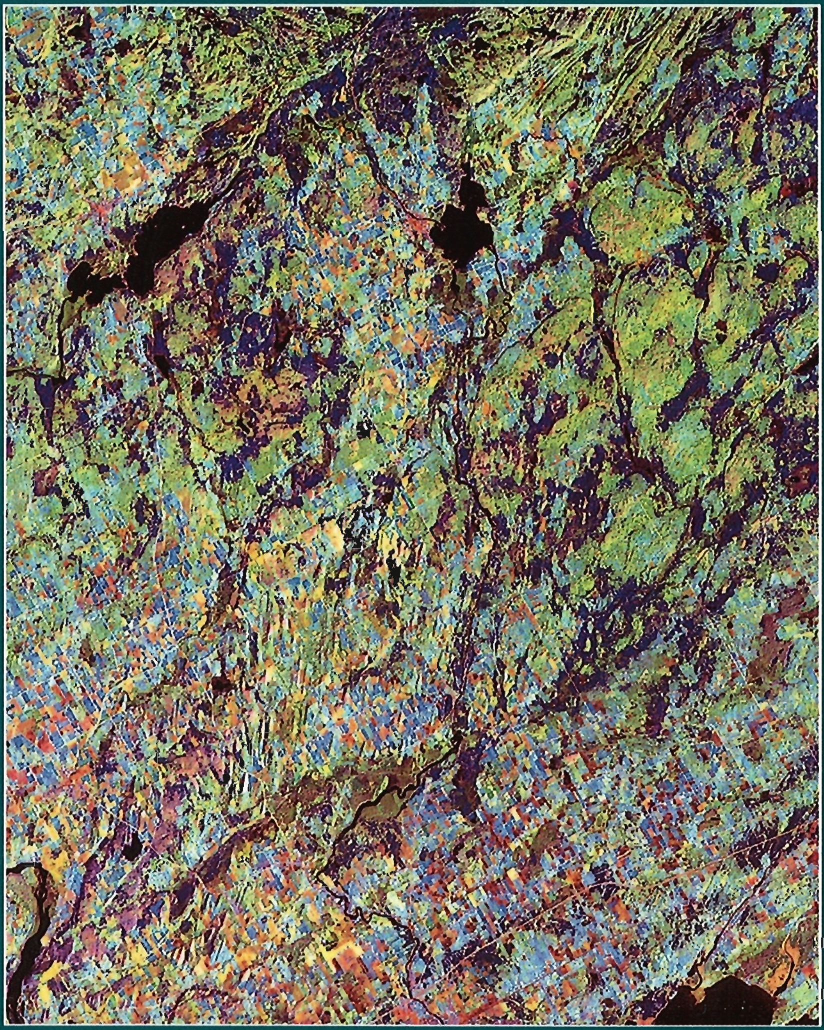
GEOLOGICAL SURVEY OF CANADA
BULLETIN 453

A FIELD GUIDE TO THE GLACIAL AND POSTGLACIAL LANDSCAPE OF SOUTHEASTERN ONTARIO AND PART OF QUEBEC

Robert Gilbert, compiler



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Cover description

Portion (approximately 42 km x 33 km) of satellite image of area north of Bay of Quinte. The thickness of sediment varies across southeastern Ontario due to erosion and deposition of glacial drift. Deformed Precambrian Shield rocks carry little drift cover in the north, whereas Paleozoic carbonate sediment thickness increases to the southwest, south of the Shield margin. Drumlins are cut into this thicker drift (lower left). Tunnel channels (purple) and eskers traverse the region. (Data provided by RADARSAT International Inc. and distributed under authority provided by the Canada Centre for Remote Sensing; reference: 80 km x 80 km, TM 16-29, bands 3, 4, 6, May 1988.)

