Article

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LATE GLACIAL SEDIMENTATION AND HISTORY OF THE LAKE NIPIGON BASIN, ONTARIO

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ABSTRACT *Late glacial sedimentation and history of the Lake Nipigon basin, Ontario.* The Lake Nipigon basin lies north of the Lake Superior basin and was the hydrological link between glacial Lake Agassiz and the Great Lakes during part of the last deglaciation. A sequence of glaciolacustrine sediments, composed mainly of silt-clay rhythmites and sand, was deposited in the offshore waters of glacial Lake Nipigon by overflow from Lake Agassiz and meltwater from the retreating glacier margin. Sections from six long sediment cores and four lake bluff exposures reveal a sandy (early deglacial) lower section that is overlain by 300 to 850 silt-clay rhythmites (varves). Deposition of these varves, as well as coarser sediment along the western shore, began after 9200 BP, as the glacial margin retreated northward along the continental divide that separated the Nipigon basin from the higher Lake Agassiz basin to the west. The absence of ice rafted clasts in the rhythmites suggests that the ice had retreated from the lake by the time they were deposited. On the basis of their elevation in relation to the lowest raised beach at West Bay, which formed about 9000 BP, most rhythmites probably were deposited between 9000 and 8000 BP. Species of arboreal pollen are present in early postglacial sediments of the Nipigon-Superior lowlands, suggesting that the Lake Nipigon region became colonized by coniferous and deciduous forests soon after deglaciation. The presence of non-arboreal pollen species suggest that these forests were interspersed with open meadows and grasslands, similar to today's floral assemblages. Fossil molluscs recovered from glaciolacustrine sand exposed along the eastern side of the basin suggest that the limnological characteristics of late glacial Lake Nipigon were similar to those of today.

RÉSUMÉ *Sédimentation tardiglaciaire et évolution du bassin du lac Nipigon, en Ontario.* Le bassin du lac Nipigon situé au nord du bassin du lac Supérieur a assuré le lien hydrologique entre le Lac Agassiz et les Grands Lacs pendant une partie de la dernière glaciation. Une séquence de sédiments glaciolacustres, surtout composés de rythmites silto-argileuses et de sable, a été déposée au large du rivage du Lac glaciaire Nipigon par les eaux de crue du Lac Agassiz et les eaux de fonte du glacier en recul. Les coupes dans six carottes de sédiments et quatre coupes naturelles dans la falaise révèlent la présence de sable dans la partie inférieure (début de la déglaciation) recouverte par 300 à 850 rythmites silto-argileuses (varves). La mise en place de ces varves, comme celle des sédiments plus grossiers le long de la rive ouest, a commencé après 9200 BP, alors que la marge glaciaire reculait vers le nord le long de la ligne de partage des eaux entre le bassin du Lac Nipigon de celui plus élevé du Lac Agassiz. L'absence de fragments glaciels dans les rythmites indique que le glacier s'était déjà retiré. Selon leur altitude par rapport à la plage perchée la moins élevée à West Bay, formée vers 9000 BP, la plupart des rythmites ont été déposées entre 9000 et 8000 BP. Le pollen arboréen présent dans les premiers sédiments postglaciaires des basses terres Nipigon-Supérieur montre que les forêts de conifères et de décidus se sont établies peu après la déglaciation. La présence de pollen non arboréen montre que ces forêts étaient parsemées de prés ouverts et de prairies, comme c'est le cas aujourd'hui. Les mollusques fossiles recueillis dans les sables glaciolacustres le long du côté est du bassin montrent que les propriétés limnologiques du Lac Nipigon sont semblables à celles d'aujourd'hui.

ZUSAMMENFASSUNG *Spätglaziale Sedimentablagerung und Geschichte des Nipigon-Seebeckens, Ontario.* Das Becken des Nipigonsees liegt nördlich von dem des Oberen Sees, und war das hydrologische Bindeglied zwischen dem glazialen Agassizsee und den großen Seen während eines Teils der letzten Enteisung. Eine Sequenz glaziallimnischer Sedimente, hauptsächlich aus Schlamm-Lehm, Rhythmiten und Sand bestehend, wurde in dem küstennahen Wasser des glazialen Nipigansees abgelagert durch überlauf vom Agassizsee und Schmelzwasser von dem zurückweichenden Gletscherrand. Abschnitte von sechs langen Sedimentbohrkernen und vier See-Steilhang-Aufschlüsse zeigen einen sandigen (Beginn der Enteisung) unteren Bereich, der von 300 bis 850 Schlamm-Lehm-Rhythmiten (Warwen) überlagert ist. Die Ablagerung dieser Warwen sowie gröberer Sedimente entlang der westlichen Küste begann nach 9200 v.u.Z., als der Eisrand nordwärts zurückwich, entlang der Wasserscheide, die das Nipiganbecken von dem höheren Agassizbecken nach Westen hin trennte. Das Fehlen von Eisfragmenten in den Rhythmiten läßt vermuten daß das Eis zum Zeitpunkt ihrer Ablagerung vom See schon zurückgewichen daß das Eis zum Zeitpunkt war. Entsprechend ihrer Erhebung in Bezug auf den niedrigsten gehobenen Strand in der West Bay, der sich um 9000 v.u.Z. herausbildete, wurden die meisten Rhythmite wohl zwischen 9000 und 8000 v.u.Z. abgelagert. Baumpol-lenarten in den frühen postglazialen Sedimenten der Superior-Nipigan-Ebenen legen nahe, daß im Nipigansee -Gebiet kurz nach der Enteisung Tannen- und Laubwald sich ausbreiteten. Das Vorkommen von baumfremden Pollenarten zeigt, daß die Wälder von offenen Weiden und Grasland durchsetzt waren, ähnlich der heutigen Bewachsung. Schalentier-Fossile, die aus dem glaziallimnischen Sand gewonnen wurden, entlang der östlichen Seite des Beckens, zeigen, daß die limnologischen Charakteristika des spätglazialen Nipigan-sees den heutigen ähnlich waren.

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INTRODUCTION

The hydrological connection of glacial Lakes Agassiz, Nipigon (Lake Kelvin), and Superior has been studied by Elson (1967), Zoltai (1965, 1967), Teller and Thorleifson (1983, 1987), Clayton (1983), Drexler *et al.* (1983), Farrand and Drexler (1985), Teller and Mahnic (1988), and Thorleifson and Kristjansson (1993). Some of these researchers have speculated on the impact Lake Agassiz overflow had on the Superior basin. Others have discussed the possible effect of outflow on the other Great Lakes and on the Ottawa-St. Lawrence River system (*e.g.* Teller, 1985, 1987; Lewis and Anderson, 1989, 1992; Anderson and Lewis, 1992; Colman *et al.,* 1994); Rodrigues and Vilks, 1994), as well as on oceanic regimes (*e.g.* Broecker *et al.*, 1988, 1989, 1990; Keigwin *et al.*, 1991; Rodrigues, 1992).

Late-glacial events in the Lake Nipigon basin played a major role in the history of Lake Agassiz and in the evolution of the Great Lakes. Located at the "crossroads" of North America, where the 2 million km^2 Agassiz basin periodically overflowed into the Great Lakes, the sedimentary record in the Lake Nipigon basin provides key information about the interaction between two of the continent's largest late glacial drainage basins, and about the history of ice retreat, climate, and isostasy in the region. This paper focuses on the sedimentary record along the western and northern sides of Lake Nipigon, which includes a thick sequence of rhythmites. The age, sedimentology, and paleoecology of these sediments are used to help interpret the late glacial history of this region and its relationship to the paleohydrology along the sourthern Laurentide Ice Sheet.

Fieldwork was conducted along the perimeter of Lake Nipigon, as well as to the west and north of the lake, and consisted of mapping, section description, and coring. A CME-750 rotary auger drill, using hollow and solid stem augers, split-spoon samplers, and Shelby tubes (75 mm diameter x 0.6 m length), allowed for the recovery of a total of 37 m of sediment core from six drill sites within the study area.

