

Hamilton Harbour, Ontario: 8300 years of limnological and environmental change inferred from microfossil and isotopic analyses

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Received 24 February 1995; accepted 4 May 1995

Key words: Hamilton Harbour, paleolimnology, diatoms, sediment isotopes, palynology

Abstract

Limnological changes in Hamilton Harbour, Lake Ontario, over the Holocene were investigated by using proxy evidence from diatoms and other siliceous microfossils in a radiometrically dated sediment core (HH26comp), together with environmental data derived from sediment pollen and oxygen and carbon isotope analyses. The evidence demonstrates that the site of Hamilton Harbour has changed over the past 8300 y from a shallow, separate waterbody, to a deep embayment of Lake Ontario. The earliest evidence, from 8300 BP to 7000 BP, is of a mesotrophic pond of moderate alkalinity, warmer than present, and probably with an extensive marginal wetland. An initial transitory connection with the rising water level of Lake Ontario was established at c. 7000 BP, possibly via a deep outlet channel. This connection is 2000 y earlier than previously estimated. Permanent confluence with Lake Ontario was established at c. 6200 BP, causing a decline in inferred trophic level and water temperatures. Microfossils reach a minimum at 4400 BP coincident with the Nipissing Flood. Decreased mixing of Lake Ontario water from about 4000 BP following the Nipissing Flood highstand is evidenced in isotopic and diatom data. Three isolated shifts in the diatom spectrum at c. 4900 BP, 4500 BP, and 3500 BP may be associated with extreme turbidity or storm deposit events. Between 3200 BP and 280 BP, Hamilton Harbour was evidently a moderately alkaline embayment of Lake Ontario, oligotrophic to mesotrophic, and relatively cooler than present. The final 280 y sedimentary record reveals the magnitude of anthropogenically induced changes to the harbour, including eutrophication and organic pollution.

Introduction

In North America there is an ever-growing concern over the impact of humans on the environment in the recent past. Much attention has focussed on the Great Lakes since many large urban centres are situated on them, a large population lives in close proximity, and they are shared by two countries. The health of the Great Lakes ecosystem has been called into question on several occasions (Gilbertson & Schneider, 1993; Sly, 1991), and, as a consequence, efforts have been made to monitor and restore many parts of them.

Hamilton Harbour is one area of particular concern since it strongly influences the water quality of Lake

Ontario. The harbour was identified as one of the most polluted sites in the Great Lakes by the Water Quality Board of the International Joint Commission (IJC, 1985). For this reason it is the subject of an intense multidisciplinary environmental remedial action plan (Rodgers, 1989). Paleocological investigations are especially important in these circumstances because the long-term history can be reconstructed for the past several centuries and millenia. This information permits hindsight and establishes ecosystem conditions before the time of human impact, thus providing the long-term data with which to gauge the state of the extant ecosystem. Management decisions can then

be made in order to enhance desirable qualities and reverse undesirable trends.

Recently, Yang *et al.* (1993) conducted a paleoecological investigation of the recent environmental history of the harbour using siliceous microfossil analyses of a short sediment core. This study clearly demonstrated striking changes over the past 330 y resulting from human activities. However, with the exception of Coakley & Karrow (1994) and Anderson & Lewis (1985), who both examined sediment cores from western Lake Ontario and Hamilton Harbour primarily from a shoreline evolution perspective, the long-term environmental history and paleolimnology of Hamilton Harbour remains largely unknown. The main objective of this study is to investigate limnological changes over the Holocene in this nearshore waterbody of Lake Ontario by using proxy evidence from siliceous microfossils in a radiometrically dated sediment core, together with environmental data derived from isotopic and elemental analyses. Pollen data are used primarily for biostratigraphic comparison. Because the littoral and nearshore zones of lakes are especially sensitive to lake system changes (UNESCO-MAB workshop, 1988), the study should also contribute to a better understanding of Holocene water-level fluctuations in western Lake Ontario.

Materials and methods

Hamilton Harbour (Burlington Bay) is located at the extreme western end of Lake Ontario. In its natural state a baymouth bar separated the bay from main Lake Ontario. Today, it is connected with Lake Ontario by a 107 m wide and 9.5 m deep ship canal. Two small streams discharge into the shallow western end of the embayment. The harbour is triangular in shape with a surface area of about 2.2×10^3 ha, and mean and maximum depths are 13 m and 26 m respectively (Poulton, 1987). According to Karrow (1987) and Coakley & Karrow (1994) glacial till deposits are a major component of the surficial sediments, and the baymouth bar (Burlington Bar) is composed of stratified sands and gravels representing a former plane of proglacial Lake Iroquois (c. 12 000 BP).

Hamilton Harbour composite core HH-26 was collected in 21 m water (43 °17'28" N, 79 °51'05" E) on June 25, 1991, from the deck of the *RV Limnos*. The location is described in Coakley & Karrow (1994). Initially, a 7.53 m length of piston core (core HH26p2) was collected, then a 2 m benthos corer (core

HH26Be2) was used to collect the top 1 m of sediment. The combined core HH26comp used in this study is a composite of these two cores. Correlation of overlap between the two cores was done by using porosity data, and the stratigraphy of five ostracode species (Delorme, 1992a). After correlation, combined core HH26comp had a total length of 855 cm.

Sectioned samples at 1 cm intervals were freeze-dried and withdrawn for various analyses, including radiocarbon dating on organic debris.

Siliceous microfossil analyses

A total of 85 samples were chosen for siliceous microfossil analysis. Siliceous microfossils were cleaned with H₂O₂, settled in plexiglass chambers and mounted in Hyrax[®]. The prepared slides were examined at 1000 × magnification with a Nikon Labophot microscope (N.A. 1.25) and all diatoms were identified to the species or variety level. The main taxonomic references used for identification included Hustedt (1930, 1937–1939, 1957), Cleve-Euler (1951–1955), Patrick & Reimer (1966, 1975) and Germain (1981).

Diatom valves were enumerated by transect until a total of approximately 600 valves had been counted per sample. Chrysophyte cysts, freshwater sponge fossils, and diatoms were counted from the same slides. Enumeration data were processed through a database management system (StatView 512+TM) and Excel. Calculated quantities include estimates of abundance (microfossils per g dry weight of sediment), estimates of relative abundance (percentage of total assemblage) and estimates of assemblage diversity (Shannon-Wiener Index).

Multivariate analyses were performed on relative abundance with the SYSTAT statistical package. All species that occurred in at least five samples and at >1% relative abundance in at least one sample were included in the statistical analyses. Cluster analysis was used in order to examine the pattern of community change with time. This analysis is based on the Euclidean distance measure and uses the average distance clustering method (Carney, 1982; Stoermer *et al.*, 1985; Yang *et al.*, 1993).

Diatom taxa were assigned to four habitat groups (euplankton, tychoplankton, epiphyton, benthos) using the criteria of Patrick & Reimer (1966, 1975), and Lowe (1974). Changes in relative abundance of these groups in core samples were assumed to reflect water level fluctuations (Yang & Duthie, 1995a).

Water level changes were inferred by diatoms using weighted average calibration (non-linear regression, Stevenson *et al.* 1989):

$$ID_i = \frac{\sum_k^m W_k P_{ik}}{\sum_k^m P_{ik}}$$

Where, ID_i = inferred water depth at core sample i of Hamilton Harbour, and W_k = weighted average (optimum) value of water depth for each diatom taxon obtained from a Lake Ontario calibration set. The Root Mean Squared Error of the calibration is 1.4 m (Yang & Duthie, 1995a).

Changes in the trophic status of Hamilton Harbour were inferred with the empirical equation of Yang & Dickman (1993):

$$LTSI = 2.643 - 7.575 \log(\text{Index D}) (SE = 0.94)$$

Where, $LTSI$ = lake trophic status index, $\text{Index D} = (O\% + OM\% + M\%) / (E\% + ME\% + M\%)$, O = oligotrophic, OM = oligomesotrophic, M = mesotrophic, ME = mesoeutrophic and E = eutrophic. Trophic preferences of Hamilton Harbour diatoms are listed in Yang (1994).

Stratigraphic changes in lake pH were inferred qualitatively by comparing the relative abundance of five diatom pH preference groups, using the criteria of Hustedt (1937–1939), Meriläinen (1967) Cholnoky (1968), Lowe (1974) and Duthie (1989).

Paleotemperature and other paleoclimatic factors are more commonly reconstructed by isotope and palynological methods; diatoms have not been an important tool, although Pienitz *et al.* (1995) have recently published transfer functions from the arctic. However, the distribution optima of many diatoms in the Great Lakes region with respect to temperature have been reported by Lowe (1974), and by Stoermer & Ladewski (1976, 1978). Based on their information, diatom taxa found in this study (Yang, 1994) were classified in five temperature categories (Hustedt, 1957): ‘Oligothermal’ (oth) – cold water forms, occurring at a water temperature no greater than 10 °C and with optima below 5 °C; ‘Oligomesothermal’ (omth) – cool water forms, occurring between 5 °C and 15 °C with optima between 5 °C and 10 °C; ‘Mesothermal’ (mth) – moderately warm water forms, occurring between 10 °C and 20 °C with optima between 10 °C and 15 °C; ‘Mesoeutothermal’ (meth) – warm water forms, occurring between 15 °C and 30 °C with optima between 15 °C and 25 °C; and, ‘Eutothermal’ (eth) – warmest water forms, found at no less than 25 °C with optima above 25 °C.

