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Deglaciation stages of the Laurentide Ice Sheet in Canada and related glaciomarine and glaciolacustrine deposits. Review of selected features

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Resumen. Durante la deglaciación de los mantos de hielo se genera una gran cantidad de agua de fusión, una gran parte de la cual queda retenida en grandes lagos en los bordes de los glaciares. Un ejemplo es el Manto de Hielo Laurentido que cubrió una gran extensión de Norteamérica y dio lugar a vastos lagos en las distintas etapas de su deglaciación. Restos de estos ambientes son los Grandes Lagos de Norteamérica así como otros lagos en el centro y noroeste de Canadá. También se formaron mares glaciales poco profundos, desecados en parte a causa del levantamiento isostático postglacial. El legado de estos ambientes son vastas planicies, actualmente fértiles áreas agrícolas en su mayor parte, situadas sobre arenas y arcillas glaciolacustres y glaciomarinas.

Los sedimentos depositados en los lagos y mares glaciales adquieren características específicas que son fácilmente preservadas en el registro geológico. Los emplazamientos glaciolacustres y glaciomarinos se caracterizan por su reletivamente amplia distribución y por la complejidad de los depósitos deltaicos, costeros y de aguas profundas. Las ritmitas son facies formadas por decantación diferencial en verano (fusión) e invierno (bajo cobertura de hielo), o por flujos turbios procedentes de los rios glaciales. Igualmente típicos son los depósitos de sedimentos gruesos de origen fluvial depositados en aguas profundas, generalmente en el frente sumergido de los glaciares, en los puntos de descarga de corrientes glaciales canalizadas. Otros depósitos de material grueso de aguas profundas son los generados por materiales transportados por bloques de hielo estacional o icebergs.

El tipo de sedimentos y los paisajes a que han dado lugar tienen importantes implicaciones para los usos del suelo. Los ambientes deltaicos y costeros constituyen valiosas fuentes de áridos, mientras las areas predominantemente limo-arcillosas son generalmente excelentes

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áreas agrícolas. Sin embargo, los depósitos de arcillas marinas son muy inestables y pueden dar lugar a deslizamientos, como ocurre con las arcillas de la formación Leda en el sureste de Canadá. Además algunas llanuras de ortigen lacustre, topográficamente bajas, en el centro de Canadá suelen inundarse durante la primavera.

Palabras clave: sedimentos glaciolacustres, sedimentos glaciomarinos, deglaciación, usos del suelo, Canadá.

ABSTRACT. A large quantity of meltwater is generated during deglaciation of ice sheets, and much of it is temporarily trapped in large lakes along the glacier margins. A case in point is the Laurentide Ice Sheet which covered a large part or North America and developed vast lakes at different stages of its deglaciation. Remnants of these environments are the Great Lakes of North America and other large lakes in central and northwestern Canada. Shallow glacial seas developed as well, later partly dried out due to postglacial isostatic uplift. The legacy of these environments are vast plains, now for the most part fertile farmland, underlain by glaciolacustrine and glaciomarine sand and clay.

Sediments deposited in glacial lakes and seas adquire characteristic features which are likely to be preserved in the geological record. Glaciolacustrine and glaciomarine settings are characterized by relatively wide distribution, and by complexes of deltaic, shore and deep water deposits. Rhythmites are characteristic features formed either by differential summer (melt and flood season) and winter (under an ice cover) settling, or by turbid flows derived from glacial rivers. Also typical are deep-water, coarse, fluvial deposits, usually formed at the submerged front of glaciers, at the discharge point of tunnelled glacial streams. Other coarse grained deep-water deposits generated by materials rafted by blocks of seasonal ice or by icebergs.

The type of sediments, and the landscapes that have developed have strong implications for landuse. Deltaic and coastal environments are valuable sandy resources. Silty clay areas are usually good farming country. However, marine clay deposits are notoriously unstable (sensitive clay) and can slump readily, as is the case with the Leda Clay in southeastern Canada. Furtermore, some low-lying, emersed lacustrine plains of central Canada are prone to spring flooding.

Keywords: glaciolacustrine deposits, glaciomarine deposits, deglaciation, land use, Canada.

Introduction

Enough water was trapped in continental ice sheets during the Pleistocene to lower sea level worldwide by about 150 m. One large glacier, the Laurentide Ice Sheet, cover great parts of North America. At his maximum, it had an ice volume of between about 30,900 and 34,800x10^6 km^3 (Fig 1; Denton and Hughes, 1981). The objective of this paper is to briefly review the stages of deglaciation of the Laurentide Ice Sheet, and to focus on characteristic types of sediments that formed in glacial lakes and seas at its southern edge.

The Laurentide Ice Sheet, like all others, developed and waned in spurts, whereas major expansions of such glaciers are generally thought to have been relatively slow (of the order of a hundred-thousand years) the retreats and collapse are considered much faster (of the order of thousands to tens of thousands of years). Furthermore, the weight of the ice sheet forced subsidence of the Earth's crust of the order of several hundred meters. Modern examples of this exist in Greenland whose central region is depressed well below sea lever (Radok et al., 1982), and in Antarctica where the continental edge is higher than the ice-covered continent interior.

A byproduct of the differential, glacially induced subsidence and postglacial isostatic uplift was the temporary trapping of meltwater in lakes along the edge of the ice sheet or the formation of outwash channels, at times parallel to the terminus of the glacier. The Laurentide Ice Sheet had significant portions of its terminus in lakes or seas, which led to considerable amounts of glaciolacustrine and glaciomarine deposits. Their proportions are further magnified in the geological record because they are the deposits most likely to be preserved, in contrast to glacial terrestrial sediments which are likely to be eroded. The distribution of glacial erosional and depositional features and dating of the materials involved allow a relatively detailed reconstruction of ice margin environments during different stages of deglaciation.

Deglaciation stages

The deglaciation stages of the Laurentide Ice Sheet have been fairly accurately reconstructed (Flint, 1943, 1971; Denton and Hughes, 1981; Dyke and Prest, 1987; Fulton, 1989). Following the excellent syntheses by Prest (1970) and Dawson (1992), a few deglaciation stages are reported here as windows to past conditions, showing the positioning of the glacier edge and the bodies of glacial meltwater.

74,000 yrs BP. This was the time of maximum extension of the North American ice sheet over the eastern seaboard. The continental shelf of eastern Canada and northeastern United States was covered by glacier ice (Fig. 1).

18,000-20,000 yrs BP. The last maximum extension of the Laurentide Ice Sheet over the continent was reached at this time (Fig. 1). However, the glacier had already retreated somewhat from the eastern marine shelves, and was not present in the cold, but arid areas of northern Alaska which lay far from open oceanic waters and rest in the shadow of mountains.

14,000 yrs BP. Ice shelves were well defined in deep embayments in the Arctic Sea area (Fig. 2). Ice covered parts of the shelf of Labrador, Nova Scotia and the Gulf of St. Lawrence near the present mouth of the St. Lawrence River. Inland, the glacier retreated to the Great Lakes and away from the Atlantic seaboard. Terminal lobes and proglacial lakes developed in the Great Lakes region and other areas along the