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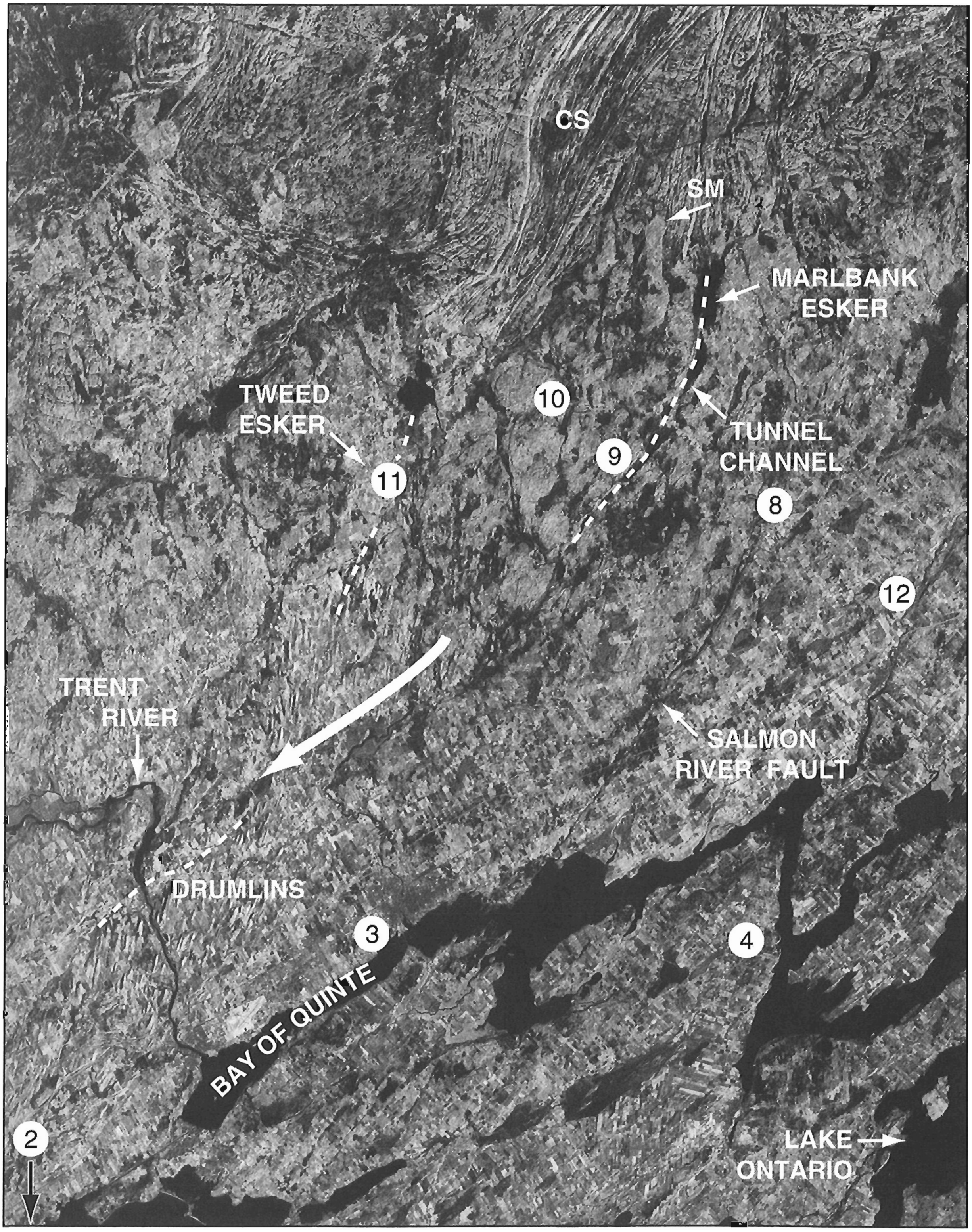


Frontispiece. *Satellite image (see cover) showing the location of stops 2-4 and 8-12 on the field trip and various features relating to the landscape north of the Bay of Quinte to the Canadian Shield (CS). South of the deformed rocks at the shield margin (SM), lie tunnel channels and eskers (Marlbank and Tweed) (arrows) and drumlins formed on the thicker sediment cover found on the underlying Paleozoic carbonates. The Trent River cuts through the thickest sediment cover in the drumlin fields. To the east thinner drift cover occurs due to erosion by glaciofluvial processes, and because of thinner sediment cover, the Salmon River Fault is visible as a striking northeast-southwest lineament.*

