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Paleoproductivity of eastern Lake Ontario over the past 10,000 years

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Abstract

We evaluated relative levels of paleo–primary productivity in eastern Lake Ontario during the past ~10,000 yr via analysis of inorganic and organic sediment from the Rochester Basin. There was significant natural variability in primary production correlative with Holocene climate change. The cold post–Younger Dryas interval (~10–9.4 ka) was a time of minimal levels of primary production. The warm Holocene Hypsithermal interval (~9.4–5.3 ka) had much higher levels of primary production but was more variable, including five well-defined cycles that have an average period of ~750 yr. The largest negative anomaly in primary productivity occurred during the 8.2-ka climate event (~8.4–8.0 ka), a time of cold, dry conditions. Another negative anomaly occurred in association with the Nipissing flood (~6.3–5.3 ka), which triggered a regional cooling event. The cool Holocene Neoglacial interval (~5.3 ka to ~1850 A.D.) was characterized by lower, but more stable, levels of primary production, as well as by a cessation of calcite precipitation and the onset of diatom productivity. During the historic interval (~1850–1940 A.D.), there was a dramatic increase in primary production to unprecedented levels over the past 10,000 yr, as well as a 30-fold increase in sediment accumulation rates. These large, abrupt changes occurred in response to regional deforestation, anthropogenic nutrient loading, and increased chemical weathering due to acid rain. We project that, during 21st century global warming, eastern Lake Ontario will evolve into an ecosystem similar to that during the Holocene Hypsithermal.

The level of primary productivity in eastern Lake Ontario is of great importance to the region's present and future economy, including sport fishing, recreation, and water quality. There is not, however, a simple relationship between the level of primary productivity in the lake and various economic and environmental interests. Proper assessment of both present and future levels of primary productivity in Lake Ontario will be necessary for addressing scientific and management issues during the 21st century.

In a series of papers published during the 1980s–1990s, C. L. Schelske and co-workers addressed the question of primary productivity in Lake Ontario during historic time of the last ~150 yr (Schelske et al. 1983, 1986, 1988; Schelske and Hodell 1991; Hodell and Schelske 1998). They used short gravity and box core samples to document significant changes in Lake Ontario productivity due to anthropogenic activity. Although these studies provide considerable insight to the historical changes in primary production in Lake Ontario, they do not address the question of long-term natural variability. Thus, at present, we do not know whether historic changes in the level of primary production in Lake Ontario are unique or if natural variability is greater than or less than that observed during historic time.

The primary objective of the present article is to document the natural variability of relative primary production in eastern Lake Ontario over the past ~10,000 yr and compare it with data from the past ~150 yr. Our approach was to collect long sediment piston cores from eastern Lake Ontario (Fig. 1) and develop a multiproxy record of primary production using established methodology. Another objective is to determine whether the historic peak (1970s–1980s) in primary productivity in eastern Lake Ontario is unprecedented in the geologic proxy record. In addition, we can evaluate inferred levels of primary production during the Holocene Hypsithermal (~9.4–5.3 ka; thousands of yr ago), when summer temperatures in the northern hemisphere were warmer than today, perhaps providing insight on the potential effect of

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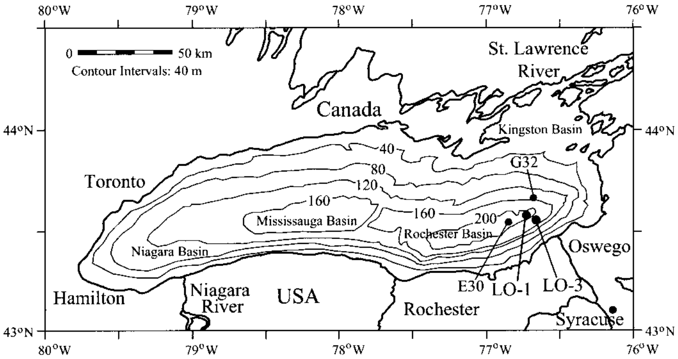


Fig. 1. Generalized bathymetric map of Lake Ontario illustrating the major deep-water basins, including the Rochester Basin to the east (based on Silliman et al. 1996). Also shown is the location of our piston cores (LO −1 and LO −3) in relation to box core G32 (Schelske and Hodell 1991) and piston core E30 (Silliman et al. 1996) discussed in text.

future global warming. We also have an opportunity to evaluate changes in paleoproductivity in response to abrupt climate change, such as the 8.2-ka and Nipissing flood events. Finally, we use these new data to broadly project the future level of primary production in eastern Lake Ontario for the 21st century.

Geologic, limnologic, and climatic setting

Lake Ontario (Fig. 1) is a hard-water (pH 8.0–8.6) lake located in a large, glacially scoured basin of Upper Ordovician shale and limestone (Hutchinson et al. 1993). The lake's watershed consists of Paleozoic carbonate bedrock to the south and Precambrian crystalline rocks to the north (Sly 1991).

Today, Lake Ontario has a surface area of ~19,450 km^2 and an average water depth of 86 m (Hutchinson et al. 1993). Two bathymetric ridges divide the lake into three main, deep-water basins, where mostly fine-grained sediment accumulated during the Holocene (Fig. 1; Carmichael et al. 1990). The Rochester Basin is the deepest, with a maximum water depth of 244 m. Water inflow to Lake Ontario is dominated by the Niagara River, which carries water and dissolved constituents eastward from the upper Great Lakes, whereas outflow is to the northeast via the St. Lawrence River (Sly 1991). Nutrient supply to the lake comes from the upper Great Lakes as well as chemical weathering in the watershed and numerous anthropogenic sources (Government of Canada and USA 1995). Lake Ontario is sensitive to environmental change, because it has a relatively small surface to watershed area ratio (~1:3, excluding the upper Great Lakes) and a water residence time of only ~8 yr (Meyers 2003). As a warm monomictic lake, Lake Ontario stratifies thermally only once per year, beginning in late spring and extending into early fall (Government of Canada and USA 1995).

Mean annual atmospheric precipitation on Lake Ontario and its watershed is ~95 cm yr^{-1}, with average monthly air temperatures ranging from −5°C in winter to +21°C in summer (Government of Canada and USA 1995). Weather conditions in the region are affected by thermal moderation of the lake as well as the polar front jet stream, which controls the advection of atmospheric air masses, storm tracks, and sources of water vapor (Eichenlaub 1979).

Materials, methods, and approaches

Sample collection—In June 2000, sediment piston cores (up to 5 m long, 3 cm diameter) were collected from eastern Lake Ontario (LO) using Hobart and William Smith's R/V *Explorer* (Fig. 1). Cores were collected in plastic liners, cut into sections, labeled on-board ship, and transported to nearby Syracuse University where they were stored in a cold room at 4°C and 100% humidity.

Magnetic susceptibility and loss on ignition—Whole core sections were analyzed for magnetic susceptibility at 3.5-cm intervals using a Barrington MS-2 magnetic susceptibility meter ring system, with values reported in SI units. Magnetic susceptibility values are largely dependent on the relative amount of terrestrial magnetic minerals present in sediment, although diagenetic overprints may be present (Gale and Hoare 1991).

Core sections were then split longitudinally, visually described, and subsampled at 2-cm intervals for detailed analyses. The loss-on-ignition (LOI) method at 550°C and 1,000°C (Dean 1974) was used to determine total weight percent organic matter (%TOM) and total weight percent carbonate (%TC), respectively. The LOI method is limited at low values of %TC, however, because clay-rich sediments may loose 3–4 wt. % water from the crystal lattice of clay minerals when they are heated to 1,000°C (Dean 1974). We consider any %TC value <5 to represent "zero" total carbonate.

%TOM values are a reflection of the amount of organic matter (OM) produced within a lake basin as well as terrestrial OM washed in from its drainage basin, minus decay (Dean 1999; Meyers and Lallier-Verges 1999). About 90% of Great Lake OM is autochthonous (Meyers 2003).