SUMMARY OF GENERAL QUATERNARY HISTORY OF THE REGION

The late glacial histories of Lake Agassiz and Lake Superior have been studied by a number of researchers, and their linkage through the Lake Nipigon basin has been discussed by Elson (1967), Zoltai (1965), Clayton and Moran (1982), Teller and Clayton (1983), and others. The chronology of overflow from Lake Agassiz through the Lake Nipigon basin, however, has not been resolved in detail, even though the timing of the linkage of Lake Agassiz with the Great Lakes, and the routing of meltwater to the oceans, is of considerable importance to global change (*e.g.*, Broecker *et al.*, 1989). During the earliest stage of Lake Agassiz (prior to about 10,800 BP), its waters overflowed southward into the Minnesota River valley to the Mississippi River and Gulf of Mexico (Fenton *et al.*, 1983). At this time, the Nipigon basin lay covered by Laurentide ice. By about 10,800 BP, the ice margin had retreated north, and Lake Agassiz water flowed directly into Lake Superior via one or more of the eastern outlets (Fig. 1), marking the onset of the Moorhead Phase of Lake Agassiz.

A readvance of Laurentide ice about 10,000 BP shut off the overflow of water into the Nipigon and Superior basins, and Lake Agassiz waters were forced to rise and again overflow out the southern outlet to the Mississippi River basin (Clayton and Moran, 1982). Some time after about 9500 BP, the eastern outlets once again became ice-free as ice retreated to the northeast, marking the onset of the Nipigon Phase of Lake Agassiz overflow. Initially, this water flowed through the Kashabowie-Seine and Dog River channels west of Thunder Bay to Lake Superior (Fig. 1). Continued retreat of the ice margin along the continental divide allowed overflow through a series of lower outlets into the Nipigon basin (Zotai, 1965, 1967; Teller and Thorleifson, 1983, 1987). From the Nipigon basin, overflow from Lake Agassiz was routed south to Lake Superior through one or more deep bedrock channels (Fig. 1) (Teller and Mahnic, 1988).

Shortly after 8500 BP, the margin of the Laurentide Ice Sheet had retreated far enough north of the Nipigon basin that overflow from Lake Agassiz was directed eastward into Lake Ojibway, by-passing the Great Lakes basins en route to the St. Lawrence valley (*e.g.* Teller and Thorleifson, 1983; Lewis and Anderson, 1989). This period of time, referred to as the Ojibway Phase, marked the end of the hydrological connection of Lake Agassiz, Lake Nipigon, and Lake Superior, which had persisted for about a thousand years (Teller and Thorleifson, 1983).

DESCRIPTION OF SEDIMENTS IN LAKE NIPIGON BASIN

INTRODUCTION

We have subdivided the bedded and laminated sands, silts, and clays of the northwestern Nipigon basin into three types. We feel these types reflect their location in relationship to the sediment source, which was either the ice margin itself or rivers carrying meltwater from the ice or proglacial lakes. Thin couplets of fine grained sediment, typically composed of alternating silt and clay laminae, are considered to have been deposited at a relatively large distance from the source (Type A rhythmites). Thicker, more complexly laminated and rippled sediments were deposited closer to the source, and form rhythmically-repeated units of sand to silt and clay (Type B rhythmites). Laminated sand and silt, without clay units or apparent rhythmic bedding, is considered to be a more proximal deposit (Type C beds). Still coarser sediments exposed to the west of the study area, including steep gravel foreset beds of Gilbert-type deltas, are thought to be the lateral facies equivalent of the Type C beds (and possibly Types B and A), which were deposited in the region of strongest current flow and highest sediment influx. Teller and Thorleifson (1983, 1987) have discussed these extremely coarse-grained (gravel to boulder sized) deposits in conjunction with catastrophic inflow of water to the Nipigon basin from glacial Lake Agassiz.

Type A rhythmites in the Nipigon basin are composed of thin silty clay laminae that abruptly overlie bioturbated clayey silt, silt, or sandy silt (Fig. 2). Laminae are typically distinct in the relatively thick, coarse part of the couplet, and flat-topped symmetrical ripples are often present (Fig. 3). As can be seen in Table I, the clay-rich part of the rhythmite remains a constant and relatively thin (2-6 mm) component in most couplets, whereas the silty to sandy portion varies more than an order of magnitude. In some locations, the thickness of the coarse and fine part of the couplet is nearly equal.

Many aspects of Type B rhythmites are similar to those of Type A, and, in fact, they form part of a continuum of varying rhythmite characteristics. The silty part of the couplet is commonly bioturbated and is abruptly overlain by a comparatively thick clay-rich unit that has no bioturbation and rarely displays lamination. The coarser, silty to sandy part is well laminated, as in the Gull Bay area where some rhythmites contain more than 35 individual laminae in the coarser part of the couplet (Fig. 4). Both in-phase and in-

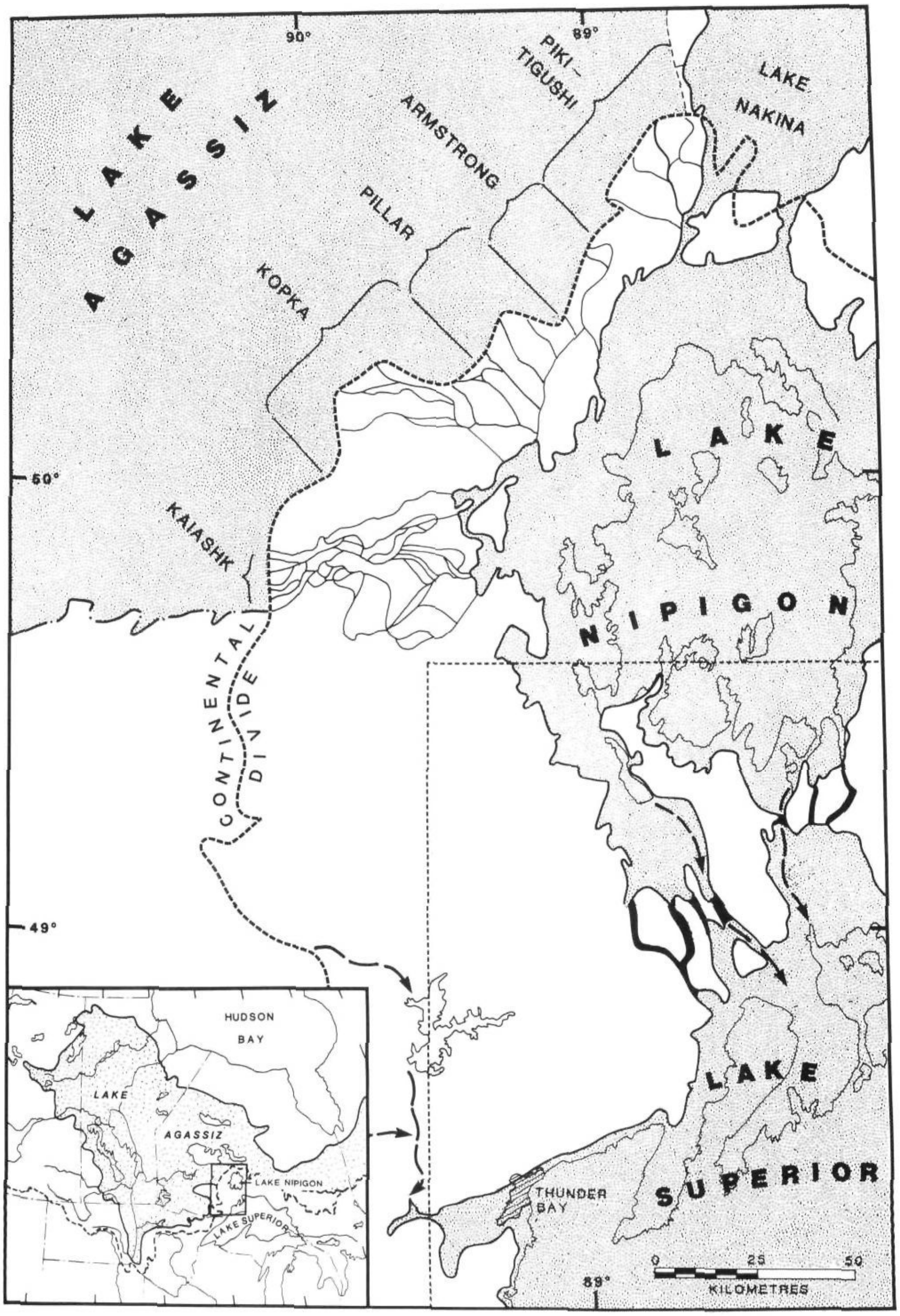

FIGURE 1. *Location of the* Nipigon basin, showing major channels that carried water from glacial Lake Agassiz to glacial Lake Nipigon and to glacial Lake Superior (former lake areas shaded) during the last deglaciation. Channel groups that linked Lake Agassiz with Lake Nipigon are named, and are progressively lower toward the north (after Teller and Thorleifson, 1983).

