

A new water-level history for Lake Ontario basin: evidence for a climate-driven early Holocene lowstand

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Received: 4 October 2010 / Accepted: 29 August 2011 / Published online: 8 October 2011
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Abstract Piston cores from deep-water bottom deposits in Lake Ontario contain shallow-water sediments such as, shell-rich sand and silt, marl, gyttja, and formerly exposed shore deposits including woody detritus, peat, sand and gravel, that are indicative of past periods of significantly lower water levels. These and other water-level indicators such as changes in rates of sedimentation, mollusc shells, pollen, and plant macrofossils were integrated to derive a new water-level history for Lake Ontario basin using an empirical model of isostatic adjustment for the Great Lakes basin to restore dated remnants of former lake levels to their original elevations. The earliest dated low-level feature is the Grimsby-Oakville bar which was constructed in the

western end of the lake during a near stillstand at 11–10.4 (12.9–12.3 cal) ka BP when Early Lake Ontario was confluent with the Champlain Sea. Rising Lake Ontario basin outlet sills, a consequence of differential isostatic rebound, severed the connection with Champlain Sea and, in combination with the switch of inflowing Lake Algonquin drainage northward to Ottawa River valley via outlets near North Bay and an early Holocene dry climate with enhanced evaporation, forced Lake Ontario into a basin-wide lowstand between 10.4 and 7.5 (12.3 and 8.3 cal) ka BP. During this time, Lake Ontario operated as a closed basin with no outlets, and sites such as Hamilton Harbour, Bay of Quinte, Henderson Harbor, and a site near Amherst Island existed as small isolated basins above the main lake characterized by shallow-water, lagoonal or marsh deposits and fossils indicative of littoral habitats and newly exposed mudflats. Rising lake levels resulting from increased atmospheric water supply brought Lake Ontario above the outlet sills into an open, overflowing state ending the closed phase of the lake by ~7.5 (8.3 cal) ka BP. Lake levels continued to rise steadily above the Thousand Islands sill through mid-to-late Holocene time culminating at the level of modern Lake Ontario. The early and middle Holocene lake-level changes are supported by temperature and precipitation trends derived from pollen-climate transfer functions applied to Roblin Lake on the north side of Lake Ontario.

Electronic supplementary material The online version of this article (doi:[10.1007/s10933-011-9551-8](https://doi.org/10.1007/s10933-011-9551-8)) contains supplementary material, which is available to authorized users.

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Keywords Lake Ontario · Laurentian Great Lakes · Holocene · Hydrologic closed-basin · Sediment stratigraphy · Pollen stratigraphy

Introduction

This study of the water-level history of Lake Ontario (18,960 km²), last in the chain of Laurentian Great Lakes draining to St. Lawrence River, employs a new method of interpretation in which the original elevations of past lake-level indicators and outlet sills are reconstructed independently to circumvent the distorting effects of on-going differential glacial rebound (Lewis et al. 2005). The relative altitudes of water levels and outlet sills can then be compared at any given time throughout the basin's lake history, and the effects of past dry climates which may have lowered lake level below outlets into a hydrologically-closed lowstand can be identified. The present study departs from the previous interpretative paradigm which assumed that abundant water supply always existed and lakes always overflowed their basin outlets. Under this paradigm, water-level indicators were connected to the overflow outlet in lake-level reconstructions, as in Karrow et al. (1961) and Anderson and Lewis (1985), with the consequence that closed-basin conditions (water level below outlet) could not be identified.

The purpose of this study is to re-examine piston cores from Lake Ontario (Anderson and Lewis 1985), many of which intersected shallow-water and terrestrial organic deposits under deep-water clastic clay sediment (Fig. 1; Tables 1, S1 in Electronic Supplementary Material [ESM]), to reconstruct Lake Ontario water-level history using the new method of interpretation, and to show that closed-basin conditions existed when early Holocene climate was known to be drier than present (Edwards and Fritz 1986; Edwards et al. 1996; Shuman et al. 2002).

Site description

The Lake Ontario basin (referred to here as Ontario Basin), comprising 82,990 km² (Government of Canada and United States Environmental Protection Agency 1995) is mostly underlain by Paleozoic sedimentary bedrock formations with Precambrian

metamorphic bedrock underlying its northern margin. The bedrock is mantled with Late Quaternary unlithified sub-glacial and glacial lake sediments. Erosion and transport of these materials have supplied sandy coastal deposits and offshore silty clay sediments in Lake Ontario. Today the length of the lake is 311 km in an east–west direction with widths ranging up to 85 km. The large main basin, which receives inflow from Lake Erie and upstream lakes via Niagara River, is 290 km long with a maximum water depth of 244 m in its eastern part. The smaller and shallower (57 m) Kingston Basin lies between the main basin and the outflowing upper St. Lawrence River (Fig. 1). The Adolphus Reach, North Channel, and Bay of Quinte are a northwest extension of Kingston Basin.

Several studies and major syntheses detail work undertaken on the bottom sediments of Lake Ontario. The surficial deposits of the lake are described in Thomas et al. (1972) and various aspects of the Quaternary sediments and water-level history of Lake Ontario basin are summarized in Lewis and McNeely (1967), Sutton et al. (1972), Lewis and Sly (1971), Sly and Prior (1984), Anderson and Lewis (1985) and Coakley and Karrow (1994). Pollen analyses were performed on a long core from the deep central basin of Lake Ontario by McAndrews (1972) and on embayment cores bordering Lake Ontario (Anderson and Lewis 1985; McCarthy and McAndrews 1988). Diatoms, ostracodes, pollen and oxygen and carbon isotope analyses formed a multidisciplinary approach (Duthie et al. 1996; Delorme 1996; Wolfe et al. 2000) to assess limnological changes of Hamilton Harbour revealing a low-level phase about 7,500 (8,300 cal) BP.

Materials and methods

Sedimentological procedures

Piston and gravity coring, in combination with seismic profiling, were originally carried out in Lake Ontario from 1966 to 1974 and the shipboard coring procedures are described in Thomas et al. (1972) and Anderson and Lewis (1985). The locations of the core sites and radiocarbon data are shown in Fig. 1; Tables 1, S4. The numerical order of the sites as discussed in Anderson and Lewis (1985) is maintained in Fig. 1; Tables 1, S4. The cores were cut lengthwise and sediment lithology and colour were described in

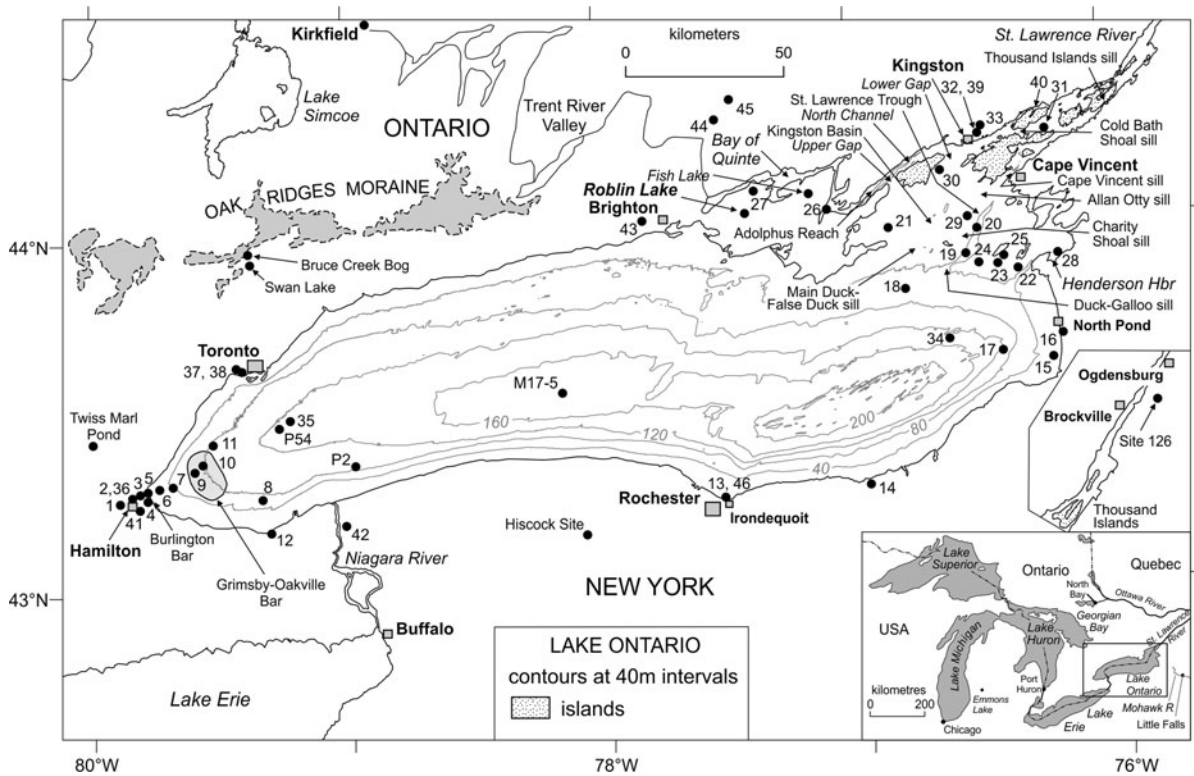


Fig. 1 Bathymetric map of Lake Ontario showing locations of cores (dots), outlet sills (arrows), islands (stippled areas), place names and other features discussed in text. Bathymetry from

Virden et al. (2000). *Upper inset* shows the upper St. Lawrence River area, Thousand Islands to Ogdensburg. *Lower inset* shows the Laurentian Great Lakes with rectangle outlining map

the laboratory (Table S1). For selected cores, water contents were measured immediately after opening up the cores and samples were removed for analysis of grain sizes, organic and inorganic carbon contents, calcite-dolomite ratios, pollen and plant macrofossil studies, and radiocarbon dating. Sand, silt, clay percents were determined using sieves and settling rates and carbon contents using the LECO IRI2 carbon determinator in the Sedimentology Laboratory of the Geological Survey of Canada. Inorganic C values multiplied by 8.3 gave % CaCO₃. Calcite and dolomite analyses were determined with the Chittick apparatus outlined in Dreimanis (1962).

