The Early Lake Ontario barrier beach: evidence for sea level about 12.8–12.5 cal. ka BP beneath western Lake Ontario in eastern North America

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During retreat of the Laurentide Ice Sheet (LIS) in eastern North America, and as the glacio-isostatically depressed continental crust was beginning to rebound, rising marine waters invaded the St. Lawrence Valley (Fig. 1). Intercalations of marine muds with glacigenic deposits indicate that continued ice retreat led to incursion of marine waters into the upper St. Lawrence Valley and the tributary valleys of Lake Champlain and lower Ottawa River to form the Champlain Sea (CS; Fig. 1; Parent & Occhietti 1988; Occhietti 1989).

The invading marine water (CS) was thought to have been related to Early Lake Ontario (ELO) (Fig. 2) on the basis that inferred water levels appeared to be at similar elevations and might be connected through a narrow strait (Fig. 3A, B; Sharpe 1979; Anderson & Lewis 1985, 2012; Pair et al. 1988; Pair & Rodrigues 1993). Abundant meltwater supply from glacial lakes Algonquin and Agassiz farther west probably supplied sufficient inflow to maintain freshwater fauna in ELO (Anderson et al. 1985; Hladyniuk & Longstaffe 2016).

In the late 1960s and early 1970s, a sand and gravel palaeo-barrier beach was recognized beneath deep-water mud of western Lake Ontario from echo-sounding records and sediment piston cores during early scientific exploration of offshore Lake Ontario sediments (Lewis & Sly 1971; Anderson & Lewis 1985). The beach was initially referred to as the Grimsby–Oakville bar or barrier beach after towns on the southwestern and northwestern lake shores by Anderson & Lewis (1985, 2012) and Coakley & Karrow (1994). Its crestal elevation was estimated from echo-sounding records as –5±5 m above present sea level (a.s.l.) or 79±5 m below the present Lake Ontario mean surface (Lewis & Sly 1971).

In this paper, we name the sand and gravel deposit the Early Lake Ontario (ELO) barrier beach after the water body in which it formed. The name ELO is restricted here to the water body that existed in the short period of a few centuries following local glacial lake phases while the lake received freshwater discharge from upstream glacial lakes and overflowed its basin before it evaporated into the closed Lake Ontario lowstand following ELO. The term ‘Early Lake Ontario’ was used previously to refer generally to the early history of Lake Ontario (Hough 1958: p. 204; Prest 1970: fig. XII-16). Once the elevation, composition and chronology of the ELO beach are presented, the question of whether and how this landform could have been constructed at the level of the CS is addressed. Further, this paper answers questions concerning the coincidental combination of mechanisms and their duration that led to the maintenance of a common level between the CS and ELO and to construction of the palaeo-barrier beach as a marine-level marker with a sea-level index point.
Regional setting and background

Lake Ontario, the lowest of the Laurentian Great Lakes with a mean elevation of 74.6 m a.s.l. (GLERL 2017) and maximum depth of 244 m (Virden et al. 1999), trends 300 km east–west with a maximum width of 85 km, to drain northeastward with a mean flow of 7000 m³ s⁻¹ (Mortsch et al. 2000) via the St. Lawrence River to the Atlantic Ocean (Figs 1, 3A). The Lake Ontario basin was excavated by Tertiary river erosion and multiple Quaternary glaciations from relatively soft lower Palaeozoic sedimentary rocks that dip gently southward into the Appalachian Basin and overlie harder crystalline rocks of the Grenville Province of the Precambrian Shield (Hough 1958; Sanford & Bauer 1983). Exposed Shield rocks extend south of the upper St. Lawrence River in the Frontenac Arch to the Adirondack Mountains, generally above 300 m a.s.l. (Franzi et al. 2016;...
Fig. 3A, B). Constrictions and rapids in the harder rocks of the Frontenac Arch in the uppermost reaches of the St. Lawrence River controlled water levels in the present Lake Ontario naturally before the river was dammed for navigation and power generation purposes.

Deglacial history

The LIS margin had retreated northeast of the Lake Ontario basin by about 12.4 $^{14}$C (14.5 cal.) ka BP, but remained in the upper St. Lawrence Valley against the northern slope of the Adirondack Mountains. This position impounded high-level glacial Lake Iroquois in the Ontario basin (Figs 2, 4A) that drained eastward via the Mohawk River Valley into the Hudson River Valley and Atlantic Ocean (Fig. 3A) (Chapman & Putnam 1984; Muller & Prest 1985; Barnett 1992; Franz et al. 2016). Lake Iroquois isobases rise north-northeastward from 110 to 300 m a.s.l. near the northern slope of the Adirondack Mountains (Fig. 3A; Bird & Kozlowski 2016). North-northeast differential uplift of the basin caused water levels to transgress south and west, and to regress north and east of the Mohawk River outlet during the life of Lake Iroquois (Figs 3A, 4A; Coleman 1936). Thus, the isobases west of the outlet are based on elevations of the youngest shoreline. Isobases north and east of the outlet are based on elevations of the earlier shoreline of the Main Iroquois phase described by Pair & Rodrigues (1993) and are older, possibly by a few hundred years.

Radiocarbon ages of Lake Iroquois (as young as 11 510 $^{14}$C (13 330 cal.) years BP) and Lake Ontario (older than 10 150 $^{14}$C (11 770 cal.) years BP) were obtained by Karrow et al. (1961). A detrital wood sample dated to 12 290 $^{14}$C years (14.3 cal. ka) BP, average of replicate analyses, and another dated to 12 100 $^{14}$C years (14.0 cal. ka) BP were obtained from Lake Iroquois beaches near Lewiston and eastward at Malloy, New York (Fig. 3A; Muller & Calkin 1993). Terasmae (1980) inferred that glacial Lake Iroquois existed from about 12 500 to 11 800 $^{14}$C years (14.7–13.6 cal. ka) BP based on pollen zonation and basal bulk gyttja dates in small inundated basins north of central Lake Ontario. Recent re-examination of these pollen data and correlation to more recent radiocarbon dates on specific plant fossils by T.W. Anderson (pers. comm. 2017) suggest the bulk gyttja dates were too old and the apparent ages of basin isolations could be up to 0.8 millennia younger. These revised younger ages for Lake Iroquois are in accordance with recently obtained accelerator mass spectrometry (AMS) radiocarbon ages of wood in till and of a peccary fossil recovered from ice-marginal sand, which indicate ice (possibly floating across Lake Iroquois in a southwest direction) last advanced onto the southern side of the Lake Ontario basin prior to 13 000 cal. years BP (Young & Owen 2017; R.A. Young, pers. comm. 2018). Possibly, floating ice over the lake broke up before the onshore grounded ice and allowed the Iroquois shore features to reform by ~13.0 cal. ka BP, the inferred youngest age for the lake. Glacial Lake Agassiz in central North America began discharging into the Lake Superior basin of the Upper Great Lakes about the same time (Fig. 1 inset; Clayton 1983; Teller & Thorleifson 1983; Teller 1985), supported by new $^{10}$Be surface exposure ages of 13.1±0.1 to 12.7±0.3 cal. ka BP for opening of ice barriers (Leydet et al. 2018). The Agassiz waters augmented glacial Lake Algonquin discharge from downstream basins of lakes Michigan and Huron already flowing through the Ontario basin (Figs 1, 3A, 4A).