CORE HH26 comp (Water depth: 21 m)

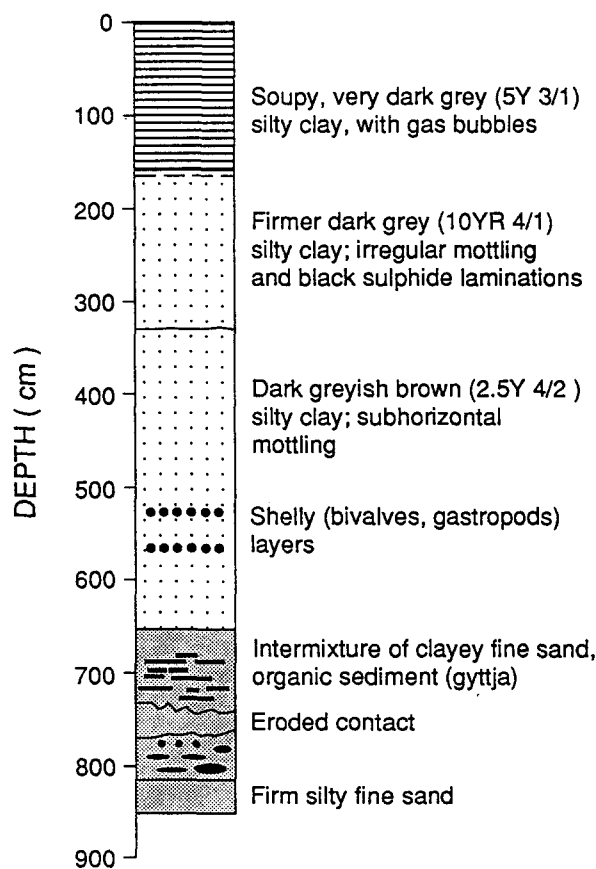


Fig. 1. Stratigraphic description of core HH26comp (from Coakley and Karrow 1994).

Pollen analyses

Subsamples of freeze-dried sediment were taken for pollen analysis and weighed. Each sample was processed following standard pollen processing procedures, as required, which included 10% HCl, concentrated HF, acetolysis, and dehydration in an ethanol series for storage in silicon oil in stoppered vials (Faegri & Iversen, 1989).

Identification and counting of palynomorphs were done under bright field at 400 × magnification, and at 1000 × magnification for critical determinations until a minimum total of 500 was attained. All calculations and pollen diagrams were prepared using the program TILIA and TILIAGRAPH (Grimm, 1992).

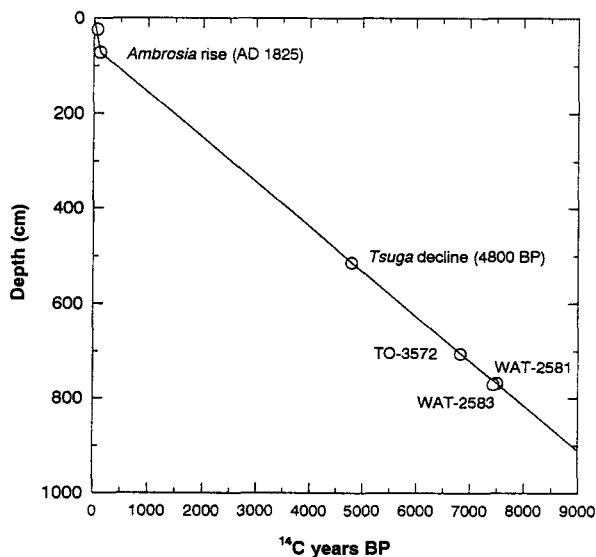


Fig. 2. Chronology of core HH26comp based on three radiocarbon dates (^{14}C y before AD 1950), the Ambrosia pollen rise (73 cm = 125 BP = AD 1825), and ^{210}Pb from porosity correlation with another core (25 cm = 70 BP = AD 1880). Radiocarbon dates, all on detrital organic debris, are as follows: WAT-2581, 768–770 cm, 7510 ± 100 y BP; WAT-2583, 770–773 cm, 7440 ± 100 y BP; TO-3572, 707 cm, 6830 ± 70 y BP. Note that the *Tsuga* pollen decline (4800 BP) fits the linear interpolation between TO-3572 and the *Ambrosia* rise. (A fourth radiocarbon date, WAT-2589, was disregarded because of inconsistency with the other three dates, presumably due to reworking of older material.)

Isotopic and elemental analyses

Stable carbon isotope ratios ($^{13}\text{C}/^{12}\text{C}$) were measured on both bulk organic matter and cellulose from the $<150 \mu\text{m}$ fraction of the sediments. Oxygen isotope ratios ($^{18}\text{O}/^{16}\text{O}$) were measured on the cellulose fraction only. Sediments were washed in warm, dilute HCl to remove carbonate mineral matter and shells. Bulk organic analysis was performed directly on the acid-washed residue, while samples for cellulose analysis received an additional three-step procedure involving solvent extraction, bleaching, and alkaline hydrolysis to remove non-cellulose organic constituents (Edwards & McAndrews, 1989). Isotopic analysis was performed on CO_2 gas prepared by conventional closed-tube combustion (Boutton, 1991) and nickel-tube pyrolysis (Edwards *et al.*, 1994a) for carbon and oxygen, respectively. Results are expressed as δ values, representing deviation in per mil (‰) from the PDB standard for carbon and V-SMOW standard for oxygen, such that

$$\delta_{\text{sample}} = [(R_{\text{sample}}/R_{\text{standard}}) - 1]1000$$

where R is the $^{13}\text{C}/^{12}\text{C}$ or $^{18}\text{O}/^{16}\text{O}$ ratio in the sample and standard. Samples were analyzed using a VG Prism II or VG MM903 mass spectrometer at the Environmental Isotope Laboratory, University of Waterloo. Analytical uncertainties are 0.3‰ for carbon and 0.5‰ for oxygen, based on repeated analyses of natural samples.

Carbon and nitrogen contents of the $<150 \mu\text{m}$ fraction of the acid-washed sediments were determined on a Carlo Erba EA1108 elemental analyzer. Analytical uncertainties are 0.2% for carbon and 0.02% for nitrogen.

The carbon isotope composition of aquatic organic matter is determined by the $\delta^{13}\text{C}$ of ambient dissolved inorganic carbon (DIC), which is controlled by a dynamic balance of factors, including isotopic exchange with atmospheric CO_2 (which tends to elevate $\delta^{13}\text{C}$), recycling of ^{13}C -depleted carbon from the decay of organic matter in the water column and bottom sediments, input of DIC having varying isotopic composition from runoff, and ^{13}C enrichment deriving from preferential uptake of ^{12}C by phytoplankton during photosynthesis. Changes in primary productivity commonly provide the dominant signal preserved in the $\delta^{13}\text{C}$ of organic matter in sediment cores (McKenzie, 1985), although this may be overprinted by changes in carbon balance arising because of hydrologic changes (Wolfe *et al.*, 1995). The $\delta^{13}\text{C}$ of undifferentiated organic matter is also sensitive to variations in the proportion of organic constituents having differing isotopic compositions, which may be related only indirectly to changes in productivity or to unrelated factors such as diagenesis. A qualitative assessment of such effects can be made by supplementing carbon isotope analysis on bulk organic matter with more selective analysis of an individual organic constituent such as cellulose, which is likely to have more direct sensitivity to changing DIC $\delta^{13}\text{C}$.

Cellulose is also a useful substrate for oxygen isotope analysis, because aquatic plant cellulose is consistently enriched by 27–28‰ relative to ambient water, independent of the temperature, the $\delta^{18}\text{O}$ of aqueous CO_2 , and the plant species (Epstein *et al.*, 1977; DeNiro & Epstein, 1981; Sternberg & DeNiro, 1983; Sternberg *et al.*, 1984, 1986). Studies of cores from a range of temperate and arctic lakes from Canada and Russia have demonstrated that cellulose preserved in the fine ($<150 \mu\text{m}$) sediment fraction is predominantly or entirely aquatic in origin, permitting direct reconstruction of lake water $\delta^{18}\text{O}$ history from sediment cellulose $\delta^{18}\text{O}$ records (Edwards & McAndrews,

1989; MacDonald *et al.*, 1993; Wolfe *et al.*, 1995; Edwards *et al.*, 1994b; Wolfe & Edwards, unpublished data). Lake water ^{18}O is in turn related to the average isotopic composition of local precipitation, which commonly varies with mean annual temperature, modified by secondary hydrologic factors such as evaporative enrichment (see also Edwards & Fritz, 1988; Edwards, 1993).

The carbon and nitrogen concentrations and C:N ratios help to characterize sedimentary organic matter and provide ancillary information about environmental changes during deposition of the sediments. Of immediate note in the Hamilton Harbour core is the existence of generally low C:N ratios in the fine fraction used for isotopic analysis, which supports the suggestion that the organic matter was derived from aquatic sources (Meyers & Ishiwatari, 1993; Meyers, 1994).