Carte de localisation du bassin du lac Nipigon montrant les principaux chenaux par où s'écoulaient les eaux du Lac Agassiz vers les lacs glaciaires Nipigon et Supérieur (tramés) au cours de la dernière glaciation. Les chenaux ont été regroupés en ensembles auxquels on a attribué des noms ; ils s'abaissent peu à peu vers le nord (selon Teller et Thorleifson, 1983).

FIGURE 2. Rhythmites in the lower part of Pike Bay section (PIKE, Fig. 5).

Les rythmites dans la partie inférieure du secteur de Pike Bay (PIKE, fig. 5).

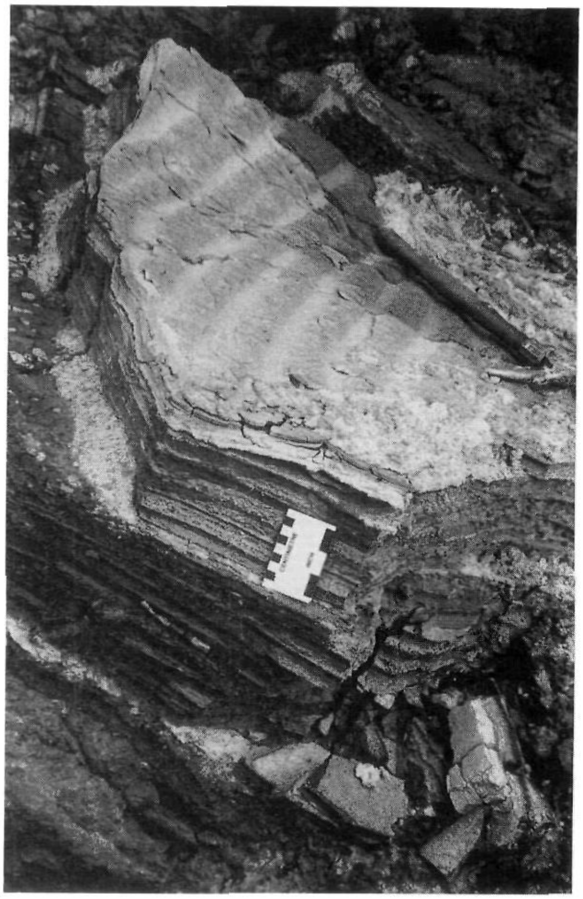

FIGURE 3. Symmetrical rippled sand within rhythmites in wave-cut cliff near mouth of Whitesand River (WHI, Fig. 5).

Sable à ondulations symétriques dans les rythmites d'une falaise érodée près de l'embouchure de la Whitesand River (WHI, fig. 5).

drift climbing ripples may be present in Type B rhythmite sequences, and distinct upward-fining silt and fine sand beds occasionally appear as "third units" at the base of a rhythmite.

Underlying the rhythmically bedded Type A and B sequences in many places are laminated beds of Type C sand to sandy silt. Core recovery in this interval was generally poor and, in many cases, sediment was examined from auger flights. No clay units are present in these sediments and although grain size tends to decrease upward in this Type C sequence, there was no repetitive variation or rhythmic change observed. There are occasional thin beds of coarser sand in some intervals, but for the most part, grain size remains uniform through many metres of section observed.

THE RHYTHMITES

The best preserved sections of rhythmically-bedded sediment are exposed along the modern wave-trimmed shoreline in the northwestern part of the Nipigon basin. Bluffs 5-20 m high at the GUL, PIKE, ENG, and WHI sites (Fig. 5) expose part of the rhythmite sequence and, in places, the overlying and underlying sediments. More complete sediment sections were obtained away from the lake shore by augering and coring at the RES, KOP, CAS, AIR, PIK, and LEE sites (Fig. 5). The general stratigraphy from these test holes is shown in Figure 6.

The largest number of rhythmites counted, 852 in total, occurs at the PIKE location (UTM 359000 m E, 5518500 m N) (Fig. 5). These clayey silt and silty clay couplets (Fig. 2) decrease in thickness from 15 mm at the base of the exposure to 10 mm near the top (see Table I) and are considered as Type A rhythmites. Bioturbation is present in the clayey silt part of the couplet, which is abruptly overlain by the thinner silty clay. The base of the rhythmite sequences is below the modern lake level. Although individual couplets are thicker, the 662 couplets at the ENG site (UTM 363800 m E, 5537000 m N) include counts extrapolated through the "covered" section (Table I), and are similar to those at the PIKE site located 20 km to the south; the 12 m sequence at the ENG site is thicker than the PIKE section.

Some of the most readily observable exposures of rhythmites occur in the wave-cut bluffs at Gull Bay (Fig. 5). Although couplets in the upper 3-4 m of these 6 m exposures are very similar to those at the PIKE section, located

TABLE I

Characteristics of rhythmites in selected exposures along the western side of Lake Nipigon; see locations in Figure 5 and general stratigraphy in Figure 6. "Unit thickness" intervals are arbitrary or were chosen to reflect changes in couplet characteristics

	Unit thickness (m)	Rhythmite type[1]	Total number of rhythmites[2]	Number of rhythmites per m	Mean thickness (mm) cg/fg[3]
GUL	1.0	clsi+cl	150	150	2/2
	2.2	clsi+cl	180	82	6/2
	3.0	si+clsi	90	30	33/3
	3.0	covered	(90)	–	–
	Lake		Total=510		
PIKE	4.0	clsi+sicl	668	167	8/2
	2.0	clsi+sicl	184	92	10/5
	Lake		Total=852		
ENGO	0.7	no rhythmites	–	–	–
	2.0	clsi+sicl	176	88	7/4
	1.0	covered	(50)	–	–
	0.8	clsi+sicl	46	58	15/2
	1.0	clsi+sicl	67	67	8/7
	3.0	clsi+sicl	150	50	15/10
	3.5	covered	(150)	–	–
	1.0	si+cl	23	23	15/15
	Lake		Total=662		
WHI	0.5	no rhythmites	–	–	–
	1.0	no rhythmites	–	–	–
	3.0	no rhythmites	–	–	–
	0.3	sdsi–cl	14	46	–
	3.8	si+cl	250	66	20/6
	1.0	si+cl	19	19	–
	1.0	sd+sdsi+cl	21	21	45/2
	3.0	covered	(60)	–	–
	1.0	sdsi+sicl	12	12	60/6
	1.0	si+cl	23	23	32/6
	1.0	si+cl	21	21	32/6
	1.0	si+cl	26	26	25/6
	1.0	clsi+cl	27	27	27/5
	1.0	clsi+cl	17	17	40/6
	2.0	covered	(40)	–	–
	Lake		Total=530		
WHIa(4)	0.5	si+sicl	47	94	12/4
	2.0	si+sicl	70	35	–
	1.5	si+sicl	24	16	50/5
	1.5	si+sicl	27	18	50/5
	1.5	sdsi+sicl	24	16	50/5
	1.5	sdsi+sicl	25	17	50/8
	1.5	sdsi+sicl	21	14	50/8
	1.5	sdsi+sicl	31	21	50/8
	4.0	covered	(84)	–	–
	Lake		Total=353		

1) cl=clay, si=silt, sd=sand, sicl=silty clay, clsi=clayey silt; 2) numbers in parentheses are extrapolated: 3) mean thickness (mm) of coarse grained (cg) and fine grained (fg) part of couplets; for rhythmites that are triplets the two coarsest parts were included in the cg count. Total of cg + fg is commonly less than the "unit thickness" divided by the "total number of rhythmites" because of occasional anomalously-thick couplets and triplets; 4) located 1 km northeast of WHI (UTM 5562250 m N, 372000 m E).

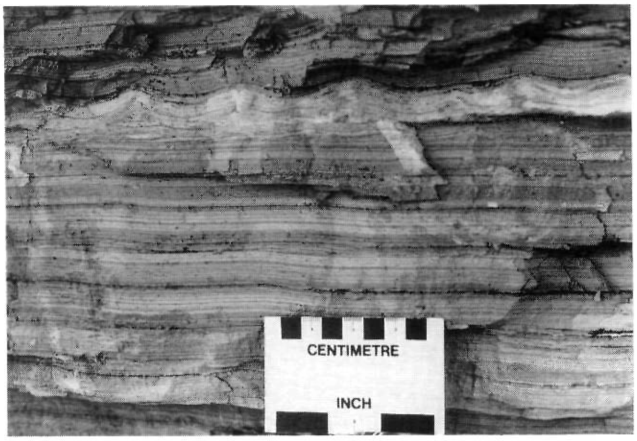

FIGURE 4. Rhythmites at Gull Bay (GUL, Fig. 5), showing multiple laminae in the coarse part; note one set of symmetrical ripples.