A FIELD GUIDE TO THE GLACIAL AND POSTGLACIAL LANDSCAPE OF SOUTHEASTERN ONTARIO AND PART OF QUEBEC

Abstract

Twenty sites between Peterborough, Lake Ontario, and the Ottawa Valley are described in this guide. Set in context by a general introduction to the geomorphic history of the region, they illustrate landscapes created by processes of glacial erosion and deposition during the Pleistocene, and by events that have occurred in the period of about 12 000 years since deglaciation. Sites especially near Lake Ontario also deal with the impact of human activity on the environment. The descriptions are intended for a nonspecialist audience, and directions are provided for road travel to each site.

Résumé

Le présent guide décrit vingt sites entre Peterborough, le lac Ontario et la vallée de l'Outaouais. Situés dans leur contexte grâce à une introduction générale à l'histoire géomorphologique de la région, ces sites constituent des exemples des paysages créés par l'érosion glaciaire et la sédimentation au Pléistocène et par des événements survenus au cours des quelque 12 000 ans depuis la déglaciation. Les répercussions des activités humaines sur l'environnement sont également mentionnées pour certains sites, particulièrement ceux près du lac Ontario. Les descriptions s'adressent à des non-spécialistes; les routes à prendre pour se rendre à chacun des sites sont indiquées.

INTRODUCTION

This field guide builds upon existing work, focusing on recent studies by the authors. It is presented in two parts: the first reviews briefly the events that shaped the landscape of southeastern Ontario (Fig. 1) and the second describes case studies from the scale of individual landforms to larger elements of landscape that illustrate the geomorphology of the region. Our perspective encompasses environments created by continental glaciers through to modern settings where human action plays the central role. Although we have directed our writing to an audience with some understanding of earth science, we have attempted to explain basic principles and concepts that will allow the nonspecialist to appreciate the relation between process and form. References to the professional literature are made throughout, so that the reader may investigate further. Other field guides to the region, including those by Jenness (1967), Fulton (1987), and McKenzie (1990), also provide insight to the landscape.

The case studies are presented as "stops" organized by geographical location rather than by topic. Because the reader may not wish to visit all stops, directions to those in the south are given from Highway 401. Directions to stops in the Ottawa region are given with respect to highways in that area. The stops and major routes are shown on Figure 2.

At the time of writing of this guide, access to stops on private property is unrestricted or is normally granted to visitors by the owner or tenant on request. In the latter cases we have provided the names of owners. Hard hats and safety boots are recommended, especially for stops located in gravel pits. Visitors are asked to respect property and avoid interfering with ongoing activity. If in doubt, please consult the owner or tenant before entering.

Parts of the work presented in this guide were carried out with research grants and scholarships to the authors from the Natural Sciences and Engineering Research Council of Canada. The Advisory Research Council of Queen's University provided a grant to assist with preparation of the guide. We appreciate the careful review of the contents by P.J. Henderson of the Geological Survey of Canada.

OVERVIEW

Bedrock geology

The Frontenac Axis of the Canadian Shield extends southeastward from the Algonquin Dome in central Ontario to the Adirondack Mountains of New York State, crossing the St. Lawrence River just east of Lake Ontario (Fig. 1). Its ancient rocks are part of the Precambrian Shield that forms the stable central core of the North American continent. They were created relatively late in the evolution of the Shield during a period of mountain building less than one billion years ago. In the southern part of the Axis and along the St. Lawrence River, intrusive igneous and metamorphic rocks including granites, diorites, and gneisses predominate. Elsewhere, clastic and carbonate metasedimentary rocks include marbles, sandstones, and shales.