Retreat of the ice front from the Adirondack Mountains opened lower routes for drainage via the Lake Champlain Valley and the Hudson River Valley. As falling glacial lake waters stabilized, they first formed glacial Lake Frontenac in the Lake Ontario basin, and then they descended through the Sydney(?), Belleville and Trenton glacial lake phases (Fig. 2; Mirynech 1962; Prest 1970; Muller & Prest 1985; Pair & Rodrigues 1993; Rayburn et al. 2005). Shoreline features below the Frontenac level, originally interpreted as the Sydney glacial lake phase by Mirynech (1962), have been attributed to intermittent, falling water levels (Pair et al. 1988). The duration of the first post-Iroquois phase, Lake Frontenac, is estimated to be about 50 years, and that of the remaining post-Iroquois time to be about 170 years, based on compositional changes in varved clay deposits and varve counts in the Lake Champlain Valley (Rayburn et al. 2011; Franz et al. 2016).

While ice blocked the lower St. Lawrence Valley near Warwick, Quebec (Parent & Occhietti 1999), the ice margin and glacial lake waters in the Ontario basin receded northeastward, towards the Ottawa River Valley to form glacial Lake St. Lawrence (Figs 1, 4B; Parent & Occhietti 1988; Occhietti 1989; Rodrigues 1992). This lake was confluent throughout the basins of Lake Ontario (glacial Lake Trenton), Lake Champlain (Fort Ann phase of glacial Lake Vermont), lower Ottawa River (glacial Lake Candona) and central St. Lawrence Lowland (glacial Lake Candona).

Near the vicinity of the ice dam impounding Lake St. Lawrence, sediments with marine fauna over freshwater glaciolacustrine varves denote entrance of seawater to the upper St. Lawrence Valley (Parent & Occhietti 1999). Failure of the ice dam, possibly facilitated by an Agassiz outburst flood (Teller 1985; Leydet et al. 2018), allowed 40–60 m of impounded lake water (Fig. 4B) to discharge rapidly to the Atlantic Ocean (Parent & Occhietti 1988, 1999; Rayburn et al. 2007, 2011) during the early part of the GS-1 stadial event in Greenland (12 900–11 700 cal. years BP; Rasmussen et al. 2014). Atlantic Ocean waters then entered the upper St. Lawrence region to establish the CS (Figs 2, 3B, 4C).

Ridge et al. (1999) and Ridge (2004) date the CS incursion at 11.1–10.6 $^{14}$C ka or 13.1–12.7 cal. ka BP based on the New England varve chronology calibrated with AMS $^{14}$C plant fossil dates and palaeomagnetic
secular variation. Richard & Occhietti (2005) date the marine incursion also at 13.1–12.7 cal. ka (11 100±100 $^{14}$C years) BP based on AMS $^{14}$C dating of terrestrial plant material and ice recession rates from the St. Lawrence Valley. The incursion of the main marine phase in the Lake Champlain Valley at 13.12–12.85 cal. ka BP (re-calibrated here to 13.07 to 12.74 cal. ka BP) is based on a wood fragment recovered from marine
The onset of ELO is coeval with the CS incursion, implied by counts of post-Iroquois varves in the Lake Champlain Valley (Rayburn et al. 2011; Franz et al. 2016), estimated about 13 000–(50+170) = 12 780 years or ~12.8 cal. ka BP assuming Lake Iroquois began ~13.0 cal. ka BP. As shown later, the ELO onset date is consistent with ages of events during and at its termination. Carbonate marine fossils have been shown to be up to 1800 years too old (Occhietti & Richard 2003) and are not used here.

**Postglacial lacustrine and marine history**

Abundant meltwater from glacial lakes Algonquin and Agassiz was draining to ELO via both the Trent River Valley and the Lake Erie basin (Fig. 4C; Eschman &...
Karrow 1985; Finamore 1985; and references in both papers), as both outlets were at similar elevations at the time of Main Lake Algonquin (Lewis & Anderson 1989). These inflows probably added sufficient freshwater outflow from the Ontario basin to resist entry of marine waters into ELO. Anderson et al. (1985), Pair & Rodrigues (1993) and Hladyniuk & Longstaffe (2016) report only freshwater molluscs and ostracodes and an absence of marine fauna in Ontario basin sediments.

In the Great Lakes region, uplifted late glacial and postglacial lake shorelines owing to glacial isostatic adjustment are distinctly concave upward indicating faster uplift towards the north in the direction of thicker and longer-lasting ice. Shoreline synthesis suggests that land rebound through time in the region can be described generally by a negative exponential function with a relaxation time of ~3700 years (Lewis et al. 2005). Eastward in Quebec, a subdued similar pattern is evident in the local isobases of glacial Lake St. Lawrence (Candona) and a predecessor glacial lake (Parent & Occhietti 1999), suggesting a continued nearby ice load and crustal depression. Eastern Lake Ontario rebound, operating faster than sea-level rise, raised the Lake Ontario basin above sea level and caused marine Atlantic waters to recede to the present brackish estuary during Late Pleistocene and Holocene time.