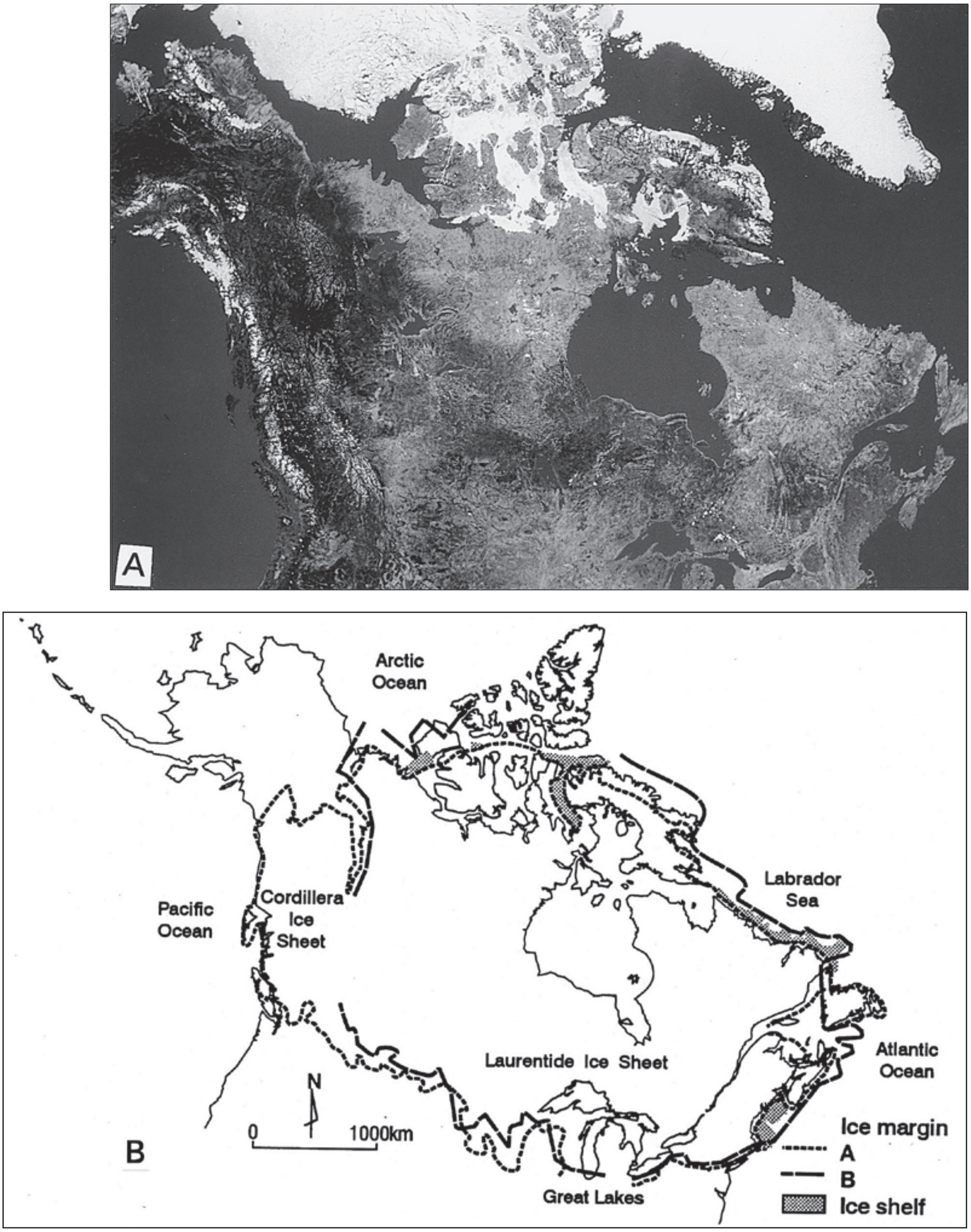

Fig. 1. Glacial and postglacial Canada. A. Physiographic map of Canada showing numerous lakes disposed in a semicircle around Hudson Bay, legacy of Pleistocene glaciation; B. The Laurentide Ice Sheet as reconstructed at: A = approximately 74,000 years B.P., and B = approximately 20,000 years B.P. (after Vincent and Prest, 1987, and Dawson, 1992).

southern margin of the ice. To the west, the Laurentide Ice Sheet started separating from the Cordilleran Ice Sheet.

13,000 yrs BP. Marine terminations still existed in the shelves of the Arctic Sea and in a narrow strip along the Labrador and Newfoundland coast (Fig. 3) and the north shore of the St. Lawrence River (Goldwait Sea: this part of the present St. Lawrence River Valley was still depressed as the glacier had just left it). At this time, the main body of the Laurentide Ice Sheet started separating from the Appalachian ice cap of the Gaspé Peninsula.

The eastern American seaboard (Nova Scotia, Maine) was abandoned and the glaciar retreated to the Appalachians, leaving only a few ice outliers scattered over Nova Scotia and Price Edward Island.

Inland, the Appalachian ice cap developed lobes into the eastern valleys. In the Great Lakes region, the terminal glacial lobes were fully developed and large lakes formed, such as Lake Whittlesey, which were wider than the present day lakes. To the west, the Laurentide Ice Sheet separated from the Cordilleran Ice Sheet and proglacial lakes started to develop.

12,000 yrs BP. Few ice shelves were preserved in the central-eastern northern areas (Fig. 4). Only part of the Labrador nearshore area was covered by ice. The glacier also retreated from the north shore of the Gulf of St. Lawrence.

Inland, a few remnant ice blocks persisted in Nova Scotia and Newfoundland. The Appalachian ice crap shrank greatly. The Laurentide Ice Sheet was rimmed to the south and southeast by lakes, from Lake Vermont on the east, to a large lake covering the St. Lawrence Lowland, to Lake Algonquin in the Great Lakes region, to Lake Agassiz (Teller and Clayton, 1983), and the precursor of the present Great Slave Lake respectively in the west and northwest.

11,000 yrs BP. Only one large ice shelf persisted in the Arctic Sea in the present Gulf of Boothia area. In the south, the main event was the penetration of sea water into the upper St. Lawrence River valley replacing the existing proglacial lake. In the northwest, the large glacial Lake McConnell was formed.

Various changes occurred in the Great Lakes and Lake Agassiz. Fluctuations in the glacier terminus triggered large, perhaps even catastrophic discharges from Lake Agassiz eastward into Lake Algonquin (this interpretation is not accepted everybody, see Barnett (1992) for a different opinion). To the east, the Newfoundland ice cap started to split into smaller blocks.

10,000 yrs BP. The ice shelf of the Gulf of Boothia persisted in the north (Fig. 5). Some lobed terminations of the ice sheet extended into the fjords of eastern Baffin Island. Ice disappeared from the central Appalachians and eastern areas, except for a few remnant of ice-blocks in Newfoundland. The precursors of the Great Lakes shrank as waters flowed out along depressed northern outlets near the glacier terminus. A

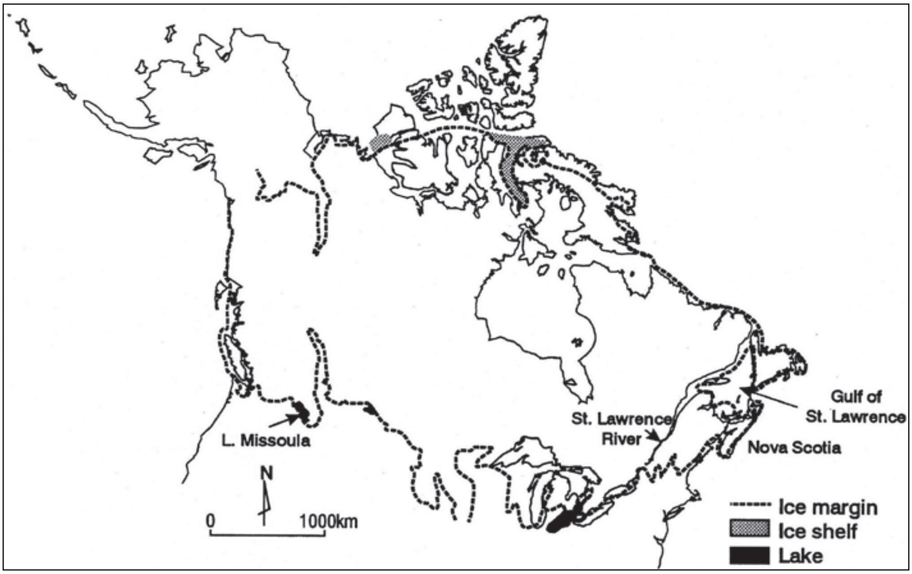

Fig. 2. The Laurentide Ice Sheet as reconstructed at approximately 14,000 years B.P. (after Vincent and Prest, 1987, and Dawson, 1992).

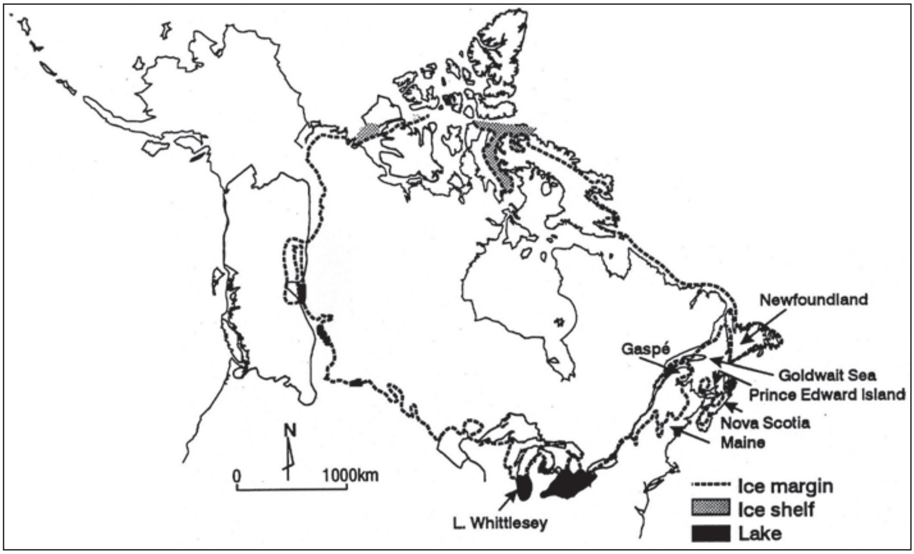

Fig. 3. The Laurentide Ice Sheet as reconstructed at approximately 13,000 years B.P. (after Vincent and Prest, 1987, and Dawson, 1992).