The precipitation of calcite in hard-water lakes is controlled by a number of complex chemical, physical, and biological processes (Kelts and Talbot 1990); the saturation state of surface waters with respect to calcite ($CaCO_3$) and temperature are the most important (Hodell et al. 1998; Lajewski et al. 2003). A high level of primary production by picoplankton may also be a critical factor (Hodell et al. 1998).

On the basis of results of visual descriptions, magnetic susceptibility and LOI, it was determined that a composite of cores LO −1 (~210 m water depth) and LO −3 (~180 m water depth) would maximize the time interval sampled. The two cores are spatially close (Fig. 1) and can readily be correlated on the basis of LOI data (Fig. 2). This "spliced" core (LO −1/3) has a total length of 514 cm and consisted of the basal 112.5 cm of core LO −3 added to the entirety of core LO −1.

Biogenic silica (BS)—Percent BS analyses were conducted on bulk sediment samples taken at a 4-cm interval using a photospectroscopy time-series dissolution method (De-

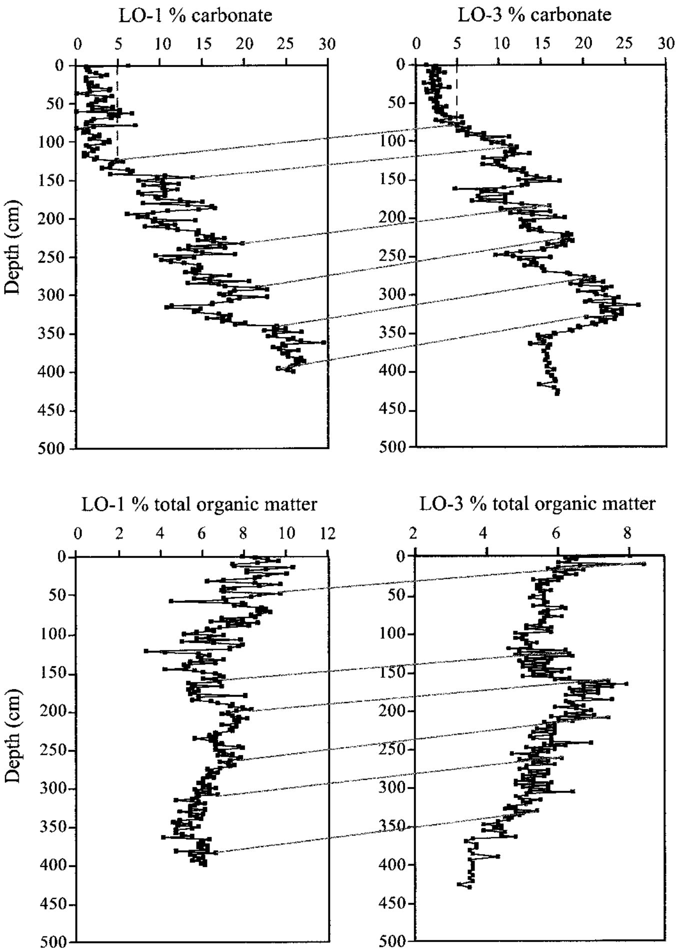


Fig. 2. Correlation of piston cores LO −1 and LO −3 (Fig. 1), based on LOI data, to develop composite ("spliced") core LO −1/3 used in the study. Note that correlation was done on a subsurface depth scale.

Master 1981). Sediment samples were first leached with 0.1 mol L^{-1} sodium hydroxide solution at 85°C for 5 h. Hourly aliquots were then taken and analyzed for silica using the reduced molybdosilicic acid spectrophotographic method of Strickland and Parson (1968). A linear regression of the time series, extrapolated to time zero, was used to correct for inorganic silica interference. Sediment samples collected every 50 cm were also examined by a scanning electron microscope (SEM) to qualitatively determine relative abundance of diatoms and degree of preservation.

Total nitrogen, organic carbon, and C : N ratios—Percent total nitrogen (%N) and percent organic carbon (%C) were measured using an elemental analyzer (EA) at the SERC Stable Isotope Laboratory at Florida International University. Subsamples (2 cm interval) were evaluated separately for nitrogen and carbon. Samples for %C were pretreated with 1 mol L^{-1} HCL for 24 h at room temperature, to remove inorganic carbonate. Molar ratios were determined using the formula

$$C_{org}/N_{total}\ \text{Ratio} = (\%C/12.01)/(\%N/14.002) \qquad (1)$$

C : N values largely reflect the ratio of autochthonous OM (primarily algae) to terrestrial OM (primarily vascular plants; Talbot and Lærdal 2000). Phytoplankton have low C : N ratios (4–10) because of low levels of carbon-rich substances, such as cellulose, whereas vascular plants typically have C : N ratios >20 (Meyers and Lallier-Verges 1999). Historic phytoplankton in Lake Ontario have an average C : N ratio of ~8 (Kemp et al. 1972). C : N values between 10 and 20 usually reflect a mixture of aquatic and terrestrial OM (Talbot and Lærdal 2000).

Stable isotope values of carbon and nitrogen in OM—Carbon and nitrogen isotope values of bulk OM were determined by standard EA isotope ratio mass spectrometry procedures at Florida International University. Elemental composition and gases for isotopic analysis were obtained using an EA to combust OM to form both N_2 and CO_2 gases, which were then measured on a Finnigan MAT Delta C mass spectrometer in continuous flow mode. Results (‰) are reported in standard delta (δ) notation relative to international standards of atmospheric nitrogen and the Vienna Peedee Belemnite (VPDB) for carbon, using the formula

$$\delta^{13}C_{OM} \text{ or } \delta^{15}N_{OM} = (R_{sample} - R_{standard}/R_{standard}) \times 1000 \qquad (2)$$

where R is $^{13}C/^{12}C$ or $^{15}N/^{14}N$.

Changes in primary productivity are recorded by the carbon isotope value of OM (Hodell and Schelske 1998) because of the preferential uptake of ^{12}C during photosynthesis (McKenzie 1985). During times of high primary production, ^{12}C becomes depleted from the dissolved inorganic carbon pool and $\delta^{13}C_{OM}$ values increase. For Lake Ontario, Hodell and Schelske (1998) found modern $\delta^{13}C_{OM}$ values to be a reliable proxy, in that increased primary production correlated with higher $\delta^{13}C_{OM}$ values.

$\delta^{15}N_{OM}$ values vary as a function of the source of OM, the level of primary production, degree of diagenesis, and the isotope value of dissolved inorganic nitrogen (DIN) in a lake (Talbot and Lærdal 2000). During times of high primary production, $\delta^{15}N_{OM}$ values increase because ^{14}N concentrations decrease by preferential uptake during photosynthesis. Terrestrial OM has low nitrogen isotope values ($\delta^{15}N \sim +0.5$‰ for C_3 land plants), compared with aquatic OM ($\delta^{15}N \sim +8$‰). Relatively high $\delta^{15}N_{OM}$ values are commonly indicative of increased levels of primary production, although changes in DIN or denitrification may also increase $\delta^{15}N_{OM}$ values. Relatively low values of $\delta^{15}N_{OM}$ reflect lower levels of primary production, an increase in nitrogen fixation by cyanophytes, an increase in the flux of terrestrial OM, or changes in the composition of DIN (Talbot and Lærdal 2000).

Stable oxygen and carbon isotope values of calcite—Oxygen and carbon isotope values of calcite were determined on fine-grained (<63 μm), nonskeletal sediment at a 2-cm interval. Analyses were conducted on a Finnigan MAT 252 gas-ratio mass spectrometer directly coupled to a Finnigan Kiel-III carbonate preparation device at Syracuse University. Prior to analysis, samples were roasted in vacuo at 200°C for 1 h to remove volatile organic components and water. Samples were then reacted with 2 drops of 103% phosphoric acid at 50°C for 90 s, to avoid reaction with dolomite. All samples were corrected for acid-water fractionation effects, ^{17}O contribution, and temperature fractionation. Results (‰) are reported in standard delta (δ) notation relative to the VPDB international standard using the same equation as Eq. 2, except that $R = {}^{18}O/{}^{16}O$ or ${}^{13}C/{}^{12}C$. NBS-18 and NBS-19 powdered carbonate standards were used to monitor calibration of the data.