Farther east, both transgressive and regressive postglacial sea-level histories are evident (Shaw et al. 2002), probably resulting from a complex deglaciation history with glaciers remaining on adjacent highlands long after a central corridor (St. Lawrence Valley and Gulf) was deglaciated (Quinlan & Beaumont 1981; Parent & Occhietti 1999; Shaw et al. 2002). Rebounded surfaces to the south in New England are tilted but flat-

Tidal ranges in the Gulf of St. Lawrence (GSL) are mainly <2 m, and thus, the Gulf is a microtidal environment. Maximum tidal ranges increase to 3.5 m entering the St. Lawrence Estuary and range up to 5.8 m in the St. Lawrence River narrows near Quebec City (Fig. 3A). However, tidal ranges are greatly reduced to 0.5 m upriver at Trois Rivieres because of dissipation of energy at the narrows and the flooding effect in a nearby widening of the river (Fisheries and Oceans Canada 2017).

A global palaeotidal model suggests a tidal amplitude of ~0.75 m in the GSL during construction of the ELO barrier (Uehara et al. 2006), supported by independent global and regional palaeotidal modelling applied to the western Atlantic Ocean by Hill et al. (2011). Both models suggest that palaeotides up the St. Lawrence Valley would have been forced by tides of <1 m.

The palaeotides probably diminished as they propagated up the St. Lawrence Valley, as do the modern tides. Some amplification may have occurred as tides passed the narrow part of the Goldthwait Sea (Fig. 1), with a subsequent decline in the much wider CS because of the basin flood effect (Shennan 2015). The tide may have amplified and dissipated again as it passed through the narrow strait connecting to ELO and probably declined significantly thereafter in the wide lake. Overall, the inferred palaeotidal range in ELO was in the low microtidal range, <0.5 m.

Sedimentological logging and palynological analysis of Lake Ontario sediment cores showed that ELO descended into hydrologically closed after constructing its barrier (Fig. 4D; Anderson & Lewis 2012; Lewis 2016). The descent of lake levels occurred in the prevailing dry climate indicated by the presence of pollen of drought-tolerant tree species (e.g. Shuman et al. 2002) and was facilitated by northward ice retreat and diversion of the Algonquin and Agassiz meltwater discharge to the Ottawa River Valley after about 10 500 (~12 500 cal.) years BP (Figs 2, 4D; Karrow et al. 1975, 1995; Eschman & Karrow 1985). By, or before, this time, relative uplift had raised and isolated the Lake Ontario basin above the CS. Early Holocene dry climate reduced water levels in most, if not all, basins of the Great Lakes (Lewis & King 2012; Lewis 2016).

After about two millennia, water level in the hydrologically closed Lake Ontario lowstand began rising with increasing meteoric water supply and overstepped the ELO barrier beach about 9500 years ago (Fig. 2; Anderson & Lewis 2012). Deepening water led to mud deposition over the former beach. Rebound of the outlet sill in the northeast was faster than the rest of the basin so that as the outlet sill uplifted and the lake ponded against it the water level continued rising in western Lake Ontario until it reached its present elevation and built the present Burlington beach.

Material and methods

Seismic survey data

Research surveys of Lake Ontario between 1968 and 1974 produced many transects across the ELO barrier using mainly Kelvin Hughes 26B sounders operating at a frequency of 14.25 kHz and broad-band boomer sources. Piston coring during those years revealed that the upper surface of the ELO barrier was composed of sorted sand and gravel (Figs 5, 6, Tables 1, 2).

A 1992 grid survey of western Lake Ontario, using a high-resolution sub-bottom boomer profiler (IKB-Seistec) and a 10 or 40 cu. in. (167.8 or 671.1 cm³) sleeve gun seismic reflection system, revealed internal reflections within the ELO barrier and the thickness of the barrier above glacial sediments (Fig. 6A, B).

Navigation for the 1992 survey used a global positioning system (GPS) with back up LORAN C (Lewis et al. 1995). An onshore GPS receiver confirmed that small errors introduced by U.S. Department of Defense availability limitation were insignificant. Line fixes and coring
stations were positioned with an estimated accuracy of 30–100 m. Earlier survey lines and coring stations were positioned relative to shore features using ship’s radar with estimated lesser accuracies of about 500 m or better. The bathymetry map compiled by the U.S. National Oceanic and Atmospheric Administration (NOAA; Virden et al. 1999) is used here as a source for water depths in Lake Ontario. Elevations of lake-bottom sites were computed by subtracting their water depth from 74.0 m, the inferred vertical datum for the NOAA bathymetry map (Lewis et al. 2018). Depths to sub-surface features of interest were measured from echograms or seismic profiles. These depths were added to the site bathymetry to determine sub-surface feature elevation with respect to present sea level.

**Sediment cores**

Sediment cores that provide information about the ELO barrier beach are described briefly in Table 1 and are located in Fig. 5. In 1992, sediments were recovered with a piston-corer with a 1364 kg head weight and barrel lengths of 12 m or longer. A 2-m gravity corer recovered near-surface sediments at piston core sites. Sediments recovered in earlier years used piston cores with 12 m barrel lengths and a 454 kg head weight. Wood in core 69-0-16-1 (Table 1, entry 10; Fig. 5) was dated by AMS. This sample and other dates in this paper were calibrated using Calib 7.0.2 (Stuiver & Reimer 1993; Reimer et al. 2013).

**Water flux**

Flows of ice-melt and precipitation were obtained and modified from Licciardi et al. (1999). Rough estimates of the height (or depth) of water needed for a given channel width by the flow of ELO through outlets were obtained by simulating flows over broad-crested contracted weirs, as used by Tinkler et al. (1992) to estimate the flux of outburst and baseline meltwater flows of lakes Agassiz and Algonquin between the Lake Erie and Lake Ontario...
Fig. 6. Seismic reflection profiles of the Early Lake Ontario (ELO) palaeo-barrier beach. Sediments of the ELO barrier beach are indicated by transparent yellow overlays. Depths on right side of profiles are metres below 74.0 m a.s.l. A. Seistec record of the northern part (Griffon 92800, day 235, 1126–1146). B. Seistec record of central part (Griffon 92800, day 238, 1300–1330) in which internal clinoform reflections illustrate progressive washover deposition events in the transgressive barrier beach. C. Kelvin Hughes 14.25 kHz sounder record of southern part (Limnos 70-0-29, day 160, 1701–1753 EDT). Profile locations indicated on Fig. 5 and listed in Table 2.
Final deviation is a combination of the deviations of beach crest and beach base elevations (0.7 and 1.4 m) and \( \frac{1}{2} \) of the uncertainty (1.9 m) of the elevation of the NOAA bathymetry map datum (Virden et al. 1999; Lewis et al. 2018), computed as \((0.7^2 + 1.9^2)^{0.5}\) and \((1.4^2 + 1.9^2)^{0.5}\), respectively. The datum uncertainty as assessed in Lewis et al. (2018) is divided equally here between uncertainty in ship’s position and water depth measurement.

