River region, isostatic uplift has tilted the plane to the south-east and it now has an average slope of 0.53 m/km (Thompson *et al*., 1989). The west-to-east progressively lower gradient of post-glacial tilt in Maine supports the concept that rapid uplift was underway when the ice margin had reached the central coastal region of Maine.

Figure 2 shows the deglaciation phase of a local relative sea-level curve for south-western Maine based on new radiocarbon-dated samples from the Casco Bay Lowland (CBL) (Retelle & Weddle, 2001). Also shown in this figure are a regional curve utilizing data from the Maine coastal lowland and offshore sediments from the Gulf of Maine (Barnhardt *et al*., 1995) and new deglacial-phase data from eastern Maine (Kreutz, 1994; Dorion, 1997; Kaplan, 1999) fit to a curve based on the projected marine limit in that region (Thompson *et al*., 1989). Two previously reported ages (Stuiver & Borns, 1975; Smith, 1985) from regressive deposits on the islands of Casco Bay also are included in the data set on Figure 2. The ages on Figure 2 have not been corrected for the marine reservoir effect.

As expected, the post-glacial emergence phase of the CBL curve roughly parallels the land-based portion of the regional sea-level curve of Barnhardt *et al*. (1995) and the sea-level curve from eastern Maine. Two ages from nearshore marine sediment cores (Hubeny, 1998), taken from terrestrial organic matter (peat) and a marine mollusc shell hash, plot close to the late Holocene transgressive portion of the Barnhardt *et al.* (1995) curve. The most similar feature in all three curves is the pattern of rapid post-glacial emergence (*cf.* Andrews, 1970). The rate of emergence determined from the CBL curve averages 30 m/ka, roughly equivalent to the regional curve. Each of the curves show a slight curvilinear trend in their uppermost segments representative of slow or restrained initial rebound (*cf.* Andrews, 1970). The main distinction of the CBL curve is that it displays both an earlier initial emergence and slightly greater total emergence than both the regional curve and the eastern Maine curve. The CBL curve represents data from only the southern coastal portion of the lowland. Greater amounts of emergence would be expected further inland due to differential post-glacial tilt.

Discussion

This study has added significant details to the history of deglaciation in Maine. Radiocarbon ages have been obtained that help define the timing of ice retreat and the age of the coastal moraine belt. The deglaciation record shows that the retreating Late Wisconsinan ice margin fluctuated along the entire coastal zone, constructing a belt of cross-cutting grounding-line end moraines, ice-contact deltas and submarine fans between approximately 14.5 and 13.0 ^{14}C ka BP. Local readvances of the ice margin occurred in the marine environment during this time, such as that described at South Pond in Warren, but no major regional readvance has been recognised.

Much of the coastal moraine belt formed during the Heinrich I Event, recognised from the deep-sea record between approximately 15.0 and 13.5 ^{14}C ka BP. Subsequently, there was rapid inland recession of the ice across central and northern Maine between about 13.0 and 11.0 ka BP. This recession is documented by extensive esker systems and the near absence of end moraines as compared to the coastal zone.

The rate of ice-margin retreat has been estimated by several investigators using radiocarbon ages for various parts of Maine. Smith & Hunter (1989) proposed a rate of 200-250 m/yr for the coastal zone. Kaplan (1999) inferred a much slower oscillatory retreat rate of 20 m/yr in the eastern coastal moraine belt, south of the Pineo Ridge Moraine System (Fig. 1). Dorion *et al.* (2001) obtained additional ages from eastern Maine and suggested that the ice recession rate was only 10 m/yr between Dennison Point (Site 6) and Pond Ridge Moraine (Fig. 1), but accelerated to 30-40 m/yr between Pond Ridge and Patrick Lake (site 22) and 150 m/yr further north as the climate warmed. Thompson (1998) estimated a rate of 70 m/yr across western Maine to the Canadian (Québec) border. The distance between successive ridges in clusters of minor moraines in the central to south-western coastal region generally averages 50-75 m (Thompson, 1982), suggesting annual deposition of these moraines. The above recession rates are based on just a few radiocarbon ages from selected parts of the state. The age control is believed to be best in eastern Maine, where a few ice margin positions such as Pond Ridge have been directly dated.

The reconstruction of ice-margin trends presented here (Fig. 1) differs in some respects from that of Davis & Jacobson (1985). These authors proposed a lobe of ice projecting south across western Maine at 13 ka, when the ice margin had retreated much further inland in the marine embayment of south-central Maine (Penobscot River lowland). However, geological mapping in south-western Maine has not revealed evidence of radial flow that would be expected in such a pronounced ice lobe. Moraines and striations indicate consistent north to north-west ice retreat across western Maine.