Interpretation

Core description and chronology

The piston and benthos cores were logged shortly after collection (Delorme, 1992a). Descriptions of core HH26comp are given in Fig. 1 and in Coakley & Karrow (1994).

A plot of sediment depth versus time for core HH26comp is shown in Fig. 2. The 4800 BP *Tsuga* pollen decline (McAndrews, 1981) at about 515 cm (natural depth) inferred from the pollen diagram (Fig. 3) plots along a best-line fit joining the three radiocarbon dates and the *Ambrosia* pollen rise. The sedimentation rate of c. 9.4 mm/yr is evidently invariant until a sharp acceleration after 1825 AD caused by human disturbance. Thus the depth scale is assumed to be linear with time, except for the post-1825 AD interval.

Palynology

The results of the palynological analyses are presented in a composite diagram (Fig. 3) which combines cores HH26p2 and HH26Be2. Two pollen zones are apparent. A lower zone represents the pre-*Ambrosia* period where pollen of coniferous and deciduous trees are important. The upper zone represents the *Ambrosia* period where proportions of pollen of tree types are reduced and pollen of herbaceous species, such as *Ambrosia*, Gramineae, Chenopodiaceae, Rumex and Plantago are noteworthy.

The pollen diagram is difficult to interpret because of the position of the coring site in a large embayment of Lake Ontario. Pollen can be expected to be derived from long distances by water transport, in addition to aerial input from vegetation in the region. McAndrews & Power (1973) and Starling & Crowder (1981) have shown that significant quantities of pollen derived from vegetation in local catchments is transported by streams and rivers flowing into Lake Ontario. Pollen derived through reworking of polleniferous deposits can enter the lake and be transported by streams, river, and by movements in long-shore currents. Turbidity flows and resuspension of sediments may be other mechanisms for mixing of pollen in the embayment sediments. For these reasons, the pollen record of Hamilton Harbour must be interpreted with some caution.

In general, the dominance of tree pollen such as *Pinus strobus*, *Tsuga*, *Quercus*, *Fagus* and *Carya* are typical of middle to late Holocene pollen records from small lakes and wetlands in the vicinity of Hamilton Harbour (Karrow & Warner 1990), and indeed, of the Holocene pollen record from elsewhere in Lake Ontario (McAndrews 1973). The *Tsuga* decline, generally placed at c. 4800 BP, evidently begins at 515 cm. The *Ambrosia* zone representative of the European period is obvious at the top of the core. Though these general trends are clear, and therefore provide good biostratigraphic markers, the curves in the pollen profile presented here are probably, at best, more smooth than would be expected in smaller and low energy lake basins in the region, for reasons given above.

Stratigraphy of siliceous microfossils

Abundances of siliceous microfossils (diatoms, chrysophytes and freshwater sponges) and data on diatom species richness and diversity for Core HH-26 comp are shown in Fig. 4. In general, samples were rich in diatoms with the exception of the portion of the core between 720 cm and 380 cm. Diatom concentrations below 750 cm are $>10^7$ valves g^{-1} dry weight, but above this level sharply decrease to $<10^5$ valves/g except for an isolated peak at 670 cm. From 380 cm, diatom concentrations gradually increase, and remain high to the surface. Stratigraphic patterns of both chrysophytes and freshwater sponge spicules fluctuate strongly throughout the core. Chrysophyte concentrations above 800 cm are higher than in the lower portion of the core, and sponge spicules are completely absent below 720 cm and above 50 cm.

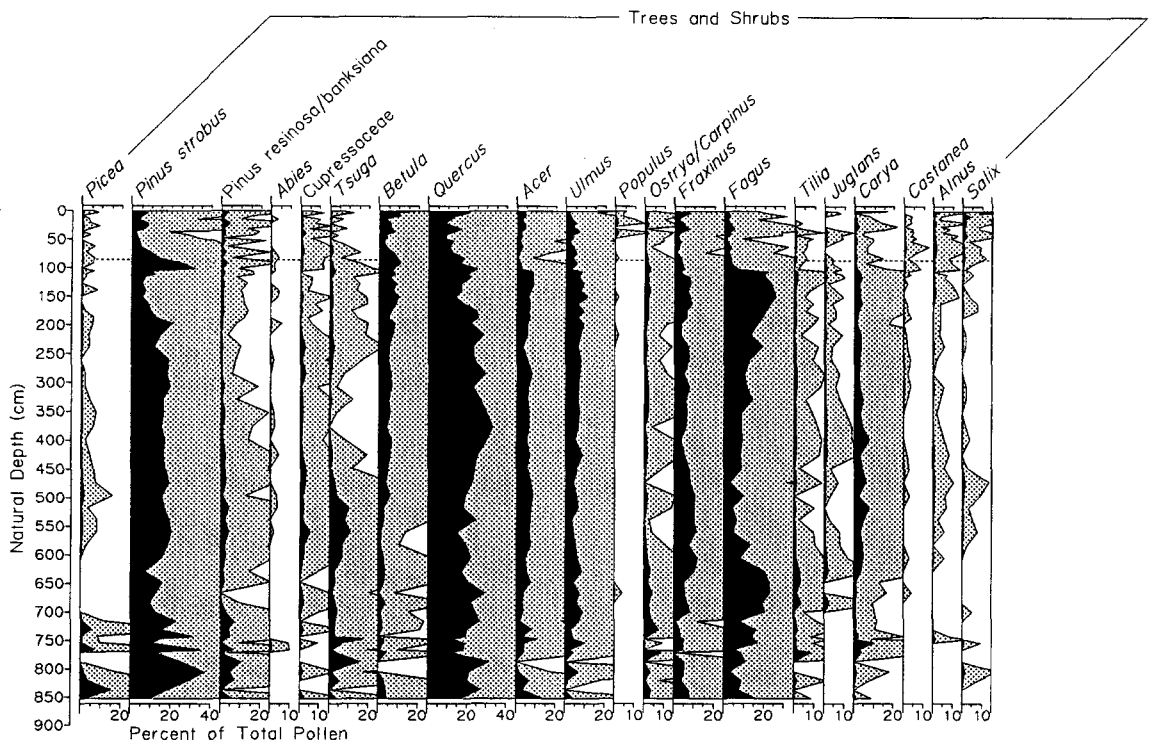


Fig. 3. Abbreviated pollen diagram for core HH26comp. Zone A denotes the pre-European settlement period, and Zone B denotes the European settlement period.

Profiles of both diatom species richness and species diversity are similar throughout the core. Both have consistently high values below 720 cm, but drop sharply above this. Above 450 cm, both increase slowly to the top of the core.

A total of 171 diatom taxa were found in the composite core (Yang, 1994). The relative abundance profiles of 35 dominant taxa are shown in Fig. 5. The following stratigraphies are described from the base of the core.

Nitzschia hantzschiana is the most abundant species in the basal samples, but disappears above 720 cm. *Navicula radiosa* is a major component of assemblages below 700 cm, becoming rare or absent after this level. *Navicula pupula* is only abundant at 840 cm. *Cyclostephanos dubius* and *Synedra acus* are the only two major euplanktonic species present in the lower portion of the core. Both *Synedra acus* and *Achnanthes minutissima* peak at 800 cm, but *Synedra acus* is absent in the bottom sample.

Profiles of *Navicula cryptocephala* and *Gomphonema longiceps* are similar, and both have an isolated peak at 485 cm. *Amphora ovalis* is the most abun-

dant species below 700 cm. Other important species in the lower portion of the core are *Navicula menisculus*, *Diploneis elliptica*, *Cocconeis placentula*, *Achnanthes clevei* and *A. linearis*.

From 700 cm there is a pronounced change in species composition; the above-mentioned species all become rare or absent, and are replaced by others. *Gyrosigma attenuatum* is the most abundant species between 720 cm and 620 cm, reaching over 58% of the total at 650 cm. It is also important at 400 cm. *Mastogloia smithii* exhibits a similar pattern. Two acidophilous species, *Eunotia pectinalis* and *Frustulia rhomboides*, reach their highest abundances 700 cm and 640 cm. *Synedra ulna* occurs throughout the core, but has a peak of >15% at 660 cm. *Aulacoseira granulata*, *Navicula scutelloides*, *Fragilaria construens* and *Martyana martyi* are occasionally abundant in middle portions of the core. *Stephanodiscus transsylvanicus* is also abundant in some samples between 720 cm and 380 cm.

