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An 8,900-year-old forest drowned by Lake Superior: hydrological and paleoecological implications

M. Boyd · J. T. Teller · Z. Yang · L. Kingsmill · C. Shultis

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Abstract Exposures along the lower Kaministiquia River (near Thunder Bay, Ontario, Canada) provide insight into early Holocene lake level fluctuations and paleoenvironmental conditions in the northwestern Lake Superior basin. These exposures show at least two large paleochannels which were downcut into offshore sediments, and were later filled with >2 m of sand, ~3 m of rhythmically laminated silt and clay, and ~6 m of interbedded silt and sand. Buried by the rhythmically laminated silty clay unit is a well-preserved organic deposit with abundant plant macrofossils from terrestrial and emergent taxa, including several upright tree trunks. Three AMS radiocarbon ages were obtained on wood and conifer cones from this deposit: 8,135 ± 25 (9,130–9,010 cal), 8,010 ± 25 (9,010–8,780 cal), and 7,990 ± 20 (8,990–8,770 cal) BP. This sequence records an early postglacial high-water phase, followed by the Houghton lowstand, and reflooding of the lower Kaministiquia River Valley. The drop in lake level associated with the Houghton phase forced the ancestral Kaministiquia River to downcut. By ~9,100 cal (~8,100) BP, older channels eroded into subaqueous underflow fan deposits in the Thunder Bay area near Fort William Historical Park (FWHP) were abandoned and colonized by a *Picea-Abies-Larix* forest. Based on stratigraphic data corrected for differential isostatic rebound, the lake was below the Sault Ste. Marie bedrock sill between at least 9,100 cal (8,100) and 8,900 cal (8,000) BP. Shortly after 8,900 cal BP, the lake quickly rose and buried in situ lowland vegetation at FWHP with varved sediments. We argue that this transgression was due to overflow from glacial Lakes Agassiz or Ojibway associated with the retreat of the Laurentide Ice Sheet from the Nakina moraine and/or the Cochrane surge margins in the Hudson Bay Lowlands. A continued rise in lake level after 6,420 ± 20 (7,400 cal) BP at FWHP may record uplift of the North Bay outlet above the Sault Ste. Marie bedrock sill and the onset of the Nipissing transgression in the Lake Superior basin.

Keywords Lake Superior · Houghton low-water phase · Boreal forest · Plant macrofossils · Paleohydrology · Holocene paleoecology

M. Boyd (✉)
Department of Anthropology, Lakehead University, 955 Oliver Rd, Thunder Bay, ON P7B 5E1, Canada
e-mail: matthew.boyd@lakeheadu.ca

J. T. Teller · Z. Yang
Department of Geological Sciences, University of Manitoba, Winnipeg, MB, Canada

L. Kingsmill · C. Shultis
Department of Anthropology, Lakehead University, Thunder Bay, ON, Canada

Introduction

The geology of the Superior basin yields crucial insight into the history of meltwater drainage from

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the interior of North America to the Atlantic Ocean. Until ~9,000 cal (8,000) BP when discharge was diverted through the Hudson Bay Lowlands (Dyke and Prest 1987; Teller and Mahnic 1988; Bajc et al. 1997; Breckenridge et al. 2004), catastrophic outbursts from Lake Agassiz over the continental divide flowed directly into the Superior basin, leaving sedimentary and geomorphic records of these events (Teller and Mahnic 1988; Breckenridge 2007). Downstream in the North Atlantic Ocean, some of these freshwater outbursts may have forced abrupt climate change through disruption of thermohaline circulation (Broecker et al. 1989; Teller et al. 2002; Leverington and Teller 2003; Yu et al. 2010). Despite its importance and possible connection to wider environmental events in the Northern Hemisphere, many aspects of the postglacial geological history of the Superior basin remain poorly understood. To a large extent, this is the result of uncertainties in the chronology of major lake level fluctuations shortly after deglaciation, as well as the exact location of the Laurentide Ice Sheet (LIS) margin which helped control overflow from glacial Lakes Agassiz and Ojibway. In many parts of the basin, radiocarbon dates on early postglacial beaches and other paleo-lake level indicators are few, unreliable, or complicated by reworking of older organics (Yu et al. 2010).

In general, the early postglacial history of the Superior basin is marked by significant changes in lake level which occurred in response to differential isostatic rebound, meltwater inflow, and outlet erosion among other factors. The focus of this paper is on the Houghton phase, a period of low-water level in the Lake Superior basin ~9,000 cal (8,000) BP that followed an earlier deep-water proglacial period (the Minong phase; Farrand and Drexler 1985; Fisher and Whitman 1999; Yu et al. 2010). The Houghton phase is important because it coincided in time with the diversion of Agassiz meltwater away from the Superior basin into the Hudson Bay Lowlands, and the onset of a dramatically reduced hydrological budget. In the Michigan and Huron basins, reduction of meltwater inflow coupled with climate change in the early mid-Holocene may have forced these lakes to fall below the level of their outlets and become hydrologically closed (Lewis et al. 2007, 2008). Details about these events in the Superior basin during this time, however, are unclear because little information exists on the exact timing and magnitude of the lake level drop associated with the Houghton phase.

In 2006, sedimentary exposures were discovered along the lower Kaministiquia River (near Thunder Bay, Ontario) which provide chronological control over the Houghton lowstand phase and a subsequent transgression. These sections expose at least two large (~250-m-wide) channels which were cut into underflow fan deposits and, later, were filled with varved silt and clay. Importantly, the varved unit is underlain by an extensive organic deposit containing plant macrofossils and in situ trees—the remains of a forest that was drowned and buried by a rise in lake level. Elsewhere in the Upper Great Lakes, buried forests have provided key insight into lake level fluctuations, glacial chronology, and paleovegetation composition during the Holocene (Lowell et al. 1999; Pregitzer et al. 2000; Lewis et al. 2005; Hunter et al. 2006; Schneider et al. 2009). In the present study, we report the first buried forest from the north shore of Lake Superior and one of only a small number of similar deposits known across the Great Lakes basin. Based on stratigraphy, plant macrofossil content, and radiocarbon ages, we reconstruct the history of early lake level fluctuations following the final retreat of the LIS from the Thunder Bay area. These data provide new insight into the paleohydrology of the Superior basin during, and immediately following, the Houghton low-water phase ~9,000 cal (>8,000) BP, while also providing an outstanding 'snapshot' of local early mid-Holocene paleovegetation.

Early postglacial lake level fluctuations in the Superior basin

Between ~11,000 and 6,000 cal (~9,650 and 5,200) BP, three major phases have been recognized in the Lake Superior basin: an initial highstand (Minong), followed by a regression during the Houghton phase, and subsequent transgression (Nipissing phase). Events after ~6,000 cal (5,200) BP, which are only briefly outlined here, are discussed in other publications (Booth et al. 2002; Johnston et al. 2007).

The Minong phase was the last period when ice-marginal lakes occupied the Superior basin, and began when the LIS re-advanced southward into the basin, reaching the northern Upper Peninsula of Michigan by 11,500 cal (10,000) BP (Drexler et al.