Radiocarbon ages support the relatively early deglaciation of the Penboscot lowland proposed by Davis & Jacobson (1985) and there is also field evidence for this marine embayment. End moraines and multiple striation sets locally indicate a late convergence of ice flow into the Penobscot Valley (Thompson & Borns, 1985). Our age evidence (Fig. 1) has established that the Late Wisconsinan ice margin retreated to central Maine by 13 ka, in contrast to an earlier model (Smith, 1985) placing the margin near the coast at this time.

There is still a large discrepancy between the deglacial chronology in southern Maine presented here (which is largely based on marine ages) and that proposed by Ridge *et al.* (2001) from varve sequence correlations in this area and elsewhere in New England. These authors suggest that a reservoir correction of 1000-2000 years is implied by the marine radiocarbon ages. Although the magnitude of the reservoir effect remains uncertain, the interpretation of the varve record may be problematic. Absolute ages assigned to the New England varve sequences hinge upon the Canoe Brook section in south-eastern Vermont, where thick (glacial Lake Hitchcock) varves at the exposed base of a dated sequence are thought to be ice-proximal (Ridge & Larsen, 1990). If the Canoe Brook section is underlain by an additional series of varves, then the deglaciation date for that locality and the many other sites correlated with it, is proportionally older.

Farther north in western Maine, near the Canadian (Québec) border, the chronology agrees with the deglaciation ages that Ridge *et al.* (1999) proposed for northern New Hampshire based on ^{14}C calibration of varve sequences deposited in glacial Lake Hitchcock. Glacial retreat from the Mahoosuc Range to Québec probably occurred between 12.0 and 11.5 ka BP. The 11.5 ka age from Lower Black Pond (Site 53 in Fig. 1) is believed to closely approximate the age of the Frontier Moraine system immediately north of there.

Flint (l95l) suggested that the Boundary Mountains of western Maine (Fig. 1) were one of many highland centres of late-glacial outflow. Borns & Calkin (1977) and Davis & Jacobson (1985) concluded that this was not true and that the late Wisconsin ice sheet simply thinned, exposing these mountains as nunataks rather than supporting a persistent ice cap as deglaciation persisted. Borns (1985) considered alternative deglaciation models by which the Frontier Moraine was deposited either by Laurentide ice or along the southern margin of a remnant ice mass in south-eastern Québec. The evidence summarised by Thompson *et al.* (1999) shows that this morainic system was deposited by the Laurentide Ice Sheet as it receded into Québec.

Basal ages from ponds suggest deglaciation as early as 12.8 ka BP in north-eastern Maine. Evidence from neighbouring New Brunswick likewise supports deglaciation beginning at about this time. Lamothe (1992) proposed an ice-margin position dating to 12.7 ka at Hillman, New Brunswick (Site 46). This interpretation was based on the radiocarbon age reported by Rampton *et al.* (1984) for organic debris overlain by diamicton. However, the Hillman site is approximately 40 km south of Youngs Lake, Maine (Site 33), where we obtained a basal age of 12.8 ka. Thus the age for one or both of these localities may not precisely indicate the time of ice retreat.

Lamothe's (1992) deglaciation map for western New Brunswick includes an east-west ice margin with estimated age of 11.5-12.5 ka that is compatible with the 12.0 ka isochron presented here. The latter isochron is determined from basal ages of pond sediments near Caribou and Presque Isle, Maine (Sites 4 and 7, respectively). These two sites are only 15 km apart, so it is proposed that a significant slowing of ice-margin retreat occurred in this area. Ponds north and west of Caribou lack the Younger Dryas lithological zone and presumably were deglaciated after 11.0 ka.

The stratigraphy of the Oxbow site suggests that ice recession in northern Maine may have briefly reversed during the Younger Dryas before final disappearance of remnant ice masses by about l0.0 ka BP. The present results show widespread evidence of Younger Dryas climatic cooling in Maine. This cooling event is recorded in pond sediments over a broad area of northern and western Maine, as well as the adjacent White Mountains in New Hampshire (Site 61). The radiocarbon ages and regional topography suggest that the northern Maine ice cap separated into two ice masses that persisted into the Younger Dryas (approximately outlined by 11.0 ka isochrons on Fig. 1).

Considering the deglaciation ages in north-eastern Maine, there are several relatively young terrestrial ^{14}C ages in the eastern part of the state. These ages range from 12.2 ka at Site 11 to 10.3 ka at Site 37. Though not differentiated on Figure 1, glacial ice may have lingered in this region long after recession to the marine limit in the Penobscot valley. Striations and a few moraines indicate a late-glacial shift in ice-flow from south-east to south in the southern part of this area. Evidence of a radial flow pattern is sparse, although striations with trends between ENE-WSW and ESE-WNW were found at two sites in Lincoln and Lee (east of the Penobscot River). These striations probably reflect late westward flow into the Penobscot valley. Directly east of here, in the vicinity of Site 59 in New Brunswick, Seaman *et al.* (1993) reconstructed a complex ice-flow history in which the latest flow directions shifted from south to east and finally east-north-east. This may be another flow component of residual ice spreading out from eastern Maine. However, there is no disruption of the regional southward trend of large esker systems, so it is doubtful that a discrete ice cap existed in this region of the state.