A second major shift in diatom species composition occurs at 635 cm, where previously important species are replaced by *Aulacoseira islandica*, *A. itali-*

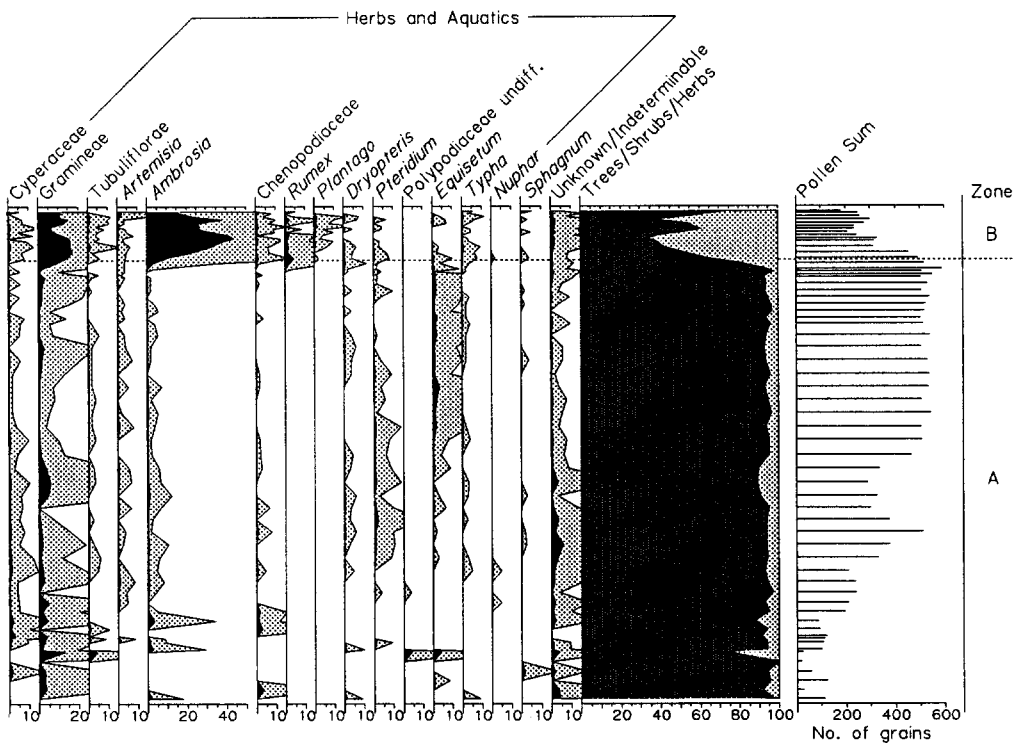


Fig. 3. Continued.

ca and *Stephanodiscus niagarae*. *Cyclotella bodanica* and *Tabellaria fenestrata* appear to fluctuate cyclically in the upper portion of the core. In the top 80 cm of the core, the flora is dominated by *Stephanodiscus parvus*, *Diatoma tenue* var. *elongatum*, *Fragilaria crotonensis*, *F. capucina*, *Cyclotella meneghiniana* and *Actinocyclus normanii* fo. *subsalsa*.

Relative abundances of four diatom habitat groups show major stratigraphic changes throughout the core (Fig. 6). Benthic diatoms completely dominate the lower portion of the core with the exception of samples between 720 cm and 705 cm. Above 650 cm they suddenly drop to a minimum. Benthic diatoms are in low abundance between 650 cm and 100 cm, except for three peaks at 530 cm, 490 cm, and 400 cm. From 100 cm the abundance of the group sharply increases, reaching over 60% at the surface. The euplanktonic stratigraphy is inverse to that of the benthic group. The epiphytic group is the second most abundant group below 720 cm, but is otherwise rare or absent. The tychoplanktonic group is a minor component of the diatom flora throughout the core.

Four major diatom zones, each of similar diatom species composition and assemblage structure, were

identified by cluster analysis (Fig. 7). Subzones B₁ and B₂ are separated by decreasing abundance or absence of some important species which occur below 370 cm, e.g., *Amphora ovalis*, *Gyrosigma attenuatum*, *Mastogloia smithii*, *Navicula scutelloides*, *Fragilaria construens* and *Martyana martyi*. Identification of the diatom zones reflects a series of major natural changes in environmental conditions in Hamilton Harbour during the past 8300 years. Changes in Zone A are a result of human disturbance.

Zone D (855 cm to 720 cm, estimated 8330 BP–7000 BP)

Diatom assemblages are dominated by the benthic diatom group suggesting that Hamilton Harbour during this period was relatively shallow compared to present. The dominant species include *Amphora ovalis*, *Nitzschia hantzschiana*, *Navicula radiosa*, *N. pupula*, *N. cryptocephala*, *N. menisculus*, *Diploneis elliptica*, and the epiphytes *Cocconeis placentula*, *Achnanthes minutissima* and *A. clevei*. All of these species were common in small, shallow lake environments in southern Ontario described by Yang & Dickman (1993).

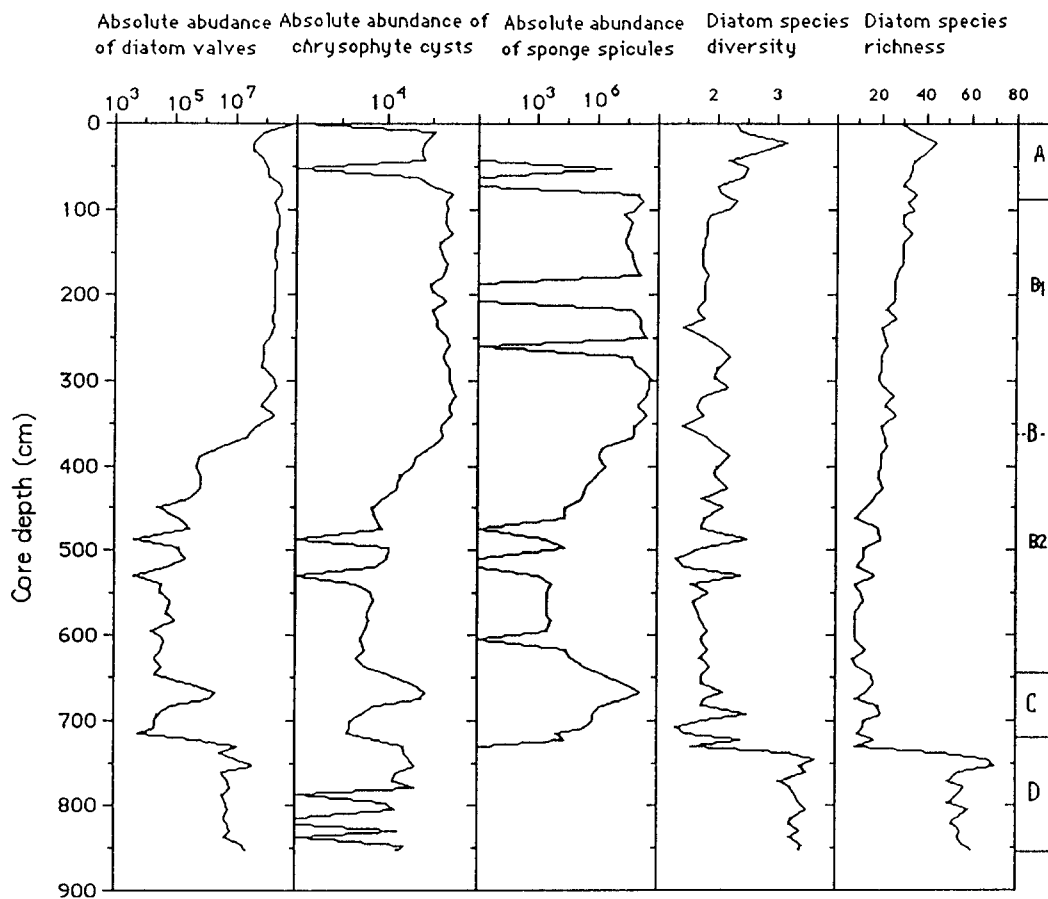


Fig. 4. Abundance (numbers/g dry wt.) of diatom valves, chrysophyte cysts and sponge spicules, and diatom species diversity (H') and species richness in core HH26comp. Diatom zones derived from cluster analysis are shown on the right.

Euplanktonic diatoms comprise only about 3% to 10% of the flora and did not include pelagic species typical of Lake Ontario. The major plankter, *Cyclostephanos dubius*, is a typical small lake species (Håkansson & Regnell 1993) and is never common in the Great Lakes. Thus the diatom evidence shows that Hamilton Harbour during this period was not confluent with Lake Ontario. The evidence for pond-like conditions at this time is in agreement with interpretation of the pollen data (Fig. 3), Ostracoda (Delorme, 1992b) and sediment size analysis (Coakley & Rukavina, 1992). The diatom flora is dominated by the mesothermal and mesoeuthermal groups, indicating relatively shallower and warmer water than present (Fig. 8).

Diatoms were abundant in most samples from this zone. High values for both diatom species richness and diversity in this zone suggest derivation from a wide diversity of habitats, typical of a shallow mesotroph-

ic, moderately alkaline waterbody with submerged macrophytes.

Zone C (720 cm to 645 cm, 7000 BP–6200 BP)

Sharp declines in diatom species richness and diversity at the end of Zone D mark a pronounced shift in diatom assemblage structure, and imply a major change in the limnetic environment. In the early part of Zone C, the abundance of diatoms and chrysophyte cysts drops sharply to a minimum. Both show recovery at 670 cm before declining again.