Oxygen isotope values of calcite are controlled by temperature and the isotopic composition of ambient waters (Kelts and Talbot 1990). Although $\delta^{18}O_{calcite}$ values are not directly influenced by changes in primary production, they are useful in defining climatic and/or environmental changes over time that may in turn affect primary production. A variety of factors can result in increased $\delta^{18}O_{calcite}$ values: (1) a decrease in water temperature, (2) an increase in the ratio of summer to winter precipitation, (3) increased evaporation, or (4) a lower latitude source of water vapor. A decrease in $\delta^{18}O_{calcite}$ values may result from the opposite of the factors listed above and/or the advective influx of water with low $\delta^{18}O$ values.

$\delta^{13}C_{calcite}$ values reflect the isotope value of dissolved inorganic carbon (DIC) in a lake, as well as biological fractionation during photosynthesis (McKenzie 1985; Hodell et al. 1998). As primary production increases, the DIC pool in the epilimnion becomes depleted in ^{12}C, which results in *higher* $\delta^{13}C_{calcite}$ values. Carbon isotope values of calcite have been successfully used to infer changes in primary production (McKenzie 1985), including Lake Ontario (Schelske and Hodell 1991). However, $\delta^{13}C_{calcite}$ values may also be influenced by carbon preservation during burial as well as concentrations of carbon dioxide (Hollander and McKenzie 1991).

Age model—Soon after core collection in 2000, 10 bulk sediment samples were sent to the University of Arizona for radiocarbon analysis of OM. When results were received, however, there were a number of age reversals and excessive ages that rendered our radiocarbon data suspect in establishing an age model. A review of the literature discovered that, in the Great Lakes, even small amounts of reworked OM from its glaciated watershed can produce radiocarbon ages that may be thousands of years too old (Colman et al. 1990; Silliman et al. 1996; Moore et al. 1998). Because of this, analysis of biogenic carbonate produces more reliable radiocarbon ages in the Great Lakes (Colman et al. 1990), provided that hard-water effects are taken into account (Rea and Colman 1995; Moore et al. 1998).

An age model for composite core LO −1/3 was established by correlation with previously dated cores from the Rochester Basin (Fig. 1). We correlated results from the top of core LO −1/3 (%TC, %BS, $\delta^{13}C_{OM}$, and $\delta^{15}N_{OM}$) with box core data (Schelske and Hodell 1991; Hodell and Schelske 1998; Hodell et al. 1998) previously dated by ^{210}Pb and ^{137}Cs techniques (Fig. 3). We also correlated core LO −1/3 with Silliman et al.'s (1996) core (radiocarbon dated using ostracod shells) via %TC, %TOC, and C : N ratio data (Fig. 4). Radiocarbon dates were corrected for hard-water effects by subtracting 500 yr (Moore et al. 1998). The %TC also drops to <5% at a depth of 124 cm in core LO −1/3, similar to that in Lake Erie (Fritz et al. 1975), as well as the nearby Finger Lakes (Fig. 5; Anderson et al. 1997; Mullins 1998), where it has been radiocarbon dated. All radiocarbon ages correlated to core LO −1/3 were then converted to calendar years using the CALIB 4.2 program of Stuiver et al. (1999). On the basis of this age model, core LO −1/3 extends from ~10,000 calendar years near its base to ~1940 A.D. at its top (Fig. 6).

Results

Sediment type versus depth—Core LO −1/3 contains exclusively fine-grained sediment. The basal 74 cm of the core consist of relatively firm, massive, gray-brown mud with centimeter-scale, oxidized, rip-up clasts (Fig. 7). The remainder of the core is made up of brown mud with millimeter- to centimeter-scale black laminations that oxidize after a few hours of exposure (Fig. 7). The black laminations are likely iron/sulfide-rich bands that formed at times of low oxygen conditions near the bottom of Lake Ontario, similar to those recently reported from Lakes Michigan and Huron (Odegaard et al. 2003).

LOI data reveal that %TOM generally increases up-core from a minimum value near the base to the top of the core, with a significant negative shift or "step" at a depth of ~200 cm (Fig. 7). The %TC values are highly asymmetric; the maximum value occurs at a depth of ~350 cm, with values dropping to "zero" at a subsurface depth of ~124 cm and continues to near the top of the core (Fig. 7).

Proxy data versus age—Using the age model in Fig. 6, we have converted subsurface depth in core LO −1/3 to calendar years. We have also superposed the major paleoclimate intervals and events of the Holocene (post–Younger Dryas climate interval, the Holocene Hypsithermal, the 8.2-ka cold event, the Nipissing flood, the Holocene Neoglacial, and the historic interval) for ease of temporal reference.

Magnetic susceptibility is low in core LO −1/3, which is not unusual for fine-grained, deep-lake sediment that is relatively rich in OM and carbonate. Overall, there is an up-core decrease in magnetic susceptibility (Fig. 8), which suggests a gradual reduction in the relative percentage of magnetic minerals, either as a result of a decreased flux or of dilution by other components. Peak values of magnetic susceptibility occur near the base of the core within the post–Younger Dryas climate interval and in association with the 8.2-ka cold event. An increase in magnetic susceptibility also occurs at the very top of the core (Fig. 8).

C : N ratios display a similar up-core decrease as magnetic susceptibility (Fig. 8). Highest values occur as single data points in the lower section of the core, whereas most of the core has C : N values < ~13; values drop to <10 at ~5.3 ka B.P. Negative anomalies occur in association with the post–Younger Dryas interval as well as the 8.2-ka event, and a positive anomaly occurs near the top of the core (Fig. 8).

The curves of %C and %N versus time (Fig. 8) are similar. Both display increases from the post–Younger Dryas into the Hypsithermal, and small negative anomalies associated with the 8.2-ka event. The most prominent feature of the %C and %N data is the abrupt shift ("step") toward lower

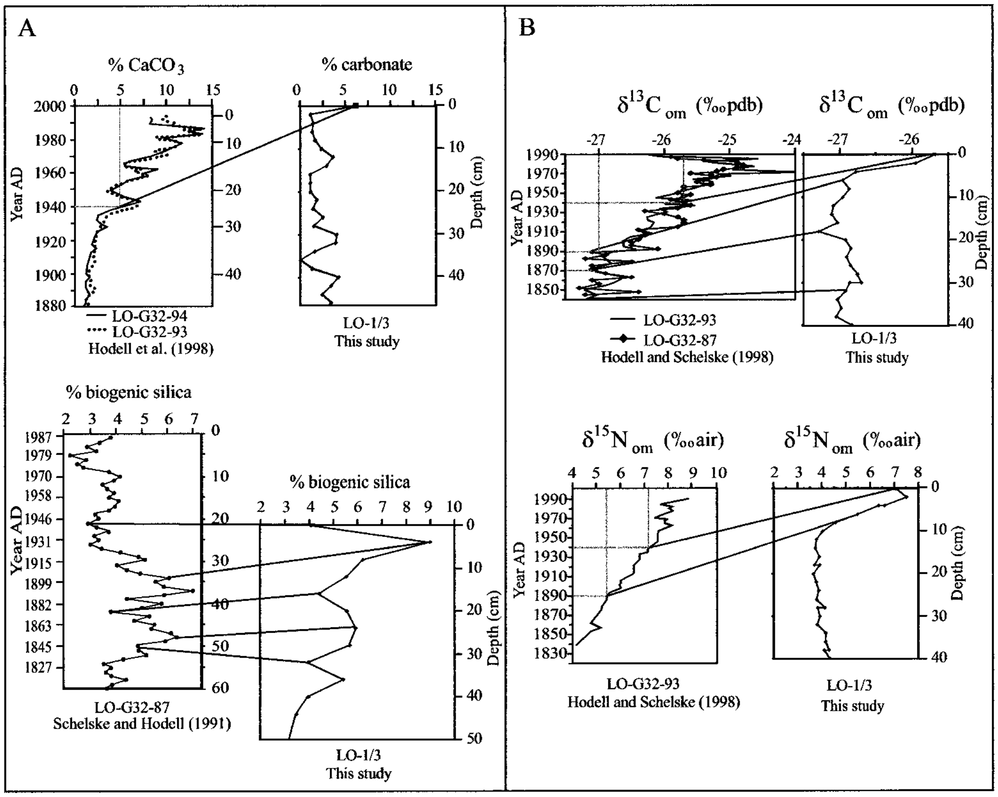


Fig. 3. (A) Correlation of core LO −1/3, based on %TC and %BS data, with previously dated box cores of Schelske and Hodell (1991) and Hodell et al. (1998) used to construct age model for the historic interval. (B) Correlation of core LO − 1/3, based on $\delta^{13}C_{OM}$ and $\delta^{15}N_{OM}$ data, with previously dated box core of Hodell and Schelske (1998) used to construct age model for the historic interval.