While considerable progress in documenting the deglaciation in Maine has been achieved, it is clear that gaps remain in our understanding of this complex subject. Additional coring and dating of basal lake sediments are needed, especially in western interior Maine, to improve the accuracy and level of resolution of the isochrons for glacial retreat. Coupled with detailed stratigraphic analysis of late-glacial organic remains, this work should improve correlations with glacial and climatic events elsewhere in the North Atlantic region.

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References

Almogi-Labin, A., Hemleben, C., Meischner, D. & Erlenkeuser, H. (1991). Palaeoenvironmental events during the last 13,000 years in the central Red Sea as recorded by Pteropoda. *Palaeoceanography*, **6 (1)**, 83-98.

Anderson, R.S., Davis, R B., Miller, N:G. & Stuckenrath, R. (1986). History of late- and post-glacial vegetation and disturbance around Upper South Branch Pond, northern Maine. *Canadian Journal of Botany*, **64**, 1977-1986.

Anderson, R.S., Jacobson, G.L., Jr., Davis, R.B. & Stuckenrath, R. (1992). Gould Pond Maine: late-glacial transitions from marine to upland environments. *Boreas*, **21**, 359-371.

Andrews, J.T. (1970). A geomorphological study of post-glacial uplift with particular reference to Arctic Canada. *Institute of British Geographers, Special Publication*, **2**, 156 pp.

Andrews, J.T. (1998). Abrupt changes (Heinrich events) in late Quaternary North Atlantic marine environments: A history and review of data and concepts. *Journal of Quaternary Science*, **13**, 3-16.

Ashley, G.A., Boothroyd, J.C. & Borns, H.W., Jr. (1991). Sedimentology of late Pleistocene (Laurentide) deglacial-phase deposits, eastern Maine; An example of a temperate marine grounded ice-sheet margin. *In:* Anderson, J.B. & Ashley, G.M. (eds), *Glacial marine sedimentation; Palaeoclimatic significance. Geological Society of America, Special Paper*, **261**, 107-125.

Barnhardt, W.A., Gehrels, W.R., Belknap, D.F. & 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.

Barnhardt, W.A., Belknap, D.F. & Kelley, J.T. (1997). Stratigraphic evolution of the inner continental shelf in response to late Quaternary relative sea-level change, north-western Gulf of Maine. *Geological Society of America, Bulletin*, **109**, 612-630.

Belknap, D.F. **and nine others (give names, please!)** (1987). Late Quaternary sea-level changes in Maine. *In:* Nummedal, D., Pilkey, O. H., Jr. & Howard, J. D. (eds), *Sea level fluctuation and coastal evolution. Society of Economic Palaeontologists and Mineralogists, Special Publication*, **41**, 71-85.

Bengtsson, D.F. & Enell, M. (1986). Chemical analysis. *In:* Berglund, B.E. (ed.), *Handbook of Holocene palaeoecology and palaeohydrology*. New York, John Wiley, 423-451.

Björck, S. **& ten others** (1996). Synchronised terrestrial-atmospheric deglacial records around the North Atlantic. Science, 274, 1155-1160.

Bloom, A.L. (1960). Late Pleistocene changes of sea level. *In:* Maine **(eds?)**. Augusta, Maine Geological Survey, 143 pp.

Bloom, A.L. (1963). Late Pleistocene fluctuations of sea level and post-glacial crustal rebound in coastal Maine. *American Journal of Science*, **261**, 862-879.

Bohncke, S.J.P. (1993). Lateglacial environmental changes in the Netherlands: spatial and temporal patterns. *Quaternary Science Reviews*, **12**, 707-717.

Bond, G. **& 13 others** (1992). Evidence for massive discharges of icebergs into the North Atlantic Ocean during the last glacial period. *Nature*, **360**, 245-249.

Bond, G., Broecker, W., Johnsen, S., McManus, J., Labeyrie, L., Jouzel, J. & Bonani, G. (1993). Correlations between climate records from North Atlantic sediments and Greenland ice. *Nature*, **365**, 143-147.

Bond, G. & Lotti, R (1995). Iceberg discharges into the North Atlantic on millennial time scales during the last glaciation. *Science*, **267**, 1005-1009.

Bondevik, S., Birks, H.H., Gulliksen, S. & Mangerud, J. (1999). Late Weichselian marine ^{14}C reservoir ages at the western coast of Norway. *Quaternary Research*, **52**, 104-114.

Borns, H.W., Jr. (1973). Late-Wisconsin fluctuations of the Laurentide Ice Sheet in southern and eastern New England. *In:* Black, R.F., Goldthwait, R. & Willman, H.B. (eds), *The Wisconsinan Stage. Geological Society of America, Memoir*, **136**, 37-45.