During this period, strong fluctuations in abundance are also observed in the profiles of major diatom species and the diatom ecological groups. Diatom assemblages in this zone are dominated by benthic species (Fig. 6), however, between 720 cm and 710 cm (c. 7000 BP) there is a sharp increase in the euplanktonic group, which is best interpreted as the

first impingement of Lake Ontario on the harbour. Strong evidence for this is the sharp increase in the euplanktonic taxa *Stephanodiscus niagarae*, *S. alpinus*, *S. transsylvanicus*, *Aulacoseira italica*, *Cyclotella bodanica* and *Tabellaria fenestrata* (Fig. 5); a pelagic assemblage typical of Lake Ontario (Stoermer & Ladewski, 1978). This date is prior to that of Anderson and Lewis (1985), who proposed that the connection between Lake Ontario and Hamilton Harbour occurred at about 5000 BP. Our data indicate at least transitory connection with Lake Ontario since about 7000 BP.

After a brief period of euplanktonic abundance, the benthic diatom group regains dominance signifying decreasing influence of Lake Ontario at the coring site. The short-lived influence of Lake Ontario was a key factor in modifying environmental conditions in the harbour. The diatom-inferred moderately warm water habitat at the beginning of Zone C declines as a result of an influx of relatively cooler Lake Ontario water, and concentrations of nutrients evidently decreased (Fig. 8). In response, the relative abundance of the oligotrophic diatom group rises to above 40%. The alkaliphilous group decreases to about 5%, but recovers to 40% when the influence of Lake Ontario decreased.

Rising water levels in western Lake Ontario, caused by isostatic rebound in the upper St. Lawrence (Anderson & Lewis, 1985; Coakley & Karow, 1994), likely breached Burlington Bar and led to the initial confluence. Subsequent re-isolation was probably dependent upon deposition of eroded material.

*Zone B (645 cm to 83 cm, estimated
6200 BP–280 BP)*

*Subzone B2 (650 cm to 370 cm, estimated
6110 BP–3230 BP)*

Another major shift in the composition of the diatom assemblages is observed at the base of Zone B. Diatom assemblages are once again dominated, except on three occasions, by euplanktonic species. The previously mentioned Lake Ontario pelagic species again become dominant, and another euplanktonic species, *Stephanodiscus transsylvanicus*, is also abundant early in the subzone (Fig. 5). This species is widely distributed in the modern upper Great Lakes system and was abundant in offshore core sediments of Lake Ontario and Lake Michigan (Stoermer *et al.*, 1985, 1990). Evidently the embayment was once again open to influence from Lake Ontario. Clearly, by the start of Zone B,

Lake Ontario water level had reached an elevation permitting permanent confluence with the harbour.

Abundances of all three groups of siliceous microfossils in subzone B2 are low. The low concentrations correspond to the period of the Nipissing Flood (Anderson & Lewis, 1982, 1985), attributed to diversion of upper Great Lakes drainage through Lakes Erie and Ontario. An accelerated water level rise at this time was reported by McCarthy & McAndrews (1988) in Grenadier Pond, Toronto, and also by Yang & Duthie (1995b) in East Lake (eastern Lake Ontario), who also reported a major reduction in the abundance of siliceous microfossils.

In samples around 530 cm (4900 BP), 485 cm (4500 BP) and 395 cm (3500 BP), sharp increases in non-planktonic diatoms and decreases in plankters are evident (Fig. 6). The dominant taxa associated with these events are epiphytes such as *Gomphonema longiceps*, and benthic forms such as *Amphora ovalis*, *Navicula cryptocephala*, *Gyrosigma attenuatum* and *Mastogloia smithii*. In addition, the abundance of the cold water habitat group decreases with increasing abundance of the warm habitat group. Clearly, either the harbour experienced three episodes of temporary isolation, or three events occurred which resulted in the deposition of large numbers of epiphytic and benthic diatoms.

*Subzone B1 (370 cm–83 cm, estimated
3200 BP–280 BP)*

Maximum abundance of microfossils is found in this sub-zone, and diatom species diversity and richness are also high. Euplanktonic taxa consistently dominate, and many benthic species became rare or absent. Diatom-inferred water depth remains at around 15 m. The trophic status is mostly between oligotrophic and mesotrophic. These data suggest that the harbour environment during this period was relatively stable and confluent with Lake Ontario. The continually high abundance of the oligothermal group taxa (about 80% of total) in this period indicates relatively cool water.

Zone A (83 cm to surface, c. 280 BP to present)

Zone A is the last page of the environmental history of the harbour recorded in this core, reflecting anthropogenic disturbances to the harbour environment.

A major shift in diatom species composition occurs in this zone. The oligotrophic and oligomesotroph-

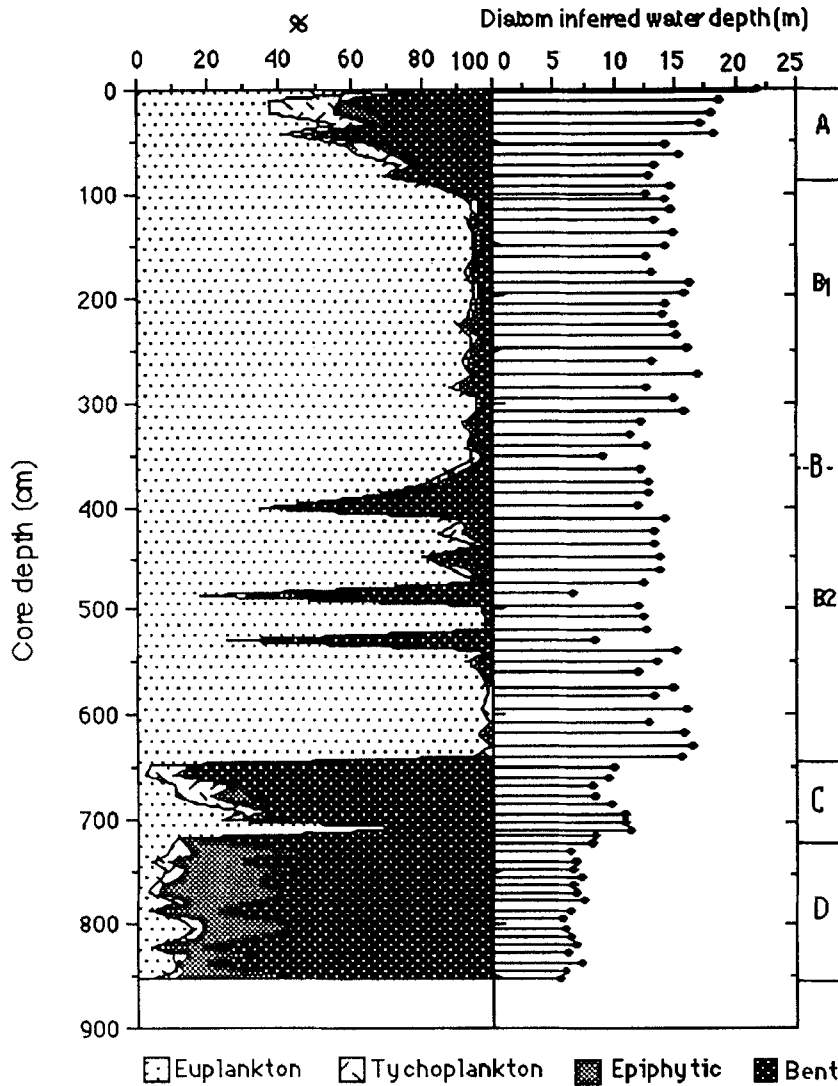


Fig. 6. Stratigraphy (% abundance) of four diatom habitat preference groups, and the diatom-inferred water depth in core HH26comp. Diatom zones are shown on the right.

ic taxa completely disappear and are replaced by a spectrum of mesotrophic and eutrophic indicators such as *Stephanodiscus niagarae*, *S. parvus*, *Cyclotella meneghiniana*, and *Fragilaria capucina* (Fig. 5). Two pollution-tolerant forms (Lowe, 1974), *Actinocyclus nomanii* fo. *subsalsa* and *Diatoma tenue* var. *elongatum*, reach maximum abundances later in the zone.

The data reflect striking changes in the diatom flora resulting from anthropogenic activities in the catchment and in the embayment. The sharp rise in both *Ambrosia* pollen (Fig. 3) and ^{13}C values (Fig. 10) at the beginning of this zone is consistent with the increase

in diatom-inferred trophic status. Chrysophyte cysts and fresh water sponges completely disappear. The interpretation of the impact of human activities in the harbour and catchment reflected in this core is consistent with the pattern described in finer resolution in core HH-137 (Yang *et al.*, 1993). Also of note is the abrupt increase in the mesoeutermal diatom group, and, conversely, the decrease in the oligothermal group (Fig. 8). The inferred increase in temperature is probably attributable to increased human influence on the harbour following deforestation, and increased industrial and urban sewage discharges (Rodgers, 1989).