Radiocarbon dating

Initially, bulk samples of basal peat and gyttja were dated at the former radiocarbon dating laboratory of the Geological Survey of Canada. For this study, the cores were re-examined for plant macrofossils and

mollusc shells for additional dating. Fossil seeds, wood, and mollusc shells were extracted for radiocarbon dating and for supportive paleoecological data according to techniques described elsewhere (Birks 1980, Anderson and Lewis 1985). The fossil seeds were identified with the aid of the seed reference collection at the Geological Survey of Canada. Appropriate seed and shell specimens were selected and carefully cleaned and forwarded to the AMS (atomic mass spectrometry) dating laboratory at the University of California, Irvine, California. Three modern mollusc shells from two localities in Lake Ontario basin were radiocarbon dated (Table S2) and the dates are used here to correct the conventional ages on the fossil shells (Tables 1 and S4) for hardwater effect (HWE). Consequently, 535 ± 15 years were subtracted from fossil shell and aquatic plant dates to account for HWE. The conventional and corrected radiocarbon ages were calibrated using the Calib 6.0.1 calibration program (Stuiver and Reimer 1993) and

Table 1 Core site locations and original elevations of newly-dated indicators of former lake levels (shallow water zones)

Site no/core no/region ^a	Lat N/long W	Lab no/ conventional date ^b	Dated material and stratigraphy	Elevation ^c / reference uplift m	1 σ -cal BP range ^d	Original elevation ^c	References
6 69-0-13-3 West LO 33.5 m	43°18.5'	UCIAMS74188	Wood pieces in silt and sand all under 5.9 m mud	35.4	3,265	33.3	Anderson and Lewis (1985), This study
	79°43.1'	3,080 ± 15		42.3	3,350	33.2	
8 69-0-13-13 West LO 71.6 m	43°16.2'	UCIAMS59738	<i>Pisidium con-ventus</i> shells in sandy clay 3.4 m depth in core	-0.2	9,000	-16.6	
	79°21.0'	8,640 ± 25		44.6	9,040	-16.9	
9 69-0-13-7 West LO 74.7 m	43°22.0'	UCIAMS74187	Wood pieces in silt and sand all under 3.9 m mud	0	10,875	-28.4	
	79°35.4'	9,940 ± 320		45.0	12,050	-39.7	
10 69-0-13-14/2 West LO 82.3 m	43°22.5'	UCIAMS74206	<i>Pisidium conventus</i> shells in silt and sand at 2.6-3.4 m depth in core	-10.6	9,150	-28.0	
	79°34.2'	8,820 ± 35		45.4	9,400	-29.3	
11a 69-0-16-1 West LO 99.4 m	43°26.0'	UCIAMS45365	Wood in 10-cm fossil zone at 3.1 m depth under mud	-30.95	12,605	-80.1	
	79°30.7'	10,775 ± 20		47.7	12,695	-81.4	
11b 69-0-16-1 West LO 99.4 m	43°26.0'	UCIAMS45366	Wood in 10-cm fossil zone at 3.1 m depth under mud	-30.95	11,305	-65.0	
	79°30.7'	9,965 ± 20		47.7	11,400	-66.0	
17 68-0-18-25 Mexico Bay 89 m	43°35.95'	UCIAMS59737	<i>P. conventus</i> shells in sand under 7.8 m silty clay	-22.0	10,420	-67.1	
	76°24.62'	9,805 ± 25		81.3	10,510	-68.3	
19d 67-0-18-28 East LO 51.8 m	43°53.6'	UCIAMS59736	<i>Pisidium</i> spp. shells in sand below clayey silt	17.8	12,450	-78.0	
	76°32.6'	11,115 ± 25		97.1	12,600	-82.1	
26a 69-0-16-2 Bay of Quinte 11.9 m	44°02.1'	GSC3441	Peat over sandy clay	59.0	8,605	20.1	Anderson and Lewis (1985)
	77°05.0'	7,920 ± 120		119.5	8,975	15.6	
26b 69-0-16-2 Bay of Quinte 11.9 m	44°02.1'	GSC3300	Gyttja over marl	59.1	6,795	36.8	
	77°05.0'	6,080 ± 100		119.5	7,155	34.1	
26c 69-0-16-2 Bay of Quinte 11.9 m	44°02.1'	GSC3460	Gyttja	59.7	5,335	46.1	
	77°05.0'	4,790 ± 80		119.5	5,600	44.8	
26d 69-0-16-2 Bay of Quinte 11.9 m	44°02.1'	GSC3464	Gyttja	60.8	3,720	54.5	
	77°05.0'	3,550 ± 80		119.5	3,960	53.7	
28 68-0-18-2 Henderson Harbor 11.6 m	43°53.68'	UCIAMS59648	<i>Potamogeton</i> seeds in shelly marl	60.1	10,515	2.9	This study
	76°11.13'	9,885 ± 25		100.3	10,640	0.8	
30b 74-00-102-20 20.4 m	44°09.3'	UCIAMS59649	Shallow-water seeds ^e under 1.29 m silty, clayey marl	53.1	9,310	8.2	
	76° 39.1'	8,350 ± 25		111.8	9,435	6.5	
31b 74-00-102-18A 20-21 m	44°14.87'	UCIAMS58061	<i>Scirpus validus</i> seeds at 1.51 m core depth	53	9,605	0.1	
	76°15.42'	8,725 ± 25		120.7	9,730	-1.9	
31c 74-00-102-18B 20-21 m	44°14.87'	UCIAMS58062	Wood twig in banded shelly marl at 1.3 m core depth	53	9,140	6.9	
	76°15.42'	8,255 ± 25		120.7	9,295	4.7	
34 74-00-102-26 Southeast LO 140 m	43°30.7'	UCIAMS58063	Wood in silty clay with sand beds	-74.8	12,740	-106.5 ^f	
	76°38.0'	10,970 ± 30		76.4	12,905	-110.4 ^f	
35 94800-26 West LO 135 m	43°29.8'	Beta92068	Small pelecypods in silty clay at 3.3 m core depth	-68.6	10,570	-99.3	
	79°10.3'	9,940 ± 60		53.1	10,710	-100.6	

 $\delta^{18}\text{O} = -2.5 \text{ PDB}$

Table 1 continued

Site no/core no/region ^a	Lat N/long W	Lab no/conventional date ^b	Dated material and stratigraphy	Elevation ^c /reference uplift m	1 σ -cal BP range ^d	Original elevation ^c	References
46c Irondequoit Bar	43°14.15' 77°31.92'	AA8637 9,300 ± 65	Peat in borehole 3 ^e	32.9 59.1	10,305 10,585	1.2 –1.5	RA Young (pers. commun.)
46d Irondequoit Bar	43°14.15' 77°31.92'	AA8635 5,590 ± 75	Organic sedi-ment, b'hole 3 ^e	35.5 59.1	6,305 6,435	26.3 25.9	Department Geological Sciences, SUNY College at Geneseo, Geneseo, NY, 14454
46e Irondequoit Bar	43°14.1' 77°32.08'	AA8630 4,530 ± 55	Peat in borehole 2 ^e	48.8 59.1	5,055 5,305	42.7 42.1	
46f Irondequoit Bar	43°14.1' 77°32.08'	AA8631 4,275 ± 55	Peat in borehole 2 ^e	39.4 59.1	4,710 4,955	34.1 33.6	
46g Irondequoit Bar	43°14.1' 77°32.08'	AA8629 3,635 ± 60	Peat in borehole 1 ^e	63.7 59.1	3,870 4,075	59.8 59.5	
46h Irondequoit Bar	43°14.1' 77°32.08'	AA8628 2,800 ± 55	Peat in borehole 1 ^e	66.9 59.1	2,800 2,965	64.5 64.3	
46i Irondequoit Bar	43°14.1' 77°32.08'	AA8627 2,415 ± 50	Peat in borehole 1 ^e	68.4 59.1	2,350 2,675	66.5 66.2	
46j Irondequoit Bar	43°14.1' 77°32.08'	AA8626 670 ± 50	Charcoal in borehole 1 ^e	70.1 59.1	560 670	69.8 69.7	

LO Lake Ontario

^a Present water depths in metres (m) below present mean Lake Ontario elevation of 74.75 m above present sea level (asl)

^b AMS dates reported in years BP (years before AD 1950)

^c Metres above present sea level for core top elevations at sites 6–35, and for sample elevations at site 46

^d Conventional ages of shells in column 3 were corrected by –535 year before calibration

^e *Scirpus validus*, *Decodon verticillatus*, *Chenopodium album* type, *Potentilla palustris*

^f Plot values; computed values are –156.5 (12,740 cal BP), 160.4 (12,905 cal BP)

^g Boreholes 1 and 2–43°14.1'N 77°32.08'W; borehole 3–43°14.15'N 77°31.92'W; top of all boreholes is 250 ft (76.2 m) asl, depth of samples (m) = 76.2-sample elevation (column 5)

the INTCAL09 dataset (Reimer et al. 2009). Means of the 1 σ age ranges were used where one-number calibrated ages were needed.