Overlying the strongly deformed and metamorphosed rocks of Canadian Shield are undeformed sedimentary rocks (Fig. 1) deposited after a hiatus of almost 500 000 000 years. A discontinuous weathered zone (paleoregolith) marking this hiatus has been exposed by subsequent erosion, for example, along the Perth Road north of Loughborough Lake and along Highway 15 at the intersection of the road to Chaffey's Locks near Elgin. These younger Paleozoic rocks include Cambrian Potsdam sandstone which directly overlies the Shield and is exposed in narrow zones especially along its eastern boundary, and Ordovician limestones and dolostones with lesser amounts of shale and sandstone. These latter form the extensive, nearly flat limestone plains of eastern Ontario. Freeman (1979) gives an overview of the geology and Sabina (1968) and Hewitt (1969) describe specific sites of interest.

Quaternary glaciation

The erosion of bedrock and the transport of sediment by the ice sheets resulted in deposition of thick sequences of drift across much of southern and southwestern Ontario. In the area west of Trenton, deposits up to tens of metres thick lie on the erodible Paleozoic carbonate rock, for example south of Rice Lake. In other areas, glacial sediment is much thinner, rarely exceeding 1-2 m thick except for isolated pockets on the Canadian Shield and the limestone plain of southeastern Ontario. This difference in drift thickness can be appreciated by comparing the settings of especially stops 19 and 20 with those of stops 4, 6, and 8 (Fig. 2).