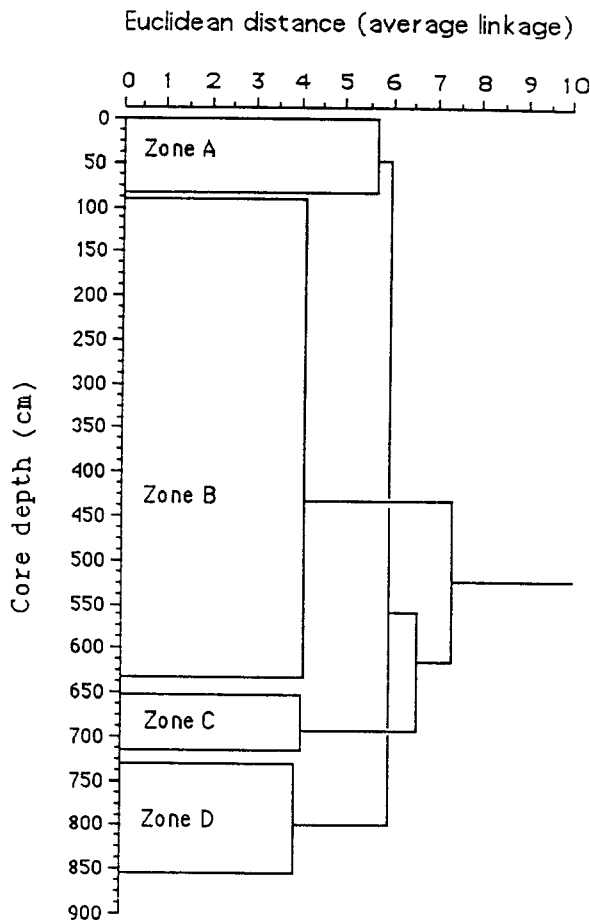


Fig. 7. Cluster analysis of diatom assemblages in core HH26comp. See text.

Isotopic and elemental analyses

The $\delta^{13}\text{C}$ profiles for bulk organic matter and cellulose from the Hamilton Harbour core are strongly correlated (Fig. 9). Although fewer samples of cellulose were analysed, it is clear that a common isotopic signal dominates both profiles and, moreover, that systematic differences between cellulose and bulk $\delta^{13}\text{C}$ values also occur within some intervals. The lowest part of the core (diatom zone D) is characterized by variability, with $\delta^{13}\text{C}$ in both materials varying between about -28 and -26‰ . Cellulose $\delta^{13}\text{C}$ in the lower part of diatom zone C rises rapidly to about -24‰ and then declines to about -30‰ . Bulk organic $\delta^{13}\text{C}$ values within this zone show the same relative changes, although offset to slightly lower values. Bulk organic and cellulose $\delta^{13}\text{C}$ diverge strongly in the lower two-thirds of diatom subzone B2, with cellulose $\delta^{13}\text{C}$ enriched up to 3‰ above

co-existing bulk organic matter, which gradually rises over the middle part of the zone from about -30.5‰ to -29.5‰ . The two records converge at the top of subzone B2, declining to about -32‰ , before diverging again in subzone B1. Bulk organic $\delta^{13}\text{C}$ reaches a marked minimum in the lower part of subzone B1 (c. -32.5‰) followed by stabilization around -31‰ in the upper two-thirds, whereas cellulose $\delta^{13}\text{C}$ oscillates systematically from below -33‰ to above -29‰ , crossing over the bulk organic profile in the middle of the subzone and converging again near the top. The uppermost samples in subzone B1 are marked by abruptly increasing $\delta^{13}\text{C}$ (by about 4‰), with the $\delta^{13}\text{C}$ offset between cellulose and co-existing bulk organic matter briefly increasing to as much as 3‰ . Bulk organic $\delta^{13}\text{C}$ reaches a local maximum (c. -26.5‰) in the lower part of diatom zone A, before stabilizing and converging with cellulose $\delta^{13}\text{C}$ around values of -28‰ to -27‰ in the uppermost part of the core.

The cellulose $\delta^{18}\text{O}$ record is characterized by substantial long-term oscillations that bear no obvious relations to the carbon-isotope profiles. Aside from the lowermost sample, relatively low values (in the range 15 to 17‰) occur in diatom zone D, rising steadily through zone C and remaining generally high (averaging over 19‰) with some variability through the lower two-thirds of zone B, before progressively dropping to a pronounced minimum (c. 16‰) in the upper part of subzone B1. Cellulose $\delta^{18}\text{O}$ rises in advance of the rapid shifts in cellulose and organic $\delta^{13}\text{C}$ to reach intermediate values in zone A, with a small additional increase to the uppermost sample at 11 cm depth (c. AD 1900). Cellulose-water isotopic fractionation in the range 27 – 28‰ yields inferred lake water $\delta^{18}\text{O}$ for this most recent sample in good agreement with the present range of measured $\delta^{18}\text{O}$ values in Hamilton Harbour (E. Harvey, University of Waterloo, personal communication, 1994).

The carbon and nitrogen contents of the sediment fine fraction covary strongly and increase progressively upward, reaching maximum values at the top of the core. The carbon profile is marked by three plateaux corresponding to diatom zone D and subzones B2 and B1, and increasing values in zones C and A, and in the B2–B1 transition. Nitrogen content covaries strongly with carbon content, although with a less pronounced increase in zone A. C:N ratio is uniformly low (in the range 5 – 8) throughout most of the core, with the notable exception of zone A, reflecting the greater relative increase in carbon in the uppermost sediment layers. An abrupt oscillation also occurs at the base

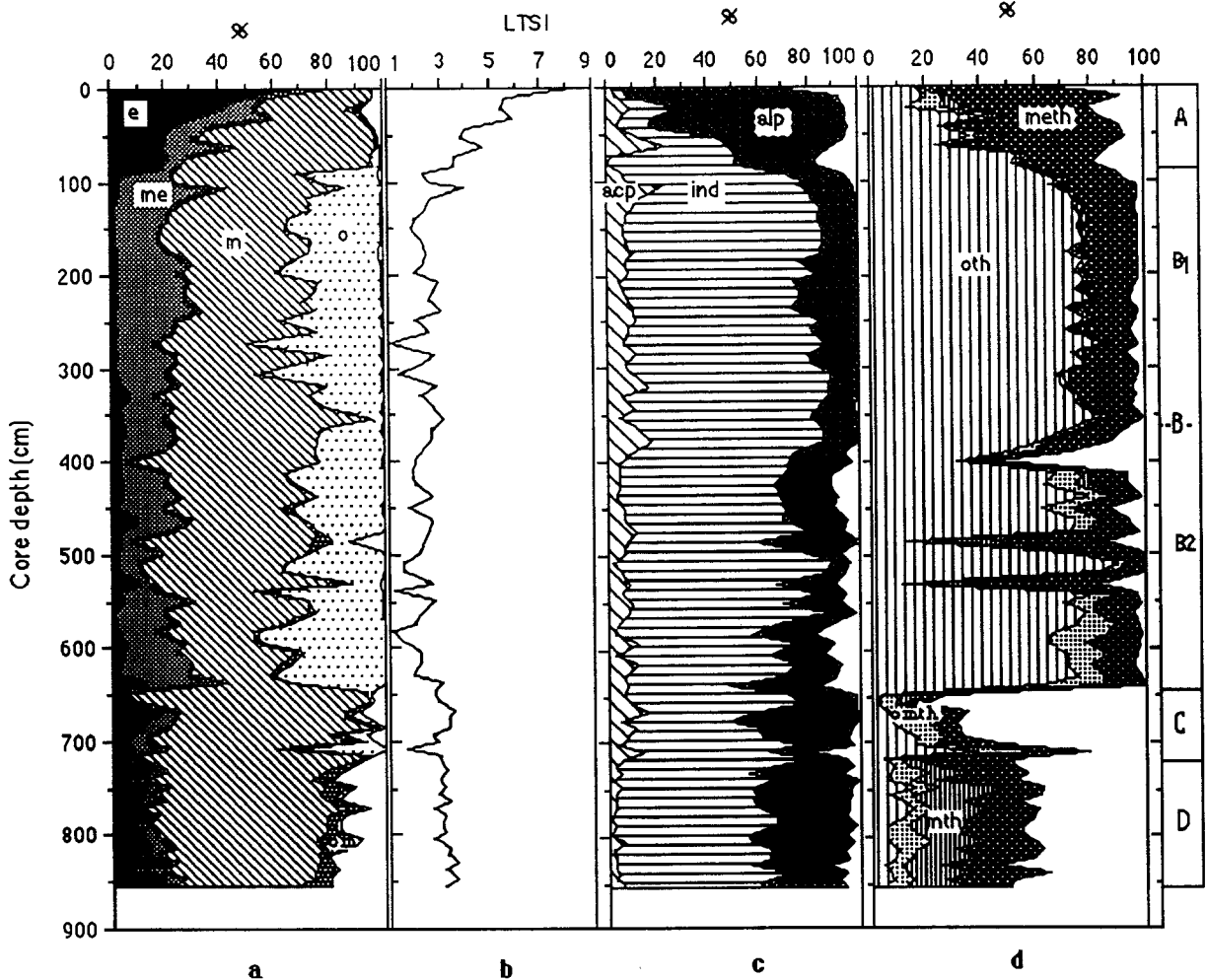


Fig. 8. Diatom-inferred conditions in core HH26comp: a) trophic categories (e=eutrophic, me=mesoeutrophic, m=mesotrophic, o=oligotrophic); b) Lake Trophic Status Index (see text); c) pH preference categories (acp = acidophilous, ind = indifferent, alp = alkaliphilous, unlabelled = unknown); d) temperature preference categories (oth = oligothermal, omth = oligomesothermal, mth = mesothermal, meth = mesoeutermal).

of zone C, although this may be exaggerated, at least in part, by the very low organic matter content at this level.