Borns, H.W., Jr. & Calkin, E., (1977). Quaternary glaciation, west-central Maine. *Geological Society of America Bulletin*, **88**, l773-l784.

Borns, H.W., Jr. (1985). Changing models of deglaciation in northern New England and adjacent Canada. *In:* Borns, H.W., Jr., LaSalle, & Thompson, W.B. (eds), *Late Pleistocene history of north-eastern New England and adjacent Quebec. Geological Society of America, Special Paper*, **197**, 135-138.

Caldwell, D.W., Hanson, L.S. & Thompson, W.B. (1985). Styles of deglaciation in central Maine. *In:* Borns, H.W., Jr., LaSalle, & Thompson, W.B. (eds), *Late Pleistocene history of north-eastern New England and adjacent Quebec. Geological Society of America, Special Paper*, **197**, 45-48.

Chappellaz, J., Blunier, T., Raynaud, D., Barnola, J. M., Schwandler, J. & Stauffer, B. (1993). Synchronous changes. *In: Atmospheric CH_4 and Greenland climate between 40 and 8 kyr BP. Nature*, **366**, 443-445.

Crossen, K.J. (1991). Structural control of deposition by Pleistocene tidewater glaciers, Gulf of Maine. *In:* Anderson, J.B. & Ashley, G.M. (eds), *Glacial marine sedimentation; palaeoclimatic significance. Geological Society of America, Special Paper*, **261**, 127-135.

Cwynar, L.C. & Levesque, A.J. (1995). Chironomid evidence for lateglacial climatic reversals in Maine, U.S.A.. *Quaternary Research*, **43**, 405-413.

Cwynar, L.C. & Watts, W.A. (1989). Accelerator-mass spectrometerages for late-glacial events at Ballybetagh, Ireland. *Quaternary Research*, **31**, 377-380.

Dansgaard, W., White, J.W.C. & Johnsen, S.J. (1989). The abrupt termination of the Younger Dryas climate event. *Nature*, **339**, 532-533.

Dansgaard, W. & **10 others** (1993). Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature*, **364**, 218-220.

Davis, T. & Davis, R.B. (1980). Interpretation of minimum-limiting radiocarbon dates for deglaciation of Mount Katahdin area, Maine. *Geology*, **8**, 396-400.

Davis, R.B., Bradstreet, T.E., Stuckenrath, R., Jr. & Borns, H.W., Jr. (1975). Vegetation and associated environments during the past 14,000 years near Moulton Pond, Maine. *Quaternary Research*, **5**, 435-465.

Davis, R.B. & Jacobson, G.L., Jr. (1985). Late glacial and early Holocene landscapes in northern New England and adjacent areas of Canada. *Quaternary Research*, **23**, 341-368.

Deevey, E.S., Jr. (1951). Late-glacial and post-glacial pollen diagrams from Maine. *American Journal of Science*, **249**, 177-207.

Denton, G.H. & Hendy, C.H. (1994). Younger Dryas age advance of Franz Josef Glacier in the Southern Alps of New Zealand. *Science*, **264**, 1434-1437.

Doner, L.A. (1996). Late-Pleistocene environments in Maine and the Younger Dryas dilemma. M.S. Thesis, University of Maine, 68 pp.

Donner, J. (1995). *The Quaternary history of Scandinavia*. Cambridge University Press 200 pp.

Donner, J.E. (1995). The effects of islands on the recession of the Late Wisconsinan ice sheet margin in the DeGeer Sea, Maine. Orono, University of Maine, M.S. thesis. **(pages?)**

Dorion, C.C. (1997a). An updated high resolution chronology of deglaciation and accompanying marine transgression in Maine. Orono, University of Maine, M.S. thesis, 147 pp.

Dorion, C.C. (1997b). Middle to Late Wisconsinan glacial chronology and palaeoenvironments along a transect from eastern coastal Maine north to New Brunswick and Quebec. *Canadian Quaternary Association, 8th Biennial Meeting, Program Abstracts*, 19-20.

Dorion, C.C., Balco, G.A., Kaplan, M.R., Kreutz, K.J., Wright, J.D. & Borns, H.W., Jr. (2001). Stratigraphy, palaeoceanography, chronology and environment during deglaciation of eastern Maine. *In:* Weddle, T. K. & Retelle, M. J. (eds), *Deglacial history and relative sea-level changes, northern New England and adjacent Canada. Geological Society of America, Special Paper*, **351**, 215-242.

Flint, R.F., (1951). Highland centers of former glacial outflow in north-eastern North America. *Geological Society of America Bulletin*, **62**, 21-37.

Gadd, N.R. (1973). Quaternary geology of south-west New Brunswick, with particular reference to Fredericton area. *Geological Survey of Canada, Paper*, **71-34**. **(pages?)**

Gosse, J.C., Evenson, E.B., Klein, J., Lawn, B. & Middleton, R. (1995). Precise cosmogenic ^{10}Be measurements in western North America: support for a global Younger Dryas cooling event. *Geology*, **23 (10)**, 877-880.