Rhythmites à Gull Bay (GUL, fig. 5), présentant de nombreux feuillets dans leur partie grossière ; noter l'ensemble d'ondulations symétriques.

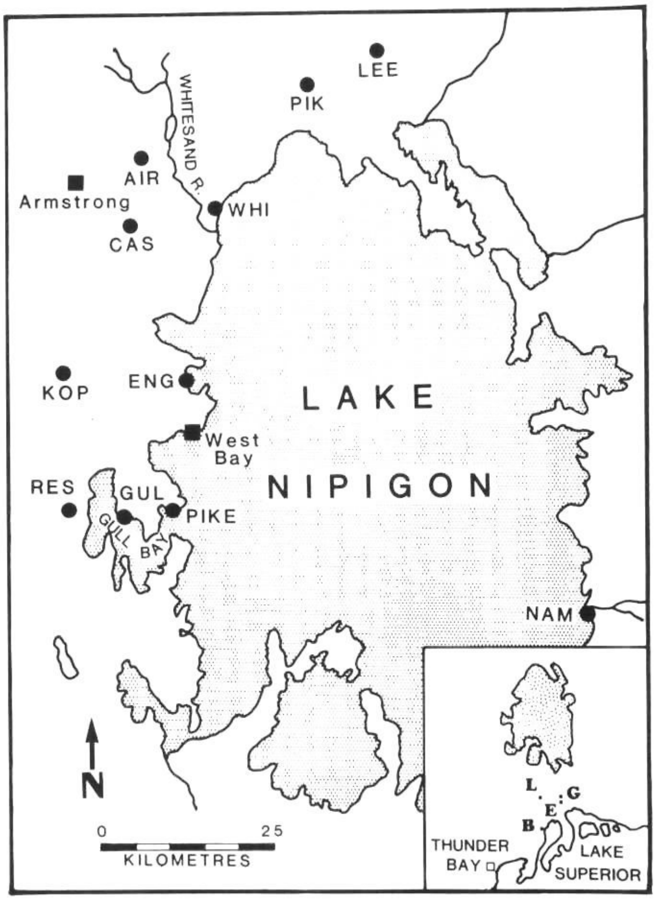

FIGURE 5. Map of Lake Nipigon basin showing locations of core sites and key exposures discussed. General stratigraphy at these sites is shown in Figure 6. Regional relationships shown in Figure 1. Inset shows other borehole locations referred to in this paper.

Carte du bassin du lac Nipigon localisant les sites de forage et les principales coupes commentées. La stratigraphie générale des sites est donnée à la figure 6. Les correspondances régionales sont données à la figure 1. Le carton montre la localisation d'autres trous de forages auxquels on fait référence.

5-10 km to the east, and to the ENG section, the lower part is more silty and contains a third, much thicker silt unit at the base of each rhythmite (Type B). This silt is laminated and, in some beds, in-drift and in-phase cross laminations are present. As a result of this additional unit, the rhythmites are comparatively thicker, averaging 36 mm. A total of 90 of these triplet rhythmites were counted at the GUL site (Table I).

The base of the rhythmite sediments was encountered in two test holes drilled to the west of the lake bluff exposures at KOP and RES (Fig. 6) (UTM 343750 m E, 5522100 m N and 347400 m E, 5518000 m N, respectively). The 10-m-thick sequence of Type B rhythmites at the RES site is laminated and ripple cross laminated, and couplets average 40 mm in thickness. The sequence overlies till. In contrast, the thinner A and B rhythmite sequence at KOP overlies more proximal Type C sand that increases in thickness with depth.

To the north, the rhythmite sequence is thicker, with 23 and >15 m present in the CAS and PIK boreholes (UTM 5560650 m N, 361700 m E and 5584900 m N, 387585 m E, respectively), and more than 18 m exposed along the extensive Whitesand River bluffs (WHI) (Figs. 5 and 6). As in other test holes, sediment recovery was incomplete and many details of sedimentary structures are obscure. In the CAS and PIK test holes, the Type B rhythmites consist mainly of triplets of clay, silt, and sandy silt, with a slightly coarser average grain size in the upper 4-5 m and in the basal few metres of the sequence. Below the upper 4-5 m, the middle silty units of each triplet are commonly bioturbated. Both rhythmite sequences overlie fine sand to silty sand, and it is possible that the poorly recovered sequence below the 15 m of rhythmites in the PIK test hole is also comprised of rhythmites. Rhythmite counts, and extrapolation through non-recovered and deformed intervals, indicate that about 600 rhythmites are present at the CAS site and at least 306 at the PIK site. The average triplet thickness is 40 mm, except near the base of the sequence, where distinct sand laminae increase the rhythmite thickness to about 60 mm.

Wave-trimmed bluffs along the northwestern corner of the lake extend for more than 4 km north of the Whitesand River (Fig. 5), exposing 10 to 20-m-high sequences of sand and silt-clay rhythmites. Although extensively slumped, more than 500 couplets were counted (or extrapolated across "covered" intervals) at the WHI site (Table I). Most of the rhythmically-bedded sequence is comprised of Type A couplets, with a laminated thick (20-60 mm) silty part that contains 1 to 40-mm-thick beds of silty fine sand that is occasionally ripple cross laminated and, in the upper part of the

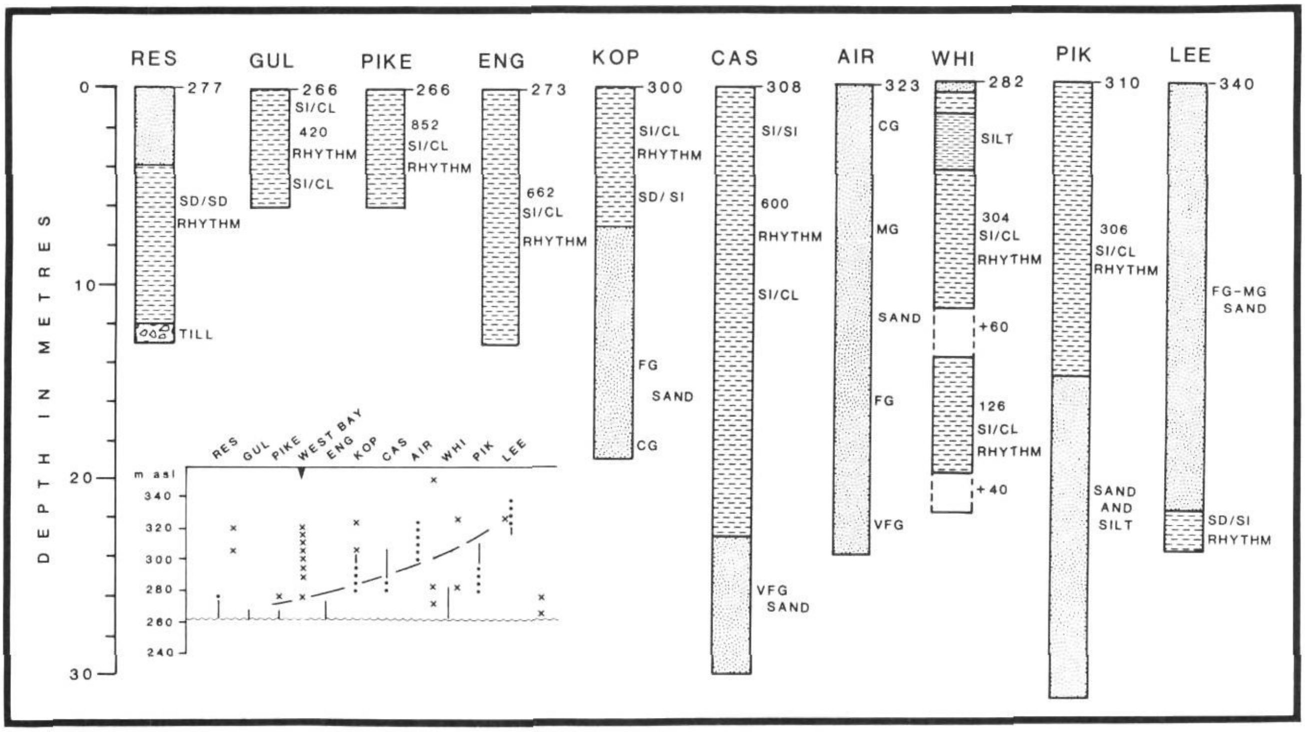

FIGURE 6. Sediment types in boreholes and natural exposures along the western and northern sides of Lake Nipigon (see locations in Fig. 5). Inset shows elevations of rhythmites (solid vertical lines) and sand (dotted line) in boreholes, and the elevations of selected beaches (x); probable correlation of radiocarbon dated beach indicated by dashed line. RHYTHM = Rhythmites; SD = Sand; SI = Silt; FG = Fine Grained Sand; CG = Coarse Grained Sand; MG = Medium Grained Sand; VFG = Very Fine Grained Sand. Numbers refer to number of rhythmites in that sequence.