The carbon isotope and elemental data can be reconciled readily with the sequence of environmental changes suggested by analysis of the siliceous microfossils and the reconstructions of lake levels and paleogeography of Anderson & Lewis (1985) and Coakley & Karrow (1994). Thus the high-frequency variability in $\delta^{13}\text{C}$ and lack of a significant long-term trend in zone D is consistent with conditions in a shallow, relatively productive lake that was not yet significantly influenced by the larger lake into which it drained. Low organic matter content in the fine fraction is attributable

to a combination of high rates of oxidation and dilution by silicate mineral detritus. Rapidly fluctuating $\delta^{13}\text{C}$ during the transitional period (zone C), as regular incursion of Lake Ontario waters began to occur, may in part reflect the changing productivity in Hamilton Harbour suggested by the diatom-based lake trophic status index, but is probably a more direct measure of the varying influx of ^{13}C -depleted DIC from the larger lake. Dominance by lake-exchange effects is suggested also by the varying abundance of diatom temperature and trophic groups in this zone, and perhaps indirectly by the abrupt fluctuations in C:N ratio.

Maximum lake-exchange effects are suggested in zone B2 by the persistent offset between bulk organic

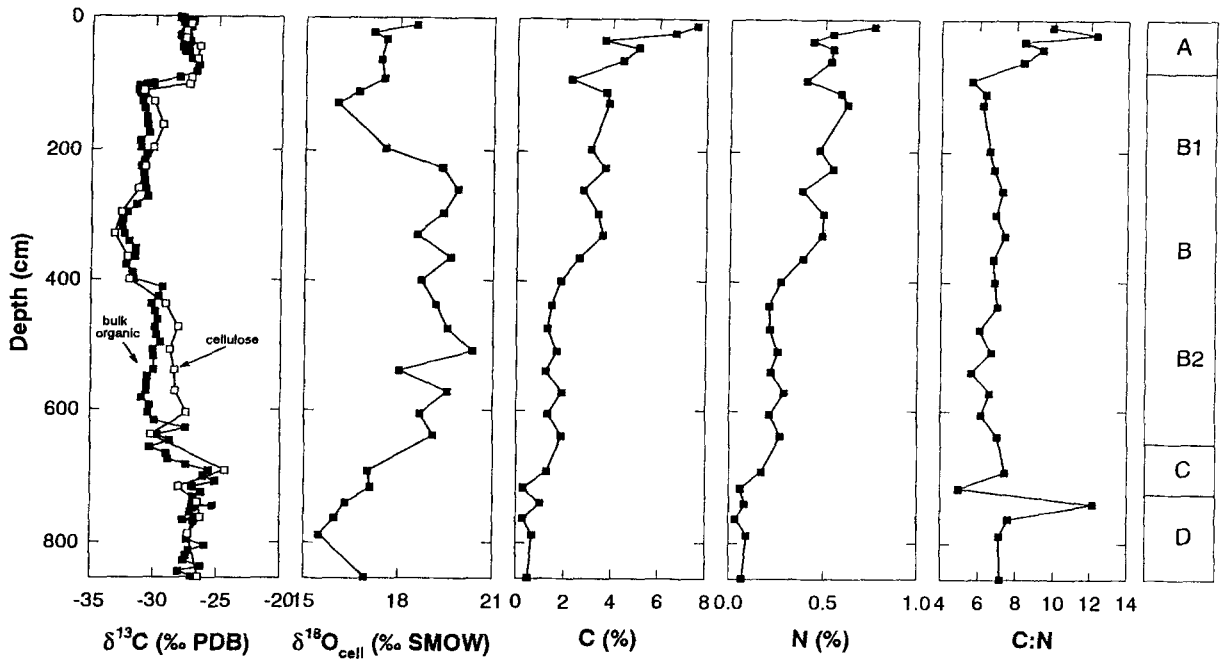


Fig. 9. Isotopic and elemental data for Hamilton Harbour core HH26.

and cellulose $\delta^{13}\text{C}$, ending suddenly at about 420 cm (c. 4000 BP), which corresponds with the end of the Nipissing Flood highstand as reconstructed by Anderson & Lewis (1985). Reduced $\delta^{13}\text{C}$ values in the lower part of zone B1 are consistent with a significant reduction in Lake Ontario influence at this time, leading to an increase in the relative importance of ^{13}C -depleted DIC derived from degradation of organic matter in harbour bottom waters and sediments. The subsequent increase in bulk organic $\delta^{13}\text{C}$ and re-establishment of less negative cellulose $\delta^{13}\text{C}$ in the upper part of zone B1 is in harmony with the subsequent strengthening of lake-exchange effects as a result of the continuing isostatically controlled increase in Lake Ontario water level. Increased organic matter accumulation in the sediments at the zone B2-B1 transition probably reflects enhanced oxygen depletion in bottom waters because of the decrease in Lake Ontario influence. The persistence of relatively high organic content in the later part of B1 may be an indication that sufficient oxygen demand developed in bottom waters and sediments to largely counteract the effects of renewed flushing. However, any such ongoing natural eutrophication of the system was radically accelerated by human activity, as shown by the abrupt shifts in bulk organic and cellulose $\delta^{13}\text{C}$, organic matter content, and C:N ratio near the top of the core.

The cellulose $\delta^{18}\text{O}$ record provides some additional insight into the paleoenvironmental history of Hamilton Harbour, as well as confirming the chronology and placing the core record within the broader framework of regional climatic and hydrologic change. Comparison of the Hamilton Harbour $\delta^{18}\text{O}$ chronology with the reconstructed history of meteoric water $\delta^{18}\text{O}$ in southern Ontario (Fig. 10) shows unequivocally that lake water $\delta^{18}\text{O}$ was fundamentally controlled by the changing isotopic composition of precipitation falling in the region, with generally minor overprinting by secondary hydrologic effects. The only strong evidence of hydrologic overprinting is the marked positive offset of inferred lake water $\delta^{18}\text{O}$ near the base of diatom zone D, which is attributable to evaporative enrichment of the water in the shallow lake that existed at this time, as occurred in other small lakes in southern Ontario under dry early-Holocene climate (see Edwards & Fritz, 1988). Subsequent decrease in lake water $\delta^{18}\text{O}$ probably reflects dampening of evaporative enrichment due to hydrologic opening of the basin in response to moistening climate. Influence from Lake Ontario (zone C) was marked by only a slight increase in lake water $\delta^{18}\text{O}$, perhaps reflecting somewhat higher evaporative enrichment in the larger water body and (or) a somewhat lower average $\delta^{18}\text{O}$ of precipitation falling on the catchment area of the

smaller lake. Subsequent changes in lake water $\delta^{18}\text{O}$ in zones B and A seem to have closely followed the general trend of changing meteoric water $\delta^{18}\text{O}$, with some small variation possibly related to moisture changes, as suggested by the greater apparent relative enrichment at present than at the height of Holocene moistness around 4000 BP, or to other hydrologic factors, such as influx of isotopically enriched surface waters during the Nipissing Flood (Edwards *et al.*, 1994b), which might account for the high $\delta^{18}\text{O}$ value at about 4700 BP. Although unambiguous interpretation of late-Holocene fluctuations in lake water $\delta^{18}\text{O}$ is not possible without a better constrained meteoric water $\delta^{18}\text{O}$ history, they are undoubtedly related to changes in climate and hydrology that have also affected Lake Ontario water levels (e.g. see McCarthy & McAndrews, 1988; Flint *et al.*, 1988; McCarthy *et al.*, 1994).

There is no reason to believe that Hamilton Harbour has unusual cellulose fractionations. The isotopic and elemental data are consistent with observations elsewhere in both small and large lakes (Edwards & McAndrews, 1989; Edwards *et al.*, 1994b; Meyers, 1994); i.e., cellulose in the fine-grained organic fraction seems to be generated entirely in the overlying water by phytoplankton, which is subsequently grazed by zooplankton and incorporated into the sediments in fecal pellets. There is nothing in our data to suggest anything different. Indeed, the remarkable correspondence between sediment cellulose oxygen isotope composition and inferred precipitation oxygen isotope composition is a striking testament to the absence of complicating factors.

Summary and conclusions

Evidence from the sedimentary diatom record and isotopic analyses clearly demonstrates that the site of the present-day Hamilton Harbour has developed over the past 8300 y from a shallow pond, through a deeper waterbody confluent with Lake Ontario, to the anthropogenically modified embayment it is today.

The environment of the earliest phase, from 8300 BP to about 7000 BP (all ^{14}C y before AD 1950), was that of a shallow mesotrophic pond of moderate alkalinity, probably warmer than present, and probably with extensive marginal wetlands and aquatic macrophytes.