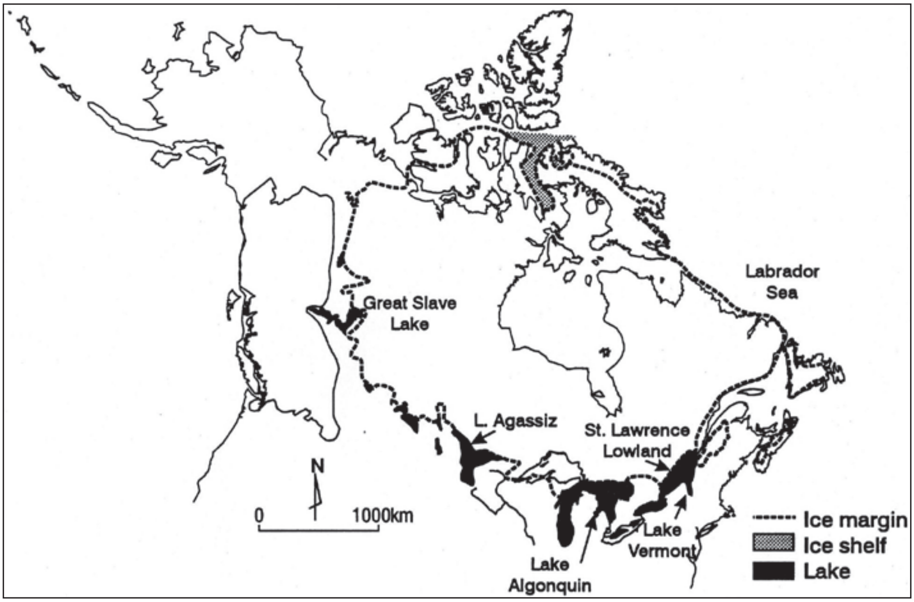

Fig. 4. The Laurentitde Ice Sheet as reconstructed at approximately 12,000 years B.P. (after Vincent and Prest, 1987, and Dawson, 1992).

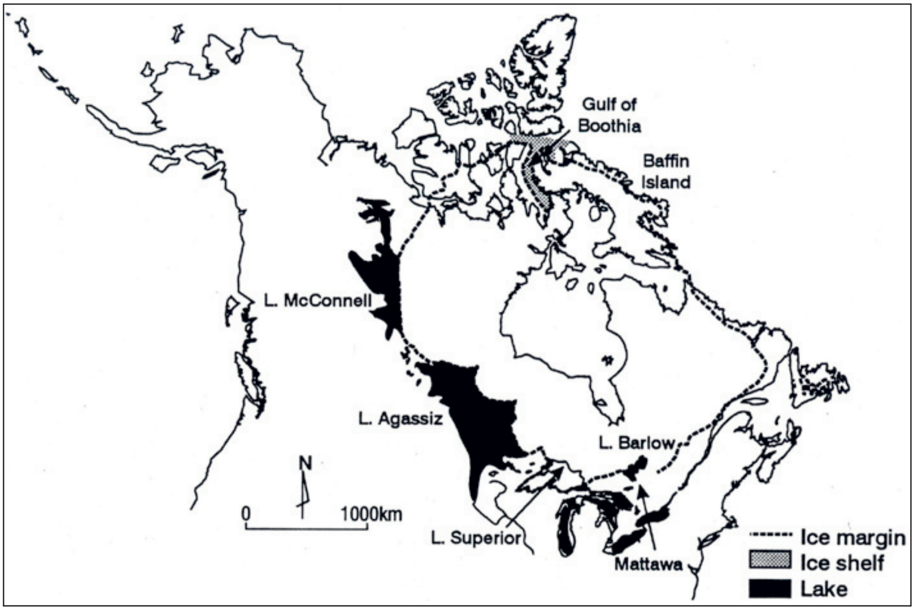

Fig. 5. The Laurentide Ice Sheet as reconstructed at approximately 10,000 years B.P. (after Vincent and Prest, 1987, and Dawson, 1992).

large delta formed at the eastern end of one of these outlets in northeastern Ontario (Mattawa). Proglacial Lake Barlow developed in northern Ontario, and Lake Agassiz and Lake McConnell enlarged in the west. A glacier advance in the Lake Superior area cut off Lake Agassiz from the eastern lakes.

9,000 yrs. BP. The Gulf of Boothia ice shelf disappeared in the north, and a large marine embayment started to open in what would later become the Hudson Strait. A large lake appeared along the ice margin in Labrador. The Great Lakes basins started to refill as the land was tilted back to the south due to differential isostatic uplift as the glaciers retreated farther northward. Because the ice thickness was greater to the north, the land was more depressed there and when the glaciar melted it rebounded more rapidly and to a greater extent than the southern lands. Lake Ojibway-Barlow enlarged in the north and received discharge from Lake Agassiz which, at this time, was slightly reduced in extent. The glacier retreated from the northwest, and Lake McConnell was split into smaller scattered lakes.

8,400 yrs BP. The glacier retreated inland in the north, but Hudson Bay was still under the ice (Fig. 6). The Hudson Strait embayment was enlarging toward Hudson Bay. Lake Agassiz enlarged to its maximum extent and discharged into Lake Ojibway, forming a continuous 3,100km-long water body along the southern margin of the Laurentide Ice Sheet. Relatively large lakes formed in the northwest, east of the remnants of Lake McConnell. The waters of the southern Great Lakes were rising slowly, but did not quite reach present day levels. The large lake in Labrador was still present.

At this time the regular retreat of the glacier was interrupted by a series of glacial surges (Dyke and Prest, 1987; Dredge and Cowan, 1989) into Lake Ojibway, into the Hudson Bay Lowland (SW of Hudson Bay), and throughout the Canadian Arctic.

8,000 yrs BP. The Laurentide Ice Sheet splits into two parts: the Labrador ice cap to the east and the Keewatin-Foxe Basin ice cap to the west and north. Possibly an ice island persisted in Hudson Bay. The sea re-entered Hudson Bay forming the Tyrrel Sea whose shores were located several hundred kilometres inland from the present ones because the land was still depressed from the recent glaciation (Fig. 7). To the south, the Mississipi River, the main collector of meltwater during the early stages of deglaciation, no longer received waters from glacial Lake Agassiz. The river changed from its early braided form when it was carrying meltwater, to a meandering form as it drained large parts of the North American continent. Lake Agassiz was greatly reduced in size and split, to become in part the precursor of the present Lake Winnipeg. The Great Lakes continued their slow adjustment toward their present levels.

The brief synthesis presented above is but one generalized view of what happened. The reader should be aware that other interpretations are possible, particularly when each area is analyzed in detail. To illustrate this to some extent, some aspects of the complex case of southwestern Ontario are presented here (Karrow andl Calkin, 1985; Barnett, 1992).

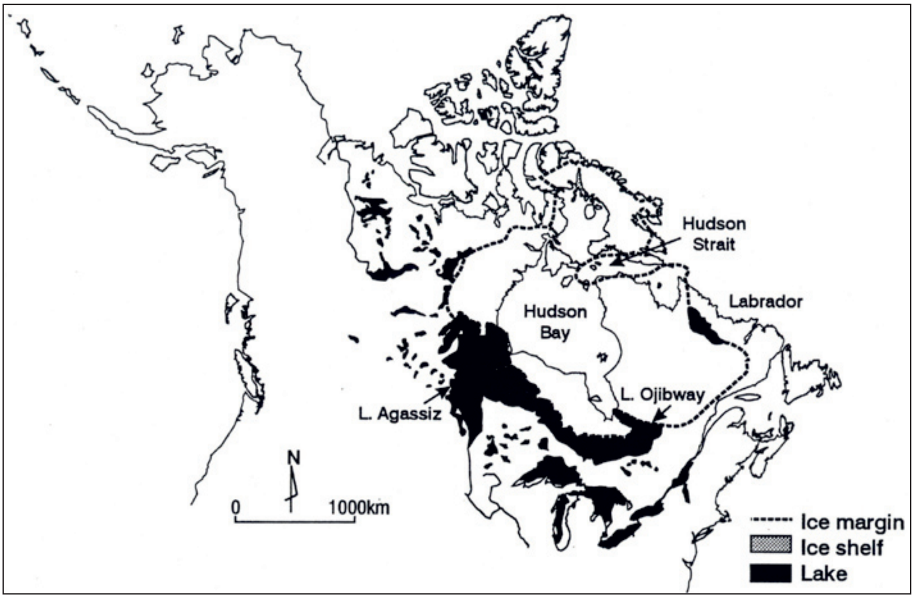

Fig. 6. The Laurentide Ice Sheet as reconstructed at approximately 8,400 years B.P. (after Vincent and Prest, 1987, and Dawson, 1992).