Pollen analysis

Samples for pollen analysis were treated with HF followed by acetolysis as described elsewhere (Anderson and Lewis 1985). Pollen percentage diagrams were created using TILIA and TILIA-GRAPH (Grimm 1993) from pollen sums of 200–250 tree, shrub and herb pollen excluding aquatic taxa.

Deductions of lake-level change from sediment, pollen, and radiocarbon data

Sites 26 and 28 are examples where one or more proxy indicators, in these cases physical properties and pollen, are combined to better demonstrate water-level

and lake-status changes and these are discussed in more detail below. Low-water levels at site 30 are mainly deduced from pollen supported by plant macrofossil presence of aquatic and marsh plants and these are elaborated further in the ESM section.

Pollen stratigraphies on the offshore sites 11, 17 and 19 provide examples of vegetative records for the glacial sediments and across the shell-rich, sand-silt zones into the overlying postglacial sediments. The shelly sand zones, intermixed in some cases with woody material and plant detritus, delineate a period of shallow-water deposition in a nearshore environment low down in the main basin and these are discussed in the ESM section.

Depth/age profiles showing changes in rates of sedimentation related to lake-level changes are based on radiocarbon dates at the sites. For sites lacking radiocarbon dates, depth/age data were determined from pollen correlation as follows: 11 and 10 (12.9 and

11.5 cal) ka BP—ages for the beginning and end of the *Picea* pollen peak correlated with the Younger Dryas (YD) cold interval (Shane 1987); 10.2 (11.9 cal) ka BP—increase in *Pinus* dominance dated 10,160 ± 100 (11,620–12,040 cal) BP (AA33214) at Emmons Lake, Michigan (LCK Shane pers. commun. 1999); 7.6 (8.4 cal) ka BP—beginning of the first peak in *Tsuga* at Roblin Lake (Terasmae 1980); 4.8 (5.5 cal) ka BP—beginning of the *Tsuga* minimum at site 26, Bay of Quinte (Anderson and Lewis 1985); 3.6 (3.8 cal) ka BP—end of the *Tsuga* minimum, beginning of the second peak in *Tsuga* at site 26, Bay of Quinte (Anderson and Lewis 1985).

Results

Physical and palynological evidence of low water levels

Here, the sequence of sediment facies from two embayments of Lake Ontario are described as examples to show substantive paleoenvironmental changes which indicate significant changes in former water levels. The examples are Site 26 in the Bay of Quinte on the north side of the Lake Ontario basin, and Site 28 in Henderson Harbor in the eastern end of the basin. In general, the evidence of low-water levels is pollen and plant macrofossils of shallow-water aquatic plants, and/or shallow-water sandy zones, commonly with mollusc shells, which are buried beneath deeper-water silty clay sediment.

Site 26 (Bay of Quinte) sediment and pollen facies

The sandy clay unit above laminated clay at the base of the site 26 core (Fig. 2a) is characterized by minimum organic and inorganic C and percent water content. However, organic C and percent water content begin to increase abruptly in the peaty marl unit and both parameters increase upward to maximum values in the overlying gyttja and remain relatively high throughout the gyttja unit to the top of the core.

The pollen record (Fig. 2a) begins with peak percentages of *Picea* (zone 1) which give way to *Pinus* dominance (zone 2) across the sandy clay into the peat units. *Pinus* then gives way to increases in *Quercus*, *Tsuga*, *Fagus*, *Acer* and pollen of other hardwoods.

The post-*Pinus* interval is distinguished by the fluctuating *Tsuga canadensis* profile (zone 3) represented as three separate subzones, the first *Tsuga* maximum (subzone 3a), the *Tsuga* minimum (subzone 3b) and the second *Tsuga* maximum (subzone 3c).

Site 26 (Bay of Quinte) inferred water level changes

The sandy clay unit was deposited during an interval of declining lake levels as the last glacial lake phase (laminated clay) receded out of Lake Ontario basin. The falling lake levels affected pollen deposition, likely by re-suspension, loss of sediment and pollen, and oxidation in shallow water, as evidenced by a decrease in pollen concentration and by the intervals of poor pollen preservation in the laminated clay and overlying sandy clay. By 10 (11.5 cal) ka BP Bay of Quinte and other sites in the eastern end of Lake Ontario became subaerially exposed as water levels fell into the main basin of Lake Ontario, resulting in little or no sedimentation.

Peat deposition (first woody peat, then marly peat) on the sandy clay means the site eventually became transformed into a wooded fen or marsh at or above lake level. Sudden increases in pollen of *Typha latifolia*, *Sparganium*, *Nymphaea*, *Myriophyllum* and taxa in the Poaceae, Cyperaceae and Ranunculaceae indicate the presence of marshy plants and shallow-water ponding. Even though deposition was extremely slow during deposition of the peat, the abrupt increase in pollen of *Pinus*, *Tsuga canadensis*, *Quercus*, *Ulmus*, *Fagus*, *Acer* and other hardwood taxa suggest the climate had warmed and became favourable for the proliferation of a pine-hemlock-mixed hardwoods forest in the immediate vicinity.

The upward change to peat with marl streaks and subsequently to gyttja indicates water levels had risen again and flooded the marshy habitat. This change is reflected by the truncated profiles of the marsh indicators *Typha latifolia*, *Sparganium*, *Ilex verticillata*, *Nymphaea* and taxa in the Poaceae, Cyperaceae and Ranunculaceae (391 cm depth level) and by abrupt jumps in percent water content and organic carbon (390 cm level) (Fig. 2a). The increase in Poaceae at 6,080 (6,795–7,155 cal) BP suggests the lake at this time was likely bordered by a grassy vegetated zone (possibly wild rice) which persisted through to late Holocene time.

Site 28 (Henderson Harbor) sediment and pollen facies

Analyses carried out at site 28 (Fig. 2b) include water content, organic and inorganic C, sand, silt, clay, calcite and dolomite measurements, and pollen. The basal laminated silty clay unit shows relatively low water contents and low or absent organic and inorganic C. The laminated silty clay is characterized by high percentages of silt and clay, with the clay size fraction increasing upward in the unit. The trends reverse in the overlying mottled silty clay, marl and black organic silt units above. Water contents and organic C increase upward reaching peak percentages in the marl and black organic silt units. Silt and sand as well as calcite and dolomite similarly increase upward to maximum values in the marl whereas clay drops to minimum values. The spike in silt and further reduction in clay fits with the description of the black organic silt unit. The plant detritus-rich silt and clay units show high but upward decreasing percents of water content and organic C. While silt and clay contents are relatively high, silt decreases whereas clay increases towards the top of the core.

Site 28 (Fig. 2b) shows a pollen zonation similar to that at site 26. A *Picea* dominance (zone 1) gives way to the *Pinus* dominance (zone 2) which coincides with the marl and black organic silt units. Like at site 26, the post-*Pinus* record is represented by the first *Tsuga* maximum (zone 3a), the *Tsuga* minimum (zone 3b), and the second *Tsuga* maximum (zone 3c).