values that occurs in association with the Nipissing flood. Following this step, %C values are fairly constant until historic times, when values increase to their maximum over the past 10,000 yr (Fig. 8). %N values display a similar trend to %C data ($r = 0.82$), with a maximum value at the top of the core (Fig. 8).

Percent BS data versus time are characterized by three very distinct parts (Fig. 9). During the post–Younger Dryas and Hypsithermal intervals, %BS values are typically <1%, which suggests either very low levels of diatom productivity or dissolution. During the Neoglacial, %BS values gradually increase to 5%. And, at the top of the core, during the historic interval, %BS increases abruptly to maximum values observed in our sediment record (Fig. 9). SEM observations (Fig. 10) indicate that, although the abundance of diatoms varies over time, preservation remains good to excellent, which indicates that diatom productivity is the major variable.

%TC results versus time also correlate very well with the major Holocene climate intervals (Fig. 9). The post–Younger Dryas is characterized by intermediate values, and the Neoglacial has essentially zero TC. The bulk of the carbonate occurs during the Hypsithermal, when there was an abrupt increase near its beginning followed by a gradual decline over a ~4,000-yr period (Fig. 9). This highly asymmetric distribution of %TC is punctuated by five distinct cycles with amplitudes of up to ~15%. These cycles have an average period of ~800 yr and a range of ~475–1,025 yr. The largest negative anomaly in the %TC data correlates with the 8.2-ka event (Fig. 9).

%TOM data versus time display an overall up-core increase from minimum values 10,000 yr ago to a maximum during historic times. A significant negative shift (“step”) in TOM also occurs in association with the Nipissing flood (Fig. 9).

Stable carbon and nitrogen isotope values of OM also vary as a function of time in accord with major Holocene climate intervals and events (Fig. 11). The lowest $\delta^{13}C_{OM}$ values occur during the post–Younger Dryas, whereas the Hypsithermal displays higher values than either the post–Younger Dryas or Neoglacial and has the greatest amount of variability. The Hypsithermal also contains five cycles of $\delta^{13}C_{OM}$ values similar to those recognized in the %TC data (Figs. 9, 11). The $\delta^{13}C_{OM}$ cycles also occur on a centennial timescale, having an average period of ~680 yr and a narrow range of ~780–560 yr. The largest negative anomaly in $\delta^{13}C_{OM}$ values occurs in association with the 8.2-ka event (Fig. 11). During the Neoglacial, $\delta^{13}C_{OM}$ values are relatively stable until the historic period, when there is an abrupt increase to the maximum value of the entire record.

$\delta^{15}N_{OM}$ trends differ somewhat from those for $\delta^{13}C_{OM}$. The most notable difference is that, during the post–Younger Dryas, $\delta^{15}N_{OM}$ values display a positive anomaly unlike $\delta^{13}C_{OM}$ values that are characterized by a negative anomaly of similar magnitude (Fig. 11). In fact, $\delta^{15}N_{OM}$ values during

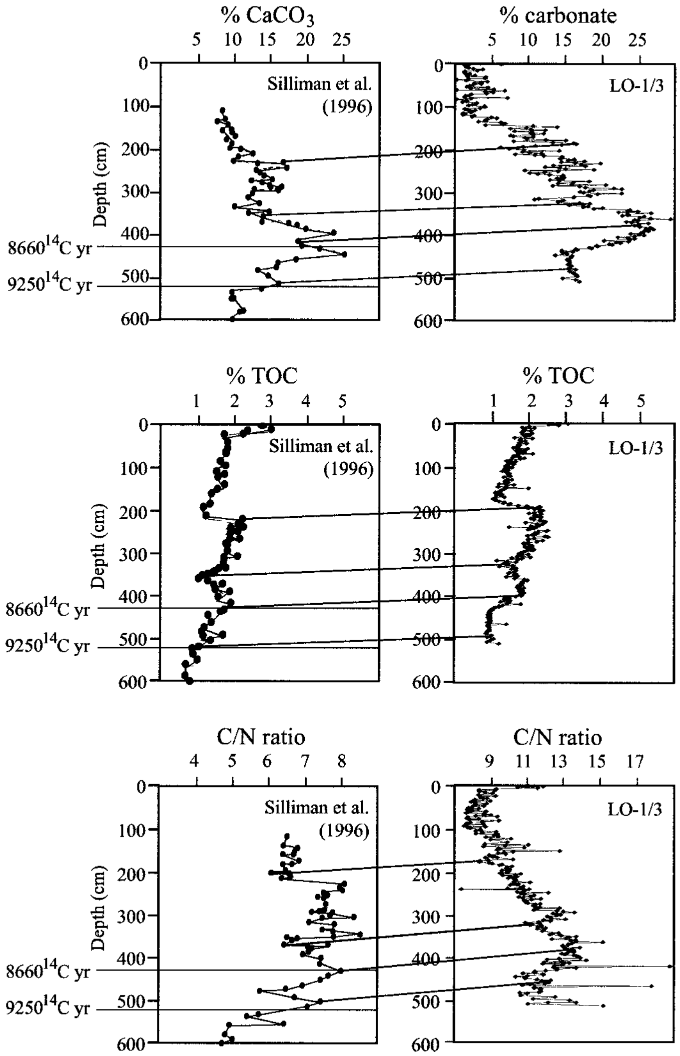


Fig. 4. Correlation of core LO −1/3, based on %TC, %C, and C : N data, with previously radiocarbon dated core of Silliman et al. (1996) from the Rochester Basin used to construct age model for the prehistoric interval in eastern Lake Ontario.

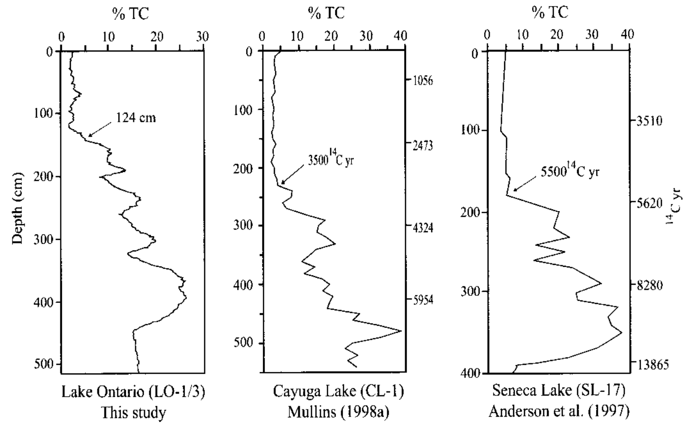
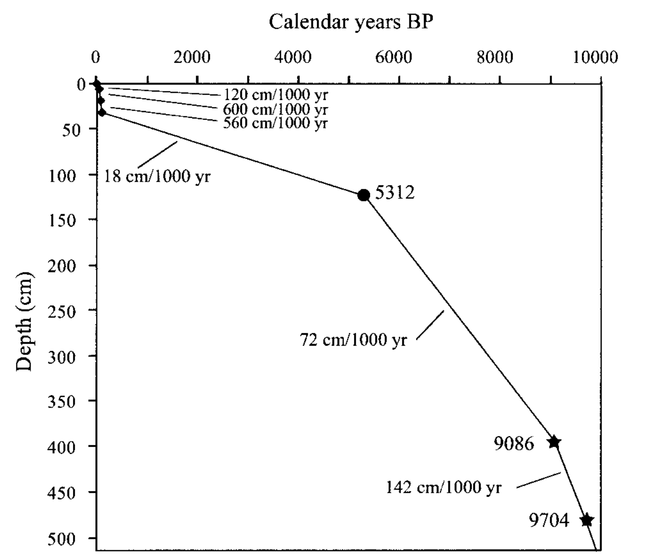