Les types de sédiments dans les trous de forage et les coupes naturelles de long des côtés ouest et nord du lac Nipigon (localisation à la fig. 5). Le carton montre l'altitude des rythmites (trait plein) et du sable (points) dans les trous de forage ainsi que l'altitude de certaines plages (x) ; la corrélation probable entre les plages datées au radiocarbone est donnée par la ligne interrompue. RHYTHM = rythmites ; SD = sable ; SI = silt ; FG = sable fin ; CG = sable grossier ; MG = sable moyen ; VFG = sable très fin. Les chiffres donnent le nombre de rythmites par séquence.

sequence, contains occasional symmetrical ripples (Fig. 3). The clay part is 5-8 mm thick throughout most of the sequence. The silty part of the couplets are commonly bioturbated. At the WHI location, the symmetrical ripples have rounded crests, and have measured wave length/height ratios of about 20/1.

RAISED SHORELINES

Raised strandlines are scattered around the margins of Lake Nipigon, at elevations as much as 88 m above the modern lake level of 262 m in the northern end of the basin (see Zoltai, 1965; Lemoine, 1989). There have been few precise measurements of elevations on these poorly developed strandlines, most of which are small erosional scarps. At West Bay (Fig. 5), however, there is a series of at least eight wave-cut relict shorelines that have been eroded into the Onaman end moraine; most of these scarps have a concentration of boulders at the base of each scarp. These strandlines were surveyed at elevations of 275, 288, 294, 299, 305, 310, 315, and 320 m, with the upper level forming the relatively flat wave-beveled crest of the moraine. Thus at this location, Lake Nipigon has fallen at least 58 m to its 262 m modern elevation because of differential isostatic rebound or, possibly, because of erosion of the lake's southern outlets and subsequent lowering of water levels. If the elevations of the southern outlets were largely eroded to their bedrock floors by the time these beaches began to form (*cf.* Teller and Mahnic, 1988), then there has been at least 58 m of differential rebound between the southern end of the lake and the central region at West Bay (Fig. 6). This differential would exceed 100 m if extrapolated to the northern end of the basin. Even if the amount of differential rebound from south to north across the lake was closer to 50 m, as indicated by Teller and Thorleifson (1983, Figs. 1-3), still >50 m of lake level lowering can be attributed to erosion of the southern outlets.

Using the northwest-southeast oriented isobases of Teller and Thorleifson (1983), the lowest raised strandline at West Bay (275 m) would fall close to the same isobase as the 272 m elevation of the raised and dated beach deposit at Namewaminikan River (NAM, Fig. 5). Because of the similarity in elevations, it seems likely that these beaches, located on opposite sides of the lake, formed at about the same time. The NAM site is the only radiocarbon dated shoreline in the Nipigon basin, and molluscs in the sands (see "Fossil Molluscan Fauna" section) gave a date of 9760 ± 180 BP (BGS-1150). Thus, all strandlines above the lowest one at West Bay must pre-date 9760 BP although, as noted later, contamination by the "hardwater effect" may mean the true age of these sediments is closer to 9000 BP. Northward extrapolation of the wave-cut scarps at West Bay, perpendicular to the isobases, makes the dated lower strandline roughly equivalent to the 325 m strandlines just north of the lake (Fig. 6, inset).

SEDIMENTOLOGY

A thick sequence of rhythmically-bedded silty and clayey sediment is found throughout the northwestern part of the Lake Nipigon basin. Where the base of these rhythmites has been observed (RES, KOP, CAS, and PIK cores), it overlies sand (Fig. 6). In the LEE core at the northern end of the basin, which lies almost entirely above the elevation of other rhythmite sequences, a thick section of fine to medium grained sand overlies the couplets. To the south, at GUL, PIKE, and ENG (Fig. 6), clay and silt rhythmites are exposed in sections down to (and probably below) present lake level. Up to 852 couplets are exposed in the PIKE section and more than 400 couplets are present in other sections (Table I, Fig. 6).

Although the rhythmically-bedded sequences all appear to have been deposited in the same body of water, detailed correlations cannot be made without chronostratigraphic data. Their overall lithostratigraphic similarity does not prove their chronostratigraphic equivalence, and the sections may form a complex off-lapping sequence toward the south, as a result of differential isostatic rebound and/or southern outlet erosion. For example, the more northerly and more elevated rhythmites of the PIK and LEE test holes may be chronostratigraphically equivalent to the lower rhythmites exposed farther south in the Gull Bay and Pike Bay sections and the RES and KOP test holes (Figs. 5 and 6).

The general upward trend toward thinner rhythmites within the sections argues for a reduced influx of sediment through time, probably as a result of a retreating ice margin. The absence of ice-rafted clasts in the sequence suggests that the glacial margin had wasted back from the edge of the lake by the time rhythmite deposition began. In addition, because there is not an upward increase in grain size and couplet thickness, the emergence of the rhythmites from the deeper waters of Lake Nipigon must have occurred rapidly, before coarser, shallow-water sediments (*i.e.* sand) could accumulate. Rapid differential isostatic rebound, southern outlet erosion, or ice barrier removal, could all result in a rapid reduction in lake level and an absence of relatively coarse-grained (shallow water) sediment at the top of the rhythmite sequences. Although symmetrical ripples in the upper part of several exposures of sediment indicate that the lake floor was at least periodically within wave base, most rhythmites appear to have been deposited in relatively deep, offshore water. In the area north of the lake, rhythmite sedimentation may have been terminated by the rapid influx of sand that, in places, overlies the varved sequence. As noted later, this sand may have been the result of a late glacial readvance into the northern Nipigon basin.

The regularity of couplet repetition and the presence of bioturbation in the silty part of the couplets argues that the rhythmites are varves. As discussed by Smith and Ashley (1985), the uniformity in thickness of the clayey part, the variable thickness of the silty part in relation to the clay, the variable lamination in the silt, and the absence of regular fining-upward trends in the silt part, all support the interpretation that these are annual couplets of sediment. The coarser, occasionally fining-upward units in some varves probably were deposited by inter-annual surge events, resulting from slumping along the basin margin or pulses of sediment supply. More than 300 years of sedimentation are recorded by the rhythmites in most sections, with 420 in

GUL, about 600 in CAS and English Bay, and 852 in the Pike Bay section (Figs. 5 and 6), if couplet counts are extrapolated through slumped and covered intervals.

Climbing ripples and thicker, coarser couplets were deposited under conditions of stronger sediment influxes to the lake (*cf.* Smith and Ashley, 1985). Although these Type B rhythmites, and laterally equivalent coarser sediment (including some Type C beds), may have resulted from shallowing water that brought sediment depocenters closer to the high energy zone at the lake margin, an increase in sediment supply from external sources (*e.g.* a change in ice margin location or influx of water from Lake Agassiz) is equally possible. Trends in sedimentation observed in some sections from Type C to B to A deposits, seem to argue for diminishing external controls, because lake level lowering by southern outlet erosion and/or differential isostasy should generate a Type A to B to C trend.