Site 28 (Henderson Harbor) inferred water level changes

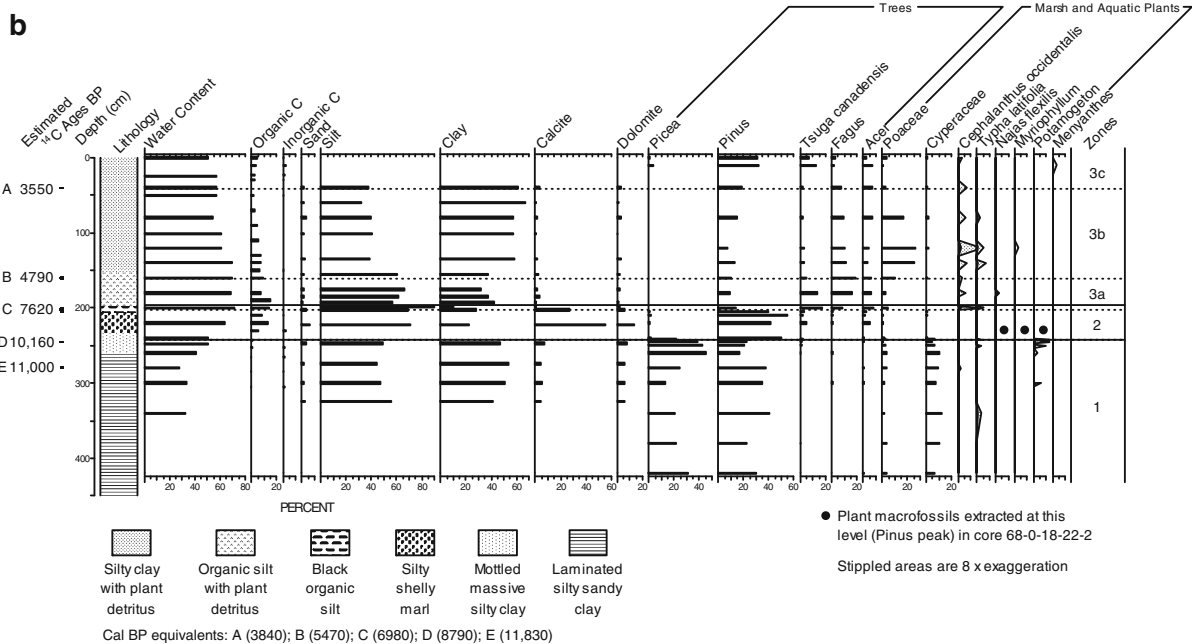
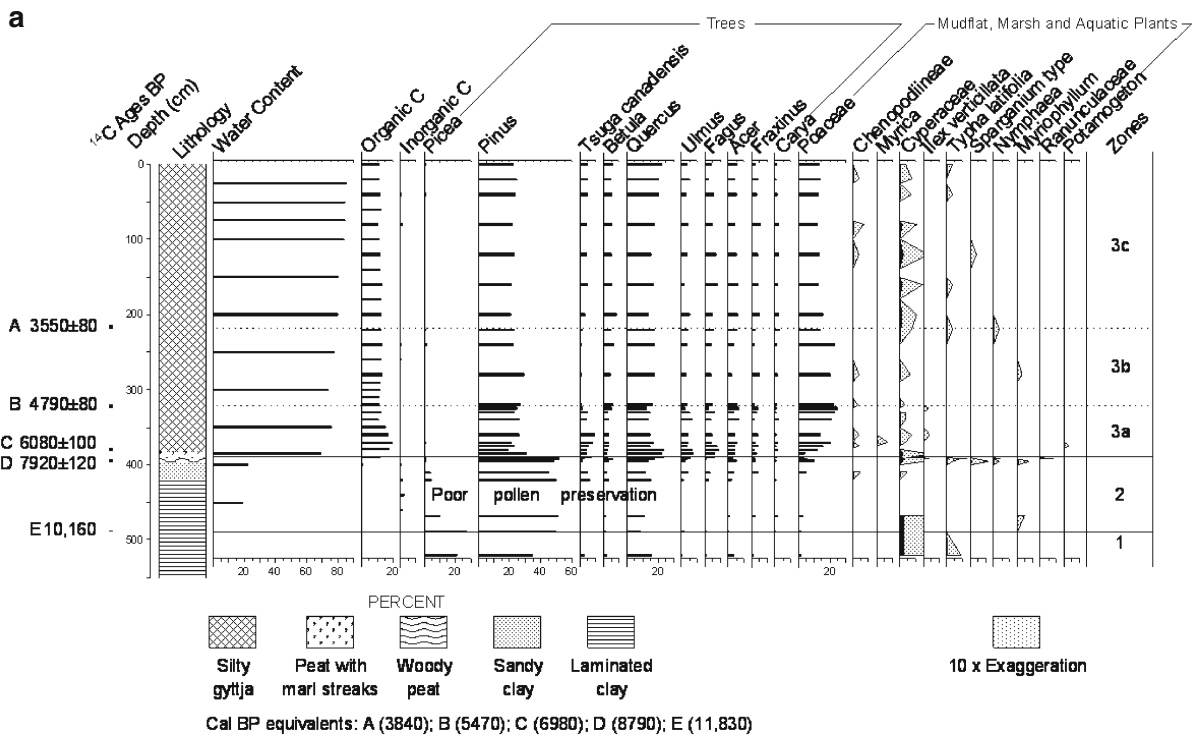
Above the laminated silty sandy clay, characteristic of deep-water glaciolacustrine deposition, the shell-bearing marl and overlying black organic silt units represent sedimentation during a prolonged low-level interval. The higher sand and silt percents in the marl and organic silt indicate shallow-water conditions prevailed with increased wave activity that winnowed out the clay leaving behind the coarser sand and silt. Some of the sand grains were dolomite in composition shown by the peak in dolomite, whereas the shell component in the marl is represented by the peak in calcite. The end of the low-level period is marked by a return to clay deposition in the overlying plant detritus-rich organic silt unit.

While sedimentation was continuous during deposition of the massive silty clay, overlying marl, and organic silt layer, sedimentation rates were severely reduced between 10,000 and 7,620 (11,470 and 8,400 cal) BP shown by the abrupt switch from high *Picea* to high *Pinus*, the compressed nature of the *Pinus* peak, and the equally abrupt replacement of *Pinus* by *Tsuga*, *Quercus* and other hardwoods which dominate to the present.

Breaks in the Poaceae, Cyperaceae, *Cephalanthus occidentalis* and *Typha latifolia* profiles across the *Pinus* maximum show that the high *Pinus* pollen masks out much of the other pollen. Shallow-water aquatics such as *Potamogeton* and *Nuphar* and marsh indicators like *Cephalanthus occidentalis* and *Typha latifolia* become almost indiscernible even though these taxa would have comprised part of the aquatic and nearshore flora. Seed occurrences of *Potamogeton* spp. and *Najas flexilis*, leaves of *Myriophyllum* spp. and an abundance of fibrous material in the shelly marl in a second core (68-0-18-22-2) from Henderson Harbor indicate the site resembled a shallow, highly eutrophic pond rich in submerged bushy pondweeds throughout the early Holocene. Yet the presence of the black organic silt layer implies the site may have dried up for a short interval (8–7.5 [8.9–8.3 cal] ka BP).

Rates of sedimentation

Depth-age curves revealing changes in rates of sediment accumulation through time were derived for 9 sites in Lake Ontario and one land site bordering Lake Ontario (Fig. 3). The chronology is based generally on radiocarbon dates, but where radiocarbon dates are not available, the chronology is based on pollen-derived ages tied to radiocarbon-dated pollen records on land. These curves show that at Lake Ontario sites of modest depth (11, 17, 26, 28, 30, 31), sediments accumulated slowly (<20 cm/kyear) in the interval 10.5–6 (12.5–6.8 cal) ka BP. These low sedimentation rates are interpreted to have occurred when falling lake levels brought the sites into an energetic coastal zone, e.g. <10–20 m water depth in Lake Ontario, through which suspended sediment was transported by currents to deeper and quieter water offshore. The deep offshore sites in Lake Ontario (M17-5, P2, P54), on the other hand, do not demonstrate evidence of early Holocene low sedimentation



rates suggesting the offshore sites were always inundated by relatively deep water and were never affected by energetic nearshore processes. The highest rates of sediment accumulation (>40 cm/year) occurred

during and following glacial lake phases in the Lake Ontario basin prior to 10.5 (12.5 cal) ka BP.

Sedimentation rates in the west end of the lake at site 11 (Fig. 3) are high (50 cm/year) in the

◀ **Fig. 2** Site examples of physical and palynological evidence of low water levels. **a, b** Figures show the plant indicators that denote changes in former water levels and tree taxa used for pollen zonation. The smallest percentage tick mark is 5%. The *solid lines* are the *upper and lower limits* of the low-level interval. **a** Sediment physical properties and pollen stratigraphy at site 26, located in outer Bay of Quinte at a water depth of 11.9 m (Fig. 1). The chronology is based on four radiocarbon ages and an estimated age of 10,160 (11,830 cal) BP for the *Picea-Pinus* pollen transition (LCK Shane pers. commun. 1999) making it one of the better dated sites. The date of $4,790 \pm 80$ (5,330–5,600 cal) BP for the *Tsuga* minimum matches almost exactly the recently determined, best estimate (4,750 [5,460 cal] BP) for the *Tsuga* decline in eastern North America (Bennett and Fuller 2002) suggesting that the other three dates are reliable. **b** Sediment physical properties and pollen stratigraphy at site 28 (core 68-0-18-22-1), located in Henderson Harbor, a bay of Kingston Basin, in the eastern end of Lake Ontario at a water depth of 11.6 m (Fig. 1). The chronology is based on pollen-derived ages shown on the left

laminated silty clay between 12 and 10.75 (13.9 and 12.6 cal) ka BP, extremely low (around 0 cm/kyear) between 10.75 and 9.9 (12.6 and 11.3 cal) ka BP in the stratified silt and sand sequence and slightly higher (7.5 cm/kyear) across the silt and sand into the overlying silty clay. Likewise, in the east end of the lake at site 17 (Fig. 3), sedimentation rates are high (109 cm/kyear) in the glacial laminated clay but fall off sharply to 5.8 cm/kyear in the shelly sand layer and increase slightly to 7.3 cm/kyear in the overlying massive silty clay.

Sites 26 and 28 show similar trends in the profiles of rates of sedimentation (Fig. 3). Both sites show similar sedimentation rates (30.5 and 35.8 cm/kyear, respectively) across the sediment interval equated with the Younger Dryas between 11 and 10 (12.9 and 11.5 cal) ka BP. Rates continued to be high in the post-YD interval at site 26 up to 8 (8.9 cal) ka BP, whereas at site 28 sedimentation rates decreased sharply (over 2-fold) to 11.4 cm/kyear between 10 and 8 (11.5 and 8.9 cal) ka BP. Both sites show low rates of sedimentation (8.3 and 3.1 cm/kyear) between 8 and 6 (8.9 and 6.8 cal) ka BP during deposition of the peat and overlying marly peat at site 26 and the marl and overlying black silt at site 28. Both sites show similar increases in rates of sedimentation after 6 (6.8 cal) ka BP corresponding with onset of gyttja deposition at site 26 (38.3 cm/kyear) and with silty clay deposition at site 28 (33.9 cm/kyear).