Fig. 5. Correlation of %TC data from Lake Ontario core LO −1/3 with previously radiocarbon dated cores from Cayuga Lake (Mullins 1998) and Seneca Lake (Anderson et al. 1997) in the nearby Finger Lakes used to develop age model for the prehistoric interval. Note that TC values were <5% in all three cores.

the post–Younger Dryas are the highest of our 10,000 year record, with the exception of historic times (Fig. 11). As with carbon isotope values, $\delta^{15}N_{OM}$ values display the largest amount of variability during the Hypsithermal. Nitrogen isotope data are not characterized by a well defined anomaly during the 8.2-ka event but do display a large negative shift (''step'') in association with the Nipissing flood (Fig. 11). During the Neoglacial, $\delta^{15}N_{OM}$ values gradually increase, whereas during historic times, there is a dramatic increase.

$\delta^{18}O_{calcite}$ and $\delta^{13}C_{calcite}$ data are only available from those portions of core LO −1/3 that contain sufficient authigenic carbonate for analysis (~10.0–5.3 ka; Figs. 7, 9). Oxygen isotope values peak during the middle of the post–Younger Dryas. The most prominent $\delta^{18}O_{calcite}$ negative anomaly during the Hypsithermal occurs in association with the Nipissing flood (Fig. 12).

The calcite carbon isotope curve differs significantly from that of oxygen. The end of the post–Younger Dryas has the lowest $\delta^{13}C_{calcite}$ value in our record and is marked by a distinct negative anomaly followed by rapid rise near the beginning of the Hypsithermal (Fig. 12). The most prominent $\delta^{13}C_{calcite}$ feature during the Hypsithermal is a large negative anomaly associated with the 8.2-ka event. Five minor peaks in $\delta^{13}C_{calcite}$ values occur throughout the Hypsithermal coin-



symbol	depth (cm)	^{14}C yr	cal yr BP	date source
◆	0		10 (1940 AD)	^{210}Pb and ^{137}Cs
◆	6		60 (1890 AD)	^{210}Pb and ^{137}Cs
◆	18		80 (1870 AD)	^{210}Pb and ^{137}Cs
◆	32		105 (1845 AD)	^{210}Pb and ^{137}Cs
●	124	4600	5312	Radiocarbon date (organic matter)
★	394	8160	9086	Radiocarbon date (ostracods), subtracted HWE 500 yrs
★	482	8750	9704	Radiocarbon date (ostracods), subtracted HWE 500 yrs

Fig. 6. Age model, in calibrated calendar yr, for eastern Lake Ontario core LO −1/3; HWE = hard-water effect. Also shown are sedimentation accumulation rates over time in cm 1,000 yr^{-1}.

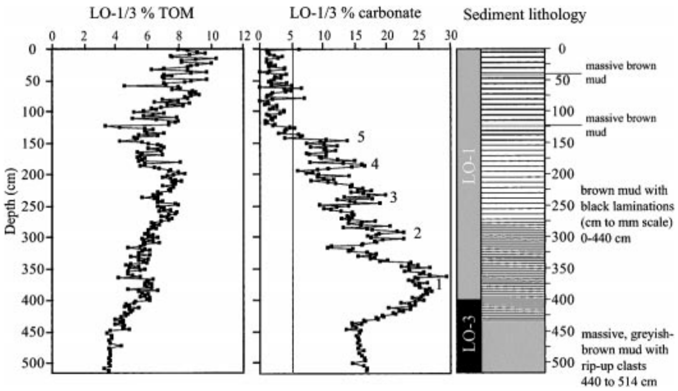


Fig. 7. Schematic stratigraphic section for core LO −1/3 (right) shown with %TOM and %TC values versus depth. Note composite nature of core, with the lower portion from core LO −3 (black) ''spliced'' onto the entirety of core LO −1 (gray). Numbers 1–5 refer to cycles in %TC data discussed in text.

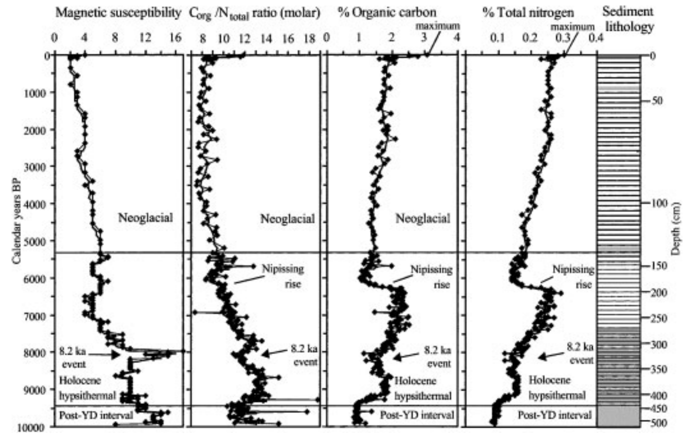


Fig. 8. Values of MS, C : N, %C, and %N in core LO −1/3 vs. age in calendar yr to the left. Also shown is the schematic stratigraphic section for the core with a nonlinear depth scale to the right. Superimposed are the boundaries between the major paleoclimate intervals of the past 10,000 yr in the northeastern United States. Also highlighted are major paleoclimate events, such as the 8.2-ka and Nipissing (''Rise'') flood. All subsequent data plots will have the same vertical axes and highlighted paleoclimate boundaries and events discussed in the text.

cident with cycles in %TC and $\delta^{13}C_{OM}$ data (Figs. 9, 11, 12). These carbon isotope cycles also occur at a centennial timescale with an average period of ~790 yr and a range of ~560–1,050 yr.

Discussion

Post–Younger Dryas climate interval (~10.0–9.4 ka)—As defined by Kirby et al. (2002), the post–Younger Dryas climate interval was the final cold period in the northeastern United States before the start of the Holocene. They argued that deep-water production in the North Atlantic after the Younger Dryas forced a southward expansion of the polar front jet stream over the northeastern United States by ~6° of latitude.

Because our record starts at ~10 ka, proxy results only pertain to the last ~600 yr of the post–Younger Dryas. In eastern Lake Ontario, %TOM, %C, %N, $\delta^{13}C_{OM}$, and $\delta^{13}C_{calcite}$ values are all at a minimum for the past 10,000 yr (Figs. 8, 9, 11, 12). Collectively, these proxies suggest relatively low levels of primary production. Cooler regional temperatures during this interval would have reduced the growing season and limited primary production. $\delta^{18}O_{calcite}$ values are high during the post–Younger Dryas (Fig. 12), consistent with cooler surface water temperatures.

Magnetic susceptibility values, C : N ratios and $\delta^{15}N_{OM}$ values are near their maximum values during the post–Younger Dryas (Figs. 8, 11), when Lake Ontario lake-levels were 70–80 m below those of the present day (Anderson and Lewis 1985). Relatively high values of magnetic susceptibility and C : N ratios suggest a greater flux of terrestrial matter to Lake Ontario at this time. TC data also suggest that surface waters were supersaturated with respect to calcite as a consequence of the chemical weathering of freshly exposed limestone after glacial retreat.