Grimm, E.C. & Jacobson, G.L., Jr. (1992). Fossil-pollen evidence for abrupt climatic change during the past 14,000 years in eastern North America. *Climate Dynamics*, **6**, 179-184.

Heinrich, H. (1988). Origin and consequences of cyclic ice rafting in the north-east North Atlantic Ocean during the past 130,000 years. *Quaternary Research*, **29**, 143-152.

Heusser, C.J. & Rabassa, J. (1987). Cold climatic episode of Younger Dryas age in Tierra del Fuego. *Nature*, **328**, 609-611.

Heusser, L.E. & Morley, J.J. (1990). Climatic change at the end of the last glaciation in Japan inferred from pollen in three cores from the north-west Pacific Ocean. *Quaternary Research*, **34**, 101-110.

Hubeny, J. B. (1997). Late Quaternary stratigraphy and geological history of The Basin, Phippsburg, Maine. Lewiston, Maine, Bates College, B. S. Thesis, 87 pp.

Johnsen, S.J., Clausen, H.B., Dansgaard, W., Fuhrer, K., Gundestrup, N., Hammer, C.U., Iversen, P., Jouzel, J., Stauffer, B. & Steffensen, J.P. (1992). Irregular glacial interstadials recorded in a new Greenland ice core. *Nature*, **359**, 311-313.

Kaplan, M.R. (1999). Retreat of a tidewater margin of the Laurentide ice sheet in eastern coastal Maine between ca. 14000 and 13000 ^{14}C yr B.P.. *Geological Society of America Bulletin*, **111**, 620-632.

Kellogg, D. (1989). Prehistoric landscapes, palaeoenvironments and archaeology of western Muscongus Bay, Maine. Orono, University of Maine, Ph.D. thesis. **(pages?)**

Korpijaakko, M.L. (1976). On the geobotanical and palaeontologic development of muskeg in New Brunswick. Fredericton, University of New Brunswick, Ph.D. thesis. **(pages?)**

Koteff, C. & Pessl, F., Jr. (1981). Systematic ice retreat in New England. *U.S. Geological Survey, Professional Paper*, **1179**, 20 pp.

Koteff, C., Robinson, G.R., Goldsmith, R. & Thompson, W.B. (1993). Delayed post-glacial uplift and synglacial sea levels in coastal central New England. *Quaternary Research*, **40**, 46-54.

Kreutz, K.J. (1994). Ice-marginal palaeoceanographic conditions during the late Wisconsin marine transgression in Maine. *American Quaternary Association, Program and abstracts of the 13th biennial meeting*, p. 116.

Kudrass, H.R., Erlenkeuser, H., Vollbrecht, R. & Weiss, W. (1991). Global nature of the Younger Dryas cooling event inferred from oxygen isotope data from Sulu Sea cores. *Nature*, **349**, 406-409.

Lamothe, M. (1992, Pleistocene stratigraphy and till geochemistry of the Miramichi Zone, New Brunswick. *Geological Survey of Canada, Bulletin*, **433**, 58 pp.

Lehman, S.J. & Keigwin, L.D. (1992). Sudden changes in North Atlantic circulation during the last deglaciation. *Nature*, **356**, 757-762.

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

Levesque, A.J., Mayle, F.E., Walker, I.R. & Cwynar, L.C. (1993b). The amphi-Atlantic oscillation: a proposed late-glacial climatic event. *Quaternary Science Reviews*, **12**, 629-643.

Levesque, A.J., Cwynar, L.C. & Walker, I.R. (1994). A multiproxy investigation of late-glacial climate and vegetation change at Pine Ridge Pond, south-west New Brunswick, Canada. *Quaternary Research*, **42**, 316-327.

Lowell, T.V. (1980). Late Wisconsinan ice extent in Maine: evidence from Mount Desert Island and the Saint John River area. Orono, University of Maine, M.S. thesis.

Lowell, T.V. (1985). Late Wisconsin ice-flow reversal and deglaciation, north-western Maine. *In:* Borns, H.W., Jr., LaSalle, & Thompson, W.B., eds., Late Pleistocene history of north-eastern New England and adjacent Quebec. *Geological Society of America, Special Paper*, **197**, 71-83.

Lowell, T.V. (1986). Late Wisconsin stratigraphy, glacial ice-flow and deglaciation style of north-western Maine. Buffalo, New York, State University of New York at Buffalo, Ph.D. thesis, 169 pp.

Lowell, T. & Calkin, E. (1987). Development, dynamics and dissipation of a late Wisconsinan ice mass over northern New England, USA. *In:* Waddington, E.D. & Walder, J.S. (eds), *The physical basis of ice sheet modeling. International Association of Hydrological Sciences Publication*, **170**, 367-375.