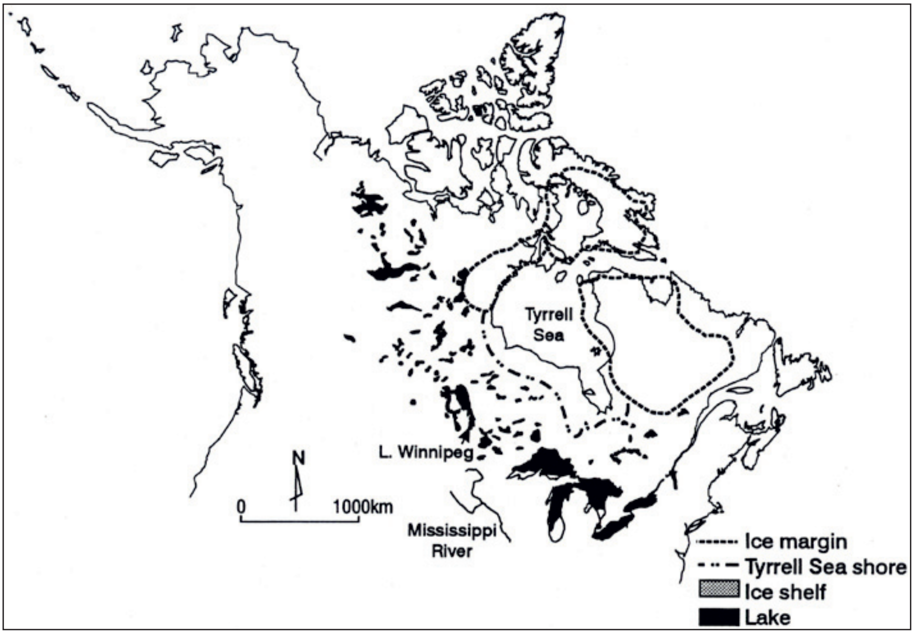

Fig. 7. The Laurentide Ice Sheet as reconstructed at approximately 8,000 years B.P. (after Vincent and Prest, 1987, and Dawson, 1992).

Deglaciation of southwestern Ontario: Great Lakes region

Ice sheets can be affected by topography to a varying degree. Toward their margin, as already noted by Chamberlain in 1883, they develop wide terminal lobes when the bedrock topography has gentle slopes and wide valleys or basins. The presence of terminal lakes fosters surging, hence lobe formation. However, lobate termini also develop on dry-land where the substratum is almost flat, such as in parts of central south Canada. These lobes reflect the tendency of glaciers to reach equilibrium, redistributing mass from the accumulation to the ablation zone, and responding in part to differential internal stresses and to basal friction. Southern Ontario is a cratonic area where a well developed geomorphological and sedimentological record of highly lobate terminations of a continental ice sheet is preserved.

The bedrock geology, preglacial fluvial erosion and the modification of the fluvial valleys by glaciers led to development of the southwestern Ontario peninsula delimited by the southern Great Lakes of North America (lakes Ontario, Erie, Huron, and Georgian Bay) (Figs. 1, 8). The peninsula is characterized by a morphotectonic high (Algonquin Arch) that has formed its SW-NE oriented backbone since Palaeozoic times (Martini and Bowly, 1991). The last ice sheet started retreating from the area about 14,000 years ago. Large lakes formed in front of the ice in the basins (modified ancient fluvial valleys) depressed by the weight of the ice. Only the highest central part of the peninsula became ice –and water– free and has consequently been called Ontario Island (Fig. 9); that is, an area that was surrounded by water, fluid to the south in lakes, and frozen to the north in glaciers. The glacier remained thick and experienced active retreats and surges in the basins, forming a series of lobes all around Ontario Island. The result has been a complex distribution of geomorphological features and sedimentary deposits which rim the island. Unravelling the deglaciation history of this area is difficult, particularly because there are insufficient numerical dates for the various sediments, and there is little chance that the new ones can be obtained from these organic-free deposits. However, a general statement can be made to the effect that the various lobes retreated from the highland toward the centre of the basins (the location of the present day lakes) in a pulsating fashion (retreats alternating with local advances), and various geomorphological units (such as drumlims, eskers, end moraines, outwash and lacustrine plains) developed in predicable successions (Martini et al., 1996).

The following examples may suffice to illustrate the difficulty in making detailed interpretations of remnant glacial features in a complex area. A first example may be that some moraines were originally defined as recessional moraine (Karrow, 1987; Martini et al., 1996), implying that they formed during staging periods of the glacier retreat. This may be true. However, more complex genetic systems have also been

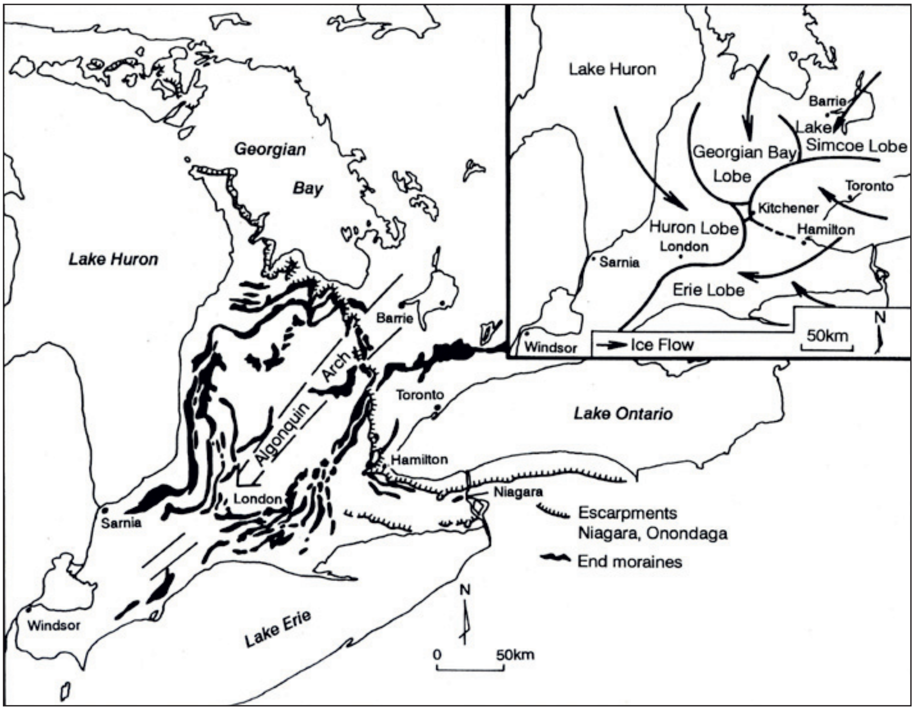

Fig. 8. Map of major end moraines of various ice lobes in southern Ontario (after Chapman and Putnam, 1984).