The only proxy indicator that is not consistent with our interpretation of relatively low levels of primary production during the post–Younger Dryas are high $\delta^{15}N_{OM}$ values (Fig. 11). The occurrence of oxidized rip-up clasts and absence of black laminations in this section of the core (Fig. 7) indicates well-oxygenated waters flowing across the bottom of eastern Lake Ontario. Under such conditions, microbes in the sediment may have converted organic nitrogenous compounds from a reduced to a more oxidized state. Nitrification increases the value of $\delta^{15}N_{OM}$, because microbes preferentially remove ^{14}N from OM (Talbot and Lærdal 2000).

Holocene hypsithermal (~9.4–5.3 ka)—The early to middle Holocene has long been known to have had warmer summers (~2–4°C) in the northern hemisphere than today (Pielou 1991). This was a result of increased summer insolation (~8%) caused by variations in orbital precession and obliquity (COHMAP 1988).

The transition from the post–Younger Dryas to the Hyp-

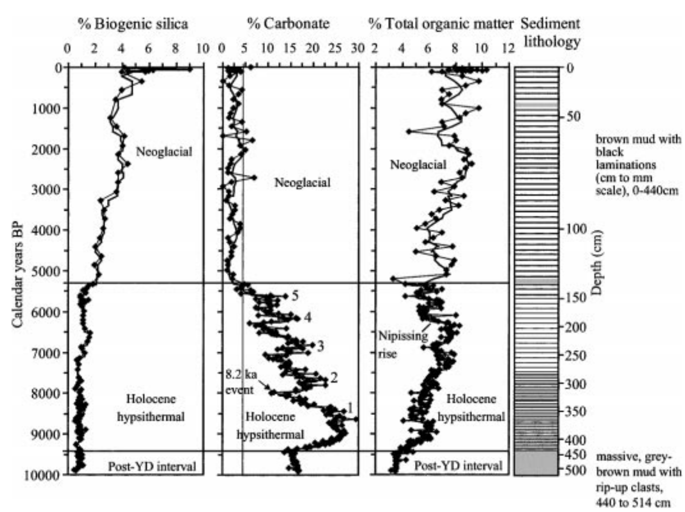


Fig. 9. Values of %BS, %TC, and %TOM vs. age in core LO −1/3 from eastern Lake Ontario.

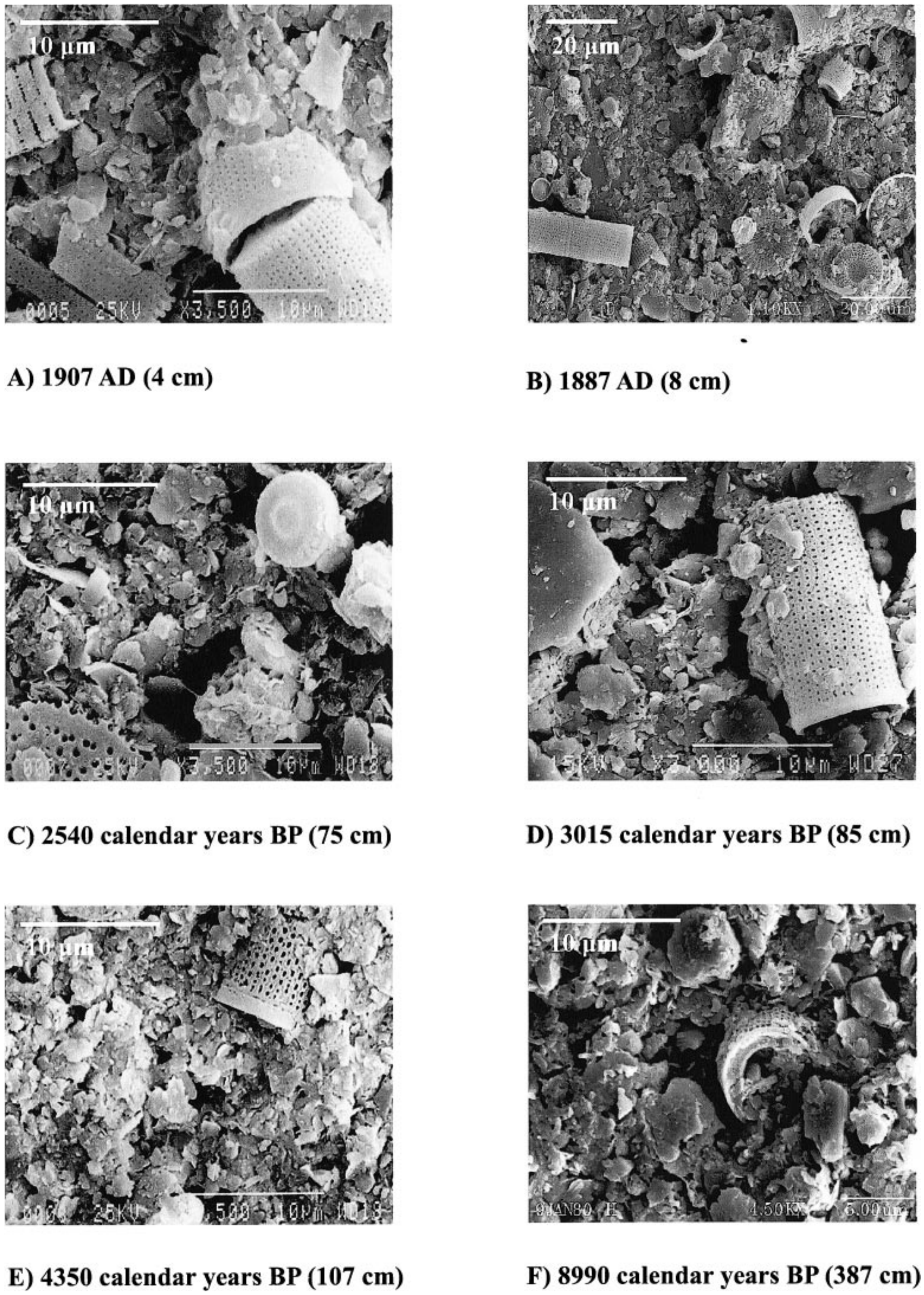


Fig. 10. Scanning electron microscope photographs from core LO −1/3 taken at various subsurface depths and ages. Note variable abundance of diatoms that display very good to excellent preservation throughout the core.

sithermal in eastern Lake Ontario is clearly delineated by changes in a number of proxies (Figs. 8, 9, 11, 12). Values of C : N, %C, %N, %TC, %TOM, $\delta^{13}C_{OM}$, and $\delta^{13}C_{calcite}$ increase dramatically across this boundary, which suggests an increase in primary production during the Hypsithermal. Higher summer temperatures would have increased the length of the summer growing season and increased chemical weathering in the watershed, leading to a greater flux of nutrients, dissolved carbonate components and terrestrial OM to the lake basin. %TC data document a dramatic increase in the precipitation of calcite from surface waters of eastern Lake Ontario during the Hypsithermal, probably due to warmer summer temperatures.

Relatively high values of magnetic susceptibility, coupled

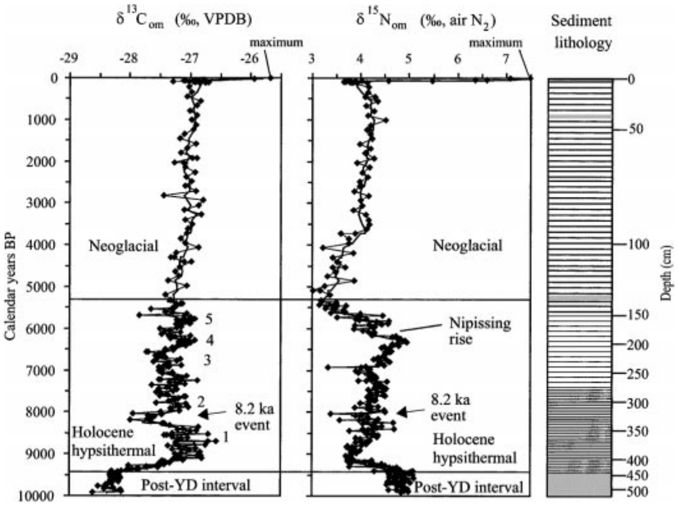


Fig. 11. Values of $\delta^{13}C_{OM}$ and $\delta^{15}N_{OM}$ vs. age in core LO $-1/3$ from eastern Lake Ontario.

with C : N values >10, support the interpretation of increased contribution of both inorganic and organic terrestrial material to the Lake at this time. Over the course of the Hypsithermal, however, C : N, MS, and %TC values decreased (Fig. 8), indicating a relative reduction in supply of terrestrial components. This may have been a response to decreases in summer temperatures, atmospheric precipitation, and/or chemical weathering over this $\sim$4,000-yr interval as orbital parameters slowly reduced summer insolation in the Northern Hemisphere.