**Table 1.** Core locations (arranged south to north) in vicinity of Early Lake Ontario palaeo-barrier beach, water depth, and simplified lithology.

| Entry no. | Core no. | Latitude (°N) | Longitude (°W) | Water depth (m) | Simplified lithology |
|-----------|----------|---------------|----------------|-----------------|---------------------|
| 1         | 69-0-13-13 | 43.269        | 79.350         | 63.8            | Silty clay, soft, black FeS bands and small mollusc shells with gritty clay in upper 12 cm (0–346 cm); stony till, hard, dark grey (346–372 cm). |
| 2         | 69-0-13-12 | 43.27         | 79.386         | 62              | Sandy-silty clay, soft, with black FeS bands and small mollusc shells (0–290 cm); stony till, calcareous (290–360 cm). |
| 3         | 69-0-13-11 | 43.2915       | 79.422         | 75              | Silty clay, soft, black FeS bands, small mollusc shells (0–350 cm); silt and sand, laminated with reddish bands, no or little FeS banding, glaciolacustrine sediment? (r350–450 cm). |
| 4         | 69-0-13-10 | 43.305        | 79.52          | 76              | Silty clay, soft, grey, FeS banding, small mollusc shells (0–475 cm); sand with clay bands (475–500 cm); sand with basal gravel, compact (500–650 cm). |
| 5         | 69-0-13-9  | 43.321        | 79.523         | 78.1            | Silty clay, soft, grey, FeS bands (0–450 cm); fine sand, compact, laminated, some plant detritus and shells (450–567 cm); sand, homogeneous (567–842 cm). |
| 6         | 69-0-13-8  | 43.339        | 79.577         | 74.6            | Silty clay, soft, dark grey, FeS bands, wood fragments and small mollusc shells (0–469 cm); medium sand, with pebbles and cobble, grey, calcareous, (469–733 cm). |
| 7         | 92800-3    | 43.362        | 79.587         | 73              | Silty clay mud, soft, dark grey, FeS bands (0–272 cm); pebbly coarse sand, shell hash (272–274 cm). |
| 8         | 69-0-13-7  | 43.368        | 79.587         | 73              | Silty clay with 11-cm sand band, soft, grey, FeS bands and wood fragments, occasional small mollusc shells (0–93 cm); fine–medium sand, fairly compact (93–360 cm); stony sand and gravel, some plant detritus (360–390 cm). |
| 9         | 69-0-13-14 | 43.376        | 79.57          | 76              | Silty clay, soft, dark grey, many black FeS bands (0–257 cm); medium sand, dark grey, few small freshwater *Pisidium conventus* clam shells (257–461 cm); stony sand and gravel (461–468 cm). |
| 10        | 69-0-16-1  | 43.433        | 79.512         | 87.5            | Silty clay, soft, dark grey, FeS bands, some shells (0–310 cm); fine sand and silt with shell and wood fragments, *Pisidium* clam shells, *P. conventus* clam shells, some shells (0–310 cm); clay and silt with shell and wood fragments, plant and insect fauna (310–325 cm); clay, greyish brown to brown, clay blebs, laminated with grit near base (325–540 cm); stony till, grey (540–580 cm). Wood from the 310–325 cm unit dated to 10 775±20 14C years BP (UC/AMS45365) (Anderson & Lewis 2012: p. 516; table 1, fifth entry). |

1Below 74.0 m a.s.l., from NOAA bathymetry map (Virden et al. 1999).
the western end of the lake basin in a general southeast–northwest direction. The present elevations of the beach surface range from −7 to −2 m a.s.l. with standard deviations of 0.7 m (crest) and 1.4 m (base) over three profiles (Table 2). In one profile, the thickness of the barrier deposit reaches 5 m (Fig. 6B). The position of the beach is delineated by dashed lines on Fig. 1 and red dashed lines in Fig. 5; it was mapped from its occurrence on acoustic and seismic survey profiles and from intersections in piston cores. Sediments of the beach are sorted sand and gravel (Table 1). All profiles of the barrier show it rising about 4 m in a westerly direction, and successive internal reflectors in one seismic profile (Fig. 6B) suggest progressive growth of the feature in the same westerly (landward) direction, similar to reflections in ground-penetrating radar profiles of other transgressive coastal landforms (Møller & Anthony 2003; O’Neal & Dunn 2003; Smith et al. 2003).

The palaeo-barrier beach is morphologically and sedimentologically comparable to a marine baymouth transgressive sandy barrier formed on a microtidal, wave-dominated coast, as described by Reinson (1992). Wave-eroded and westward-drifted sand is thought to have accumulated across the end of the Ontario basin forming a barrier beach, similar to formation of the Burlington barrier beach enclosing Hamilton Harbour at present. Sediments, eroded mainly from the southwestern shore and nearshore of Lake Ontario where un lithified Quaternary sediments predominate, nourish the Burlington beach (Rukavina 1976; Davidson-Arnott & Ollerhead 1995). Accumulation in the ELO barrier beach would have been similar also to that of the earlier Lake Iroquois sand and gravel barrier that underlies the city of Hamilton (Fig. 3A) about 36 m above the present lake (Karrow et al. 1961; Coakley & Karrow 1994). Strong easterly winds and waves driven by glacial anticyclone circulation off the Laurentide Ice Sheet might have aided construction of the ELO barrier. Such winds are inferred from sand dune orientations at the time of the CS incursion (David 1988) to the east of the Ontario basin near Montreal, and from spit orientations in glacial Lake Algonquin to the northwest (Krisk & Schaetzl 2001; Schaetzl et al. 2016). The westward and upward movement of successive washover reflectors within the barrier seismic profile (Fig. 6B) is consistent with a transgressive origin during a period of slowly rising lake level while ELO was connected with the rising CS. Fine sand, usually deposited on beaches as aeolian dunes, is absent, probably removed later by waves of Lake Ontario while overstepping the beach. Thus, the existing palaeo-beach surface as shown by acoustic reflections is close to the lake water level during its formation.

Nearshore shallow-water sediment deposition, preceding offshore deep-water deposition, is indicated by fine sand and silt under silty clay mud in a core deeper in the basin at 87.5 m water depth offshore of the ELO beach (Table 1, entry 10). The fine sand and silt are probably related to wave action in the Early Holocene lowstand of hydrologically closed Lake Ontario (Anderson & Lewis 2012) that followed formation of the ELO barrier beach.