Lowell, T.V., Kite, J.S., Calkin, E. & Halter, E.F. (1990). Analysis of small-scale erosional data and a sequence of Late Pleistocene flow reversal, northern New England. *Geological Society of America Bulletin*, **102**, 74-85.

Lowell, T. **and eight others** (1995). Interhemispheric correlation of Late Pleistocene glacial events. *Science*, **269**, 1541-1549.

Lurvey, L.K. (1999). An investigation of a glacial landform transition in south-eastern Maine. Orono, University of Maine, M.S. Thesis, 79 pp.

Mangerud, J., Andersen, S.T., Berglund, B.E. & Donner, J.J. (1974). Quaternary stratigraphy of Norden, a proposal for terminology and classification. *Boreas*, **3**, 109-128.

Markgraf, V. (1993). Younger Dryas in southernmost South America - an update. *Quaternary Science Reviews*, **12**, 351-355.

Mathewes, R.W., Heusser, L.E. & Patterson, R.T. (1993). Evidence for a Younger Dryas-like cooling event on the British Columbia coast. *Geology*, **21**, 101-104.

Mayewski, A., Meeker, L.D., Whitlow, S., Twickler, M.S., Morrison, M.C., Alley, R.B., Bloomfield, & Taylor, K. (1993). The atmosphere during the Younger Dryas. *Science*, **261**, 195-197.

Mayle, F.E. & Cwynar, L.C. (1995). Impact of the Younger Dryas cooling event upon lowland vegetation of Maritime Canada. *Ecological Monographs*, **65**, 129-154.

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

Mayle, F.E., Levesque, A.J. & Cwynar, L.C. (1993b). Alnus as an indicator taxon of the Younger Dryas cooling in eastern North America. *Quaternary Science Reviews*, **12**, 295-305.

Mott, R.J. (1975). Palynological studies of lake sediment profiles from south-western New Brunswick. *Canadian Journal of Earth Sciences*, **12**, 273-288.

Mott, R.J. & Stea, R.R. (1993). Late-glacial (Allerød /Younger Dryas) buried organic deposits, Nova Scotia, Canada. *Quaternary Science Reviews*, **12**, 645-657.

Mott, R.J., Grant, D.R.G., Stea, R.R. & Occhietti, S. (1986). Late-glacial climate oscillation in Atlantic Canada equivalent to the Allerød-Younger Dryas event. *Nature*, **323**, 247-250.

Newman, W.A., Genes, A.N. & Brewer, T. (1985). Pleistocene geology of north-eastern Maine. *In:* Borns, H.W., Jr., LaSalle, & Thompson, W.B. (eds), *Late Pleistocene history of north-eastern New England and adjacent Quebec. Geological Society of America, Special Paper*, **197**, 59-70.

Ogden, J.G., III (1987). Vegetational and climatic history of Nova Scotia. I. Radiocarbon-dated pollen profiles from Halifax, Nova Scotia. *Canadian Journal of Botany*, **65**, 1482-1490.

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

Pennington, W. (1975). A chronostratigraphic comparison of Late-Weichselian and Late-Devensian subdivisions, illustrated by two radiocarbon-dated profiles from western Britain. *Boreas*, **4**, 157-171.

Peteet, D.M., Daniels, R.A., Heusser, L.E., Vogel, J.S., Southon, J.R. & Nelson, D.E. (1993). Late-glacial pollen, macrofossils and fish remains in north-eastern U.S.A. - the Younger Dryas oscillation. *Quaternary Science Reviews*, **12**, 597-612.

Powell, R.D. (1990). Processes at glacial grounding-line fans and their growth to ice-contact deltas. *In:* Dowdeswell, J.A. & Scourse, J.D. (eds), Glacimarine environments; processes and sediments. *Geological Society of London Special Publication*, **53**, 53-73.

Rampton, N., Gauthier, R.C., Thibault, J. & Seaman, A.A. (1984). Quaternary geology of New Brunswick. *Geological Survey of Canada, Memoir*, **416**, 77 pp.

Rampton, N. & Paradis, S. (1981). Quaternary geology of Woodstock map area (21 J). *New Brunswick Department of Natural Resources, Map Report*, **81-1**. **(pages?)**

Rawlence, D.J. (1988). The post-glacial diatom history of Splan Lake, New Brunswick. *Journal of Palaeolimnology*, **1**, 51-60.

Reasoner, M.A., Osborn, G. & Rutter, N.W. (1994). Age of the Crowfoot advance in the Canadian Rocky Mountains: a glacial coeval with the Younger Dryas oscillation. *Geology*, **22**, 439-442.

Retelle, M.J. & Weddle, T.K. (2001). Deglaciation and relative sea-level chronology, Casco Bay Lowland and lower Androscoggin River valley, Maine. *In:* Weddle, T.K. & Retelle, M.J. (eds), *Deglacial history and relative sea-level changes, northern New England and adjacent Canada. Geological Society of America, Special Paper*, **351**, 191-214.