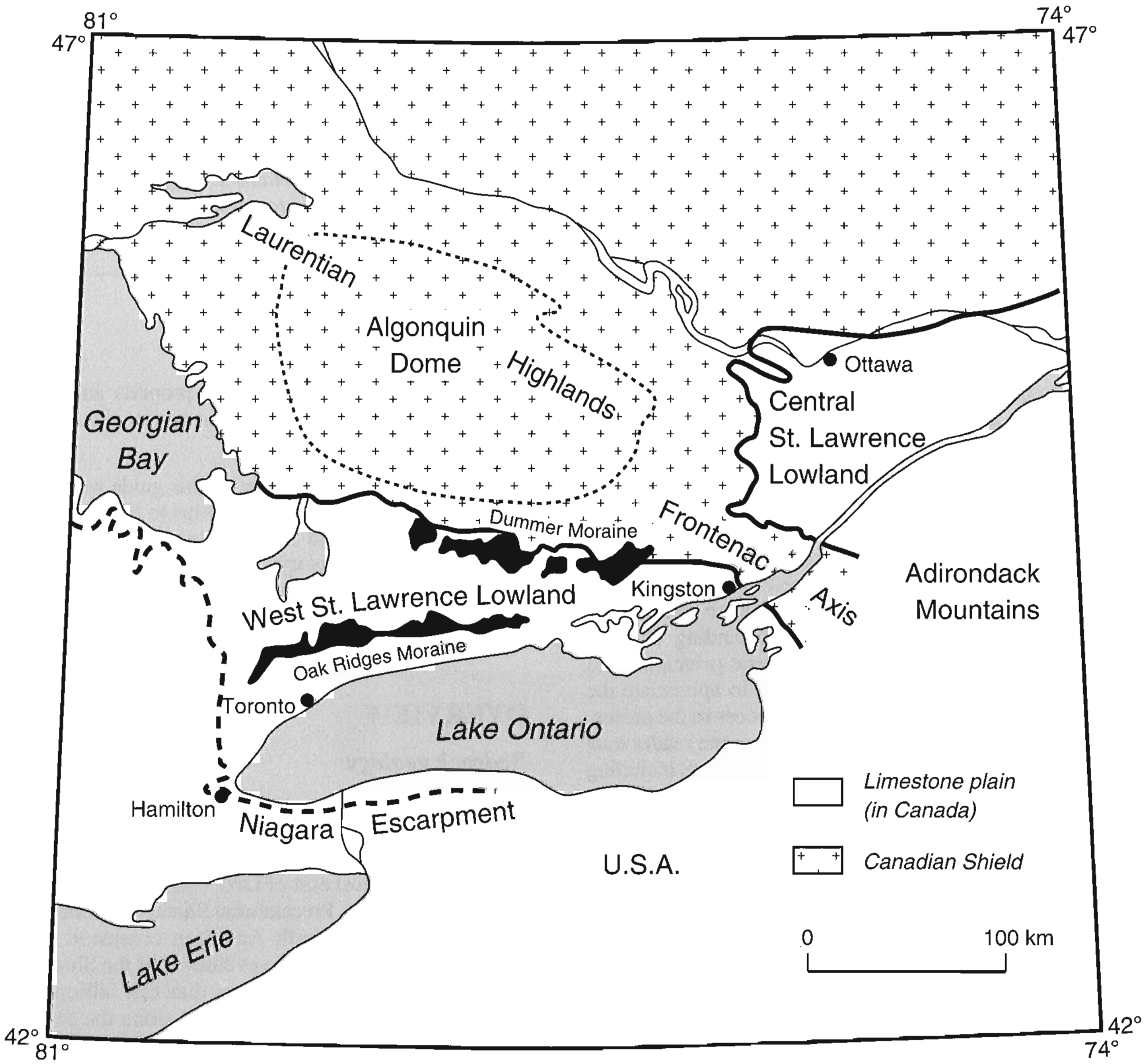


Figure 1. Generalized physiography and bedrock geology of southern Ontario.