The initial transitory connection with the rising water level of Lake Ontario at about 7000 BP had a major impact on the lacustrine environment; the water

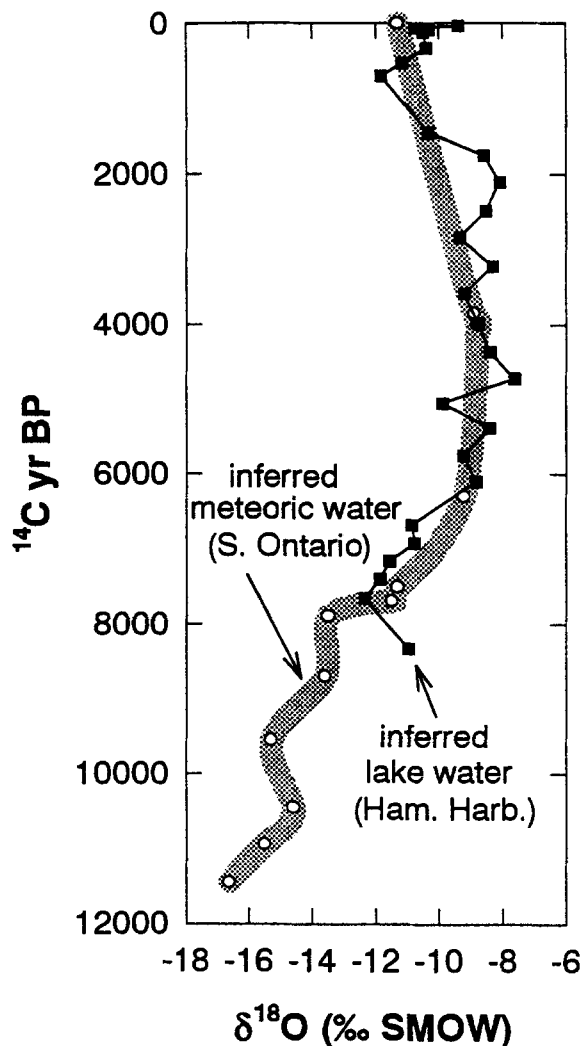


Fig. 10. Comparison of inferred lake water $\delta^{18}\text{O}$ in Hamilton Harbour with the meteoric water $\delta^{18}\text{O}$ history for southern Ontario reconstructed from fossil wood cellulose and lake sediments (Edwards & Fritz, 1986, 1988). The lake water record assumes a cellulose-water fractionation of 1.0282 (Edwards & McAndrews, 1989). Note that the meteoric water $\delta^{18}\text{O}$ history provides a good first-order proxy for precipitation contributing to Hamilton Harbour, but that detailed quantitative comparison of the records would also require consideration of the increasing effective catchment area contributing to the harbour as drainage changed from local to regional in scale.

became deeper and cooler, and trophic level and alkalinity declined. This connection is 2000 years earlier than proposed by Anderson & Lewis (1985). A possible explanation for this may be that the outlet channel connecting the harbour with Lake Ontario was deeper than previously assumed. Blockage of this outlet channel with material eroded by the rising waters of

Lake Ontario would account for the temporary nature of the connection. A somewhat parallel situation exists today in the outlet channel connecting East Lake with eastern Lake Ontario; here backflow occurs in periods of high Lake Ontario water level, carrying in euplanktonic diatoms (Yang & Duthie, 1995b).

Permanent connection with the continuously rising water levels of western Lake Ontario was established at around 6000 BP. Trophic level and water temperatures decreased relative to previous conditions.

An unstable environment in Hamilton Harbour during the period of the Nipissing Flood are clearly evident in the diatom assemblages and the isotopic record. Diatom species abundance and species richness reach a minimum at 4400 BP and productivity fluctuates. Decreased mixing of Lake Ontario water from about 3800 BP following the Nipissing Flood highstand is evidenced in the isotope data and a recovering diatom abundance.

The diatom record shows three exceptional events at around 4900 BP, 4500 BP and 3500 BP. During these events benthic and epiphytic forms replace the dominant plankters, and diatom-inferred water depths decrease and diatom-inferred temperatures increase. These events may be attributed to brief periods of low water levels, or changing sedimentation patterns, or storm deposits, or extreme turbidity events reducing the relative importance of planktonic diatoms. Low water level is the least likely explanation, since the events occurred during the Nipissing Flood when the climate was rather moist (Edwards & Fritz, 1988) and average lake levels high. Given the low abundance and diversity of diatoms in the sedimentary record relating to the Nipissing Flood, an explanation involving periods of high turbidity must be regarded as the most plausible. Yang & Duthie (1995b) also reported a greatly reduced diatom record in East Lake coincident with the Nipissing Flood.

Maximum numbers of microfossils were found in the sedimentary record between 3200 BP and 280 BP. Diatom-inferred evidence indicates deposition in a moderately alkaline, oligotrophic to mesotrophic embayment, with relatively cool water and dominance by euplanktonic diatoms indicating major interaction with Lake Ontario.

Profound changes to the Hamilton Harbour environment are evidenced in the final 280 years of sedimentary record, encompassing the period of anthropogenic influence. A significant increase in trophic level is shown by changes in the diatom flora and by the increase in both bulk and cellulose ^{13}C . Chryso-

phyte cysts and sponges disappear from the record. Diatoms tolerant of organic pollution reach maximum abundance in sediments after 1850 AD. A finer resolution of changes to the Hamilton Harbour environment during the past 330 years was described by Yang *et al.* (1993) from the analysis of a short core from a different site in the harbour.

Water depth changes in Hamilton Harbour and western Lake Ontario are summarized in Fig. 11. The depth and age of the permanent connection (*c.* 6000 BP) agrees with the lake-level curve of Anderson & Lewis (1985). As noted above, the difference in timing from their estimate of 5000 BP can be accounted for by the existence of a deeper outlet channel. According to Anderson and Lewis, at the time of connection Lake Ontario was about 26 m below present level. Thus in the last 6000 y water level increased 26 m but about 6 m of sediment accumulated, so the actual increase in water depth was about 20 m. Given the modern depth of 21 m, this suggests that the depth at the coring site was only about 1 m, agreeing with the estimate of Coakley & Karrow (1994) and Delorme (1992b).

The water depth history of Hamilton Harbour suggested by our data may be summarized as follows:

8300 BP–6000 BP: The depth prior to and at the time of permanent connection must have been constant around 1 m–2 m, less sediment accumulation.

6000 BP–c. 4500 BP: By the time of the Nipissing Flood highstand water levels were probably around the modern level; i.e. maximum water depth at the coring site rose rapidly to 25 m–26 m (21 m modern depth plus 4 m–5 m sediment).

c. 4500 BP–3000 BP: Lake level dropped to about 8 m below present by 3000 BP. Accounting for about 3 m sediment that subsequently accumulated, water depth must have been about 16 m.

3000 BP–present: Water levels have been rising faster than the rate of sedimentation for the past 3000 y, resulting in a net gain of about 5 m to the present depth of 21 m.

The above water depth estimates derived from physical data permit a test of diatom-inferred water depths. It is evident that diatom-inferred depths may be misleading in dynamic systems or in periods of transition. However, diatom-inferred depths during periods of relative stability, such as in diatom zone B1, are a good approximation. Furthermore, the diatom-inferred surface depth of 21 m agrees with the actual measured depth. Clearly, diatom-inferred water depths should be interpreted with caution, nevertheless, changes in

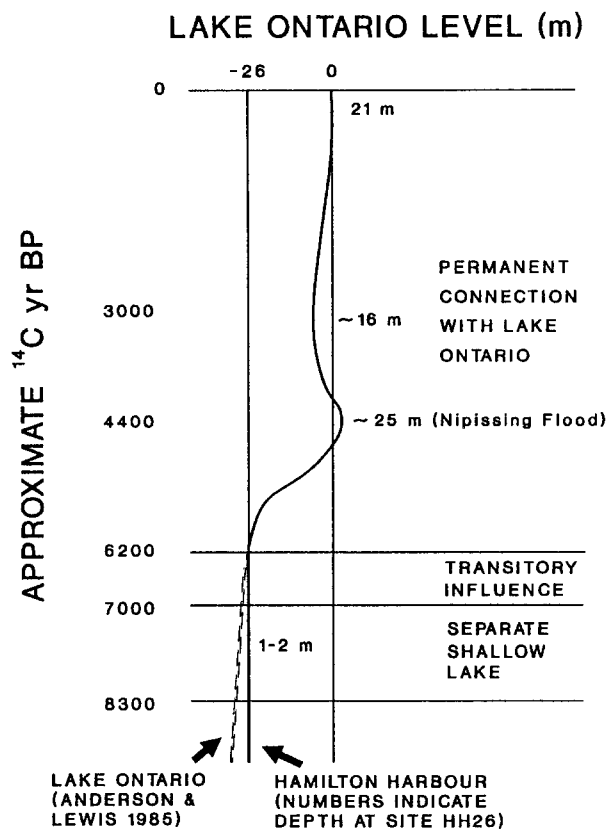


Fig. 11. Sketch of water levels in Lake Ontario and Hamilton Harbour from 8300 y BP to present. Lake Ontario water levels from Anderson & Lewis (1985). See text for explanation.

diatom-inferred depth are useful markers of changes in the physical environment.

Acknowledgments

We thank L. D. Delorme and N. Rukavina for providing the material from core HH26comp, J. P. Coakley and P. F. Karrow for the stratigraphic description, S. Marsters for the pollen analysis, and R. J. Elgood, H. Weatherly, D. Cieslewicz, and P. Van Wesenbeeck for isotopic analyses. The research was funded by a grant from the Great Lakes University Research Fund.

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