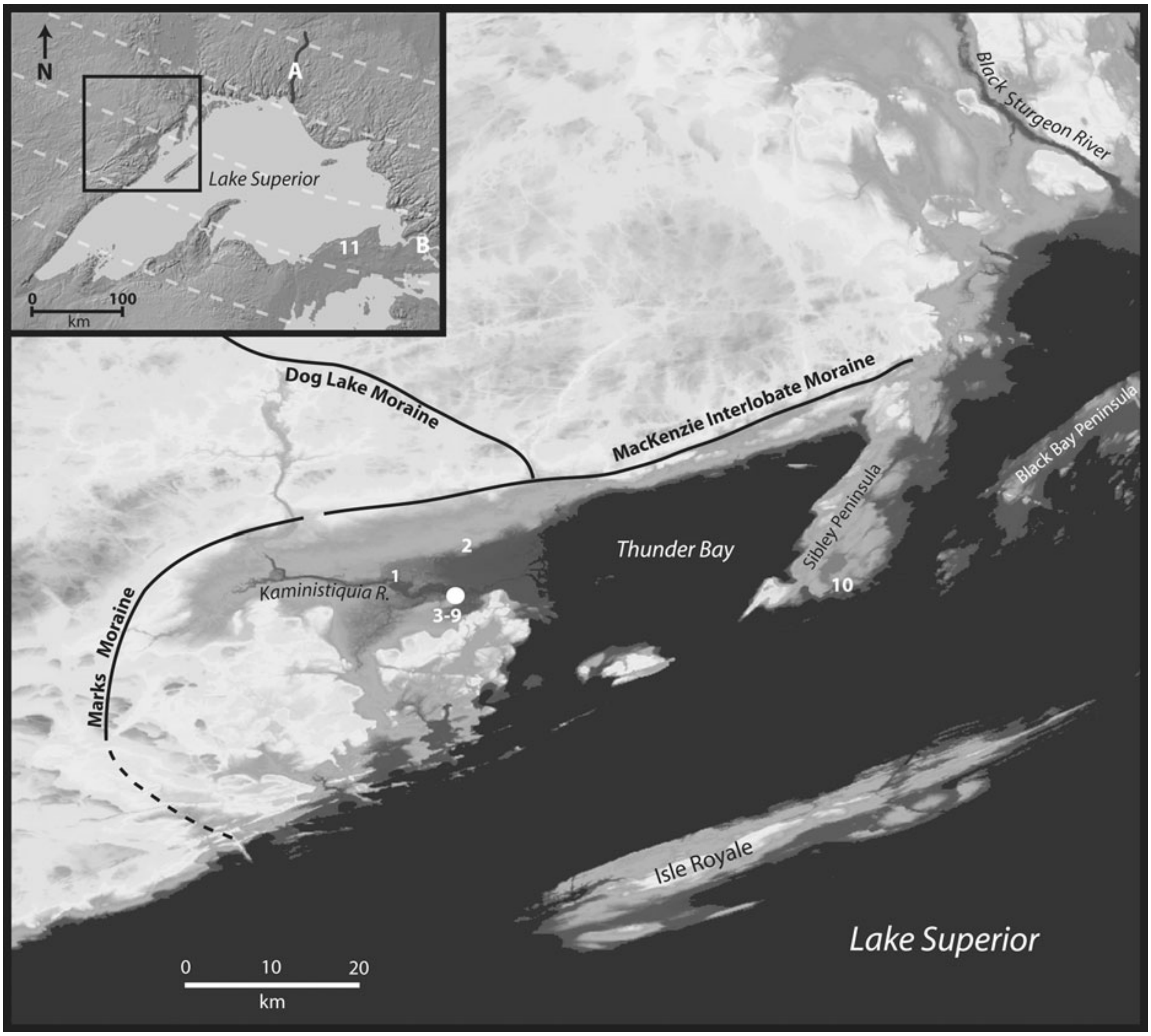

Fig. 1 Modern digital elevation model (*DEM*) of study area showing location of Fort William Historical Park (*white dot*), and moraines in the northwestern Lake Superior basin (Thunder Bay area). *Numbers* refer to radiocarbon dates and sites listed in Table 1. *A* Pic River spillway channel; *B* Sault Ste. Marie outlet. *Dashed lines* on inset map are isobases (Teller and Thorleifson 1983)

1983; Lowell et al. 1999). This advance (Marquette) briefly divided the basin into two lakes, Duluth and Minong (Farrand and Drexler 1985). In the Thunder Bay area, the Marks and Dog Lake moraines (Fig. 1) may represent the position of the LIS during this time (Clayton 1983; Drexler et al. 1983; Teller and Thorleifson 1983). As the ice margin retreated from the Keweenaw Peninsula (largest peninsula on the south shore [Fig. 1, inset]) ~10,500 cal (9,300) BP, Lakes Minong and Duluth coalesced. Elevation of the Minong phase lake level was controlled by a morainal sill crossing from Nadoway Point (Michigan) to Gross Cap (Ontario) at the eastern end of the basin (Farrand and Drexler 1985; Yu et al. 2010). Near Thunder Bay, the elevation of the Minong level is ~230 m asl (47 m above the modern water plane). A conventional radiocarbon age of 9,380 ± 150 (~10,500 cal) BP (Table 1) was obtained on wood from the base of the Minong beach at Rosslyn (Dyck et al. 1966), roughly 6 km upstream of the study area.

Table 1 Radiocarbon ages and other relevant data associated with early postglacial sites in the Thunder Bay area. Radiocarbon years were calibrated using the IntCal04 terrestrial dataset (Reimer et al. 2004), with the 2σ range and means shown. Location of sites and dates are shown in Fig. 1

	Site name	Lab number	^{14}C years BP	Calibrated 2σ range and mean (cal BP)	Material dated	Context	Modern elevation (m asl)	Latitude (N)	Longitude (W)	Reference
1	Rosslyn Pit	GSC-287	9,380 ± 150	11,094–10,248 (10,671)	Wood	Base of Minong beach	227	48°21.8′	89°27.3′	Dyck et al. (1966)
2	Cummins Pond	TO-547	9,260 ± 170	11,092–9,948 (10,500)	Conifer wood	Basal sand (Minong beach)	230	48°24.3′	89°20.9′	Julig et al. (1990)
3	Old Fort William	UCIAMS-26800	8,135 ± 25	9,127–9,009 (9,070)	Wood	Tree in life position extending into silty clay	186	48°20.7′	89°21.1′	This study
4	Boyd Cut	ETH-31437	8,070 ± 70	9,046–8,658 (8,850)	Charcoal	Base of interlaminated clayey silt and sand	199	48°20.4′	89°21.5′	Loope (2006)
5	Old Fort William	UCIAMS-26801	8,010 ± 25	9,005–8,776 (8,891)	*Picea* sp. cone	Located at base of silty clay in organic laminae	186	48°20.7′	89°21.1′	This study
6	Boyd Cut	ETH-31438	7,995 ± 65	9,014–8,641 (8,828)	Charcoal	Base of interlaminated clayey silt and sand	199	48°20.4′	89°21.5′	Loope (2006)
7	Upstream paleochannel	UCIAMS-61732	7,990 ± 20	8,994–8,774 (8,884)	*Pinus banksiana* cone	Located at base of silty clay in organic laminae	186	48°20.5′	89°21.2′	This study
8	Old Fort William	UCIAMS-26802	7,970 ± 30	8,993–8,659 (8,826)	Wood	Log in silty clay ~2.5 m above base of unit	189	48°20.7′	89°21.1′	This study
9	Old Fort William	UCIAMS-61733	6,420 ± 20	7,421–7,291 (7,356)	*Scirpus* sp. achenes	In upper organic deposit	195	48°20.7′	89°21.1′	This study
10	Surprise Lake	BETA-230960	8,170 ± 40	9,010–9,260 (9,135)	*Betula* leaves and scales	Lake isolation	187	48°20.1′	88°49.3′	Yu et al. (2010)
11	Little Harbor	ETH-32328	8,365 ± 100	9,090–9,540 (9,315)	*Betula* leaves and twigs	Lake isolation	183	46°49.2′	85°21.7′	Yu et al. (2010)