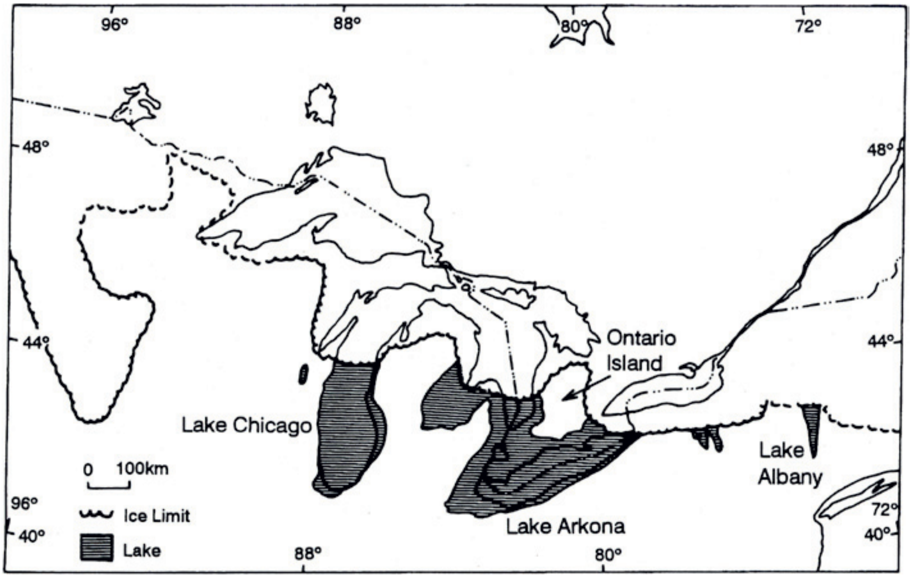

Fig. 9. Ontario Island and first Great Lakes (after Prest, 1970).

devised illustrating the possibility that glaciers may have at times retreated for up to 100 to 300 km into the lake basins and even further north of them, and then reavanced, depositing new tills and forming end moraines, including those previously interpreted as recessional features (Barnett, 1992). A second example may be the formation of the drumlin fields of the region. The conventional interpretation is that drumlins form behind moraines under active glaciers where ice is thick enough to move some of the basal sediments around obstructions (bedrock or cluster or non-dilatant deposits) (Smalley and Unwin, 1968; Embleton and King, 1975; Boulton, 1987; Menzies and Rose, 1987, 1989; Menzies, 1996). According to this hypothesis, the drumlin fields of southwestern Ontario were formed at different times as the terminus of the glacier retreated from one place to another. A recent hypothesis is that meltwater megafloods have occurred in this area, some under the glacier itself (Shaw et al., 1989; Shaw and Gilbert, 1990). These megafloods would have been responsible for the quasi-contemporaneous formation of all drumlims. Partial justifications for this hypothesis are the streamlined form of the drumlins which resemble similar, smaller water-formed features, and the presence of stratified drift inside some of the hills.

The glaciomarine and glaciolacustrine sedimentary record. Highlights

An effect of the last Pleistocene deglaciation has been the development of glaciomarine and glaciolacustrine deposits in large areas of Canada, many exposed after the draining of the water bodies (Fig. 10). These areas are of considerable economic importance, including farming, but also present hazards for real estate development and for infrastructure such as roads, bridges and railways because of recurring flooding and slumping. The glaciomarine and glaciolacustrine settings, hence their deposits, have, however, some fundamental differences.

1. The two systems have waters with different salinity. This affects the distribution of meltwater and entrained sediments in front of the glaciers (Powell, 1990). Fine particles are more likely to be kept in suspension in the sea and, at the same time, clays are less likely to be separated from silt and fine sand because they can flocculate and settle together.

2. The size of the basins are generally much different. Many seas are connected with oceans and subject to tides. Lakes are restricted basins and their extent and water level are strongly and rapidly affected, directly or indirectly, by glacier advances and retreats. During advances, the basins may be filled by ice. During retreats, basins and outlets depressed by the weight of the glacier may become ice-free and much of the water may drain from part of the lake. Subsequently, as the glacier retreats farther away, differential isostatic rebound may occur, raising the outlets and allowing re-inundation of the lake basins.

In both lacustrine and marine settings, glaciers are generally more active than inland. One reason is that parts of the ice float (ice shelves) or have wet bases, and

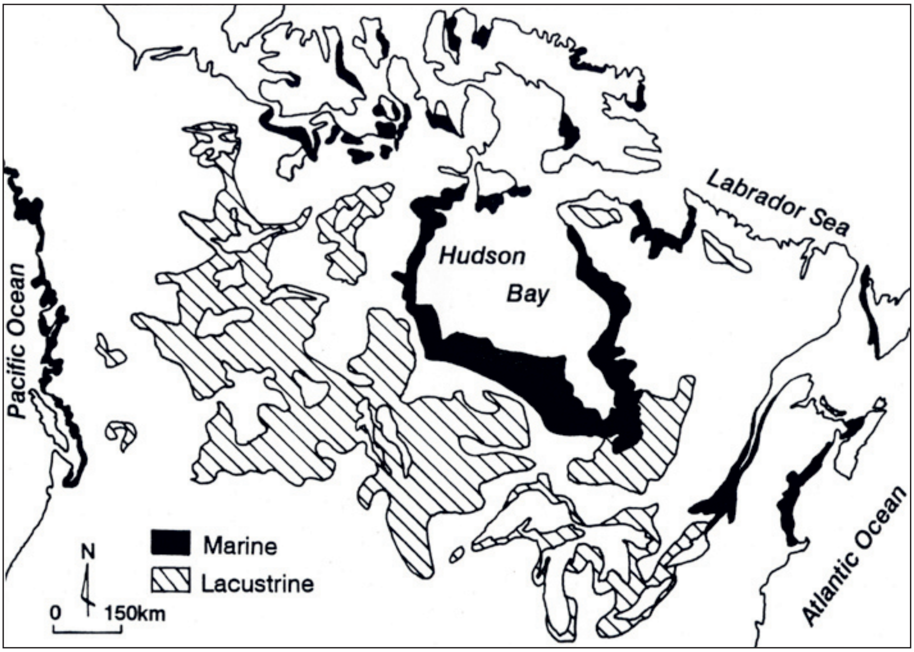

Fig. 10. Extent of glaciomarine and glaciolacustrine deposits in Canada (after Eyles and Menzies, 1983).

thus are subject to more frequent surges, ice loss through calving, and the formation of icebergs. The glacier termini, particularly those ending in deep water bodies, change rapidly in space and time as follows.

a. Advancing glaciers usually show steep terminal inland and ice cliffs in water; retreating glaciers show gentle sloping, rotting termini inland and ice ramps in water. At any one time, all these may be present at different locations because the advances and retreats or different parts of the ice margin are not synchronous.

b. Through time, in the same area, changes may occur from glaciolacustrine to glaciomarine termination when the land is still depressed by the nearby glacier and intercommunicating channels are opened between the lake and the sea. At times, the lacustrine-marine transition occurs rapidly, in a matter of a few weeks or months, as was probably the case in the southern part of the James Bay basin (southwestern elongated appendage of Hudson Bay) in northern Canada (Skinner, 1973).

Glaciomarine settings and deposits

As reported previously, the Laurentide Ice Sheet terminated for the most part along the eastern and northern seaboard during its maximum extent, and continued to have sea-termini in embayments in the south, far north and, later, for a brief period in Hudson Bay and adjacent Canadian inland seas.

1. At its maximum, the Laurentide Ice Sheet covered the Atlantic shelf of Canada and the Gulf of St. Lawrence. This led to disruption of the shelf ecozones, which were replaced by subglacial settings, glacial drift deposits, moraines and other glacial features (Grant, 1989).

2. An approximate 150m drop in sea level placed the shoreline at or below the shelf-edge, and, as a result, large quantities of coarse sediment were carried to the continental-shelf edge. These sediments fed lowstand fans at the shelf toe by way of various sediment-gravity flows including turbidity currents. A deep channel across the Gulf of St. Lawrence (Fig. 11; Grant, 1989) may have conveyed deposits to the edge of the continental shelf between Nova Scotia and Newfoundland providing material for several deep-sea turbidity flows into the Atlantic Ocean.

3. Large quantities of marine silt and clay were deposited in shallow, temporary seas which developed during deglaciation in those areas still depressed by the weight of the glaciers. One such deposit is the infamous Leda Clay of the Champlain Sea in the St. Lawrence Lowland (Fig. 4; Gadd, 1986; Karrow and Occhietti, 1989). After the area was uplifted by postglacial isostatic readjustment, these thick marine clays proved to be very sensitive if infiltrated by fresh water; that is, if bonds between particles are disrupted by washing off some of the binding ions, the clay loses strength and flows readily. This has led to numerous and at times disastrous slope failures whereby entire villages or parts of towns were disrupted. A similar marine deposit was formed along the southern and southwestern shores of Hudson Bay in the glacial Tyrrell Sea (Figs. 7, 10). However, the major impact of the Tyrrell Sea Clay in that remote, undeveloped region has been primarily to foster development of vast weatlands. The Hudson Bay Lowland, southest of Hudson Bay, underlain by this clay, contains, in fact, the second largest unconfined peatland (Radforth and Brawner, 1977; Martini, 1989) in the world after that of the Irsch and Ob' rivers area in Siberia.