Notably, the Hypsithermal displays a relatively high degree of variability in proxy data values (Figs. 8, 9, 11, 12). This variability indicates that, despite its summer warmth, the Hypsithermal was not stable and was subject to forcing mechanisms other than those associated solely with orbital parameters. Overall, $\delta^{13}C_{OM}$, $\delta^{15}N_{OM}$, and $\delta^{13}C_{calcite}$ data indicate higher levels of primary productivity for eastern Lake Ontario during the Hypsithermal than during either the post–Younger Dryas or the Neoglacial (Fig. 12).

The 8.2-ka cold event—The largest amplitude global climate anomaly during the Holocene occurred $\sim$8.4–8.0 ka (Alley et al. 1997). Stager and Majewski (1997) argued that atmospheric circulation changed abruptly at this time, producing cold, dry conditions over North America. This short-lived climate event has been attributed to the catastrophic drainage of large proglacial lakes into the North Atlantic, a reduction of thermohaline circulation, and reduced northward heat transport (Barber et al. 1999).

In eastern Lake Ontario, the 8.2-ka event is characterized by negative anomalies in C : N ratios, %C, %TC, $\delta^{13}C_{OM}$, and $\delta^{13}C_{calcite}$ values (Figs. 8, 9, 11 12). Collectively, decreases in these values argue strongly for lower levels of primary production in eastern Lake Ontario at this time due to lower surface water temperatures as well as a decrease of nutrient input from the watershed.

The largest negative anomaly in %TC values over the past 10,000 yr in eastern Lake Ontario occurred during the 8.2-ka event (Fig. 9). Cooler temperatures would have increased

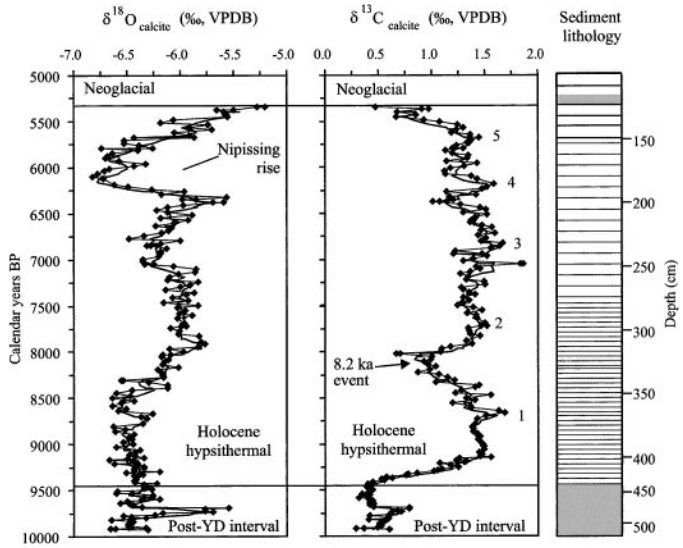


Fig. 12. Values of $\delta^{18}O_{calcite}$ and $\delta^{13}C_{calcite}$ vs. age in core LO $-1/3$ from eastern Lake Ontario. Note that time scale extends only from 10 to 5.3 ka, which is the interval in which there is sufficient calcite for analysis. Also note the large anomalies associated with the Nipissing (''Rise'') flood and 8.2-ka event.

the solubility of calcite while also reducing chemical weathering in the watershed and, thus, the saturation-state of surface water with respect to calcite (Lajewski et al. 2003). A reduction in the relative productivity of picoplankton may also have contributed to the observed decline in %TC (Hodell et al. 1998).

The 8.2-ka reduction in %TC in eastern Lake Ontario is but one part of five cycles observed throughout the Hypsithermal (Fig. 9). The period of TC cycles, as well as similar cycles in $\delta^{13}C_{OM}$ and $\delta^{13}C_{calcite}$ (Figs. 9, 11, 12), is $\sim$750 yr. This suggests that the TC cycles in eastern Lake Ontario may be a harmonic of the well-known $\sim$1,500-yr Holocene climate cycle, which is driven by solar variability (Bond et al. 2001) and/or a negative feedback in the North Atlantic hydrologic cycle (Mullins et al. 2003).

The Nipissing flood—The highest lake levels resulting from the Nipissing flood occurred $\sim$5,000 calendar years ago, according to radiocarbon dating of near-shore deposits above modern day Great Lake lake levels (Anderson and Lewis 1985). This flood has been interpreted as a redirection of upper Great Lakes outflow from the North Bay Outlet of Lake Huron to the lower Great Lakes in response to differential isostatic rebound (Anderson and Lewis 1985; Rea et al. 1994).

Others have argued that the Nipissing flood began as early as 6.8 ka, with peak flow occurring between $\sim$6.3 and 6.0 ka (Tinkler et al. 1994; Pengelly et al. 1997). Because of their higher latitude and greater distance from oceanic sources of water vapor, the upper Great Lakes have lower $\delta^{18}O$ values of outflow waters (Lake Superior = $-9.2‰$) than the lower Great Lakes (Lake Ontario = $-6.8‰$; Patterson et al. 1993; Gat et al. 1994). Our $\delta^{18}O_{calcite}$ data (Fig. 12) display a dramatic negative anomaly of $\sim$1.5‰ beginning $\sim$6.3 ka and ending $\sim$6.0 ka.

In eastern Lake Ontario, there are abrupt negative shifts in %C, %N, %TOM, and $\delta^{15}N_{OM}$ values between ~6.3 and 6.0 ka (Figs. 8, 9, 11). All of these proxies suggest reduced levels of primary production at this time, perhaps as a consequence of regional cooling triggered by the Nipissing flood. Colder temperatures would have limited the summer growing season and reduced chemical weathering in the Lake Ontario watershed. The fact that %TC values declined during this interval (Fig. 9) further supports the interpretation of lower rates of chemical weathering in the watershed and perhaps picoplankton productivity. The steplike negative shift in the values of a number of proxies (Figs. 8, 11) suggests that the onset of this event was abrupt.

The Holocene Neoglacial (~5.3 ka–1850 A.D.)—The start of the Neoglacial is defined by the readvance of mountain glaciers that had been receding during the early to middle Holocene Hypsithermal (Pielou 1991). Glacial advances were controlled by a decrease in Northern Hemisphere summer insolation which resulted in a 1–2°C drop in summer temperatures (COHMAP 1988).

In eastern Lake Ontario, the Neoglacial is characterized by a number of very dramatic changes. Average sediment accumulation rates drop from ~72 cm 1,000 yr^{-1} to only 18 cm 1,000 yr^{-1} (Fig. 6), which is consistent with an ~5-fold decrease observed in Lake Michigan (Colman et al. 1990). The decrease in Lake Ontario was, in part, due to a decrease in the input of terrestrial material (Silliman et al. 1996). This is supported by magnetic susceptibility data that display their lowest values of the entire 10,000-yr record as well as C : N values <10, indicating autochthonous OM production (Fig. 8). A second factor is the fact that calcite precipitation in eastern Lake Ontario effectively ceased ~5.3 ka (Fig. 9). This was likely a consequence of surface waters becoming undersaturated with respect to calcite because of reduced levels of chemical weathering in the watershed and cooler surface waters during the Neoglacial.