This date is in line with other chronological estimates for the Minong phase in the Superior basin (Drexler et al. 1983; Julig et al. 1990; Fisher and Whitman 1999), and indicates that final deglaciation of the Thunder Bay area was completed less than 1,000 years after the Marquette re-advance had reached its maximum on the Upper Peninsula of Michigan.

During the Minong phase (~10,500 cal [~9,300] BP), meltwater from glacial Lake Agassiz was diverted into the Superior basin through a series of outlets located on the western side of Lake Nipigon (Teller and Thorleifson 1983; Teller and Mahnic 1988; Leverington and Teller 2003; Breckenridge et al. 2004, 2010). This event may have been due to a re-advance of the LIS which closed the Clearwater-Athabaska spillway in Saskatchewan and diverted Agassiz overflow to its eastern outlet (Teller and Boyd 2006). Increased flow via the Nipigon basin may have briefly raised the level of Lake Minong (by hydraulic damming at the outlet [Tinkler and Pengelly 1995]); in Lake Huron, this event initiated the early Mattawa highstand (Lewis and Anderson 1989; Lewis et al. 2007). The exact sequence of events in the Superior basin between ~10,300 and 9,000 cal (~9,150 and 8,000) BP is unclear, although several fluctuations in lake level may have occurred during this time in response to changes in Agassiz outflow, differential isostatic round, the location of the LIS margin, or a combination of these factors (Farrand 1960; Saarnisto 1974; Slattery et al. 2007; Breckenridge et al. 2010; Yu et al. 2010).

Between ~9,400 and 9,000 cal (~8,400 and >8,000) BP, multiple sites from the Superior basin record a significant drop in lake level (Farrand and Drexler 1985; Fisher and Whitman 1999; Saarnisto 1974; Yu et al. 2010). The initiation of this phase, the "Houghton Low", has been attributed to erosion of the sill at Nadoway Point (Prest 1970; Saarnisto 1974; Farrand and Drexler 1985; Fisher and Whitman 1999; Yu et al. 2010). Yu et al. (2010) propose that the large volume of freshwater released into the North Atlantic Ocean by this event triggered the 9,300 (8,300) cal BP cold event, which is widely recorded across the Northern Hemisphere. Additionally, diversion of Lake Agassiz overflow away from the Superior basin by ~8,000 (~9,000 cal) BP would have dramatically changed the hydrological budget and made the lake more prone to climate-induced drawdown after this time (Breckenridge et al. 2004; Lewis et al. 2007, 2008).

At the end of the Houghton phase, the lake level began to rise in response to differential isostatic rebound of the North Bay outlet that carried overflow to the Ottawa River drainage system (Lewis and Anderson 1989; Booth et al. 2002; Lewis et al. 2007). This is recorded by the Nipissing transgression which culminated 6,800–5,700 cal (~5,950 and 5,000) BP with the Nipissing I highstand (Lewis 1970; Fisher and Whitman 1999; Booth et al. 2002). Continued uplift of the North Bay outlet made it inactive by 5,500 cal (4,800) BP, transferring control to the Port Huron and Chicago outlets (Lewis 1969; Lewis and Anderson 1989; Booth et al. 2002). A second high-water phase, Nipissing II, dates to between 4,600 and 4,400 cal (4,100–4,000) BP, and the drop in lake level following these two highstands may be due to erosion of the Port Huron outlet (located at the southern end of Lake Huron; Booth et al. 2002; Johnston et al. 2007). By at least 1,200 cal (1,300) BP, the Sault Ste. Marie (SSM) outlet rebounded above the elevation of the Port Huron outlet and Lake Superior became separated from Lakes Michigan and Huron (Johnston et al. 2007).

Materials and methods

The focus of this study is a 1 km section along the lower Kaministiquia River across from Fort William Historical Park (FWHP). Along this stretch of the river, cutbanks are upwards of 14 m high and expose multiple fine-grained sedimentary units deposited following the retreat of the LIS from the Marks moraine (Fig. 1). At FWHP, the elevation at river level is roughly 4 m above the present (183 m) water plane of Lake Superior. As a result of its low elevation, the lower Kaministiquia Valley would have been highly sensitive to past fluctuations in lake level. These fluctuations are largely recorded, and mediated, by fluvial sedimentary processes operating within the valley, and can be seen in episodes of channel downcutting and aggradation.

Paleotopographic reconstructions

Three paleotopographic time slices (10,400, 8,900, and 7,400 cal [9,250, 8,000 and 6,500] BP) were generated using GIS software. Modern digital elevation data (90 m resolution) was obtained from the USGS Seamless Data Distribution System (http://seamless.usgs.gov/index.php). Lake Superior bathymetric data (1 km resolution) was downloaded from The Lake Superior Decision Support Project (Natural Resources Research Institute, University of Minnesota Duluth) website (http://www.nrri.umn.edu/lsgis/databases.htm). After the downloaded data were imported into ArcGIS, the bathymetric data were resampled to get 90 m resolution data, and then the DEM mosaics for the terrestrial and lake portions of the Superior basin were merged using ArcGIS Spatial Analyst. Isostatic rebound for three time points (based on dates and elevations at #2, 5 and 9 in Table 1) was calculated using an empirical model developed by Lewis and Thorleifson (2003), starting with isobase lines shown in Teller and Thorleifson (1983). The isostatic rebound values of isobase lines were then interpolated and a rebound surface was generated for the three time periods using ArcGIS 3D Analyst. Finally, three paleotopographic maps were generated by subtracting isostatic rebound values for the three time slices from the modern DEM mosaic. Comparison with the isobases drawn by Farrand and Drexler (1985) on the Minong and Washburn beaches in the Superior basin shows a close match in trend and value with those in our paleotopographic reconstructions.

Stratigraphy and AMS radiocarbon dating

Exposed sediments along the Kaministiquia River in the Thunder Bay region near Fort William Historical Park (FWHP) were described and sampled over four successive field seasons. The majority of this work, however, focused on one section, named the "Old Fort William" (OFW) site, where in situ buried trees were first observed in 2006 (Fig. 1).

Six slab samples were collected from the rhythmically laminated unit and underlying organic deposit for thin-section preparation. These samples were desiccated in a vacuum oven for 7 days and impregnated under low vacuum pressure using a low viscosity Spurr resin kit composed of nonenyl succinic anhydride, vinylcyclohexene dioxide, diglycidyl ether of polypropylene glycol, and 2-dimethylaminoethanol. The ratios of these chemicals used to produce the resin, the embedding technique, and other procedures followed a standard recipe for lake sediments (Myrbo 2005).