Ridge, J.C. & Larsen, F.D. (1990). Re-evaluation of Antevs' New England varve chronology and new radiocarbon dates of sediments from glacial Lake Hitchcock. *Geological Society of America Bulletin*, **102**, 889-899.

Ridge, J.C. **and seven others** (1999). Varve, palaeomagnetic and ^{14}C chronologies for Late Pleistocene events in New Hampshire and Vermont. *In:* Thompson, W. B., Fowler, B. K. & Davis, T., eds., Late Quaternary history of the White Mountains, New Hampshire and adjacent south-eastern Québec. *Géographie physique et Quaternaire*, **53**, 79-106.

Ridge, J.C., Canwell, B.A., Kelly, M.A. & Kelley, S.Z. (2001). An atmospheric ^{14}C chronology for Late Wisconsinan deglaciation and sea level change in eastern New England using varve and palaeomagnetic records. *In:* Weddle, T. K. & Retelle, M. J., eds., Deglacial history and relative sea-level changes, northern New England and adjacent Canada. *Geological Society of America, Special Paper*, **351**, 171-189.

Roberts, N., Taieb, M., Barker, P., Damnati, B., Icole, M. & Williamson, D. (1993). Timing of the Younger Dryas event in East Africa from lake-level changes. *Nature*, **366**, 146-148.

Rodrigues, C.G. (1992). Successions of invertebrate microfossils and the late Quaternary deglaciation of the central St. Lawrence Lowland, Canada and United States. *Quaternary Sciences Reviews*, **11**, 503-534.

Ruddiman, W.F., Sancetta, C.D. & McIntyre, A. (1977). Glacial/interglacial response rate of subpolar North Atlantic waters to climatic change: the record in oceanic sediments. *Philosophical Transactions of the Royal Society of London*, **B 280**, 119-142.

Seaman, A.A., Broster, B.E., Cwynar, L.C., Lamothe, M., Miller, R.F. & Thibault, J.J. (1993). Field guide to the Quaternary geology of south-western New Brunswick. *Fredericton, New Brunswick Department of Natural Resources and Energy*, *Open File Report*, **93-1**, 102 pp.

Shane, L.C.K. & Anderson, K. (1993). Intensity, gradients and reversals in late-glacial environmental change in east-central North America. *Quaternary Science Reviews*, **12**, 307-320.

Shilts, W.W. (1981). Superficial geology of the Lac-Megantic area, Québec. *Geological Survey of Canada, Memoir*, **397**, 102 pp.

Smith, G.W. (1980). *End moraines and glaciofluvial deposits, Cumberland and York Counties, Maine*. Augusta, Maine, Maine Geological Survey, 1:250,000-scale map.

Smith, G.W. (1981). Kennebunk glacial advance: A reappraisal. *Geology*, **9**, 250-253.

Smith, G.W. (1982). End moraines and the pattern of last ice retreat from central and south coastal Maine. *In:* Larson, G.J. & Stone, B.D. (eds), *Late Wisconsinan glaciation in New England*. Dubuque, Iowa, Kendall/Hunt, 195-209.

Smith, G.W. (1985). Chronology of Late Wisconsinan deglaciation of coastal Maine. *In:* Borns, H.W., Jr., LaSalle, & Thompson, W.B. (eds), *Late Pleistocene history of north-eastern New England and adjacent Quebec. Geological Society of America, Special Paper*, **197**, 29-44.

Smith, G.W & Hunter, L E. (1989). Late Wisconsinan deglaciation of coastal Maine. *In:* Tucker, R.D. & Marvinney, R.G. (eds), *Studies in Maine geology, Volume 6 - Quaternary geology*. Augusta, Maine Geological Survey, 13-32.

Spear, R.W. (1989). Late-Quaternary history of high-elevation vegetation in the White Mountains of New Hampshire. *Ecological Monographs*, **59 (2)**, 125-151.

Stea, R.R. & Mott, R.J. (1989). Deglaciation environments and evidence for glaciers of Younger Dryas age in Nova Scotia, Canada. *Boreas*, **18**, 169-187.

Stea, R.R. & Mott, R.M. (1998). Deglaciation of Nova Scotia: Stratigraphy and chronology of lake sediment cores and buried organic sections. *Géographie physique et Quaternaire*, **52**, 3-21.

Stuiver, M. (1980). Workshop on ^{14}C data reporting. *Radiocarbon*, **22**, 964-966.

Stuiver, M. & Borns, H.W., Jr. (1975). Late Quaternary marine invasion in Maine: its chronology and associated crustal movement. *Geological Society of America Bulletin*, **86**, 99-104.

Stuiver, M. & Polach, H.A. (1977). Discussion: Reporting of ^{14}C data. *Radiocarbon*, **19**, 355-363.

Taylor, K.C., Lamorey, G.W., Doyle, G.A., Alley, R.B., Grootes, M., Mayewski, A., White, J.W.C. & Barlow, L.K. (1993). The 'flickering switch' of Late Pleistocene climate change. *Nature*, **361**, 432-436.