Although the age of the exposed and cored rhythmites is not known with any certainty, the absence of ice-rafted clasts suggests that they post-date the retreat of active ice from the basin. Because water depths of at least 10 m were probably needed to produce the Type A and B fine-grained rhythmites, those sediments exposed at elevations at or above the modern (262 m) shoreline in the West Bay area or to the south, are likely to have been deposited at or before the time when the 275 m (lowest) shoreline beach in the area formed. The age of the Namewaminikan beach deposit, which is correlated with the 275 m scarp at West Bay (see "Raised Strandlines" section), is about 9000 BP. Thus it seems likely that rhythmites in all exposures and cores south of the central part of the basin (*i.e.* from West Bay south) were deposited in the early postglacial period. To the north, where the elevation of this dated strandline rises by 50 or more metres, rhythmites at elevations below this strandline will be younger, and it is possible that some rhythmites, as well as sandy beds on the surface that lie near the elevation of the modern lake, have only recently emerged from the waters of Lake Nipigon. It seems likely that exposed lacustrine sediments in the northern part of the basin form an off-lapping sequence, with older units exposed at progressively higher elevations northward from the lake.

FOSSIL BIOTA IN LAKE NIPIGON BASIN

MOLLUSCA

Fossil molluscan assemblages in the Nipigon and adjacent Superior lowlands, have been recognized by a number of researchers. Coleman (1922), following up on research by R. Bell in 1870 in the Pic River basin in the northern Lake Superior region, discovered fossil gastropods and bivalves in the sediments of that area. Farrand (1960) identified the shell remains of 17 species of freshwater molluscs in the White Otter River basin north of Lake Superior. Bajc (1986) studied fossil molluscs in a large area along the northeastern shore of Lake Superior, near Marathon, Ontario, and he recognized 84 freshwater and terrestrial taxa. He identified two "immigrational surges" into the Superior basin related to the confluence of ice-marginal lakes at 9500 to 8000 BP, and later at 7000 to 4000 BP. Zoltai and Herrington (1966), in a study along the north shore of Lake Superior from Thunder Bay to Marathon, identified 32 species of fossil molluscs, comparing them to the modern assemblage of Lake Nipigon. To the west of the Nipigon basin, in northwestern Ontario, Zoltai (1969), Nielsen *et al.* (1982), and Bajc (1991) identified molluscs in sediments deposited in Lake Agassiz.

Immediately to the south of the Nipigon basin, near the town of Dorion, E. Pip (*in* Teller and Mahnic, 1988), identified a mixed assemblage of 19 lacustrine and terrestrial molluscs (Table II, Dorion B site, Fig. 5) that were radiocarbon dated at 7790 ± 380 BP (GX-10856). Just to the north, along the Black Sturgeon River, 14 species of molluscs (Table II, Black L site), dated at 9670 ± 270 BP (GX-10856), were also identified (Teller and Mahnic, 1988).

In this study, more than 200 specimens of freshwater bivalves and gastropods were collected from an 11-m-high wave-cut terrace in lacustrine beach sands at the mouth of

TABLE II

Composition of Fossil Mollusc Assemblage in the Superior Basin (sites B and L, Fig. 5; after Pip in *Teller and Mahnic, 1988) and the Nipigon basin (NAM, Fig.5)*

Species	NAM	DORION (B)	BLACK (L)
Amnicola limosa	–	+	+
A. walkeri	–	+	–
Anodonta grandis	–	–	+
Cincinnatia cincinnatiensis	–	–	+
Discus cronkhitei	–	+	–
D. cf *macclintocki*	+	–	–
Fossaria dalli	+	+	–
F. decampi	–	+	+
Gyraulis circumstriatus	+	–	–
G. parvus	–	+	–
Helisoma anceps	+	+	+
H. campanulatum	–	–	+
Lampsilis radiata	+	–	–
Lymnaea stagnalis	+	+	–
Marstonia decepta	–	–	+
Nesovitrea electrina	+	–	–
Oxyloma retusa	–	+	–
Pisidium casertanum	–	+	–
P. fallax	–	+	–
P. variable	+	–	+
Physa gyrina	–	–	+
Promenetus umbilicatellus	+	–	–
Sphaerium lacustre	–	+	–
S. simile	+	–	–
S. striatinum	+	+	+
S. transversum	+	–	–
Stagnicola catascopium	+	–	–
S. elodes	+	+	+
Succinea avara	+	+	–
Valvata tricarinata	+	+	+
V. sincera	–	+	+
Zonitoides arboreus	–	+	–

+ Present – Not Present

the Namewaminikan River on the eastern shoreline of Lake Nipigon (Fig. 5, NAM site; UTM 5501500 m N, 421000 m E). These molluscs were radiocarbon dated at 9760 ± 180 BP (BGS-1150) but, as noted elsewhere, this date appears to be too old because of the hard water effect. Table II lists the 16 species identified by E. Pip within the Namewamikan River (NAM) assemblage, as well as those species identified within the Dorion and Black Sturgeon valley assemblages, noted previously. All of the species identified in the Namewaminikan River fossil assemblage, with the exception of *Discus macclintocki*, can be found within the Lake Nipigon basin at the present time, as indicated by the results of the work of Adamstone (1923, 1924).

Many of the molluscs identified to the south of the Nipigon basin (Table II), and eight of those in the Zoltai and Herrington (1966) study are present within the Namewaminikan River assemblage. Ashworth and Cvancara (1983), in studying the paleoecology of the southern region of the Lake Agassiz basin, identified 84 species of molluscs, 13 of which overlap with the taxa of the Namewaminikan River assemblage. Late Quaternary sediments of the Lake Huron basin yielded 61 species of molluscs (Miller *et al.*, 1985), of which 13 species are common to those present in the Lake Nipigon (NAM) fossil assemblage.

TRACE FOSSILS

Trace fossils (burrows) were found throughout the sequences of silt and clay rhythmites studied within the Lake Nipigon region. In cross sectional view, the burrows are semi-circular to ovoid. In plan view, the traces appear as discontinuous, mm-scale, radiating "feather-like" ridges. The burrows are confined to the silty laminae or located on the bedding plane surfaces of the clayey rhythmite laminae. Teller and Mahnic (1988, p. 1666) describe similar traces in early Lake Superior rhythmites. The nature of these trace fossils suggest that they are grazing traces (*Repichnia*) and that the organism(s) which produced them were motile deposit feeders, and not attached to the substrate (Gibbard and Stuart, 1974). The lack of trace fossils within the clay laminae suggests that the organism responsible for producing the traces may not have been able to survive in environments conducive to the deposition of the finer, clayey sediments. Smith and Ashley (1985) suggest that couplets in which trace fossils are confined to the silty portion, may reflect an annual cycle, where conditions during the winter are unsuitable for trace fossil activity. Thus, the absence of traces within the clay layers of the Lake Nipigon rhythmites suggests that these rhythmites are varves.

PALYNOLOGY

Detailed pollen analyses have not been conducted on sediment from the Nipigon basin. To the south, in the Thunder Bay area, where two archaeological sites were investigated for pollen, radiocarbon ages range from 12,000 BP (on marl) to 9200 BP (on wood) to 4200 BP (on gyttja) (Julig *et al.*, 1990). The older part of the sequence is dominated by spruce, shrubs, and herbs, much like sediment of this age elsewhere in the region (Julig *et al.*, 1990). Closed spruce forest conditions closely followed this early period, continuing until about 8000 BP, after which pine-dominated forests gradually replaced the spruce forests.

Selected subsamples of three cores south of the Nipigon basin near Lake Superior (G, E, B, Fig. 5 inset; see also Teller and Mahnic, 1988), and one core from near Gull Bay (GUL) along the western side of Lake Nipigon (Fig. 5), were submitted to J.H. McAndrews (Royal Ontario Museum) for pollen analyses. According to Teller and Mahnic (1988), these sediments were deposited between 9500 and 8000 BP. The analyses revealed the presence of numerous species of both arboreal pollen and non-arboreal pollen in the sediments. *Pinus* (pine) and *Picea* (spruce) are the dominant arboreal species. Other significant species of arboreal pollen identified were *Abies* (fir), *Betula* (birch), *Ulmus* (elm), and *Alnus* (alder). Significant non-arboreal pollen species identified are Graminaea (grasses) and *Artemisia* (sage).

PALEOECOLOGY OF EARLY POSTGLACIAL FAUNA AND FLORA

The occurrence of many similar molluscan taxa within the sediments of the Lake Agassiz, Lake Nipigon, and Great Lakes basins, results from the hydrological link between these basins during deglaciation. The similarity of the fossil taxa with the modern assemblage suggests that the ecological and environmental characteristics of the glacial Lake Nipigon (Lake Kelvin) setting was comparable to that of the present lacustrine environment. Pelecypods were able to colonize shallow depths of the lake, and the presence of terrestrial gastropods suggests that suitable habitats had also developed along the shores. Although broad sandy and gravelly plains fringed the basin to the north and west, and the offshore environment may have been turbid, soft-bodied organisms were able to colonize the late glacial lake floor.