Chronological control was established by AMS radiocarbon dating of hand-picked, terrestrial, macrofossils or wood from in situ trees (Table 1). Samples selected for radiocarbon dating were submitted to the Keck Carbon Cycle AMS Facility (UCIAMS). Radiocarbon ages were calibrated using Calib (version 6.0) and the Reimer et al. (2004) IntCal04 terrestrial dataset. These dates and their locations, as well as conversions to calibrated ages, are presented in Table 1. One-number calibrated ages, when given, represent the mean of the 2σ calibrated range.

Plant macrofossils

A total of 14 bulk samples from Unit D were collected at 5 m horizontal intervals across the 70 m-long exposure at OFW. In the lab, 100 cc subsamples were deflocculated in 5% sodium hexametaphosphate for 24 h and gently washed through a column of screens (Beaudoin 2007). The remaining material was then air-dried for 48 h before being examined under a stereomicroscope. Plant macrofossil identifications used comparative collections in addition to keys in Lévesque et al. (1988) and other sources. In this paper, we present a summary of initial macrofossil recoveries from the 8,900 cal (8,000) BP organic unit (Unit D). Detailed discussion of the full paleoecological dataset from the OFW site will be presented in a subsequent paper.

Results

Fluvial sequence (Units F and G)

The lower part of the sedimentary sequence exposed in cuts along the lower Kaministiquia River Valley generally consists of >12 m of massive to horizontally- or cross-laminated, well-sorted fine sand to sandy silt (*Unit F*; Fig. 2). This unit exhibits no obvious vertical trends in grain size, bed thickness, or other sedimentary characteristics within the sequence. Within Unit F there are units (10–25 cm thick) of silty rhythmites and interbedded (often ripple-laminated) sand (Fig. 2). Climbing ripples are commonly observed in sandy units within Unit F, and zones of convoluted laminae/beds are occasionally visible. This unit is also described by Loope (2006: 28–32), who

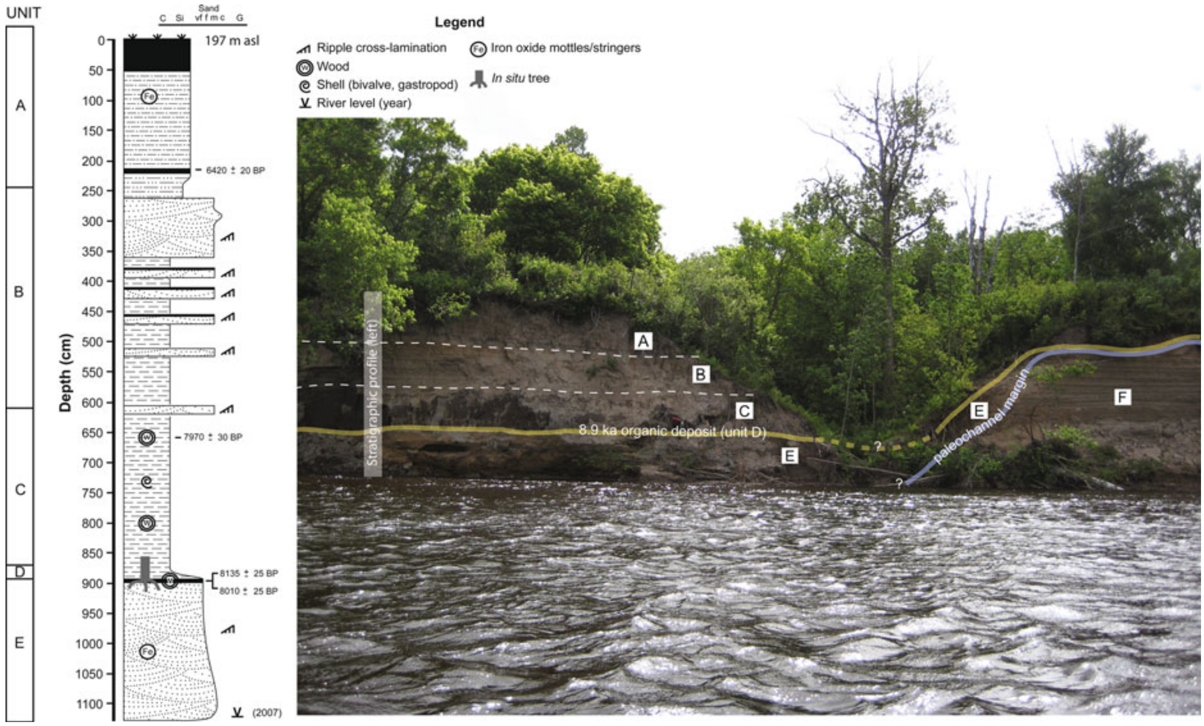

Fig. 2 Photo (looking towards the south) of Old Fort William (*OFW*) cutbank site showing ~8,900 cal BP organic deposit (Unit D), western paleochannel margin, major lithologic units, and composite stratigraphic column (*left*)

observed rhythmic beds of clayey silt and sand in a 4 m interval at the Boyd Cut (located 700 m upstream of the OFW paleochannel). Approximately 1.2 km upstream of FWHP, the basal contact of this unit is exposed above river level and is sharply underlain by 50+ cm of weakly imbricated, rounded, gravel and coarse sand (*Unit G*).

Paleochannel sequence (Units A–E)

In the FWHP study area, two large paleochannels (~250 m wide, ~12 m deep) are exposed in cross-section across a continuous series of cutbanks on the southern side of the Kaministiquia River. Both channels are incised into, and truncate, Unit F (Fig. 2), and dip below the modern river level (<186 m in this area). The infill sequence within both channels is similar and consists, in general, of a fining-upward sequence of sand and laminated silty clay separated by a thin and nearly-continuous organic (peat) deposit (Unit D). A more detailed description of this sedimentary sequence follows.

Basal sediments within the paleochannel at the OFW site consist of <0.5–2.0+ m of massive to laminated fine to medium sand in the sample area (*Unit E*; Fig. 2). Asymmetrical ripples (including climbing ripples) are common throughout. This sandy unit can be traced laterally along the bottom of both channels, forming the contact with Unit F (Fig. 2). Along the higher channel margin, laminae were deposited parallel to the contact between Units E and F, and at least one example of a lateral accretion/point bar deposit (composed of fine to coarse sandy laminae that dip towards the channel thalweg) was observed. In the paleochannel farthest upstream in the FWHP study area, massive to weakly bedded coarse sand and gravel were observed mid-channel in the same stratigraphic position as Unit E.