**Radiocarbon dates.** – A wood fragment from the lowstand nearshore fine sand and silt unit between 310 and 325 cm in core 69-0-16-1 (entry 10 in Table 1, and Fig. 5) returned a mean age of 10 775±20 14C years BP (Anderson & Lewis 2012: table 1, site 11). Ostracode fossils recovered from the same interval of lowstand sediment indicated that water depth at this site could have been as shallow as 8 m (L.D. Delorme in Anderson & Lewis 1985: p. 239 and table II, site 11). As interpreted later, the date mentioned in the first sentence of this paragraph is taken as an indication of the period in which waves and water-level processes formed the ELO barrier beach. The maximum age of the beach (about 12.78 or −12.8 cal. ka BP) allows for the preceding last ice advance, its break-up over the lake, drainage of glacial lakes, and incursion of the CS, described above in the ‘Regional setting and background’ section. Other plant fragment dates indicate ages of plant growth on or near the ELO barrier during the Lake Ontario lowstand. These ages are 9940±320 14C years BP or 11.5 cal. ka BP (UCIAM74187) on plant material from surficial stony sand and gravel of the barrier beach in core 69-0-13-7 (Table 1, entry 8), and 9965±20 14C years BP or 11.4 cal. ka BP (UCIAM45366) in nearshore sand and silt at 3.1 m depth in core 69-0-16-1 (Table 1, entry 10). An age of 8820±35 14C years BP or 9.9 cal. ka BP (UCIAM74206) on freshwater mollusc shells in sand over stony sand and gravel of core 69-0-13-14 (Table 1, entry 9) indicates deposition in shallow water over the ELO barrier during the lake rise from its lowstand (Anderson & Lewis 2012: table 1, sites 9, 11b and 10).

**Champlain Sea marine limits**

Figure 1 illustrates the shoreline of the CS north and east of the Lake Ontario outlet as inferred by Parent & Occhietti (1988) and Occhietti (1989). Specific elevation measurements of a marine shore on the northern slope of the Adirondack Mountains (Figs 3A, 7) were obtained from the studies of Clark (1980), Pair et al. (1988) and Pair & Rodrigues (1993) and as listed in appendix B of Rayburn (2004) from D. Pair (Table S1). Elevations and locations used for the CS in the Lake Champlain Valley (Figs 3A, 7) were checked in the field and listed by Rayburn (2004: appendix A) (Table S1).

Highest marine elevation limits along the western boundary and northern region of the CS (Fig. 7) were also selected from the literature (Table S1). These limits consisted commonly of the highest occurrences of
marine mollusc shells in beach gravel in the western CS, as at sites 30 and 31 (Table S1), or as consistently aligned coastal landforms such as beaches, deltas and terraces on the CS shore on the southern side of the St. Lawrence Valley and in the Lake Champlain Valley. Marine limits used for isobase construction were assumed to be contemporaneous or quasi-contemporaneous and to represent the marine level at the time of the CS incursion. Some selected limits were confirmed on topographical maps, and all are listed with their position coordinates in Table S1. Errors in identification of marine limits could affect the configuration of the inferred CS surface. Recognizing marine beaches of the western shore of the sea is difficult as they occur in rough terrain and sediment-poor areas of Precambrian bedrock (Sharpe 1979).

Isobases of the CS surface were fitted to the above data using trend surface analysis software in ESR1 ArcGIS, similar to trend surface modelling of isobases on the raised shoreline of Lake Saimaa, Finland (Saarnisto & Huhn 1973). A second-degree trend surface was selected to best represent the Champlain isobases (Fig. 7), neglecting the three northernmost marine limits, which may have been ice-covered during marine incursion. The trend of these isobases is similar to those of the earlier glacial Lake St. Lawrence where they overlap in the Covey Hill area – only slightly closer to east–west, and rising ~0.8 m km\(^{-1}\), slightly less than 1.0–1.1 m km\(^{-1}\) for the earlier lake (Parent & Occhietti 1999; Fig. 5).

**Interpretation**

**Comparison of ELO barrier beach and CS elevations**

Two profiles of CS isobases consisting of isobase elevation vs. distance (Fig. 8A, B) were developed from Fig. 7 to show trends of the isobase surface in relation to the central barrier position (latitude 43.362°N, longitude 79.589°W) as shown in Fig. 6B and Table 2. Several lower isobases of glacial Lake Iroquois from Bird & Kozlowski (2016) were also plotted on the first two profiles. These isobases, being southwest of the lake outlet in a differentially uplifted basin to the northeast, describe the youngest surface of Lake Iroquois and guide projection of the ELO barrier elevations onto profile 1 (Fig. 8A). A regression line fitted to the intersection points of the marine isobases on profile 1 (Fig. 8A) projects toward the ELO barrier beach. Also, a second-degree polynomial fits the intersection points of marine isobases on profile 2 (Fig. 8B) and projects toward the ELO barrier. A third profile (not illustrated) was prepared to verify the relatively flat uplifted profile of the CS surface in the Lake Champlain Valley, as previously noted by Rayburn (2004), Rayburn *et al.* (2005) and others.

Completion of the lower ends of the extended CS–ELO surface in each of profiles 1 and 2 (Fig. 8A, B) mimics the gentler slope of the lower end of the Lake Iroquois shoreline defined by intersections of its isobases.
in the profiles. Portions of the younger CS–ELO profiles (~12.8 cal. ka BP) were computed from the Iroquois isobase intersections in each profile. For each horizontal interval between successive Iroquois isobases, indicating 5-m uplift, from 110 m to 135 m, a new uplift for the same intervals, \( U_2 \), was computed from the expression:

\[
U_2 = \frac{\text{age}}{3.7} - 5 \times (12,780 / \text{age}) - 1 = 0.
\]