4. Sand and gravel deposits were formed along the shores of these ancient seas, but similar materials were also deposited in deeper offshore settings in front of the glacier terminus, by meltwater floods derived directly from the glacier and flowing within and under the ice itself. These deposits have characteristics similar of those of fluvial and deltaic materials, but are invariably overlain by fossiliferous, marine sediments. They are called "subaqueous outwash" (Rust, 1997), and they are good indicators of past environment conditions during deglaciation, as well as good aggregate resources in areas otherwise covered by fine basinal materials. Subaqueous outwash is not confined to seas. It is also found in several large lakes such as in the precursors of Lake Ontario (Brookfield et al., 1983).

5. Glacial-seas are affected by numerous icebergs. Recent examples are those where icebergs are carried by the circular marine current around Antarctica and those

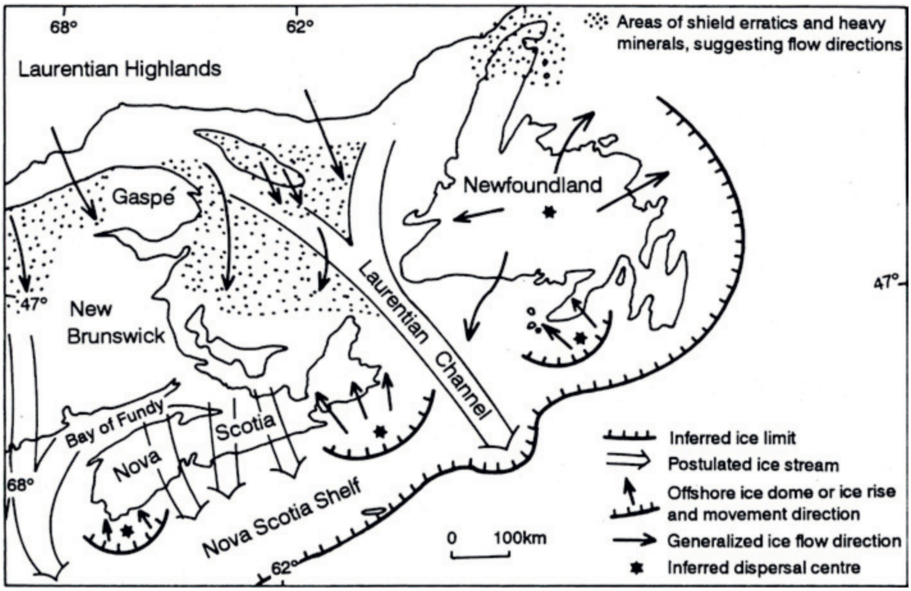

Fig. 11. Map of eastern Canada showing channel in the Gulf of St. Lawrence (after Grant, 1989).

carried southward along the coast of Labrador. These latter ones establish the so called iceberg alley along the northeastern seaboard of Canada in Labrador and Newfoundland. Icebergs can leave a legacy of scour marks, sometime at depths of several hundred meters, and rafted sediment accumulations. These accumulations are more intense in fjords where icebergs may be trapped, but they are also present, more scattered, in open sea. Indeed, ice rafted sediments have been used to recognize repeated frigid conditions and glacial advances in the northern Atlantic (Heinrich, 1988).

6. Rhythmic deposits of silt and clay are formed in front of tidewater glaciers (cyclopsams and cyclopels; Powell and Molina, 1989). They are formed by turbid meltwater currents.

7. Marine deposits may be also affected indirectly by glaciers. For example, some continental ice shelves, such as those of Nova Scotia and the northeastern seaboard of the United States of America were at the margin of the ice, suffered little subsidence, hence little isostatic post-glacial rebound. As a consequence, during deglaciation they were rapidly flooded by rising seawater. Northern coasts, instead, such as those of Hudson Bay, were affected by strong isostatic rebound and much emersion is still occurring forming a sequence of raised beaches extending hundreds of kilometres inland (Hillaire-Marcel and Fairbridge, 1978; Dredge and Cowan, 1989). The St. Lawrence River valley is an intermediate area. The land was inundated by the sea while still depressed by the nearby glacier, but later emerged due to isostatic uplift. However, the uplift was non strong enough to leave very wide successions of raised beach ridges.

Glaciolacustrine environments and deposits

Glaciolacustrine deposits can vary greatly. Only three cases are presented here.

1. Varves from one of the most characteristic glaciolacustrine deposit and a variety of them exists (Ashley, 1975). True varves generally form by separation of coarse silt and sand from fine clay during settling. The material is injected into the lake from rivers during spring and summer floods. The coarse material settles rapidly, but the fine, charged clay particles repel each other (that is, the clays do not flocculate in fresh water) and have a very low settling velocity; they remain in suspension, until waves and current motion are dampened by winter-ice cover over the lake. Thus, a layer-couplet or varve is genetated, where the coarser, usually lighter coloured material represents summer sedimentation and the darker, finer material, the winter layer (Fig. 12A). As such, they have been used to establish the time a cold lake has persisted in an area, counting the number of varves in deposit.

When analyzed in detail, though, many varves show several laminations of coarse and fine sediments within the coarse part of the couplet. Furthermore, some of the rhythmic deposits may show coarse sand and may have thicknesses on the order of metres. Thus, the coarser portion of the rhythmites is not formed by settling of material from suspended plumes, but from bottom-hugging, sediment-gravity flows such as turbidity currents. These turbidity currents can originate both from high sediment-concentration fluvial floods (turbid flows) entering the lakes, or from slumps of shallow-water, generally deltaic deposits. The fine interlaminations in part represent the tail end deposits of these turbid currents, and in part by fine clay deposited during the winter under the ice-cover. This winter layer can usually be recognized because of its extremely good sorting, fine grain size, and by the presence of tiny traces left by crawling organisms, which indicates a considerable lapse of time before deposition of the next coarse layer

Very thick sandy rhythmites are locally exposed in some glaciolacustrine deposits of the Great Lakes, such as along the north-shore bluffs of Lake Erie (Barnett, 1987) and Lake Ontario (Fig. 12B). The sandy rhythmites vary from tens of centimetres to several metres in thickness. They have sharp upper contacts with fine clay beds which are generally less than 5-10 cm. The sandy layers are composite, generally with abundant ripple cross-laminations. The ripple cross-laminations locally show a climbing tendency, at various, but generally shallow climb angle. The sand is relatively well sorted, but may also contain disseminated clay chips, up to clay balls 10 cm in diameter. These clay balls indicate that the transporting density-flow had sufficient energy to carry coarse material, but such material was not available. Some rhythmites composed of apparently massive sands, have a fining upward tendency shown only by clay chips that are more common toward the base of the layers and disappear upwards

(Fig. 12C). These thick rhythmites are taken to indicate the re-occurrence of large floods, perhaps even catastrophic ones discharging from nearby glaciers.

2. Proglacial lakes have, by definition, glacier ice bounding them at least along one shore. Thus, the lake may receive water directly from the glaciers as well as from overland rivers which themselves can be in part fed by meltwater from a glacier with terrestrial termination. This leads to spatial and temporal, highly variable sedimentation, but, in general, several vertically stacked successions occur with a common motif with a diamict or true till at the base, changing upward into distal clay-rich rhythmites, into proximal sand-rich rhythmites, and into well washed, deltaic and/or coastal sands (Fig. 12D).

When analyzed in detail, the glaciolacustrine sediments are seen to be composed of two main assemblages: one related to overland glaciofluvial derived materials (lacustrofluvial assemblage), and in some, generally deep parts of the lake, an interlayering of units of the two assemblages (Fig. 13; Ashley, 1975; Ashley et al., 1985; Martini and Brookfield, 1995). The lacustrofluvial assemblage reflects pretty much the generalized succession mentioned above. The lacustroglacial assemblage has more lensing units. For example, (a) at times, the lower diamict is folded and faulted reflecting ice push features of surging, grounding glaciers; (b) the fine grained rhythmites contain whitish silty clasts formed by material abraded under the glaciers, compressed and deposited in front of the ice; (c) in parts of some successions, gravelly units are found interbedded between the basal diamict and the distal rhythmites or within other lacustrine layers, and they are interpreted as subaqueous outwash-fan deposits. Deltaic and coastal deposits are not present in lacustroglacial successions, although they may develop as caps during the lacustrofluvial phase, when the front of the glacier retreats from the area.