Above Unit E, several sets of thin (mm-scale), horizontally-bedded, organic laminae were observed in both paleochannels. This distinct, peaty, organic deposit (*Unit D*) is ~10 cm thick and consists of well preserved plant macrofossils; in the field, bryophyte mats, conifer needles, conifer and alder cones, as well as in situ tree trunks (between 3 and 17 cm diameter), were observed in this unit over three successive field seasons (Fig. 3). Although discontinuous in places, this thin peat deposit was traced over almost 200 m. In the centre of both channels, Unit D dips below the modern river level. Closer to the channel margins, it can be traced laterally along the margin to an elevation roughly 11 m above the modern river (Fig. 2). The thin organic laminae of Unit D generally show little evidence of post-depositional disturbance, although convolutions and other structures indicative of loading or post-depositional reworking were observed in one small (~1 m) horizontal zone. Woody debris, frequently alternating with zones dominated by moss litter, is common in Unit D and much of this wood material consists of small to medium, well-preserved, roots which are oriented parallel to the bedding plane. Significant root penetration into the underlying sterile sand of Unit E was not observed along the length of this deposit. In at least one example, roots oriented in this manner were connected to an upright (3-cm-diameter) tree stem indicating that much, if not all, of this material is in its original (life) position (Fig. 3).

All in situ trees observed in Unit D have stems that extend upwards (up to 1 m) into the overlying silt and clay of Unit C. Three AMS radiocarbon dates were obtained from this unit on hand-picked, terrestrial macrofossils (Table 1). One of these dates (7,990 ± 20 BP, UCIAMS-61732) was obtained on a jack pine (*Pinus banksiana*) cone recovered from the basal organic zone in the upstream paleochannel, while the other two dates (UCIAMS-26800 and 26801) were obtained from an 18-cm-diameter tree stem in life position (8,135 ± 25 BP) and spruce (*Picea* sp.) cone (8,010 ± 25 BP) recovered from the same stratigraphic unit at the OFW site (Table 1). This indicates that the basal organics (Unit D) likely were contemporaneous in both paleochannels exposed at FWHP. The combined 1-σ range for all three basal radiocarbon ages is 8,780–9,089 cal BP (mean = 8,935 cal BP). However, based on the radiocarbon date for the in situ tree (UCIAMS-26800, 8,135 ± 25 [~9,070 cal] BP), and allowing some time for growth (36 tree rings counted in one log), organic accumulation in Unit D probably began no later than ~9,100 cal (~8,100) BP. Loope (2006) reports two radiocarbon ages on charcoal

Fig. 3 Photos of the 8,900 cal BP organic deposit: **a** in situ tree (3 cm diameter) with preserved bark on lower portion of stem extending upwards into varved sediment (Unit C), and horizontal roots within ~8,900 cal BP organic deposit (Unit D); **b** spruce (*Picea*) cone associated with bryophyte (moss) litter; and **c** thin section showing contact between Units C and D (1 mm scale bar located in *lower right corner*)

from the base of an interlaminated silty clay and sand unit at the Boyd Cut: 8,070 ± 70 BP (ETH-31437) and 7,995 ± 65 BP (ETH-31438; Table 1). We observe that the organic bed dated by Loope (2006) correlates stratigraphically with the peat deposit near the bottom of the OFW and upstream paleochannels.

Plant macroremains are abundant in Unit D; seeds and other debris from a variety of terrestrial, aquatic and emergent, and bryophyte taxa were commonly recovered in the sieved fractions from nearly all samples at the OFW site (Table 2). In general, macroremains from conifer species were the most abundant, with *Abies balsamea* (fir), *Larix laricina* (tamarack), and *Picea glauca* (white spruce) being the most common taxa identified. Occasional *Pinus* (pine) needles were also recovered in addition to a jack pine cone from the same deposit at the upstream site. In addition to these trees, macroremains from *Betula papyrifera* (paper birch), *Alnus crispa* (green alder) were common, and occasional seeds from *Arctostaphylos uva-ursi* (kinnikinnick), and *Rubus idaeus* (raspberry) were found in some of the samples. Several emergent taxa were also identified, including *Scirpus americanus* (bulrush), and *Carex* sp. (sedge). In some samples, mosses were very abundant although only two species have been identified: *Pleurozium schreberi*, and *Dicranum polysetum* (Table 2).

Table 2 List of plant and other macroremains recovered from ~8,900 cal (8,000) BP peat deposit (Unit D), OFW site

Trees and shrubs
Abies balsamea needles
Abies balsamea seeds
Abies balsamea twigs
Larix laricina needles
Larix laricina seeds
Larix laricina twigs
Larix laricina cones
Picea needles
Picea seeds
Picea twigs
Picea glauca cone
Pinus needles
Pinus seeds
Pinus banksiana cone
Alnus crispa nutlets
Arctostaphylos uva-ursi seeds
Betula papyrifera nutlets
Betula papyrifera fruit scales
Rubus idaeus seeds
Aquatics and emergents
Nymphaeaceae seeds
Scirpus americanus achenes
Carex sp. (trigonous) achene
Bryophytes
Pleurozium schreberi
Dicranum polysetum
Bryozoans
Cristatella mucedo statoblasts
Plumatella sp. statoblasts
Cladocera
Daphnia sp. ephippia
Fungi
Cenococcum geophilum sclerotia

Conformably overlying Unit D is a ~2.75-m-thick deposit of predominantly silt and clay (*Unit C*) which is overlain by silty clay containing thin sandy ripple- to horizontally-laminated units (Unit B; Fig. 2). In thin sections obtained from the contact between Unit C and the underlying organic deposit we observed: small-scale convolutions and other deformations of the laminae in the main zone of organic accumulation, and small-scale rip-ups and occasional fine sand at the contact with Unit C (Fig. 3c). In general, Unit C is composed of silty clay to clayey silt with faint to distinct light–dark couplets (rhythmites) that are common throughout. These couplets vary in thickness from ~1 to 2 mm (per couplet) to 2 cm. Analysis of thin sections indicates that each couplet contains a thin (<1 mm) clay lamina, one or more overlying normally graded silt laminae of variable thickness, and a zone above this which is dominated by coarse silt sized particles embedded in a finer clayey matrix. Occasional very fine to fine sand laminae and couplets are also present, and these coarser laminae are more common towards the top of this unit. The rhythmically laminated sediments of Unit C are interpreted as varves due primarily to their uniformity and repetitiveness; the sandy laminae which are occasionally observed in this sequence may have been deposited by sediment gravity flows and/or suspension settling of eolian sand. The normally-graded zone which occurs between the clay (winter) and coarse silt (summer) laminae in each couplet may represent density underflow currents associated with the spring melt. Occasional bivalve (up to 7 cm diameter) and gastropod shells occur throughout this unit. Within the lower ~1–2 m of Unit C, occasional logs are visible; the largest of these is 30 cm in diameter and is oriented horizontally. Other than the in situ trees extending upward from Unit D (see above), all logs observed in Unit C were redeposited. One of these horizontal logs, deposited ~2.5 m from the base of Unit C, yielded a radiocarbon age of 7,970 ± 30 (8,990–8,660 cal) BP (UCIAMS-26802). Based on extrapolation from zones where light–dark couplets are distinct (average sedimentation rate = 0.74 cm/year), we estimate that Unit C at the OFW site represents roughly 200 years of deposition. For the same unit in the upstream paleochannel we counted an average of 192 couplets (range = 182–203) directly from the outcrop. The upper contact with Unit B is generally wavy and erosional in appearance.