Site 30 became exposed and plant detritus started accumulating by 8,350 (9,370 cal) BP. Sedimentation rates (Fig. 3) averaged 4.0 cm/kyear between 8,350

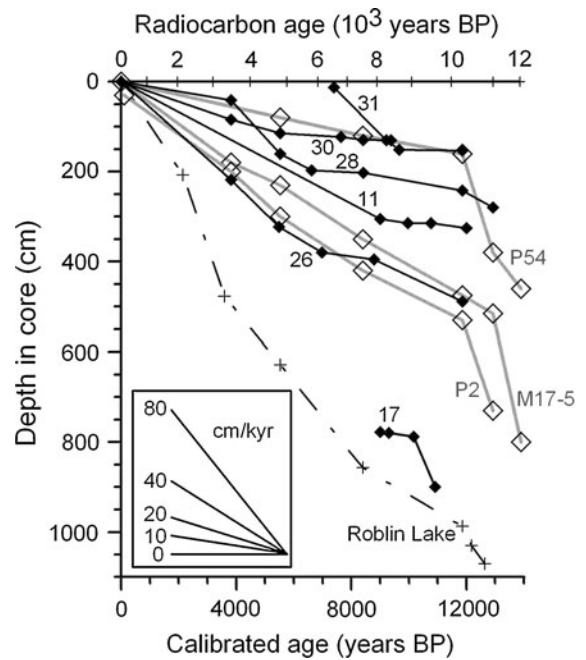


Fig. 3 Depth/age curves and rates of sedimentation (cm/kyear) at ten sites in and bordering Lake Ontario basin. Site locations shown in Fig. 1 and listed/described in Tables 1, S1, and S4. Note that relatively shallow offshore Lake Ontario sites (*solid diamonds* and *heavy lines*) all reveal low sedimentation rates (<20 cm/kyear) within the interval 10.5–6.0 (12.5–6.8 cal) ka BP, whereas deeper sites (*open diamonds* and *light lines*) and onshore Roblin Lake (*plus signs* and *broken line*) accumulated sediment throughout the Holocene at relatively constant and generally higher rates. The highest sedimentation rates (>40 cm/kyear) generally occurred during and following glacial lake phases in the basin

and 6,800 (9,370 and 7,640 cal) BP, during deposition of the plant detritus layer, but increased over fourfold to over 17 cm/kyear after 4,800 (5,540 cal) BP during deposition of the silty marly clay.

Sedimentation rates for site 31 (Fig. 3) were low (0.7 cm/kyear) between 10,000 and 8,725 (11,470 and 9,660 cal) BP, indicating that this site became exposed as an isolated sub-basin during the interval. A hiatus in sedimentation likely occurred between the time the site became exposed and initiation of peat growth at 8,725 (9,660 cal) BP. Sedimentation rates increased noticeably to 48 cm/kyear when peat deposition changed to marl after 8,725 (9,660 cal) BP.

By contrast, rates of sedimentation at site M17-5 (Fig. 1, core 68-0-17-5), a deep-water site in central Lake Ontario (McAndrews 1972), are steep (291 cm/kyear) prior to 10.2 (11.9 cal) ka BP but become practically linear with an average rate of sedimentation

of 44 cm/kyear after 10.2 (11.9 cal) ka BP (Fig. 3). Similar trends occur at core sites P2 and P54 (Fig. 1) in offshore western Lake Ontario (Pippert et al. 1995) where the sedimentation rates decrease substantially across the glacial-postglacial boundary and become noticeably linear (averaging 13 and 44 cm/kyear, respectively) between 10.2 (11.9 cal) ka BP and present (Fig. 3). The lack of a period of near-zero sedimentation rate in these cores implies that these sites were always in deep water and were little affected by reductions in lake level. For comparison, Roblin Lake shows a rate of sedimentation profile (Fig. 3) that is relatively constant throughout the Holocene, like the deep-water sites.

Interpretation of lake-level history

The original elevations of radiocarbon-dated water-level indicators (Tables 1 and S4) were plotted on a graph of original elevation versus age (Fig. 4). Original elevations (E_t) of indicators or sills were computed as their present elevation (E_p) minus their uplift since formation (indicators) or since any given age (sills), or $E_t = E_p - A(\exp(t/T) - 1)$ where the factor $A = (\text{reference uplift since } 12,500 \text{ cal BP}) / (\exp(12,500/3,700) - 1)$ and T (tau) = 3,700 years. The reference uplift for each site is provided by an isostatic response surface developed from the observed deformation of former shorelines (Lewis et al. 2005).

Former water levels in the main Lake Ontario basin were interpreted on this plot in accordance with the meaning of each plotted indicator. The interpreted water levels were connected to produce an inferred curve of lake-level history. Trajectories of possible original elevations for lake-outlet sills and undated indicators were computed (Tables S3 and S5) and plotted on the same graph to evaluate their relationship with the inferred lake-level history (Fig. 4).

Transition from glacial Lake Iroquois to early Lake Ontario

While ice blocked the St. Lawrence River valley, high-level glacial Lake Iroquois filled the Ontario basin and overflowed southeastward down the Mohawk and Hudson River valleys to the Atlantic Ocean (Muller and Prest 1985). The Iroquois level is defined here by

the elevations and ages of organic material recovered from its beaches and nearshore sediments in the western end and central northern side of the Lake Ontario basin (sites 41–45, Table S4 and star symbols on Fig. 4), and by the uplift curve (Table S5) for the Lake Iroquois outlet sill at Little Falls, NY (155 m asl, Fullerton 1980; Pair and Rodrigues 1993) in the Mohawk River valley (Figs. 1, 4). The intersection of uplift trends of these dates (excluding S4: 44–45) suggests Lake Iroquois ended about 11.7 (13.5 cal) ka BP. After this time, during the grey interval marked '3' on Fig. 4, lake level in Ontario basin fell through a series of lake stages (Frontenac, Sydney(?), Belleville-Sandy Creek, and Trenton-Skinner Creek) as the impounding Laurentide Ice Sheet (LIS) retreated from the St. Lawrence River valley (Miryneck 1962; Sutton et al. 1972; Clark and Karrow 1984; Muller and Prest 1985; Pair et al. 1988; Pair and Rodrigues 1993). The Sydney(?) shore deposits were reinterpreted as fluvial deposits of glacial Lake Algonquin drainage from the Lake Huron basin (Leyland 1984). The Belleville-Sandy Creek and Trenton-Skinner Creek levels were confluent with Fort Ann phases of Lake Vermont in the Lake Champlain basin (~230 km east of Ontario Basin) which overflowed to Hudson River valley, and the entire confluent water body throughout all basins is termed Lake St. Lawrence by Pair and Rodrigues (1993) and Lake Candona by Parent and Occhietti (1999) and Rayburn et al. (2007). Uplift curves for selected beaches of these lakes (Table S5) are shown on Fig. 4 by thin black lines. The figure shows that organic (peaty) sediments (x symbols labelled S4: 46a,b) beneath Irondequoit Bar near Rochester, NY, dated by Young and Sirkin (1994) correlate with the Trenton-Skinner Creek level.

About 11.0 (12.9 cal) ka BP, continuing LIS retreat in the lower St. Lawrence valley simultaneously drained Lake Trenton to the Early Lake Ontario level and allowed seawater to inundate the upper St. Lawrence River valley (Occhietti and Richard 2003; Richard and Occhietti 2005; Rayburn et al. 2007). The Champlain Sea, defined here by an uplift curve of one of its beaches in the upper St. Lawrence valley (Site 126 on Fig. 1; S5:6 on Fig. 4) was confluent with Early Lake Ontario. A near stillstand ensued in the basin as both sea level and basin were rising, though land uplift was somewhat faster. During this near stillstand, wave-drifted sand constructed a bayhead barrier bar across the western end of the basin between