3. Indirect effects of the glacier on lacustrine deposits are due to rapid changes in lake level due to differential isostatic uplift and opening and closing of outlets. The geomorphological and sedimentological record of resulting transgressive and regressive events is well recorded in the Great Lakes region of North America. This includes well developed coastal lagoons protected from rising transgressive waters by gravel and sand barriers, usually pinned at a locality by bedrock or glacial landforms such as end moraines, everything partially capped by dunes which may have developed during a subsequent slight regression when beach-sand became available to the blown inland (Fig. 14; Martini, 1975, 1981). Another type of development is the succession of features and sediments that develop during the progressive drying out of parts of the proglacial lakes due to postglacial isostatic rebound. In one instance at least, in southwestern Ontario, a series of nested deltas developed whereby older deltas were cut by rivers, and a series of nested deltas were formed at lower lake levels (Fig. 15; Chapman and Putnam, 1984)

Fig. 12.
Pictures of several glaciolacustrine features. A. Fine grained varves. B. Thick sandy rhythmites exposed along the Lake Ontario Bluffs, probably formed by a single floods, showing, from the bottom up, a progressively increasing flow velocity from climbing ripple cross-laminations (sr) to plane beds (pl) and a subsequent decrease in velocity from pl to sr; C. Veritcal grading in a turbidite layers shown by distribution of clay chips; D. Vertical glaciolacustrine succession of sediments showing fine rhythmite deposits (f) at the base, grading upward into deltaic sands (s), sealed, at the top, by a till layer (t).

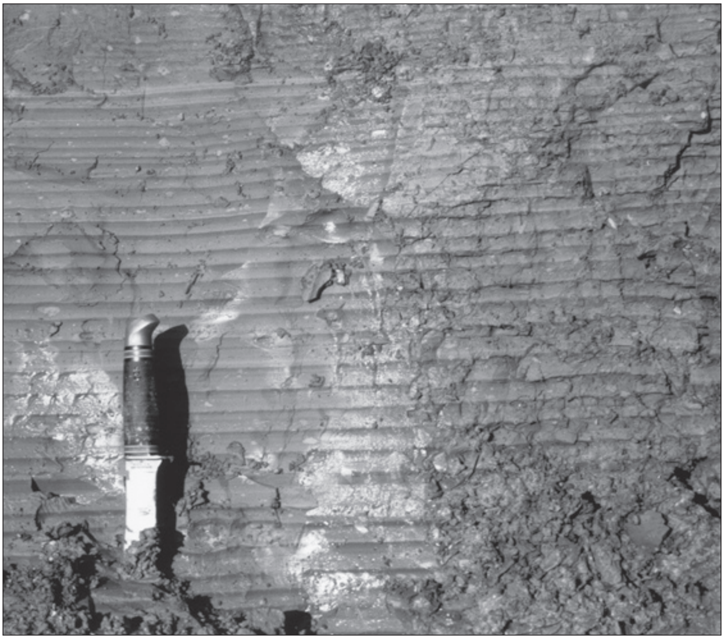

Fig. 12 A.

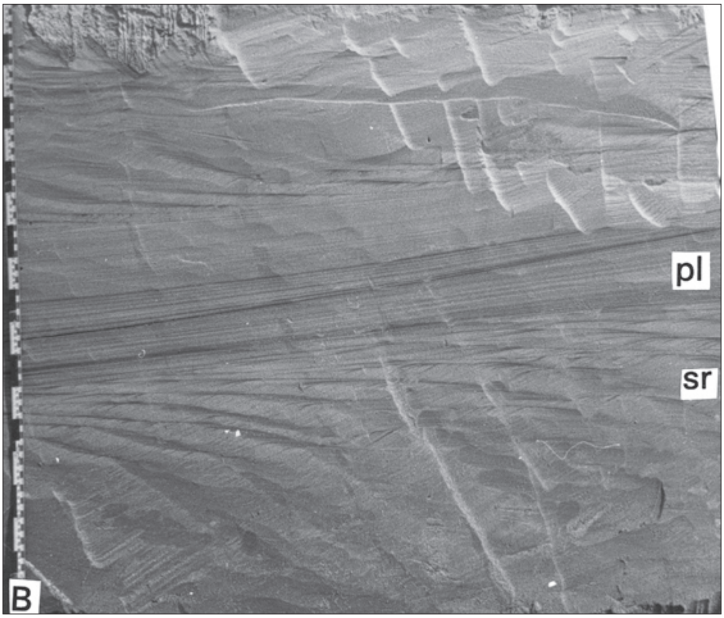

Fig. 12 B.

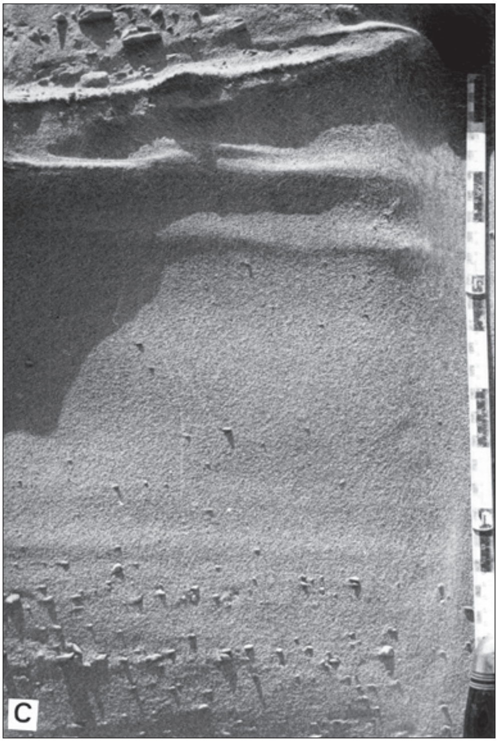

Fig. 12 C.

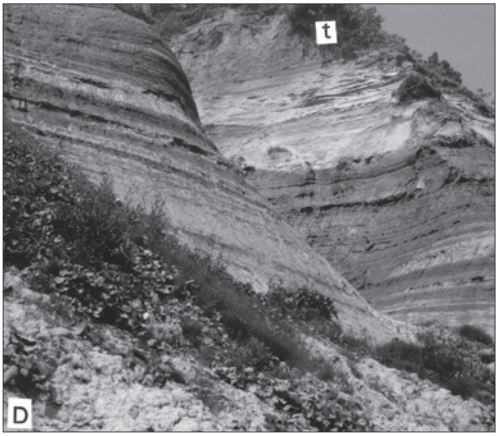

Fig. 12 D.

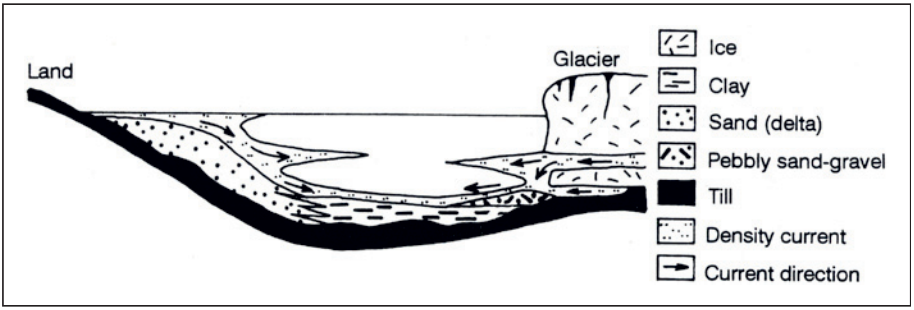

Fig. 13. Schematic diagram of a glacial lake depositional systems. The lacustrofluvial assemblage develops from the land side, the lacustrological assemblage from the glacier side (after Martini and Brookfield, 1995).