Unit B is distinguished from the underlying unit by the presence of multiple fine to very fine sand beds with massive to faintly laminated silty clay (Fig. 2). In places, laminae form couplets. The sandy beds are commonly laminated and contain ripples and climbing ripples, and thicken (from 13 to 100 cm) and coarsen upward. In three of the sand beds, fibrous/woody organics are present.

Unit A is massive to poorly laminated sandy silt with a soil developed in the upper part (Fig. 2). Approximately 2 m below surface, a zone of organic laminae (5–15 cm thick in total) is visible. Individual organic laminae are ~1–5 mm thick and alternate with silt laminae that are about the same thickness. Plant macrofossils recovered from this organic zone are dominantly sedge (*Carex* sp.) and bulrush (*Scirpus* sp.) seeds, suggesting a shallow pond environment with emergent vegetation. Several of the *Carex* achenes still had their fragile perigynium attached, indicating little or no postdepositional reworking and redeposition of macroremains. *Scirpus* seeds sieved from this laminated organic deposit yielded an average age of 6,420 ± 20 (7,400 cal) BP (UCIAMS-61733).

Interpretations

Sedimentary exposures in the FWHP study area document the following series of events: (1) deposition of gravel (Unit G) and a sandy subaqueous underflow fan (Units F) following retreat of the Superior Lobe from the Marks moraine; (2) erosion of channels into these sediments when the level of Lake Superior fell, and partial infilling of channels by fluvial sand (Unit E); (3) local channel abandonment and colonization of the channel by a spruce-fir-tamarack forest (Unit D); (4) rapid reflooding of the lower Kaministiquia River valley and onset of a deep water, low-energy, depositional environment with high sediment flux (Unit C); (5) episodic fluvial and lacustrine sedimentation and deposition of Units B and A; and (6) incision of the modern Kaministiquia River to its present elevation.

Unit F records a period of rapid sedimentation in a subaqueous underflow fan environment. Evidence for this is the presence of climbing ripples, loading structures, overall fine grain size with occasional silty rhythmites, lack of fossils, and similarity with underflow fan sequences described elsewhere (Klassen 1983; Sun 1993; Boyd 2007). Presence of an underlying, coarse, gravel deposit (Unit G) indicates that deposition of Unit F was preceded by a higher-energy fluvial event. Due to the position of the FWHP sequence inside the Marks moraine and absence of till in the exposed sequence, these sediments must have been deposited after the Marquette re-advance reached its maximum ~11,500 cal (9,900) BP (Dyck et al. 1966; Drexler et al. 1983; Julig et al. 1990; Lowell et al. 1999; Lowell et al. 2009). With the retreat of the Superior Lobe, proglacial lakes that had been impounded between the Marks moraine and continental divide to the west would have drained rapidly to the east through the Kaministiquia River valley (Loope 2006). Proglacial lake sediments are extensive west of this moraine, and these deposits have been attributed to a string of short-lived ice-marginal lakes (e.g., glacial Lakes O'Connor, Baldy, Kaministiquia; Zoltai 1963; Teller and Thorleifson 1983; Phillips and Fralick 1994). The thick sequence exposed in the study area indicates that a large amount of sediment was rapidly deposited in the lower Kaministiquia River valley following deglaciation, and meltwater discharge from one or more proglacial lakes west of the study area, plus meltwater from the retreating LIS, are the likely sources of this sediment. Consequently, we suggest that Unit F records a period of underflow fan construction in Lake Minong, the precursor to Lake Superior, that occurred sometime after ice retreated (~11,000 cal (9,650) BP) but before 9,100 cal (8,100) BP (the oldest age of basal organics at OFW).

The paleochannels at FWHP, which are incised into subaqueous underflow fan deposits to below the modern river level, record a significant drop in the base level of the lake. Radiocarbon ages on plant macrofossils inside the channel indicate that low water levels were established no later than 9,100 cal (8,100) BP. In general, development of the organic deposit (Unit D) may reflect a substantial decrease in discharge through the channel, which allowed vegetation to colonize a previously active alluvial surface. By 8,900 cal (8,000) BP, this vegetation was a mix of coniferous (boreal) forest and wetland elements. Spruce (*Picea*), tamarack (*Larix*) and fir (*Abies*) formed the dominant tree cover along with some boreal deciduous tree species such as paper birch (*Betula papyrifera*) and green alder (*Alnus crispa*). The paucity of pine macroremains suggests that pine trees did not grow in the wetter channel bottom; instead, jackpine (*Pinus banksiana*) was probably confined to drier upland zones in the local area (Julig 1984).

In the channel at OFW, ground cover was dominated by mosses (especially *Pleurozium schreberi* and *Dicranum polysetum*), in addition to occasional low shrubs such as kinnikinnick (*Arctostaphylos uva-ursi*) and raspberry (*Rubus idaeus*). Presence of wetland taxa such as sedges and bulrush may imply occasional above-ground water levels in the channel bottom or presence of discrete wetland habitats.

The absence of carbonized wood, needles or other charred plant macroremains indicates that fire frequencies were locally very low between ~9,100 and 8,900 cal (~8,100 and 8,000) BP. This agrees with the ecology of *Pleurozium schreberi* and *Dicranum polysetum*; in the boreal forest today, these moss species are dominant in mature forests (80–120 year) with little or no fire disturbance (Boudreault et al. 2002). *Abies balsamea*, the most abundant conifer inside the channel (based on total number of macrofossils), is also a late-successional, fire-intolerant, species that is commonly associated with *Picea glauca* (white spruce), tamarack (*Larix laricina*) and *Pleurozium* and *Dicranum* in lowland areas of the boreal forest (Uchytil 1991). Low fire frequencies during the period of peat accumulation at OFW may have been due to climate or, more likely, topography. It is reasonable that the relatively moist alluvial surface would have inhibited local wildfires, enabling late successional species to dominate in this setting.

The drop in lake level before 9,100 cal (8,100) BP, recorded at FWHP by fluvial downcutting and local channel abandonment, as well as elsewhere in the basin (Yu et al. 2010), represents the regression of Lake Superior during the Houghton phase. This event was due to rapid erosion of the Nadoway morainal sill near Sault Ste. Marie (Yu et al. 2010). Yu et al. (2010) concluded that this level was 45 m below the Nadoway barrier. Because both paleochannels in the FWHP study area dip below the modern river level, we infer that the relative lake level at this time lay at or below the modern water plane ($\leq$183 m asl). Yu et al. (2010; Fig. 3) report similar results from lake cores collected at Little Harbor, Michigan and Lake Surprise, Ontario; these sites, which today lie between 183 and 187 m asl, were isolated beginning between 9,100 and 9,400 cal (8,170 and 8,370) BP (Table 1), as lake level dropped ~45 m below the overflow channel/sill and previous level of the lake (Yu et al. 2010). Significantly, when corrected for differential isostatic rebound, one of these sites (Little Harbor) was disconnected from Lake Superior ~10 m below the elevation of the SSM bedrock sill between 9,200 and 9,400 cal (8,400) BP (Yu et al. 2010: Fig. 3). Our isostatic rebound model also shows that overflow across the sill at SSM could not yet have occurred at 8,900 cal (8,000) BP (Fig. 4c, d). By interpolating between isobases 6 and 7 in Fig. 4c, d, where isobase 7 is 100 km north of isobase 6 and is 32 m higher at this time, it can be seen that SSM lies ~30 km south (on isobase 6.5) of the isobase that extends through FWHP, which lies on isobase 6.8. This means that their difference in elevation was $30/100 \times 32 = 9.6$ m. So, when the organics at OFW were flooded, the floor of the overflow channel at SSM still lay almost 10 m above that. This was confirmed by more detailed processing of DEM data of Fig. 4d in the channel (not shown). Thus, we suggest that the level of Lake Superior had dropped below the elevation of the SSM bedrock sill between at least 9,100 and 8,900 cal (8,100 and 8,000) BP. In the Michigan, Huron, and Georgian Bay basins, Lewis et al. (2008) describe a closed-lake phase between 8,770 and 8,300 cal (~7,900 and 7,500) BP which they attribute to high rates of evaporation and decreased hydrological inputs associated with early mid-Holocene climate change and the northward retreat of the LIS. In the Superior basin, early postglacial paleoclimatic records are either too few or too coarse to compare with the timing of the Houghton Low. However, in southern