Thompson, L.G., Mosely-Thompson, E., Davis, M.E., Lin, P.-N., Henderson, K.A., Cole-Dai, J., Bolzan, J.F. & Liu, K.-b. (1995). Late Glacial Stage and Holocene tropical ice core records from Huascaròn, Peru. *Science*, **269**, 46-50.

Thompson, W.B. (1982). Recession of the Late Wisconsinan ice sheet in coastal Maine. *In:* Larson, G.J. & Stone, B.D. (eds), *Late Wisconsinan glaciation in New England*. Dubuque, Iowa, Kendall/Hunt, 211-228.

Thompson, W.B. (2001). Deglaciation of western Maine. *In:* Weddle, T.K. & Retelle, M.J. (eds), *Deglacial history and relative sea-level changes, northern New England and adjacent Canada. Geological Society of America, Special Paper*, **351**, 109-123.

Thompson, W.B & Borns, H.W., Jr. (1985). *Surficial geological map of Maine*. Augusta, Maine, Maine Geological Survey 1:500,000-scale map.

Thompson, W.B. & Fowler, B.K. (1989). Deglaciation of the upper Androscoggin River valley and north-eastern White Mountains, Maine and New Hampshire. *In:* Tucker, R.D. & Marvinney, R.G., (eds), *Studies in Maine geology, Volume 6 - Quaternary geology*. Augusta, Maine Geological Survey, 71-88.

Thompson, W.B., Crossen, K.J., Borns, H.W., Jr. & Andersen, B.G. (1989). Glaciomarine deltas of Maine and their relation to late Pleistocene-Holocene crustal movements. *In:* Anderson, W.A. & Borns, H.W., Jr. (eds), *Neotectonics of Maine. Augusta, Maine, Maine Geological Survey, Bulletin*, **40**, 43-67.

Thompson, W.B., Davis, T., Gosse, J.C., Johnston, R.A. & Newton, R. (1995). *Late Wisconsinan glacial deposits in the Portland - Sebago Lake - Ossipee Valley Region, south-western Maine*. Augusta, Maine, Maine Geological survey. Guidebook for 58th field conference of the north-eastern friends of the Pleistocene, 71pp.

Thompson, W.B., Fowler, B.K. & Dorion, C.C. (1999). Deglaciation of the north-western White Mountains, New Hampshire. *In:* Thompson, W.B., Fowler, B.K. & Davis, T. (eds), *Late Quaternary history of the White Mountains, New Hampshire and adjacent south-eastern Québec. Géographie physique et Quaternaire*, **53**, 59-77.

Voisin, D.T. (1998). Late Quaternary post-glacial history of Kaler's Pond, Waldoboro, Maine. Lewiston, Maine, Bates College, B.S. honors thesis, 85 p.

Walker, M.J.C. & Harkness, D.D. (1990). Radiocarbon dating the Devensian lateglacial in Britain: new evidence from Llanilid, South Wales. *Journal of Quaternary Science*, **5**, 135-144.

Walker, I.R., Mott, R.J. & Smol, J.P. (1991). Allerod-Younger Dryas lake temperatures from midge fossils in Atlantic Canada. *Science*, **253**, 1010-1012.

Walker, M.J.C., Coope, G.R. & Lowe, J.J. (1993). The Devensian (Weichselian) lateglacial palaeoenvironmental record from Gransmoor, East Yorkshire, England. *Quaternary Science Reviews*, **12**, 659-680.

Watts, W.A. (1980). Regional variation in the response of vegetation to lateglacial climatic events in Europe. *In:* Lowe, J.J., Gray, J.M. & Robinson, J.E. (eds), *Studies in the lateglacial of north-west Europe*. Oxford, Pergamon Press, 1-21.

Weddle, T.K., Koteff, C., Thompson, W.B., Retelle, M.J. & Marvinney, C.L. (1993). The late-glacial marine invasions of coastal central New England (north-eastern Massachusetts-south-western Maine): its ups and downs: Field Trip I. *In:* Cheney, J.T. & Hepburn, J.C. (eds), *Field trip guidebook for the North-eastern United States. 1993 Geological Society of America Annual Meeting, Boston, Massachusetts*, I-1 to I-31.

Weddle, T.K. & Retelle, M.J. (1995). Glaciomarine deposits of the Late Wisconsinan Casco Bay Sublobe of the Laurentide Ice Sheet. *In:* Hussey, A.M., II & Johnston, R.A. (eds), *Guidebook to Field Trips in southern Maine and adjacent New Hampshire. New England Intercollegiate Geological Conference, 87th Annual Meeting, Brunswick, Maine*, 173-194.

Wolfe, A. & Butler, D.L. (1994). Late-glacial and early Holocene environments at Pine Hill Pond, Newfoundland, Canada: evidence from pollen and diatoms. *Boreas*, **23**, 53-65.

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