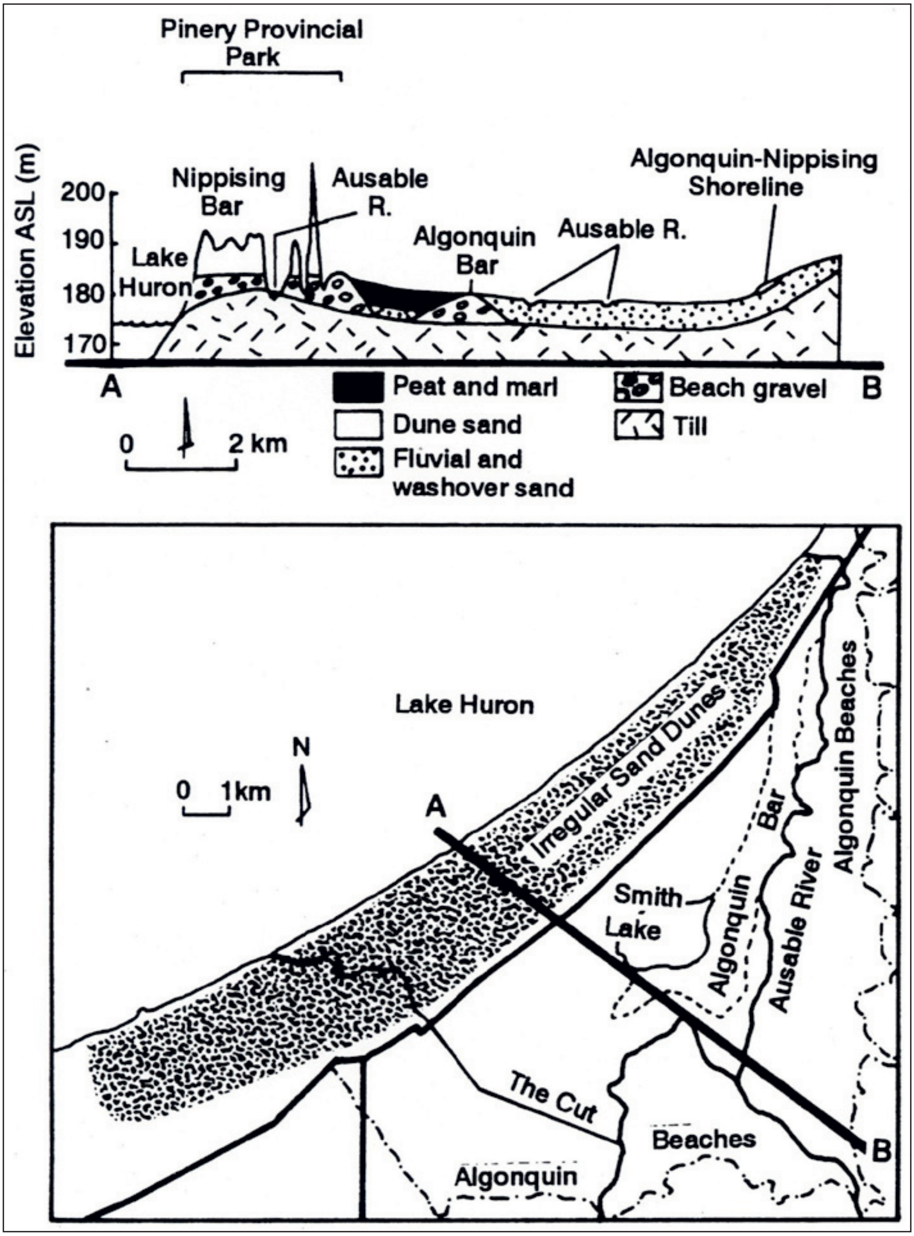

Fig. 14. Example of coastal barrier, dunes an lagoonal (with peat and marl) complex in the Great Lakes (after Ontario Ministry of Natural Resources, 1977).

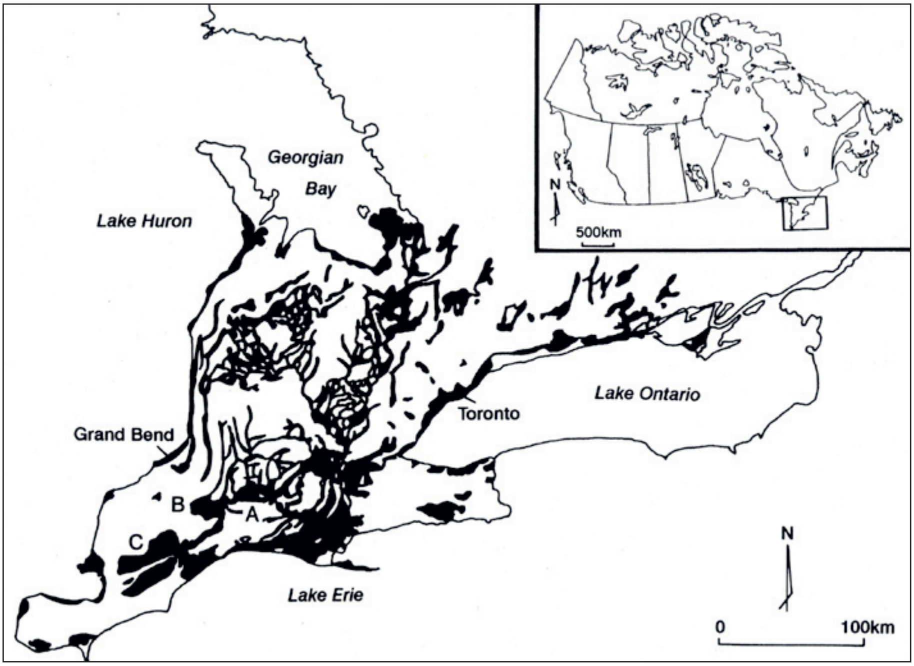

Fig. 15. Map of the outwash systems of southwestern Ontario with indicated (A, B, C) the nested deltas formed during lake retreat due to isostatic rebound (after Chapman and Putnam, 1984).

Conclusions

1. Deglaciation of an ice sheet is a relatively fast (order of a thousand years), complex event. The glacier terminus retreats in spurts, alternating with local surges. As the glacier melts and thins, its margin is increasingly affected by local topography. The glacier may float and advance into basins occupied by meltwater, whereas it retreats on the intervening highlands. Whenever topographic variation exists, even if small, the terminus of the glacier develops lobes.

2. As it retreats, the glacier leaves behind a series of landforms. Some, such as moraines and rock scarps, constrain the meltwater flow, and glacial lakes may develop. Because of the isostatic depression of the land due to the weight of the ice, the lakes usually abut the glacier itself. As the glacier retreats, the lakes change in depth and dimension due to the interplay between (a) opening of lower elevation outlets closer to the glaciers, hence draining of some lake water, and (b) differential isostatic rebound with (i) some distal shores (opposite or away from the glacier) raising faster at first,

hence forcing retreat of the waterline, and, (ii) later, when the glacier disappears, a faster and greater rebound of the northern shores closer to the centres of glaciation. This forces a tiling of the lake waters southward, with transgressions occurring again onto the southern shores. The legacy of these regressions and transgressions is left along some lake shores in the form of barrier systems with well developed lagoons and often capped by large aeolian sand dunes. The change in glacial lakes through time is well illustrated along the southern border of the Laurentide Ice Sheet, by the precursors of the Great Lakes and of some western lakes like Lake Winnipeg in central Canada.

3. Parallel to this change in setting, is the development of the lacustrine sediments. On the whole, the sedimentary sequence is characterized by a basal erosional surface locally covered by basal till, in turn overlain by variable sedimentary successions. At the distal margin of the lake (opposite or away from the glacier) there may be coarsening (shallowing) upward successions (lacustrofluvial assemblage) associated with prograding deltas or shore deposits. Toward the basin centre, there is a sudden upward change from the lodgement (basal) till, to meltout till, to deep water rhythmites. In places, closer to the glacier terminus where intraglacial streams discharge sediments, there may be coarse sand or even gravel deposited over the basal till. This sand and gravel is formed in subaqueous outwash (part of a lacustroglacial assemblage). These coarse, deep water deposits are usually overlain by deep water rhythmites. The fine glaciomarine deposits show frequent synsedimentary slumping. Ice push features due to local glacial readvance are present as well. Ice rafting deposits occur but usually they are small volumetric.

4. The glaciomarine deposits have different structures from the glaciolacustrine sediments. Notorious is the fact that glaciomarine silty clays have weak structure, easily compromised when infiltrated by freshwater, and, thus, prone to slumping when emersed. This is the case of the sensitive glaciomarine silty clay deposits of southeastern Canada (St. Lawrence Lowland) and central-east Canada (Hudson Bay Lowland).

5. Meltwater floods are a common feature during deglaciation in front and under ice sheets. The floods may discharge directly into water bodies generating subaqueous outwash, or in front of the glacier at its land termination generating erosional features or subaerial outwash. It is debated whether megafloods have occurred with discharges of the order or a hundred-times that of regular floods, and wheter these floods have had a significant effect on the moulding of the landscape. Such floods could have been generated by rapid discharge from subglacial lakes when the icy dams failed. Large (order of hundred meters), long (order of tens kilometres) tunnel valleys under the glacier may have been formed almost instantaneously during these floods. Similarly, some authors contend, streamlined erosional features on bedrock and even drumlins could have been formed over wide areas, almost instantaneously by these floods.

Byproduct of the megafloods would have been large amount of sediment carried into depositional basins where thick sedimentary layers may have formed. Some evidence of very large floods exists in exposed bluff sections of the Great Lakes, where sandy rhythmites several metres thick occur. However, the importance of megafloods on the development of a deglaciated landscape still needs further assessment.

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