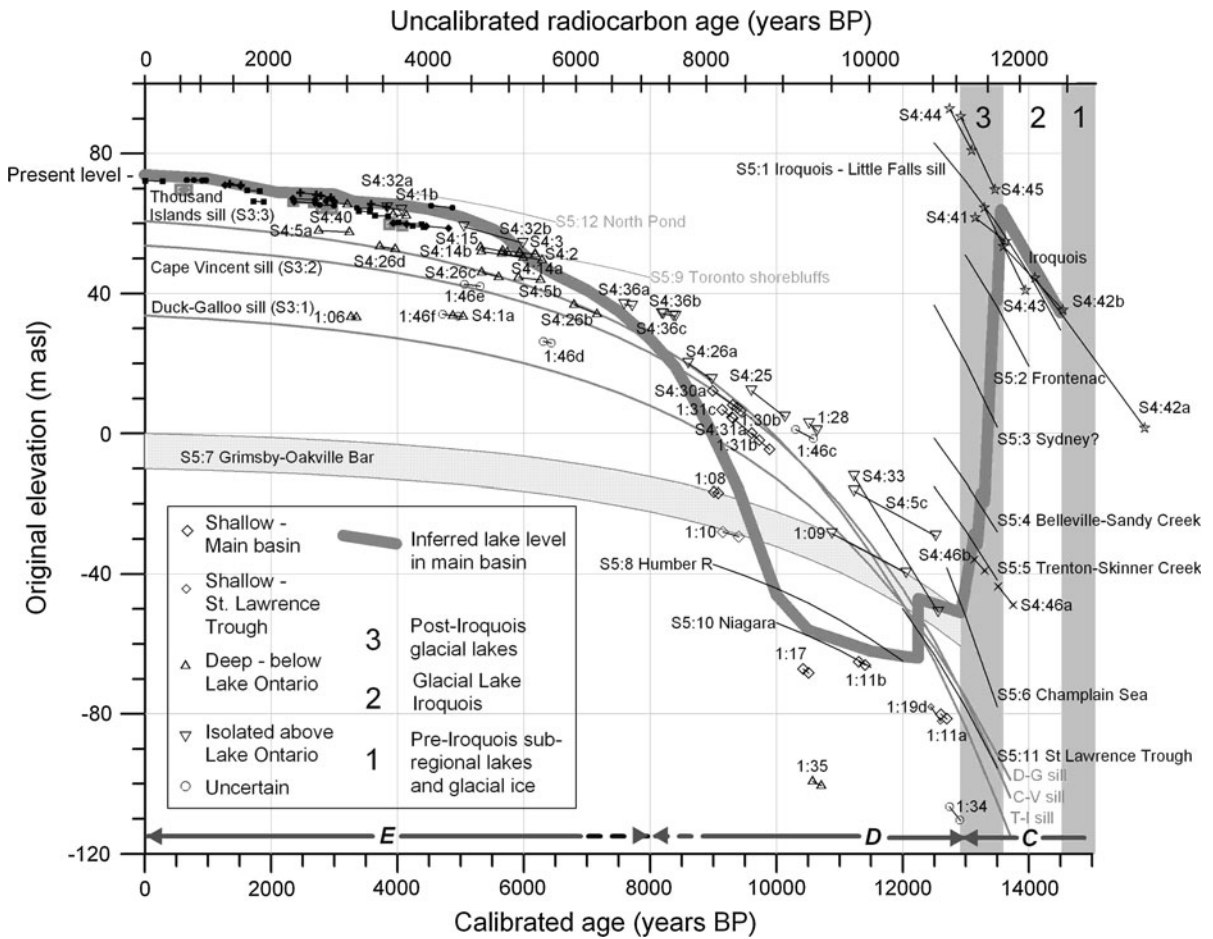


Fig. 4 Lake level (*thick grey line*) in the main Ontario basin inferred from the positions of lake-level indicators on a plot of original elevation (m above present sea level) versus age. Most plotted features are identified by table and site numbers. Indicators younger than ~4,430 (~5,000 cal) BP are identified by symbols: *black squares* (Table S4:38a–g), *black circles* (Table S4:12a–e), *black diamonds* (Table S4:37a–f), *black plus signs* (Table S4:39a–d), and *grey squares* (Table 1:46g–j). *Three grey lines* which span 0–11,900 (0–13,700 cal) BP trace the uplift history of outlet sills in the eastern Ontario basin and upper St. Lawrence Valley. Note the period of closed-basin conditions when lake level fell below sills during early Holocene dry climate. The now-submerged and mud-buried

Grimsby-Oakville bar (S5:7) was constructed by wave-drift across the western end of the basin by earliest Lake Ontario when it was confluent with sea level (Champlain Sea in St. Lawrence Valley). Radiocarbon-dated indicators of shallow-water environments are each plotted with symbols at end points of their calibrated age range; the slope of the line joining symbols reflects uplift during the age range. *Thin black lines* and *light grey curves* are partial uplift trajectories of undated or poorly-dated indicators. *Lettered arrows* at base indicate periods of deposition of offshore seismostratigraphic units (Hutchinson et al. 1993) which correlate with elements of the lake-level history

Grimsby and Oakville in the same manner as bars were formed earlier in Lake Iroquois at Hamilton, Ontario, and later by the present lake (Burlington Bar). The present elevation of the Grimsby-Oakville Bar which is now buried beneath silty clay mud of Lake Ontario is given as -5 ± 5 m asl by Lewis and Sly (1971). The Early Lake Ontario-Champlain Sea confluent water level is inferred to have remained near the elevation of

the bar for several centuries until the more rapidly uplifting sills at the head of St. Lawrence River valley raised the Ontario basin above sea level (Fig. 4). During the ‘stillstand’ period, precipitation and melt-water inflow from glacial Lake Algonquin in Huron basin supplied enough water, via Trent River valley and possibly via the Lake Erie basin, for the lake to resist penetration of Champlain seawater from about

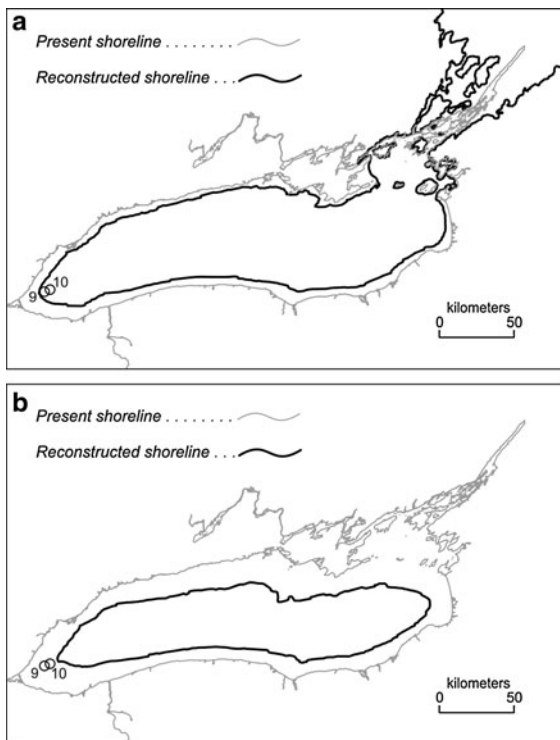


Fig. 5 Reconstructions of the Lake Ontario shoreline at **a** ~ 10.5 (~ 12.5 cal) ka BP and **b** ~ 9.3 (~ 10.5 cal) ka BP. In **a** early Lake Ontario is confluent with Champlain Sea which has inundated the St. Lawrence valley. At this stage, early Lake Ontario constructed the Grimsby-Oakville sand and gravel bar across the western end of its basin, sampled at core sites 9 and 10. In **b** Lake Ontario is at its lowstand, a hydrologically-closed lake with reduced water level such that its shorelines receded from the Grimsby-Oakville bar in the west (marked by sites 9 and 10) and from the lake overflow outlets to the northeast. Colour versions with more detail of these reconstructions are available as Figs. S5a and S5b in Electronic Supplementary Material

11.0 (12.9 cal) ka BP to about 10.4 (12.3 cal) ka BP. A reconstruction of the early Lake Ontario shoreline while confluent with Champlain Sea at about 10.5 (12.5 cal) ka BP is shown in Figs. 5a and S5a.

Closed basin conditions (10.4–7.5 [12.3–8.3 cal] ka BP)

At about 10.4 (12.3 cal) ka BP, rising Ontario Basin sills severed the connection with Champlain Sea. At the same time, Lake Algonquin drainage from Huron Basin switched from southern outlets (Kirkfield and Trent Valley route to Ontario Basin) to northern outlets (near North Bay, Ontario) (Eschman and

Karrow 1985). Loss of this meltwater supply combined with enhanced evaporation in the early Holocene dry climate (Edwards and Fritz 1986; Edwards et al. 1996; Shuman et al. 2002) forced lake level in the main Ontario basin to fall below the Duck-Galoo sill, as indicated by evidence of shallow-water conditions in a fossiliferous sand bed with pelecypod and ostracode faunas (Table 1: 11b), and driftwood (Table 1: 11a) at site 11 (Fig. 1). At the same time, water level fell in St. Lawrence Trough, a depression between the Duck-Galoo and Cape Vincent sills, indicated by shallow water at site 1:19d (Table 1; Fig. 4) where sand and silt accumulated with a pelecypod (*Pisidium conventus*) fauna. Wave action on this low water level could have formed the erosion surface (S5:11[Table S5, entry 11]) in the St. Lawrence Trough reported by Sly and Prior (1984).

Low water levels continued for about two millennia from 10.4 to 8.9 (12.3–10 cal) ka BP to account for a shallow-water sandy environment with a *P. conventus* molluscan fauna at site 1:17 (Table 1; Fig. 1). High evaporative stress on this low-level phase is suggested by a relatively high $\delta^{18}\text{O}$ isotopic composition of -2.5 per mil (PDB) in a pelecypod fauna at the deepwater site 1:35 (Table 1; Fig. 1), similar to $\delta^{18}\text{O}$ values in the closed phase of Lake Erie at the same time (Lewis et al., Lake levels in the Erie Basin of the Laurentian Great Lakes, submitted to Journal of Paleolimnology). During this phase the water surface was low enough to account for fluvial erosion of the Humber River channel (S5:8) (Table S5) south of Toronto (Lewis and Sly 1971), and wave abrasion of the low erosion surface (S5:10) (Table S5) north of the Niagara River mouth (Sly and Prior 1984). A reconstruction of the low-level Lake Ontario shoreline at about 9.3 (10.5 cal) ka BP is shown in Figs. 5b and S5b. A shallow-water sandy zone with a molluscan fauna at sites 1:10 and 1:8 requires lake level to rise quickly from about 9.2 to 8 (10.3–8.9 cal) ka BP though the water surface was still below outlet sills. Shortly after, the main and Kingston basin water levels were connected over Main Duck-False Duck sill (S3:4, not shown in Fig. 4), and by 7.5 (8.3 cal) ka BP the rising lake inundated and terminated peat and marl deposition in small basins in the upper St. Lawrence River bed and off Amherst Island (sites 30 and 31). During the rise in lake level, older peat (1:46c, 9.3 [10.5 cal] ka BP), possibly eroded from a deposit elsewhere, was buried in the Irondequoit Bar.