Another major proxy change during the Neoglacial was %BS (Fig. 9). Before the Neoglacial, BS values were <1%, indicating minimal primary production by diatoms. However, at ~5.3 ka, %BS values begin to increase and gradually build toward a maximum near the end of the Neoglacial (Fig. 9). Increased wind during the Neoglacial (Mullins and Halfman 2001) likely led to increased upwelling and nutrient supply to surface waters, which would have led to increased levels of diatom productivity.

In general, primary productivity in eastern Lake Ontario during the Neoglacial was lower than during the Hypsithermal but higher than during the post–Younger Dryas. The Neoglacial also appears to have been more stable, in terms of primary production, because proxy values (Figs. 8, 11) display less variability than at any other time over the past 10,000 yr. Collectively, all these results indicate a fundamental change in the physical, chemical, and biological limnology of Lake Ontario during the Neoglacial.

The historic interval (~1850–1940 A.D.)—This relatively short period of time is characterized by some of the largest proxy anomalies of the past 10,000 yr. Sediment accumulation rates increase from 18 cm 1,000 yr^{-1} during the Neoglacial to values as high as 600 cm 1,000 yr^{-1} since ~1850 A.D. (Fig. 6). This dramatic increase in sedimentation correlates with historic records of regional deforestation of the Lake Ontario watershed between ~1820 and 1850 A.D. (Schelske et al. 1983). Deforestation destabilized soils and resulted in a major increase in the contribution of terrestrial material, including nutrients, to the lake (Schelske 1991; Schelske and Hodell 1991; Silliman et al. 1996; Hodell and Schelske 1998), which resulted in higher levels of primary production. Further anthropogenic disturbance, including agricultural, municipal, and industrial activity, loaded additional nutrients to Lake Ontario and led to the historic peak in primary productivity during the 1970s to the early 1980s.

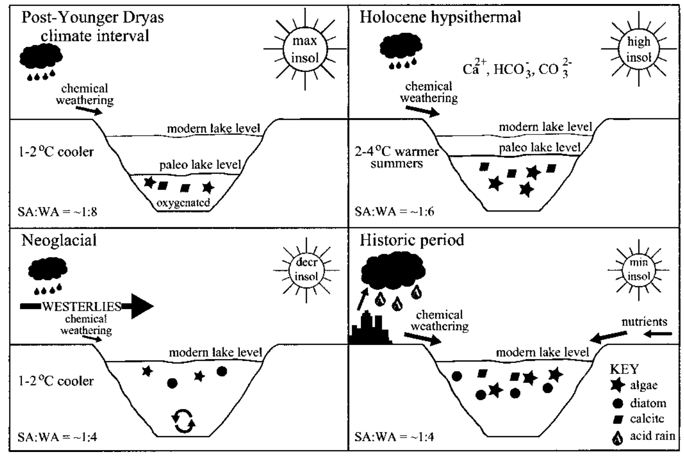


Fig. 13. Schematic cartoon illustrating major controls of paleoproductivity in eastern Lake Ontario during the major paleoclimate intervals of the past 10,000 yr. Note interaction of temperature, lake level, summer insolation, westerly wind, chemical weathering, atmospheric precipitation, lake surface area (SA) : watershed area (WA) ratio, acid rain and anthropogenic nutrient loading that collectively control primary production in eastern Lake Ontario.

Percent C, %N, and %TOM (Figs. 8, 9), as well as $\delta^{13}C_{OM}$ and $\delta^{15}N_{OM}$ (Fig. 11), data from the historic period all indicate an abrupt increase in primary production, and all are at their maximum values for the past 10,000 yr. These results document that primary production in eastern Lake Ontario during the past ~150 yr was not only elevated but at unprecedented levels. %BS values during historic time (Fig. 9) are also ~9-fold greater than during the post–Younger Dryas or Hypsithermal intervals and are ~2 times greater than the Neoglacial, indicating a dramatic increase of diatom productivity.

There is also evidence in %TC values (Fig. 9) at the very top of our core (~1940 A.D.), as well as direct observations (Hodell et al. 1998), that calcite precipitation returned to Lake Ontario after an ~5,000-yr hiatus, as it did in the nearby Finger Lakes (Mullins 1998). Lajewski et al. (2003) argued that the return of calcite precipitation during the historic period was the result of acid rain falling on carbonate-rich areas in the Lake Ontario watershed, which led to ''alkalization'' of lake waters (Fig. 13).

The future—The accurate prediction of future levels of primary production in eastern Lake Ontario is not possible at this time (if ever) because of our limited understanding of aquatic ecosystems, as well as a lack of knowledge about future anthropogenic impacts, including invasive species. However, the projection of future levels of primary production is reasonable with our current understanding of global and regional environmental change, combined with knowledge on long-term natural variability of primary productivity. Today, Lake Ontario appears to be in transition from a highly productive lake, a few decades ago, to a lake with moderate or even low levels of primary production.

We know that global and regional climates are changing (IPCC 2001) and that these changes are affecting the Great Lakes in a variety of complex ways (Kling et al. 2003). During the 20th century, mean global surface temperatures increased by ~0.6°C, and precipitation increased by as much as 1% per decade over most midlatitudes of the northern hemisphere (IPCC 2001). Also during the 20th century, the temporal extent of winter ice cover in the Great Lakes decreased by ~12 d (Magnuson et al. 2000), and the length of the summer growing season increased by ~14–18 d (McCormick and Fahnenstiel 1999). Climate models suggest that, by the end of the 21st century, winter temperatures in the Great Lakes region may increase by as much as 3–7°C and summer temperatures may warm by as much as 3–11°C (Kling et al. 2003). During the 21st century, Lake Ontario may become similar to what it was during the Holocene Hypsithermal, when summer temperatures were ~2–4°C warmer than today. Longer ice-free periods and higher surface water temperatures, along with greater chemical weathering and nutrient supply, are likely to increase future levels of primary production in Lake Ontario (Kling et al. 2003). We project that, during the 21st century, levels of primary productivity in eastern Lake Ontario will become similar to what they were during the Holocene Hypsithermal. Although levels of primary production during the Hypsithermal were less than the historic peak of the 1970s to the early 1980s, they were higher than levels that characterized either the post–Younger Dryas or Neoglacial. If this projection is correct, the current transition of Lake Ontario from a highly productive to less productive lake may reverse. However, considering contemporary mitigation efforts to reduce nutrient loading of Lake Ontario, it seems unlikely that there would be a return to peak levels of primary production that characterized the mid- to late 20th century.

This multiproxy study has documented that natural processes have resulted in considerable variability of primary production in eastern Lake Ontario. This variability occurred in phase with known climate intervals and events of the past 10,000 yr. Cooler climatic conditions resulted in lower levels of primary production related to lower rates of chemical weathering in the watershed which controls the natural supply of nutrients to the lake basin. The Hypsithermal interval had higher levels of primary production as a result of warmer summer temperatures as well as an increased supply of nutrients.

These new proxy data from eastern Lake Ontario also indicate that, despite its summer warmth, the Hypsithermal was relatively unstable, which may have implications for the stability of future levels of primary production in a warmer world. Evidence from relatively short paleoclimate intervals, such as the 8.2-ka and Nipissing flood events, further document that levels of primary productivity in eastern Lake Ontario are capable of abrupt and relatively large amplitude change.

The most dramatic result of this study is the unprecedented rise in levels of primary production over the past 10,000 yr during the brief historic interval. This clearly demonstrates the large anthropogenic effect on primary productivity in eastern Lake Ontario. Future levels of primary production in eastern Lake Ontario will vary as a function of both natural and anthropogenic forcing. We project that current trends toward lower levels of primary production in eastern Lake Ontario will reverse during the 21st century and may become similar to what it was during the warm Hypsithermal. However, it is unlikely that peak levels of primary production that occurred during the 1970s–1980s because of cultural eutrophication will return because of mitigation efforts to reduce the anthropogenic loading of nutrients to the lake basin.

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