The results of the palynological analyses conducted on samples of ancient lacustrine sediments in the region suggests the presence of a spruce-dominated forest during most of the deglacial period, with some pine and mixed coniferous and deciduous trees present. Minor amounts of grass and sage pollen suggest that these forested areas were interspersed with open meadows and grasslands. Similar vegetation assemblages today colonize the Lake Nipigon basin. Pollen from a depth of 4.5 to 4.9 m in a core 7.2 km northwest of the RES site (Fig. 5) is anomalous, because of the high content of *Abies* and *Alnus*, which are generally associated with later stages of forest succession. In northwestern Ontario, Bjorck (1985) found that a significant content of this pollen only occurs after 7000 BP, although the age of the sediment in the Nipigon basin containing the *Abies* and *Alnus* appears, by correlation, to be much older.

HISTORY OF DEGLACIATION

About 9500 BP, the glacial margin stood near the mouth of the Kaiashk outlet system along the continental divide west of the Nipigon basin (Fig. 1). Failure of the ice dam in this area resulted in the catastrophic discharge of water into the Nipigon basin from Lake Agassiz (Teller and Thorleifson, 1983, 1987). The configuration of the glacial boundary at this time may have been similar to that proposed by Teller (1987, Fig. 20), which suggests that at least the southwestern corner of the Nipigon basin was free of ice. In this situation, overflow waters from Lake Agassiz would have been routed through channels of the Kaiashk system, and then have overflowed from the southwestern corner of the Nipigon basin into the Superior basin (Fig. 1). Although most have suggested that this Agassiz-Nipigon-Superior linkage occurred about 9500 BP (*e.g.* Teller and Thorleifson, 1983; Teller and Mahnic, 1988), a new interpretation by Thorleifson and Kristjansson argues that ice may not have retreated from the Nipigon basin for the last time until several hundred years later. During the early stages of overflow from Lake Nipigon, water in the Superior basin probably stood at the Minong level, as depicted by the outline of the lakes in Figure 1, and lake water elevations in both the Nipigon and Superior basins were likely only a few metres apart (Teller and Thorleifson, 1983).

On the basis of a detailed study of the Beardmore-Geraldton area, immediately to the east of the Lake Nipigon basin, Thorleifson and Kristjansson (1993) concluded that the southern Nipigon basin became ice free after about 9200 BP, following a rapid glacial readvance from the northeast to the Nipigon Moraine, which lies just to the south of the basin. Using the Dorian varve section in the Superior basin (B, Fig. 5 inset), described by Teller and Mahnic (1988), and the sequence of calcareous clays in the Superior basin that overlie the red, non-calcareous clays related to the Marquette advance at 10,000 BP, Thorleifson and Kristjansson (1993) reasoned that the Nipigon basin must have been re-invaded after 9500 BP by an ice-stream emanating from the Hudson Bay Lowland, where a source of calcareous drift is available. They attribute the two dramatic increases in varve thickness described by Teller and Mahnic (1988) from the Superior basin Dorion section (and considered by Teller and Mahnic to be the result of catastrophic influxes of Lake Agassiz overflow) to this Nipigon basin readvance of ice, and to a later readvance to the Nakina Moraine, which lies to the northeast of the Nipigon basin. Varve counts in the Dorion core indicate that the Nakina advance occurred about 900 years after retreat from the Nipigon Moraine, or at 8300 BP (Thorleifson and Kristjansson, 1993). Thus, after 9200 BP, ice retreated from the Nipigon basin for the last time, opening a succession of Lake Agassiz outlets along the western side of the basin, which allowed a series of floods from Lake Agassiz to pass through to the Superior basin. The erosional and depositional impact of these, and preceding floods in the region west of Lake Nipigon have been described by Teller and Thorleifson (1983, 1987). The laminated sands to sandy silts (Type C beds) that are found in the lower part of cores along the western side of the lake, such as in the RES, KOP, and CAS boreholes and, possibly in the PIK borehole (Fig. 6), including climbing ripples that indicate a high influx of sediment, probably represent the distal deposits of the Lake Agassiz floods. These beds are probably correlative with some of the bouldery and gravelly flood sediments described by Teller and Thorleifson (1983) that were deposited near the mouth of the eastern outlets of Lake Agassiz when the areal extent of Lake Nipigon was much greater, perhaps like that shown in Figure 1.

Continued ice retreat allowed Lake Agassiz overflow to be routed through more northerly channels into the Nipigon basin (Fig. 1) and the influx of sandy sediment to the southwestern part of basin declined. Rhythmic deposition of clayey and silty couplets (Types A and B) was initiated. We believe these couplets are annual increments of deposition, perhaps controlled in part by underflow (density) currents from continuing inflow of Lake Agassiz waters or meltwater generated at the ice margin.

In response to differential isostatic rebound of the Nipigon basin, and erosion of the channels linking this basin with Lake Superior, water levels declined through this period. A series of eight water levels are recorded by strandlines at West Bay (Fig. 6), the highest more than 50 m above the modern lake level. When the lake stood at the uppermost beach, during the early stages of deglaciation, the main shoreline of Lake Nipigon would have been about 20 km west of its present position. The Lake Nipigon strandlines at West Bay suggest that there may have been a number of relatively abrupt changes in lake level in the early history of the lake, possibly related to ice retreat or outlet erosion. The presence of sand over clay and silt rhythmites at levels well above the modern lake in the northern part of the Nipigon basin (*e.g.* LEE core, Fig. 6), and possibly the second recorded increase in varve thickness in the Superior basin Dorion core (Teller and Mahnic, 1988; Thorleifson and Kristjansson, 1993), may be related to a readvance of ice to the Nakina Moraine, northeast of Lake Nipigon. If this is the case, varve sedimentation in the northern end of the Nipigon basin may have ended about 8300 BP, the time Thorleifson and Kristjansson (1993) assigned to the readvance of this ice.

The lowermost raised strandline at West Bay (275 m) is correlated with the sands at Namewaminikan River (NAM, Fig. 5), which have a radiocarbon date on mollusc shells of 9760 ± 180 BP (BGS-1150). This date is considered too old because of the "hard water effect" (*cf.* Karrow and Geddes, 1987). On the basis of discussions of radiocarbon dates on carbonate shells from elsewhere in the region (Zoltai and Herrington, 1966; Nielsen *et al.*, 1982), we conclude that the true age of the mollusc shells and the associated beach deposit at this location (and its correlative strandline at West Bay) is 400-1000 years younger than the radiocarbon date. If, as Thorleifson and Kristjansson (1993) suggest, ice retreated from the Nipigon basin for the last time about 9200 BP, then the age of the mollusc-bearing sediments is likely to be about 9000 BP.

Varved sequences observed in cores and lake bluffs along the southwestern side of Lake Nipigon that lie between the elevation of the dated raised strandline at West Bay and the modern level of the lake (at RES, KOP, GUL, PIKE and ENG, Fig. 6), are believed to have been deposited in early postglacial time. The absence of ice rafted clasts in these Type A and B rhythmites, however, indicates that ice already had retreated from the lake. Thus, the more than 850 varves at the PIKE section, as well as most of the varves in other sections discussed in this paper, were deposited between 9000 and 8000 years BP. To the north of West Bay, rhythmites of similar age must lie at progressively higher elevations because of differential isostatic rebound, and varves exposed in bluffs along the modern lake, such as at WHI (Fig. 6), probably were deposited later in the Holocene. Sediments near lake level in the northernmost end of the basin, which are dominantly sand, may only recently have emerged from the shallow waters of the lake as a result of isostasy.

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REFERENCES

Adamstone, F.B., 1923. The distribution and economic importance of Mollusca in Lake Nipigon. Toronto University Studies, Ontario Fisheries Research Laboratory, 14: 69-119.

——— 1924. The distribution and economic importance of the bottom fauna of Lake Nipigon, with an appendix on the bottom fauna of Lake Ontario. Toronto University Studies, Ontario Fisheries Research Laboratory, 24: 35-100.

Anderson, T.W. and Lewis, C.F.M., 1992. Climate influences of deglacial drainage changes in southern Canada at 10 to 8 ka suggested by pollen evidence. Géographie physique et Quaternaire, 46: 255-272.