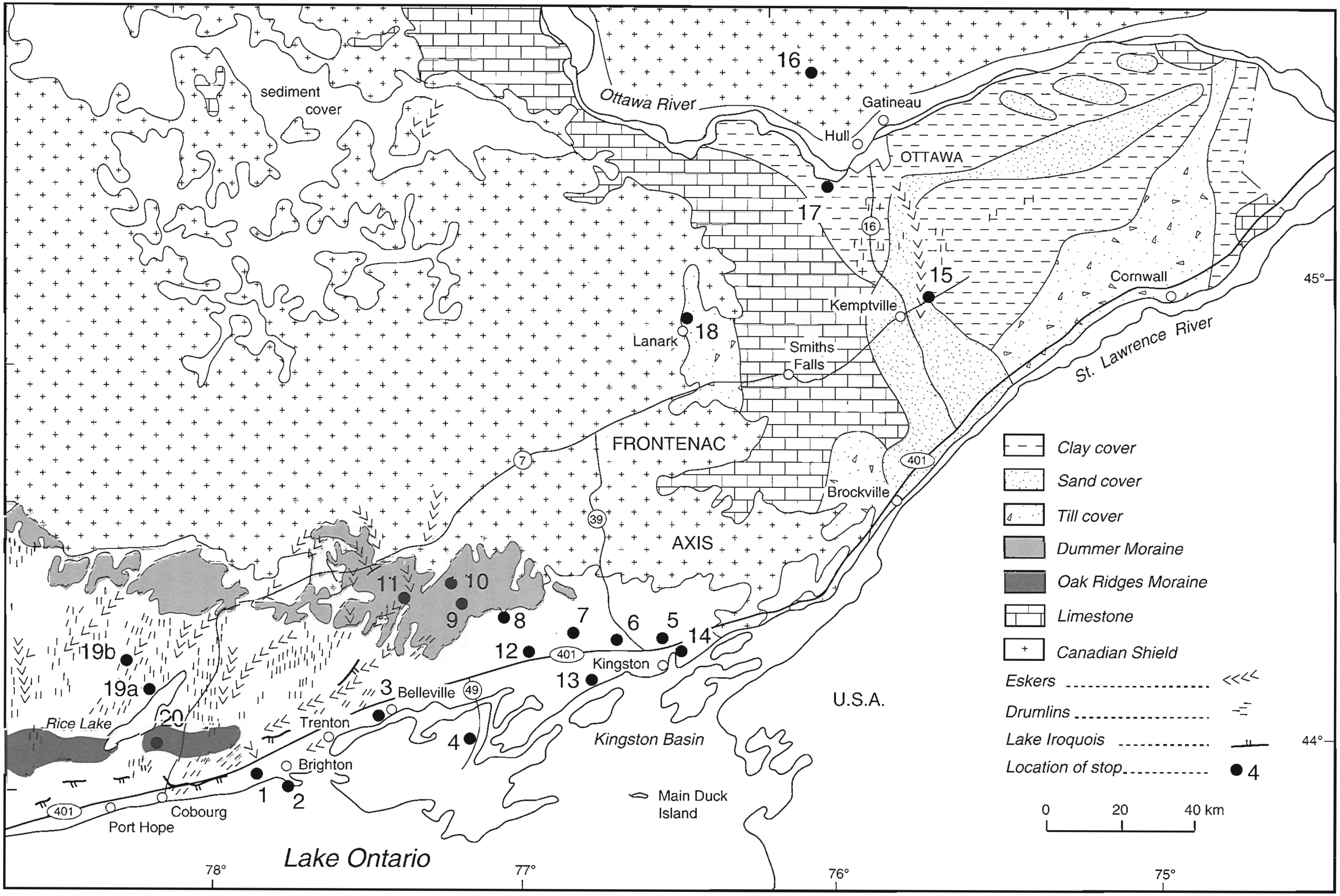


Figure 2. *Glacial landforms of southeastern Ontario (after Chapman and Putnam, 1972a, b; Barnett et al., 1991) showing locations of stops described in this guide.*

Therefore, unlike southern and southwestern Ontario where the thick sedimentary deposits provide important information on glacial and interglacial environments (Karrow, 1967; Eyles, 1987), there is little sedimentary evidence in eastern Ontario (except in the lower Ottawa and St. Lawrence valleys, Anderson et al., 1990) of any but the late stages of the last glacial episode. The Wisconsinan Glaciation dominated southern Ontario from about 75 000 years before the present (BP) until the glaciers finally retreated from eastern Ontario about 12 000 BP. Its influences continue to the present.

During much of the Wisconsinan, the margins of the Laurentide Ice Sheet that covered northern North America probably fluctuated across western Quebec and eastern Ontario, although there is considerable debate about their extent in southern and southwestern Ontario (Karrow and Ochietti, 1989). About 25 000 BP the ice sheet began a period of growth which culminated about 18 000 BP when the terminus reached its maximum southward extent near New York City, and through northern Pennsylvania and central Ohio (Dyke and Prest, 1987). At the maximum, ice thickness may have exceeded 2 km over eastern Ontario.

On the Canadian Shield, glaciers accentuated structural features in the bedrock by selective erosion of softer materials. Myriad lakes occupy the depressions that were created in this way, especially in the most structurally complex and softer clastic and carbonate metasedimentary rocks in the southeastern portions of the Shield. On the Palaeozoic rocks, erosion sculpted small escarpments and accentuated previously existing features such as river valleys.

Nevertheless, the total amount of glacial erosion in all of the glaciations probably did not exceed several tens of metres (Gravenor, 1975; Sugden, 1976). This conclusion is based on the amount of sediment in terminal moraines and in marine basins, and on inferences about the preservation of preglacial landforms, especially widespread erosion surfaces.

It is most often reported that erosion by ice sheets is the result of abrasion by the load of sediment carried in the sole of the glacier, and of plucking associated with freezing in cracks in bedrock as the glacier slides over the surface (Boulton, 1974; Sugden and John, 1976). Even complex grooves and channels scoured in bedrock have been ascribed to direct glacial erosion (Goldthwait, 1979). However, a growing body of evidence indicates that many of the landforms created beneath the Laurentide ice sheet were formed by streams or sheets of water flowing at high velocity for relatively short periods of time. These landforms (seen at stops 6, 7, 12, and 13; Fig. 2) vary from scales of millimetres to metres (Shaw, 1988), through tens of metres (Sharpe and Shaw, 1989), to kilometres (Gilbert, 1990; Gilbert and Shaw, 1992). They occur both as isolated features and as assemblages of landforms covering large regions (Kor et al., 1991; Shaw and Gilbert, 1990). In fact, the small subglacial fluvially-eroded forms were probably much more widespread, and subaerial erosion of exposed bedrock surfaces since deglaciation has obliterated or muted many of them.