This expression is derived from equation 3 in Lewis et al. (2005) in which unknown uplift of a dated shoreline is computed from another shoreline of known age and uplift in the same horizontal interval of a transect or profile. The numerals, 13 000 and 12 780, are ages of the youngest Lake Iroquois and earliest CS shorelines, respectively, and \( \tau = 3700 \) years is the relaxation time of Great Lakes area uplift described by a negative exponential function (Lewis et al. 2005). The computed uplifts for each interval were cumulated in succession to define the lower ends of the CS–ELO profiles (Fig. 8A, B). These profile segments were aligned with the extrapolated CS–ELO profiles 1 and 2 to show their general connection with the lowest part of the ELO barrier beach (Fig. 8A, B). These connections illustrate the anticipated relation between sea level and the barrier beach in ELO following the incursion of the CS about 12.8 cal. ka BP. Although vertical offset of slightly >2 m is expected between the extrapolated Champlain isobase surface and the ELO barrier because of sea-level rise during its construction, the observed offset is in the order of 2 m more, possibly because of oscillations and variation in sea level (see standard deviation of sea level in Table S2), or contributions due to wave run-up on high water levels during wind set up on high tides. Relative sea level is thought to have oscillated somewhat during the CS inundation, for example, variations in relative sea level in the ranges of 10–20 m are postulated in the northwest CS (Barnett 1988). These alternations in the rates of isostatic land rebound vs. global sea-level rise may have caused short-term variations of the cross-sectional area of the connection between CS and ELO, as well as fluctuations of unknown amplitude of the water level in the Ontario basin.

**ELO outlet hydraulic constrictions, outlet uplift and sea-level rise**

The ELO level was constrained mainly by the rate of sea-level rise, tidal and other oscillations, and possibly resistance of water flow through outlet constrictions. The outlet constrictions are narrow or shallow reaches of the head of the St. Lawrence River, as well as the upstream Duck–Gallow (DG) Ridge between the main basin and eastern (Kingston) sub-basin of Lake Ontario as illustrated in Pair et al. (1988). Inflow to ELO was mostly local runoff and discharge of meltwater from upstream proglacial lakes Agassiz and Algonquin via the Trent River Valley and Lake Erie basin routes (Fig. 4C). Licciardi et al. (1999) estimated baseline freshwater outflow, neglecting short-term outburst floods from upstream glacial lakes, to the North Atlantic Ocean via the St. Lawrence Valley at 0.15 Sv (1 Sv = 10^6 m^3 s^(-1)) of ice-melt and precipitation. This sum was reduced here to 0.1 Sv or 100 000 m^3 s^(-1) for the Ontario basin by removing contributions to the downstream St. Lawrence Valley specified by Licciardi et al. (1999). For the eastern Great Lakes area, the precipitation was considered by Licciardi et al. (1999) to be overestimated by a factor of 2 after comparison with modern data in the region. Similarly, the effective precipitation was found to be reduced by 30% by evapotranspiration. Accordingly for this paper, the baseline outflow used for ELO was reduced to 0.055 Sv or 55 000 m^3 s^(-1). This reduction in precipitation is reasonable in the light of the known dry climate of the region at this time (Shuman et al. 2002) that...
evaporated ELO into a multi-millennial lowstand after ~12,500 cal. years BP (Anderson & Lewis 2012). Also at this time during the Younger Dryas cold period, meltwater output from the Laurentide Ice Sheet was reduced owing to re-advection or stabilization of the ice margin.

As shown in Fig. 9A, the initial CS surface was ~25 m above the shallowest part of the thalweg of the St. Lawrence River near its head (see baseline distance 0 to ~6 km in Fig. 9A) and rapidly diverged from it downstream. The width of this outlet control, termed ‘Head of St. Lawrence’ constriction, ranged from ~3 km at the initial ELO–CS surface to ~1.5 km after the constriction had been uplifted ~18 m. The remaining cross-sectional area, about 7 m deep, below this level would have been completely used in conveying the full 55,000 m$^3$ s$^{-1}$ outflow according to flow depth estimates based on contracted weir calculations (Pacific Northwest Irrigation 2018). Thus, the CS and ELO could maintain a common level and were essentially confluent while the ELO was raised above the CS by ongoing isostatic rebound.

However, a part of the ELO flow that was routed by the Lake Erie basin (Fig. 4C) was clearly guided by gaps in the upstream DG Ridge where the projected CS surface dipped to ~58–56 m a.s.l., below most crests of the northeast-facing bedrock cuestas that constitute the ridge (Fig. 9A, B). Neglecting leakages through small gaps, this outflowing water was mainly conveyed by a large gap between Main Duck and Galloo islands, and to a lesser extent through the Black River Channel adjacent to Stony Point (Fig. 9B). The DG gap was ~7 km wide at the palaeo-lake surface (~57 m a.s.l.) and tapered to ~1.8 km at its bedrock base near 31 m a.s.l. (Virden et al. 1999). Calculations of depth of flow suggested that this constriction could have conveyed all fluxes up to 55,000 m$^3$ s$^{-1}$ without altering the common level of ELO and CS. Figure 10A schematically illustrates the progression of basin outlet uplift while the ELO was confluent with CS and building its barrier beach in the western end of the Lake Ontario basin.

At the beginning of confluence, while the lake and sea level were at maximum clearance over the outlet sill, CS marine bottom water might have entered the Ontario basin and mixed with ELO bottom water by estuarine circulation. However, any tendency for marine water to penetrate the Ontario basin would have been thwarted generally by the outflowing freshwater flux of 55 000 m$^3$ s$^{-1}$ supplied by the Agassiz–Algonquin inflows. Steady freshwater flows of this magnitude and less were found capable of freshening the western CS despite possible occurrences of high palaeocides in studies using modelled estuarine circulation dynamics for the St. Lawrence River Valley (Katz et al. 2011). This conclusion is supported by low salinity foraminifera and ostracode assemblages in the upper St. Lawrence River Valley (Rodrigues 1992) and by the east-to-west decline in foraminiferal diversity in the CS that is interpreted as a trend to lower salinity in the western part of the CS (Guilbault 1989, 1993). Also, the nearest marine macrofossils to ELO have only been found about 100 km downstream in the St. Lawrence River Valley from the lake outlet (Henderson 1970).

The average uplift rates for the ELO outlet control near the head of the St. Lawrence River were calculated and contrasted with far-field tropical (eustatic) sea-level rise (Table S2 and Fig. 10B). This table and figure show palaeo-sea levels at 100-year intervals from 13,000 to 12,200 cal. years BP from Bard et al. (2010). Averaged sea levels from Tahiti, New Guinea and Barbados are shown with their standard deviations.