Transition to overflow conditions (7.5–4.5 [8.3–5.2 cal] ka BP)

Continuation of the rise brought Lake Ontario slightly above Cape Vincent and Thousand Islands sills into an open, overflowing state, ending the closed phase of the lake by about 7.5 (8.3 cal) ka BP. This rise in lake level is consistent with increasing precipitation and water supply after 7.6 (8.4 cal) ka BP, as inferred from pollen diagrams that indicate increasing moisture availability by the increasing presence of moisture-favouring trees such as hemlock (*Tsuga*), beech (*Fagus*) and maple (*Acer*) in the drainage basin (Fig. 6a; Shuman et al. 2002). A similar increase in moisture at this time is indicated by the $\delta^{18}\text{O}$ isotope climate proxy in cellulose (Edwards et al. 1996).

After 7.5 (8.3 cal) ka BP, the lake continued to rise gradually above its Thousand Islands sill to deepen water at the Bay of Quinte site (26b) where it converted the site from one of marl accumulation to gyttja sedimentation at about 6.0 (6.8 cal) ka BP. The rise continued on a similar trajectory to inundate peat environments in Mexico Bay (15) and Sodus Bay (14). During this part of the rise, the main lake connected with and began circulating into Hamilton Harbour

about 6.2 (7.1 cal) ka BP in accordance with the paleolimnological evidence from diatoms (Duthie et al. 1996), ostracodes (Delorme 1996), and carbon isotope data (Wolfe et al. 2000) at site 36.

Dated sites above the inferred early Holocene water level (Fig. 4), which are now submerged by the lake, were mostly small isolated basins above the main lake. The isolated basins include Bay of Quinte (S4:26a), a small basin near Galloo Island (S4:25), Henderson Harbor (1:28), Cataraqui River marsh near Kingston (S4:33), a small basin in the upper St. Lawrence River (S4:31a, 1:31b,c), and a site in Kingston Basin near Amherst Island (S4:30a, 1:30b). The wood date S4:5c also plots above lake level. This sample in silt and sand beneath Burlington Bar is interpreted here as detrital driftwood that was deposited elsewhere before being transported to and re-deposited in the Bar after 7.5 (8.3 cal) ka BP when lake level had risen above the elevation of S4:5c. Weathering of the till surface 4.6 m below S4:5c, which Karrow et al. (1961) attributed to subaerial exposure, could have occurred during the previous millennia, e.g. 11–8 (12.9–8.9 cal) ka BP when Lake Ontario was below the site. Dated wood samples (S4:5a,b) higher in Burlington Bar were deposited in Lake Ontario as originally suggested (Karrow et al. 1961).

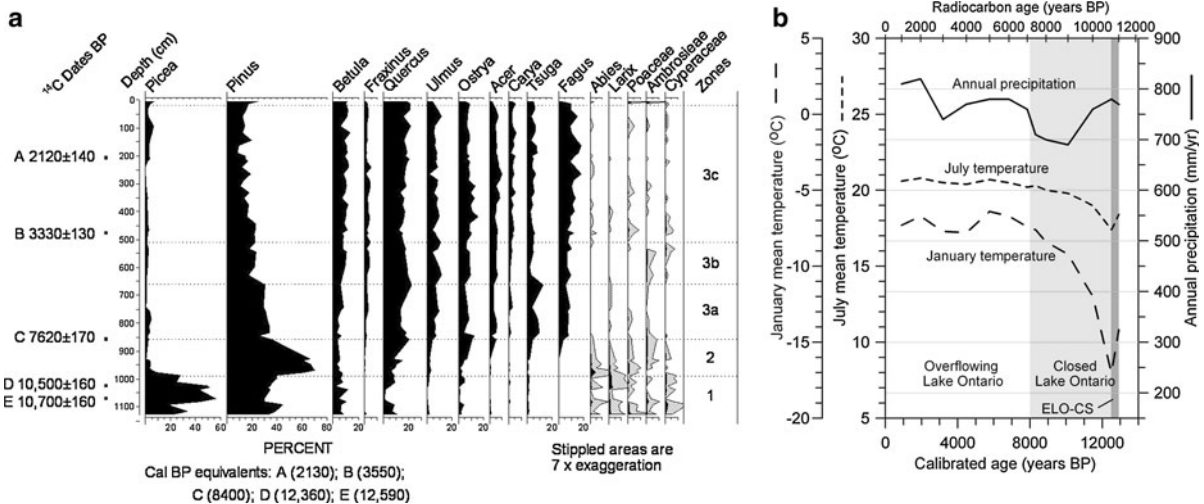


Fig. 6 Pollen stratigraphy (Fig. 6a) and paleo temperature and precipitation data (Fig. 6b) from Roblin Lake, located 25 km west of Bay of Quinte, at elevation 110 m asl (Fig. 1). Figure 6a shows the pollen profiles (redrawn from Terasmae 1980) used to derive the temperature and precipitation data courtesy of JH McAndrews (pers. commun. 2010). Estimated ages of the zone

boundaries are as follows: 1–2 (10.2 [11.9 cal] ka BP), 2–3a (7.6 [8.4 cal] ka BP), 3a–3b (4.8 [5.5 cal] ka BP), 3b–3c (3.3 [3.5 cal] ka BP). Figure 6b shows mean July and January temperatures and precipitation values superimposed on the lake status changes for Lake Ontario. ELO-CS = Early Lake Ontario-Champlain Sea

After the end of marl deposition in Bay of Quinte at site 26 and connection with Hamilton Harbour, Lake Ontario continued its rise above the outlet sills to flood and trigger the onset of silty clay mud deposition over peat deposits in Sodus Bay (S4:14a,b) and in the shallow area of Mexico Bay in southeastern Ontario basin (S4:15). This part of the lake rise from about 5.5 to 4.5 (6.3 to 5.2 cal) ka BP was likely caused by increased water supply from the upper Great Lakes as discharge from the Nipissing phase began to flow through Erie Basin into the Niagara River and Lake Ontario. The rise flooded Cataraqui River mouth (S4:32b), and continued to flood the Bay of Quinte (S4:26c,d), Hamilton Harbour and Cootes Paradise (S4:1a, 2, 3). The rise also raised water level above a series of notches or steps presently in about 15 m water depth on the lakebed south of Toronto summarized by Lewis and Sly (1971); these features (S5:9) may have formed around 5.2 (6 cal) ka BP as wave-eroded shorebluffs just prior to and during the rapid lake-level rise.

Modern Lake Ontario (4.5 [5.2 cal] ka BP to present)

Mean lake level throughout the late Holocene is shown at its present altitude (about 14 m) above the Thousand Islands sill, the highest of the three basin overflow controls. Interpretation of lake level during the final 4 millennia is based on data from the sedimentary records of Cataraqui River, near Kingston, Grenadier Pond and Humber River floodplain ponds in Toronto, and the Fifteen and Sixteen Mile Creek embayments between Hamilton and Niagara River (sites 12, 37, 38, 39 in Table S4), shown by the filled black symbols on Fig. 4, and from diatom and pollen evidence in a wetland site at Cootes Paradise (Finkelstein et al. 2005). The uppermost (and youngest) four peat and charcoal dates from Irondequoit Bar (1:46g–j in Table 1 and grey square symbols on Fig. 4) also track late Holocene lake-level rise, similar to the foregoing data. Three dates (1:46d–f) on peaty sediment recovered in borehole samples from Irondequoit Bar are positioned below the inferred lake-level curve. These positions are difficult to explain. Though material for dating was selected carefully from borehole samples (RA Young pers. commun.), possibly these materials were originally deposited at higher (younger) elevations and fell or were pushed down in

the boring process before being retrieved. The inferred lake-level generally follows the upper envelope of all other data points with solid black and open grey symbols in Fig. 4; low amplitude oscillations of water level are evident with minor peaks at about 5,000, 3,000, and 1,000 cal BP, similar to variations in western Lake Ontario presented by Flint et al. (1988), challenged by Bishop and Coakley (1989) but refuted by Flint et al. (1989), and presented again from eastern Lake Ontario by Dalrymple and Carey (1990).

Discussion

Relation of lake-level changes to regional climate

Independent climate reconstructions indicate that regional climate played a significant role in the history of water levels in Lake Ontario basin, as determined in this paper, particularly between 11 and 4.5 (12.9 and 5.2 cal) ka BP. For example, climate parameters derived by transfer function analysis (JH McAndrews pers. commun.) from the Roblin Lake pollen record (Fig. 6a) are consistent with the inferred lake level. Mean July and mean January temperatures increased abruptly by 1.4°C and 4.5°C, respectively, and precipitation decreased by as much as 70 mm/annum in the 10–7.5 (11.5–8.3 cal) ka BP interval corresponding to the closed phase or lowstand of Lake Ontario (Fig. 6b). Then, all three climate parameters (Fig. 6b) show parallel increases during the open-overflow phase 7.5–4.5 (8.3–5.2 cal) ka BP when increased precipitation of 70 mm/annum supplemented the rise in lake levels bringing an end to the closed phase.