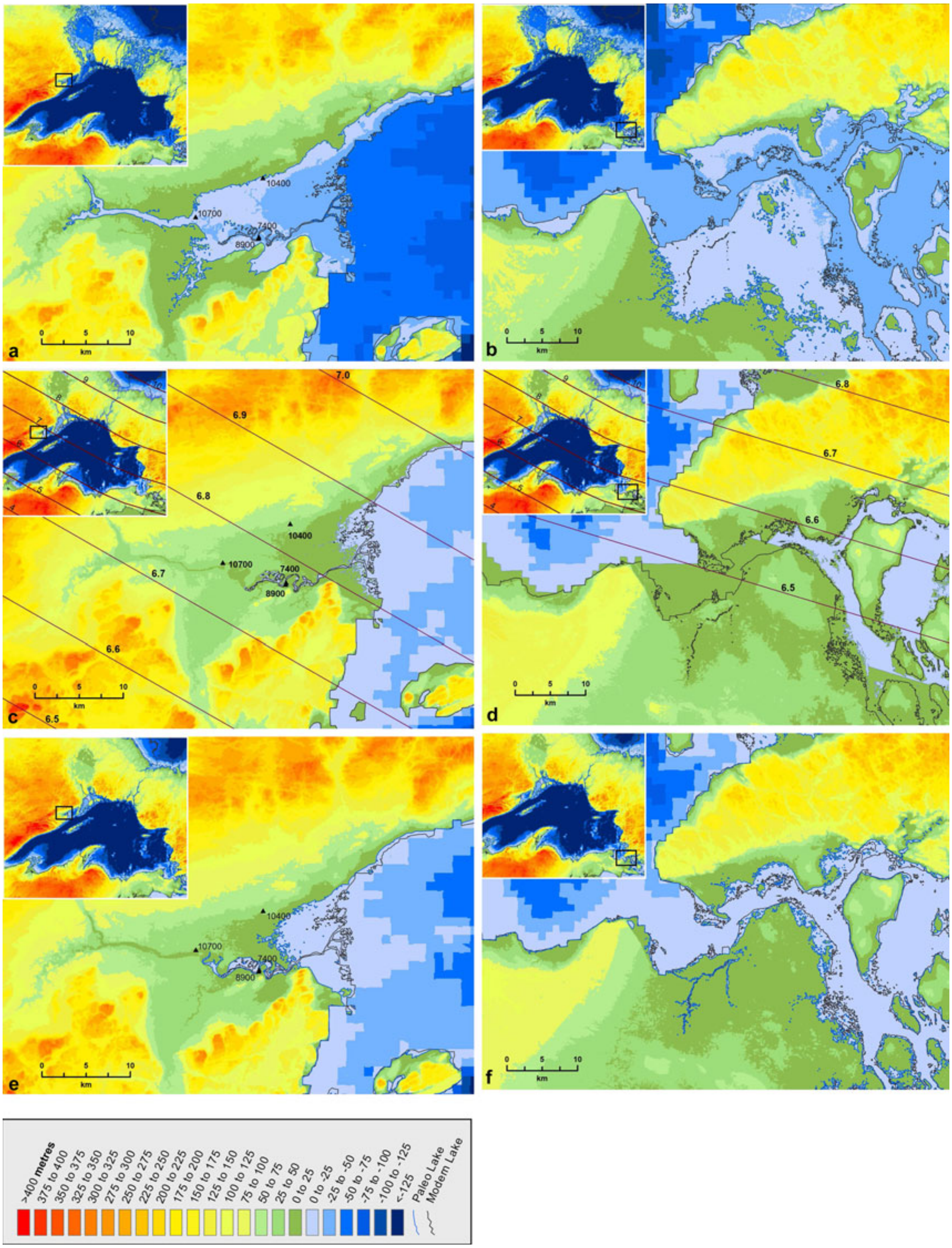

Fig. 4 Paleotopography of the Thunder Bay and Sault Ste. Marie areas, with inset maps showing the Lake Superior basin, at: 10,400 cal (9,250) BP (*a* + *b*); 8,900 cal (8,000) BP (*c* + *d*); and 7,400 cal (6,500) BP (*e* + *f*), calculated from modern DEM topography adjusted for differential isostatic rebound at the times shown as described in the text. Isobases for the Thunder Bay area are shown in (**c**) and can be correlated to the outlet channel at Sault Ste. Marie shown in (**d**). Dates and their locations used to control the DEM paleotopography model are shown, and include the Cummins site (10,400 cal BP; Julig et al. 1990; #2 in Table 1), and select dates from the OFW section (#5 and 9 in Table 1)

Minnesota, the time-transgressive nature of mid-Holocene aridification, and its broad overlap in time with the Houghton low-water phase, is indicated by the rapid eastward movement of the prairie-forest border between 10,000 and 8,000 cal (8,900 and 7,200) BP (Williams et al. 2009). The shift to more arid climatic conditions, in combination with diversion of meltwater into the Hudson Bay Lowlands (Dyke and Prest 1987; Teller and Mahnic 1988; Bajc et al. 1997; Breckenridge et al. 2004), may have produced a negative hydrological budget within the Superior basin and drawdown of the lake below the elevation of the SSM bedrock sill by ~9,100 cal (~8,100) BP.

The rhythmically laminated (varved) fine-grained sediment of Unit C that overlies the organic zone of Unit D records the onset of a subsequent deep water lake phase in the lower Kaministiquia River valley. We argue that the transition from a subaerial, terrestrial, environment during the period of peat formation to a deep water depositional environment occurred rapidly, and very shortly after 8,900 cal (8,000) BP. This is based on: (1) presence of in situ trees with intact bark that extend up to at least 1 m into the overlying varves; (2) presence of well-preserved organics with minimal reworking; and (3) fine-grained sediment (silt/clay) immediately above the basal peat with no evidence of intervening nearshore deposits. Based on varve counts, Unit C represents at least 200 years of continuous lacustrine deposition. Importantly, a sustained rise in relative lake level beginning ~8,900 cal (~8,000) BP is also observed at other sites in the Superior basin (Yu et al. 2010). In the Thunder Bay region, this is most clearly seen at Surprise Lake, Ontario (#10 in Fig. 1 and Table 1), where re-connection with Lake Superior began sometime between 8,730 and 9,010 cal (8,010 ± 40) BP, and lasted until ~1,840 cal (1,900 ± 65) BP (Yu et al. 2010).