Data of the Lake Iroquois and CS phases are used for estimates of outlet uplift rates relative to the ELO barrier beach. Lake Iroquois isobases at the ELO outlet in Bird & Kozlowski (2016) are based on shoreline elevations of the Main Iroquois phase (Pair & Rodrigues 1993); their age is inferred to be ~13,200 cal. years BP, somewhat older than isobases south and west of the Mohawk River outlet. The approximate total post-Iroquois uplift of the ‘Head of St. Lawrence’ constriction, relative to the ELO barrier, is the difference between Iroquois isobase values at the two sites, or 225 ± 1 m, the ELO barrier elevation at profile B (Fig. 6) (Table 2) = 78.1 ± 2 m. The average rate of uplift is estimated by the difference in Iroquois and Champlain uplifts divided by the difference in their ages, that is (ui – uELO)/(ti – tELO) = [(107 ± 2) – (78.1 ± 2)]/(13,200 – 12,780) = 28.9 ± 4/420 = 0.069 m a$^{-1}$ on average. At this rate, constriction uplift through 18 m could have progressed for (18/0.069) = 261 years while the ELO level was confluent with the slowly rising sea level, or until about 12,780–261 = 12,519 cal. years BP when ongoing isostatic uplift of the constriction would have raised ELO above the CS (Fig. 10B). Shortly after, the lake fell to its lowstand, revised in this paper from the previous estimate of 12.2–12.3 cal. ka BP by Anderson & Lewis (2012), considering the age of the diversion of meltwater inflow from ELO.

Shortly after 12,500 cal. years BP, discharge of lakes Algonquin and Agassiz was routed to lower outlets in the North Bay area because of ice recession (Karrow et al. 1975, 1995; Eschman & Karrow 1985). At that time, prevailing dry climate evaporated some of ELO, and the water level descended to the Lake Ontario lowstand (Anderson & Lewis 2012). The descent of lake level stranded the ELO barrier beach in the western Lake Ontario basin. Thus, the ELO barrier beach is a marker for sea level in western Lake Ontario from ~12.8–12.5 cal. ka BP.

While the Ontario basin outlets were rising isostatically under the influence of differential rebound within the post-12,780 cal. years BP period, the ELO level was confluent with the CS level and would have risen slowly
with long term sea-level rise rather than by the faster outlet uplift (Fig. 10B). Short-term variations in level due to tides, wind set up and variable water flow through the constrictions may have been superposed on the sea-level-driven water-level rise and its possible oscillations. This slower lake-level rise combined with the coincidental terminal lake-level descent to a lowstand as the basin was raised above the CS are the mechanisms that stabilized lake level relative to its shore zone, and facilitated the accumulation and preservation of swash-zone, wave-drifted sand in the ELO barrier beach.

Subsequent transgression of lake level from the lowstand (Anderson & Lewis 2012) apparently was rapid enough to overstep coarser beach sediment and preserve the ELO barrier. Other than the present barrier of western Lake Ontario, no other large continuous submerged barriers were observed from numerous sounding transects in the area. Apparently, waves reworked sandy sediment upslope and alongshore through Middle and Late Holocene time into the present Burlington barrier beach.

At present, other segments of the ELO shore, other than the barrier beach at the western end of the Ontario basin, are unknown. However, both depositional beaches and erosional notches that date to this lake phase probably do exist and remain to be discovered.

**Interpretation of a sea-level index point**

The central profile of the barrier beach (Figs 6B, 11) and its complementary core (Table 1, entry number 7, core 92800-3) are the key data for interpretation of the palaeo-barrier beach as they were collected last, were most accurately positioned, and the profile shows the greatest amount of detail. Successive overwash reflections at successively higher elevations in the seismic profile (Figs 6B, 11) show that the ELO beach was formed under conditions of rising lake level, and is a transgressive barrier beach deposit. Transgression is indicated also by the rising tropical sea level at the time of barrier formation (Fig. 10B).

Amongst the various transgressive barrier systems reviewed by Reinson (1992), the ELO barrier that formed under microtidal conditions would be characterized probably as ‘wave-dominated with numerous storm washover features but few tidal channels’. This characterization is supported by the apparent absence of known cross-barrier channels and by five or six washover reflections evident in the cross-barrier seismic profile image (Figs 6B, 11). These reflections may be from coarse-grained sediment washed over the beach into a back-barrier lagoon by extreme waves during large storms at high tide. Core 92800-3 nearest this site (Table 1, entry 7) encountered only deep-water silty clay mud over pebbly coarse sand at the surface of the barrier. The absence of a sand unit (possible foreshore or beach-face sediment) over the gravel, as in other cores (Table 1, entries 4, 5, 6, 8), supports the interpretation that the pebbly coarse sand is a former washover surface.
The highest washover reflection in Figs 6B and 11 at elevation \(-3.1 (2.0 \times 1.96) = -3.1 \pm 3.9 \text{ m a.s.l. (95% confidence limits)}\) and location 43.362° N, 79.589° W is interpreted as evidence of the last accumulation on the barrier at \(-12.5 \text{ cal. ka BP} and is targeted as a sea-level index point (Shennan 2015). The 95% confidence limits (standard deviation \(\times 1.96\)) of this elevation are a combination of the variation amongst crestal elevations of the barriers listed in Table 2 and a part of the uncertainty of the bathymetry datum as explained in footnote 1 to Table 2. The ultimate crestal elevation of the barrier (Figs 6B, 11) is avoided as this may have been deposited during later overstepping of the beach. Lake level was rising with sea level at the time so the tendency of this sea-level index point is upward and landward.

The age of final barrier deposition and the age of this index point are approximately coincident with diversion of significant meltwater inflow to the Ontario basin and abrupt descent of water level from ELO to a lowstand. The diversion is attributed to ice recession and the stranding of the Main Lake Algonquin beach in the Lake Huron basin as lake level there dropped to lower outlets. Two pieces of spruce wood in a river terrace graded to the Main Algonquin lake shore with \(^{14}C\) ages of \(10.500 \pm 150\) and \(10.600 \pm 150\) years BP (GSC11126 and GSC11127; Karrow 1986) were pooled and calibrated to \(12.121 \pm 12.704 \) (2\(\sigma\) range) to provide an age with 95% confidence limits for this index point. Similar ages for the Algonquin shore were obtained by Karrow et al. (1975).