Supporting evidence for an early Holocene dry interval exists at other sites in, north and south of Lake Ontario basin. The presence of sediment/pollen hiatuses dating just under 9 (10.1 cal) ka BP at Fish Lake, Ontario (JH McAndrews pers. commun.) and 8.7 (9.6 cal) ka BP at the Hiscock site, NY (Miller 1988) suggests these sites may have dried up at the peak of the dry interval.

The initial high levels of summer-precipitated calcite in eastern Lake Ontario sediment about 8.2–7.7 (9.2–8.5 cal) ka BP (McFadden et al. 2004, 2005) are consistent with the Roblin Lake pollen inferences that indicate summer (July) temperatures as warm or warmer than present even while winter

(January) temperatures were colder than present. Increased precipitation indicated at Roblin Lake after 7.7 (8.5 cal) ka BP may have increased soil stability by vegetative growth and decreased the flux of carbonate and other terrestrial components to the lake while lake productivity, indicated by declining C:N ratios, and increasing organic matter, C, and N, increased during the 8.4–4.6 (9.4–5.3 cal) ka period (McFadden et al. 2004, 2005). The permanent diversion of upper Great Lakes waters through Lake Erie and Niagara River by 4.6 (5.3 cal) ka BP together with late Holocene declines of July and January temperatures (Fig. 6b) could have been significant factors in the loss of carbonate, rise of biogenic silica, and changes in other components of eastern Lake Ontario sediments (McFadden et al. 2004, 2005).

The climate is described as being cold and dry between 10 and 7.5 (11.5 and 8.3 cal) ka BP and warm and dry between 7.5 and 6 (8.3 and 6.8 cal) ka BP (Edwards et al. 1996; Edwards and McAndrews 1989). Similar episodes of early Holocene dryness are known farther west in the Great Lakes region (Davis et al. 2000) and farther east in the St. Lawrence lowlands (Muller et al. 2003). The period of maximum dryness (10–6 [11.5–6.8 cal] ka BP) would have greatly enhanced evapotranspiration and evaporation in Lake Ontario basin which would have had the effect of maintaining low or minimum water levels, i.e., a lowstand.

The period of reduced lake levels in Lake Ontario coincided with a similar period in the Oak Ridges Moraine (Fig. 1) when lake levels were low or lakes had started to infill. For example, shallowing was evident at Swan Lake by 9–8 (10.1–8.9 cal) ka BP and intermittent to 6 (6.8 cal) ka BP, and Bruce Creek Bog transformed from a lake with marl deposition to a bog with peat deposition at 8.3 (9.3 cal) ka BP (Anderson 1997). A lowstand in Rice Lake dated 9.5–8.8 (11.1–9.6 cal) ka BP (Sonnenburg et al. 2011), east of the Moraine in the Trent River valley (Fig. 1), is also consistent with the Lake Ontario lowstand.

Revision of the Nipissing Flood

In 1982 and 1985 Anderson and Lewis accepted evidence of a ‘transgression’ in North Pond adjacent to eastern Lake Ontario which was inferred to have followed a lower-level ‘Dune’ stage dated about 4,810

(5,540 cal) BP. The ‘transgression’ accounted for a peat deposit about 1 m above the present lake (Sutton et al. 1972). Because the transgression was close in time to the return of upper Great Lakes drainage to the Erie and Ontario basins during their late Nipissing phase, and because it was supported by corresponding higher-than-present water levels in the Lake Erie basin dated $3,900 \pm 100$ (4,160–4,500 cal) BP (BGS-898) (Coakley and Lewis 1985), a higher-than-present water level was proposed for the Lake Ontario basin and termed the Nipissing Flood (Anderson and Lewis 1982). However, several later studies of late Holocene Ontario basin lake level have not confirmed the rise (Flint et al. 1988, McCarthy and McAndrews 1988; Weninger and McAndrews 1989; Dalrymple and Carey 1990). In addition, the higher-than-present water level in Lake Erie has now been related to a higher sill across the Niagara River which has since been eroded by recession of the Niagara cataract (Pengelly et al. 1997, Lewis et al., Lake levels in the Erie Basin of the Laurentian Great Lakes, submitted to Journal of Paleolimnology). Consequently, the present interpretation of Lake Ontario water level ignores the North Pond ‘transgression’ and presents a progressive rise in level once the upper Great Lakes drainage began passing through lakes Erie and Ontario ~5.0 (5.8 cal) ka BP. The term ‘Nipissing Flood’ should not imply a short-term peak in lake level, but should now just refer to this modest progressive rise and the increased flow through Lake Ontario. The uplift trajectory of the North Pond peat is shown 3–5 m above the inferred lake level between 5.7 and 3.7 (6.5 and 4 cal) ka BP on Fig. 4.

Correlation of lake-level history to style of offshore sedimentation

A synthesis of seismic reflection profiles and sediment cores led to recognition of distinct sediment units and a regional stratigraphic framework for the Ontario Basin (Hutchinson et al. 1993). In addition to bedrock surface, glacial diamict (unit A) and ice-marginal units (unit B), the synthesis identified three lacustrine units which correlate with the lake-level history inferred in this paper (Fig. 4). A well-laminated unit C with numerous parallel reflections of uniform thickness which drapes underlying topography was related to glacial lake deposition between 12.5 and 11.0 (14.3 and 12.9 cal) ka BP. Unit C ties well to period 3 in

Fig. 4 when glacial lakes Iroquois to Trenton occupied the basin. An overlying moderately laminated, silty clay unit *D* with basal sandy beds was attributed to early Holocene deposition; this unit correlates to low-level Lake Ontario and its rise to an overflowing state about 7.5 (8.3 cal) ka BP. The uppermost weakly laminated unit *E* is silty clay of variable thickness, thin on topographic highs, that relates to the modern style of sedimentation since the lake became an open, overflowing water body (Fig. 4).

Conclusions

1. From 10.4 to 7.5 (12.3–8.3 cal) ka BP, dried-up embayments in the Lake Ontario basin are indicated in sediment cores by intervals of peat, sand, and marl with shells; former nearshore zones deeper than present in the main basin are indicated by intervals of silt, sand and mollusc fossils. Both sets of low-water indicators are buried beneath ‘deep-water’ silty clay mud. These low-water indicators are supported by the presence of pollen and macrofossils of aquatic plants and species favouring recently dewatered surfaces, such as *Polygonum lapathifolium* and *Chenopodium album* type which Birks (1980) refers to as “drawdown species”. Sediment accumulation rates also reflect major changes in lake level. Pauses or reductions in sedimentation generally accompany sedimentary zones indicating former low-water or dried-up conditions, whereas cores which were always in relatively deep water revealed little change in sedimentation rate.
2. A new water level history was developed for Lake Ontario basin utilizing an empirical model of isostatic adjustment for the Great Lakes basin to restore and compare original elevations of dated indicators of former lake levels with original elevations of basin outlets. This approach revealed early Holocene periods when lake level fell below outlets into a hydrologically-closed condition.
3. Drainage of the last glacial lake phase in the Lake Ontario basin to sea level just after 11 (12.9 cal) ka BP resulted in a common water level for Early Lake Ontario and the Champlain Sea. During this time early Lake Ontario constructed the Grimsby-Oakville sand and gravel bar across its western end about 80 m below present lake level. Strong outflow, supported by meltwater from glacial Lake Algonquin, prevented the development of marine environments in the Lake Ontario basin.
4. A prolonged period of low-water levels and low rates of sedimentation between 10.4 and 7.5 (12.3 and 8.3 cal) ka BP occurred after cessation of meltwater inflow and severance from Champlain Sea by rising outlets. In this interval, lake levels fell below the Duck Galloo and Cape Vincent outlet sills in the eastern end of the lake and peat, woody detritus, and marl accumulated in separate sub-basins in the inter-islands area of eastern Lake Ontario; shell-rich sand zones accumulated in relatively energetic (shallow) littoral environments at low elevations in the main basin.
5. Lake Ontario basin operated as a closed basin with no outlets for up to ~4 kyear (10.4–7.5 [12.3–8.3 cal] ka BP); it was maintained at minimum levels under the influence of a dry climate with enhanced evapotranspiration and evaporation. The closed lake phase ceased after 7.5 (8.3) ka BP with the transition to higher lake levels and outflow over the Thousand Islands sill.

Acknowledgments We are grateful to G. Mackie (University of Guelph) and M. Smith (formerly of the National Museum Natural Sciences, Ottawa) for the mollusc identifications and to R. J. Mott (Geological Survey of Canada, Ottawa) for wood identifications. R.A. Young (State University of New York at Geneseo) kindly provided borehole logs and radiocarbon dates for bar sediments at the mouth of Irondequoit Bay, New York. Early radiocarbon determinations were provided by the former Radiocarbon Laboratory, Geological Survey of Canada, Ottawa. We thank Geological Survey of Canada Atlantic (GSCA) for support for later AMS radiocarbon dating, and J. H. McAndrews (University of Toronto) for deriving temperature and precipitation profiles using pollen transfer functions for Roblin Lake, Ontario and providing unpublished data from Fish Lake, Ontario. Grain size, carbon contents, and calcite and dolomite analyses were carried out in the Sedimentology Laboratory, Geological Survey of Canada, Ottawa. Dr. J-M. Gagnon, Canadian Museum of Nature, identified molluscs, provided habitat information, and contributed shell specimens for determination of the hardwater effect in Lake Ontario. We thank P. O’Regan (GSCA) for assistance with illustrations. We are grateful to reviewers for the Geological Survey of Canada and for the journal for their comments which have helped to improve the paper.

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