Ashworth, A.C. and Cvancara, A.M., 1983. Paleoecology of the southern part of the Lake Agassiz basin, p. 133-156. *In* J.T. Teller and L. Clayton, eds., Glacial Lake Agassiz. Geological Association of Canada Special Paper 26.

Bajc, A.F., 1986. Molluscan paleoecology and Superior basin water levels, Marathon, Ontario. M.Sc. thesis, University of Waterloo, 271 p.

——— 1991. Late Glacial and glaciolacustrine history of the Fort Frances — Rainy River area, Ontario, Canada. Ph.D. thesis, University of Waterloo, 309 p.

Broecker, W.S., Andre, M., Wolfli, W., Oeschger, H., Bonani, G., Kennet, J. and Peteet, D., 1988. The chronology of the last deglaciation: Implications to the cause of the Younger Dryas Event. Paleoecology, 3: 1-19.

Broecker, W.S., Kennett, J., Flower, B., Teller, J., Trumbore, S., Bonani, G. and Wolfli, W.,1989. Routing of meltwater from the Laurentide Ice Sheet during the Younger Dryas cold episode. Nature, 341: 318-321.

Broecker, W. S., Bond, G., Klas, M., Bonani, G. and Wolfli, W., 1990. A salt oscillator in the Atlantic Ocean? I. The concept. Paleoceanography, 5: 469-477.

Clayton, L., 1983. Chronology of Lake Agassiz drainage to Lake Superior, p. 291-307. *In* J.T. Teller and L. Clayton, eds., Glacial Lake Agassiz. Geological Association of Canada Special Paper 26.

Clayton, L. and Moran, S.R., 1982. Chronology of Late Wisconsinan glaciation in middle North America. Quaternary Science Reviews, 1 (1): 55-82.

Coleman, A.P., 1922. Glacial and post-glacial lakes in Ontario. Toronto University Studies, Ontario Fisheries Research Laboratory, 10: 1-76.

Colman, S.M., Keigwin, L.D. and Forester, R.M., 1994. Two episodes of meltwater influx from glacial Lake Agassiz into the Lake Michigan basin: Contrasts in climatic and oceanographic effects. Geology, 22: 547-550.

Drexler, C.W., Farrand, W.R. and Hughes, J.D., 1983. Correlation of glacial lakes in the Superior basin with eastward discharge events from Lake Agassiz, p. 309-329. *In* J.T. Teller and L. Clayton, eds., Glacial Lake Agassiz. Geological Association of Canada Special Paper 26.

Elson, J.A., 1967. Geology of glacial Lake Agassiz, p. 36-95. *In* W. Mayer-Oakes, ed., Life, Land, and Water. University of Manitoba Press, Winnipeg.

Farrand, W.R., 1960. Former shorelines in western and northern Lake Superior basin. Ph.D. thesis, University of Michigan, Ann Arbor, 226 p.

Farrand, W.R. and Drexler, C.W., 1985. Late Wisconsinan and Holocene history of the Lake Superior basin, p. 17-32. *In* P.F. Karrow, and P.E. Calkin, eds., Quaternary evolution of the Great Lakes. Geological Association of Canada Special Paper 30.

Franklin, J.M., McIlwaine, W.H., Poulsen, K.A. and Wanless, R.K., 1980. Stratigraphy and depositional setting of the Sibley Group, Thunder Bay district, Ontario, Canada. Canadian Journal of Earth Sciences, 17: 633-651.

Fenton, M.M., Moran, S. R., Teller, J. T. and Clayton, L., 1983. Quaternary stratigraphy and history in the southern part of the Lake Agassiz basin, p. 49-74. *In* J.T. Teller and L. Clayton, eds., Glacial Lake Agassiz. Geological Association of Canada Special Paper 26.

Gibbard, P.L. and Stuart, A.J., 1978. Trace fossils from proglacial lake sediments. Boreas, 3: 69-74.

Julig, P.J., McAndrews, J. H. and Mahaney, W.C., 1990. Geoarchaeology of the Cummins site on the beach of proglacial Lake Minong, Lake Superior Basin, Canada, p. 21-50. *In* N.P. Lasca and J. Donahue, eds., Archaeological geology of North America. Geological Society of America, Centennial Special Volume 4.

Lemoine, R. M.J., 1989. Late glacial history, paleoecology, and sedimentation in the Lake Nipigon basin, Ontario. M.Sc. thesis, University of Manitoba, Winnipeg, 218 p.

Lewis, C.F.M. and Anderson, T.W., 1989. Oscillations of levels of cool phases of the Laurentide Great Lakes caused by inflows from glacial Lakes Agassiz and Barlow-Ojibway. Journal of Paleolimnology, 2: 99-146.

——— 1992. Stable isotope (O and C) and pollen trends in eastern Lake Erie, evidence for a locally induced climatic reversal of Younger Dryas age in the Great Lakes basin. Climate Dynamics, 6: 241-250.

Miller, B.B., Karrow, P.F. and Mackie, G.L., 1985. Late Quaternary molluscan faunal changes in the Huron basin, p. 95-107. *In* P.F. Karrow and P.E. Calkin, eds., Quaternary evolution of the Great Lakes. Geological Association of Canada Special Paper 30.

Nielson, E., McKillop W. and McCoy, J., 1982. The age of the Hartman moraine and the Campbell beach of Lake Agassiz in northwestern Ontario. Canadian Journal of Earth Sciences, 19: 1933-1937.

Rodriques, C.G., 1992. Successions of invertebrate microfossils and the Late Quaternary deglaciation of the central St. Lawrence Lowland, Canada and the United States. Quaternary Science Reviews, 11: 503-534.

Rodriques, C.G. and Vilks, G.,1994. The impact of glacial lake runoff on Goldthwait and Champlain Seas: The relationship between glacial Lake Agassiz runoff and the Younger Dryas. Quaternary Science Reviews, 13 : 923-944.

Smith, N.D. and Ashley, G.M., 1985. Proglacial lacustrine environments, p. 135-215. *In* G.M. Ashley, J. Shaw and N.D. Smith, eds., Glacial sedimentary environments. SEPM Short Course No. 16, Society of Economic Paleontologists and Mineralogists.

Teller, J.T., 1985. Glacial Lake Agassiz and its influence on the Great Lakes, p. 1-16. *In* P.F. Karrow and P.E. Calkin, eds., Quaternary evolution of the Great Lakes. Geological Association of Canada Special Paper 30.

——1987. Proglacial lakes and the southern margin of the Laurentide Ice Sheet, p. 39-69. *In* W.F. Ruddiman and H.E. Wright, eds., North America and adjacent oceans during the last deglaciation. Geological Society of America, The Geology of North America, K-3.

Teller, J. T. and Clayton, L. eds., 1983. Glacial Lake Agassiz. Geological Association of Canada Special Paper 26, 451 p.

Teller, J.T. and Thorleifson, L.H., 1983. The Lake Agassiz-Lake Superior connection, p. 261-290. *In* J.T. Teller and L. Clayton, eds., Glacial Lake Agassiz. Geological Association of Canada Special Paper 26.

——1987. Catastrophic flooding into the Great Lakes from Lake Agassiz, p. 121-138. *In* L. Mayer and D. Nash, eds., Catastrophic flooding. Allen and Unwin, Boston.

Teller, J.T. and Mahnic, P., 1988. History of Sedimentation in the northwestern Lake Superior basin and its relation to Lake Agassiz overflow. Canadian Journal of Earth Sciences, 25: 1660-1673.

Thorleifson, L. H. and Kristjansson, F. J., 1993. Quaternary geology and drift prospecting, Beardmore-Geraldton area, Ontario. Geological Survey of Canada, Memoir 435, 146 p.

Zoltai, S.C., 1965. Glacial features of the Quetico-Nipigon area, Ontario. Canadian Journal of Earth Sciences, 2: 247-269.

——1967. Eastern outlets of Lake Agassiz, p. 107-120. *In* W. Mayer-Oakes, ed., Life, land, and water. University of Manitoba Press, Winnipeg.

——1969. Sampling fossil molluscs from glacial Lake Agassiz sediments. Journal of Paleontology, 43: 534-537.

Zoltai, S.C. and Herrington, H.B., 1966. Late glacial molluscan fauna north of Lake Superior, Ontario. Journal of Paleontology, 40: 439-446.