The indicative meaning and indicative range of this index point comprise the elevation and range of its reference water level in a comparable modern environment. The closest modern microtidal transgressive beaches to which the ELO palaeo-barrier can be compared are located along the south and southwestern shores of the microtidal GSL (same area in which Reinson (1992) identified and analysed barrier systems) at Kouchibouguac, Bouctouche, and Stanhope Lane (Fig. 3A; Forbes et al. 2004; O’Carroll et al. 2006). In this region of ongoing submergence and barrier overwash, crestal elevations of barriers at washover locations, measured mainly by light distance and ranging instrumentation (LiDAR) and other modern survey instruments, range from a low of 1.3 m a.s.l. at Kouchibouguac to a high of 3.7 m a.s.l. at Stanhope Lane, with an average elevation of 2.5 \(\pm 1.2 \text{ m a.s.l. (Forbes et al. 2004; O’Carroll et al. 2006).} \) Elevations are relative to mean sea level (average of hourly water level readings at select Canadian Atlantic and Pacific tidal stations over a 19-year cycle; Natural Resources Canada 2017). Thus, the indicative range of the reference water level for this sea-level index point is 2.4 or \(\pm 1.2 \text{ m, the range of elevation of the modern barrier crests at washover sites in the southwestern GSL.} \)

Discussion

Chronology

The foregoing estimated period of ELO barrier beach construction from \(-12.8\) to about 12.5 cal. ka BP is
based on the established age of the Champlain Sea, the inferred youngest age of glacial Lake Iroquois, varve composition and varve counts for the duration of post-Iroquois time, and the radiocarbon-dated age for the decline of glacial Lake Algonquin in Huron basin that signalled termination of a large water supply to the Lake Ontario basin and the end of the ELO phase. The period of barrier beach construction arising from these events compares reasonably well with the 12 737–12 679 (2σ) cal. years BP age for a wood fragment (UC1AMS45365) found in core 69-0-16-1 (Fig. 5 and Table 1, entry 10) in the lowstand nearshore sediments deposited following the ELO barrier formation. We infer that shore or tributary stream erosion during ELO beach formation may have floated woody vegetation into the lake. This possibility would have caused plant death during beach construction and ultimate transportation and deposition of the dated wood fragment in nearshore lowstand sediment. Overall, the end of the chain of post-Iroquois events, as portrayed here, fits reasonably well with the age of wood in sediments near the ELO barrier beach.

Formation of the ELO palaeo-barrier beach

The ELO beach resembles the formation of large beaches in glacial Lake Agassiz, north-central North America, as explained by Teller (2001). He concluded that as water level and waves in a differentially uplifting lake basin transgress the shore opposite a rising and overflowing outlet, a large and extensive beach is likely to develop, and that this beach will be stranded when a lower outlet becomes operable. In the ELO situation, transgression of water level up the western margin of the basin was forced mainly by rising sea level. Stranding of the resulting ELO beach occurred when water level descended to an evaporative lowstand (Anderson & Lewis 2012).

Variation in style of rebound

The southwest to northeast profile 2 (Fig. 8B) from the ELO barrier beach in western Lake Ontario through the alignment of marine limits on the southern side of the St. Lawrence Valley is slightly convex upward and downturned towards the east. This profile would be most influenced by uplift in the region east of Lake Ontario, in a region and during post-Iroquois time where this study and previous studies have detected rotation of isobases from east-southeast to east (Prest 1970; Sly & Prior 1984; Muller & Prest 1985; Pair et al. 1988; Rayburn 2004; Fig. 7). Also, in this region and farther east in New England, uplifted surfaces of this age are mainly flat (Kotth & Larsen 1989; Kotth et al. 1993; Rayburn 2004; Rayburn et al. 2005; Hooke & Ridge 2016). They are not concave upward as in the Great Lakes region as described in the ‘Regional setting and background’ section. Such concave-upward palaeo-shorelines in the latter region indicate more rapid uplift to the northeast, and this trend is consistent with the strong northeasterly upward gradient of the CS–ELO surface inferred in this paper. Possibly, the CS limits on its southern shore have been influenced more by the eastern style, but it is beyond the scope of this paper to explore reasons for these disparate styles of rebound.

Conclusions

A mud-buried, well-formed ELO palaeo-barrier beach with its average surface elevation ranging from −6 to −2.8 m a.s.l. beneath western Lake Ontario, consists of a distinct concave-eastward deposit of sorted sand and gravel up to 5 m thick. The beach was probably constructed by longshore drift of sandy littoral sediment operating under waves driven by easterly winds while ELO was confluent with the CS from about 12.8 to 12.5 cal. ka BP. The CS was an embayment of the Atlantic Ocean that had inundated the St. Lawrence
River Valley after deglaciation while it was isostatically depressed by the former ice load.

Construction of the ELO barrier beach occurred within a few centuries. ELO was then rising slowly at a common level with sea level while the CS–ELO surface ranged from ~25 to ~7 m above the more rapidly rising outlet sills. This situation reduced water-level change relative to the shore zone in the Lake Ontario basin, enhanced shore erosion, and facilitated the accumulation of wave-drifted sand and gravel sediment in a transgressive barrier beach in the western end of the Lake Ontario basin. The connection of ELO water level with slowly rising sea level and the coincidental stranding of the beach as the lake fell into a lowstand after beach construction are the mechanisms that led to formation and preservation of this palaeo-barrier beneath western Lake Ontario. A reduced water supply because of meltwater rerouting and enhanced evaporation in a dry climate led to the lowstand. The Ontario basin outlet was uplifted coincidentally above sea level slightly before the Lake Ontario water level descended to its lowstand after about 12.5 cal. ka BP.

Although the CS and ELO were confluent for several centuries, the Ontario basin remained a freshwater environment owing to the passage of abundant meltwater through it. The absence of a marine influence is supported by the presence of a freshwater fossil fauna (freshwater molluscs and an ostracode species) in the sediments of ELO and the absence of marine fossils.

The ELO barrier beach elevations are palaeo-sea-level markers in the western Lake Ontario basin for the period from ~12.8 to ~12.5 cal. ka BP. Abandoned after about 12.5 cal. ka BP, the beach is the farthest inland evidence of palaeo-sea-level in eastern North America north of the Gulf of Mexico.

The upper part of the ELO barrier beach at 43.362° N, 79.589° W with an elevation of ~3.1±3.9 m a.s.l. represents a palaeo-washover surface with a reference water level 2.5±1.2 m above present mean sea level and is considered a sea-level index point dated 12 121–12 704 (2σ) years BP.

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Supporting Information

Additional Supporting Information may be found in the online version of this article at http://www.boreas.dk.

Table S1. Champlain Sea limits (highest elevations).

Table S2. Far-field sea level for 13 000 to 12 200 cal. years BP from Bard et al. (2010).