An increase in lake level shortly after 8,900 cal (8,000) BP may have occurred due to: (1) isostatic rebound at the outlet; (2) damming the outlet by a glacial readvance; (3) climate change; or (4) an increase in meltwater flux entering the lake. We rule out #1 because lake transgressions due to isostatic rebound are gradual, occurring often over hundreds or thousands of years as is the case for the Nipissing rise (Booth et al. 2002; Fisher and Whitman 1999; Lewis 1969, 1970). As argued above, the switch to a deep water depositional environment at FWHP (represented by the rhythmically laminated silt and clay of Unit C) occurred rapidly, with no intervening nearshore facies. Scenario #2 is unlikely because the ice margin was well north of the SSM outlet by this time (Dyke 2004). We rule out #3 because the timing of this transgression in FWHP occurred during one of the driest intervals in the Holocene (Bartlein et al. 1998; Lewis et al. 2008; Williams et al. 2009). Instead, we suggest that the rise in water level that occurred shortly after 8,900 cal (8,000) BP was due to a renewed influx of meltwater, possibly from Lakes Ojibway and/or Agassiz or directly from the LIS, which raised the level of Lake Superior for at least 200 years.

Supporting evidence for a late influx of water from Lakes Ojibway and/or Agassiz comes from:

(1) Glaciodeltaic deposits at the mouth of the Black River valley, ~15 m above the Houghton low-water level, which date to 8,070 ± 180 (~9,000 cal) BP (Bajc et al. 1997: 689).
(2) Glaciofluvial sediments with paleoflow directions toward the southwest in the Mullet Outlet-Pic River lowland, which links the Superior basin with the Hudson Bay Lowland. These sediments record meltwater overflow into the Superior basin following the retreat of the LIS from the Nakina II moraine (Slattery et al. 2007); this moraine formed ~9,200 cal (8,200) BP (Teller and Thorleifson 1983; Teller et al. 1996; Dyke 2004). The subsequent retreat of the ice margin from this moraine after 9,200 cal (8,200) BP would have allowed overflow from Lakes Agassiz or Ojibway to be directed into the Pic River and into a post-Minong phase of the Superior basin (Slattery et al. 2007: 344).
(3) Instability of the southern margin of the LIS in the Hudson Bay Lowlands, with frequent surges into glacial Lake Ojibway after 9,000 cal (~8,000) BP (Dyke and Dredge 1989). These surges, including the Cochrane readvances (Prest 1970; Dyke and Dredge 1989), are important because they may have raised the level of Lake Agassiz-Ojibway and caused meltwater to overflow into the Superior basin at the end of the Houghton Low. The Pic River, which is situated in an area of low topography across the continental divide (Fig. 1), could have served as a conduit for meltwater discharge into the Superior basin

during this time. Significantly, ostracodes from lakes Huron and Michigan record negative, and abrupt, oxygen isotopic excursions that are dated to about 8,900 cal (8,000) BP (Rea et al. 1994; Moore et al. 2000; Breckenridge and Johnson 2009). Because this change occurs at almost the same time as the reflooding of the lower Kaministiquia River valley, it is possible that overflow from the same source was responsible for both.

AMS dating of Unit A indicates that rising lake level and sediment accumulation continued until sometime after 7,400 cal BP. This subsequent rise in lake level at FWHP probably records the Nipissing transgression, which was due to differential uplift of the North Bay outlet (Lewis and Anderson 1989; Booth et al. 2002; Lewis et al. 2007). This event slowly raised the water level in the Superior basin above the SSM bedrock sill until Lake Superior was confluent with Lakes Michigan and Huron (Lewis and Anderson 1989; Lewis et al. 2007). Comparison of uplift histories at North Bay and Sault Ste. Marie indicates that the North Bay outlet was lifted above the previous outlet at SSM beginning ~7,400 cal (6,400) BP (Lewis and Anderson 1989: Figure 7c); our paleotopographic reconstructions also suggest that the water level was above the SSM bedrock sill by the time that Unit A was being deposited (beginning ~7,400 cal (~6,400) BP; Fig. 4f). Continued uplift of the North Bay outlet, in combination with a wetter climate (Lewis et al. 2008), would have resulted in a slow transgression at FWHP to the Nipissing level until the outlet switched to Chicago/Port Huron. In the Thunder Bay region, the Nipissing level is represented by a prominent wave-cut scarp ~210 m asl (Julig et al. 1990), 27 m above the modern water plane and 24 m above Unit D at FWHP. Sometime after 7,400 cal (6,400) BP, however, the modern Kaministiquia River downcut to its present elevation. We suggest that this was due to the overall decline in lake level that occurred at the end of the Nipissing highstand, in combination with differential isostatic rebound. Following the Nipissing high water phases (~6,800 and 4,400 cal [~5,950 and 3,950] BP), the lake level in the Superior basin dropped gradually due, perhaps, to erosion of the overflow outlet at Port Huron (Booth et al. 2002; Johnston et al. 2007). In addition to an absolute drop in lake level, differential isostatic rebound between FWHP and the outlets at Port Huron and, later, Sault Ste. Marie (see isobases in Fig. 4c, d), would have uplifted FWHP at a faster rate, causing downcutting of the Kaministiquia River and emergence of the region from the lake.

Conclusions

Profound environmental and hydrological changes occurred in the Lake Superior basin in the first three millennia following deglaciation. One of the most significant events during this time period was the large drop in lake level during the Houghton phase. Multiple sites in the Superior basin, including the one reported in this paper, indicate that the level of the lake was no higher than the SSM bedrock sill between >9,100 and 8,900 cal (>8,100 and 8,000) BP, and was therefore hydrologically closed during this time. By 9,100 cal (8,140) BP, abandoned river channels near FWHP were colonized by a lowland boreal plant association. In this protected setting, fire-intolerant, late-successional conifer species (e.g., *Abies balsamea*) dominated the overstory in association with common boreal deciduous trees and shrubs (*Betula papyrifera, Rubus idaeus, Arctostaphylos uva-ursi*), emergent (*Carex* sp., *Scirpus americanus*), and bryophyte (*Pleurozium schreberi*, *Dicranum polysetum*) taxa. Shortly after 8,900 cal (8,000) BP, the lake level rapidly rose, drowning and burying lowland vegetation in the lower Kaministiquia River valley in a deep-water depositional environment. We suggest that this rapid transgression was due to overflow from glacial Lake Agassiz-Ojibway, perhaps related to surging of the margin of the LIS in the Hudson Bay Lowlands, which routed water into the Superior basin through one or more spillway channels on the northern side of the lake. A continued rise in lake level and sedimentation after ~7,400 cal (6,400) BP likely records the beginning of the Nipissing transgression, when the North Bay outlet was differentially lifted above the SSM bedrock sill. The subsequent entrenchment of the Kaministiquia River to its modern elevation (after 7,400 cal (6,400) BP) was probably initiated by the basin-wide drop in water level that occurred at the end of the Nipissing highstand in combination with differential isostatic rebound.

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