The water that created these forms most likely originated in reservoirs within the central regions of the ice sheet (Shoemaker, 1992). Periodic failure of the ice damming these water bodies resulted in large outbursts that flowed beneath the glacier following pre-existing depressions, in some cases lifting large segments of the glacier off its bed, and flooding and eroding much of the land surface. These flows may have been short-lived, but the velocities were high (to tens of metres per second) due to the pressure of the overlying ice, and the volumes of water involved were sufficient to effect responses throughout the world's oceans (Shaw, 1989).

The timing of all of these events cannot be determined except to suggest that they occurred in all the periods of glaciation throughout the Pleistocene. Based on the orientation of subglacial fluvial features, Shaw and Gilbert (1990) concluded that in southern Ontario, earlier floods from the north (which they called the Algonquin event) spread across lakes Ontario and Erie and through the northern New York State. These were followed by floods from the northeast along the length of the basins of lakes Ontario and Erie (the Ontarian event).

The final phases of glaciation

The Wisconsinan ice sheet began to shrink from its maximum southward extent about 18 000 BP. The subglacial, fluvial processes that were important in creating landforms in the bedrock of southern Ontario continued during this period as well, shaping or reshaping landforms in the glacial sediments. The large drumlin field in the Peterborough and Trenton areas and the smaller field near Ottawa, as well as the extensive esker and tunnel channel networks (Shaw and Gorrell, 1991) were most likely formed or modified in the very early stages of general retreat (by the Algonquin and Ontario events). These features are the subjects of stops 1, 8, 10, 11, 13, 18, 19, and 20 (Fig. 2).

Landforms such as the Oak Ridges Moraine (Stop 20; Duckworth, 1979; Fig. 2) were formed after the drumlins in a period when meltwater was carried in conduits and formed eskers. The exact origin of the Oak Ridges Moraine is still not well understood.

During the 1000 years beginning about 12 700 BP, the landscape of eastern Ontario changed dramatically from ice covered to ice free (Prest, 1976; Fig. 3a, b). First, small lakes formed around the northern and western edges of the retreating ice in the Ontario basin (including one in the Peterborough area). Within 400 years the ice had withdrawn completely from the Ontario basin and the retreating front formed a dam that created Lake Iroquois (Fig. 3b), the first of a series of precursors of modern Lake Ontario.

In order to understand the size and level of this and the other lakes in the Great Lakes basins and the drainage networks that connected them, it is necessary to appreciate the effect of the isostatic depression of the Earth's crust due to the weight of the ice sheets. The maximum depression varied depending on the thickness and duration of the ice cover, and the characteristics of the crust and mantle (Andrews and Peltier, 1989). In eastern Ontario the maximum depression resulting from the last phases of the Wisconsinan glaciation was about 175 m in the St. Lawrence valley near Lake Ontario, and about 180-230 m in the Ottawa and middle

St. Lawrence valleys (Andrews, 1970; Fulton and Richard, 1987). Rebound of the land surface occurred most rapidly immediately following deglaciation and has continued at a slower rate to the present. In eastern Ontario the rates averaged about 20 mm/a at 8000 BP and 10 mm/a at 6000 BP, compared with about 3 mm/a at present (Andrews, 1970).

In terms of the evolution of the Great Lakes, the most important aspect of this depression and rebound is that it was greatest in the northeast and least in the southwest. This has resulted in an upward tipping of the land surface toward the northeast throughout the Holocene (Fig. 4). For example, the eastern outlet of Lake Ontario to the St. Lawrence River is still rising with respect to the inlet at the western end at a rate of about 1.8 mm/a (Clark and Persoage, 1970).

As a result of this tilting of the Earth's surface, Lake Iroquois stood less than 40 m above the present level of Lake Ontario at the western end, but more than 140 m above the present level at the ice dam near Kingston in the east (Anderson and Lewis, 1985). A series of clearly defined shorelines with their sedimentary deposits marks the northern shore of the lake (Fig. 3b) as far east as the inferred position of the ice dam. Waters flowed from the lake through an outlet at Rome,

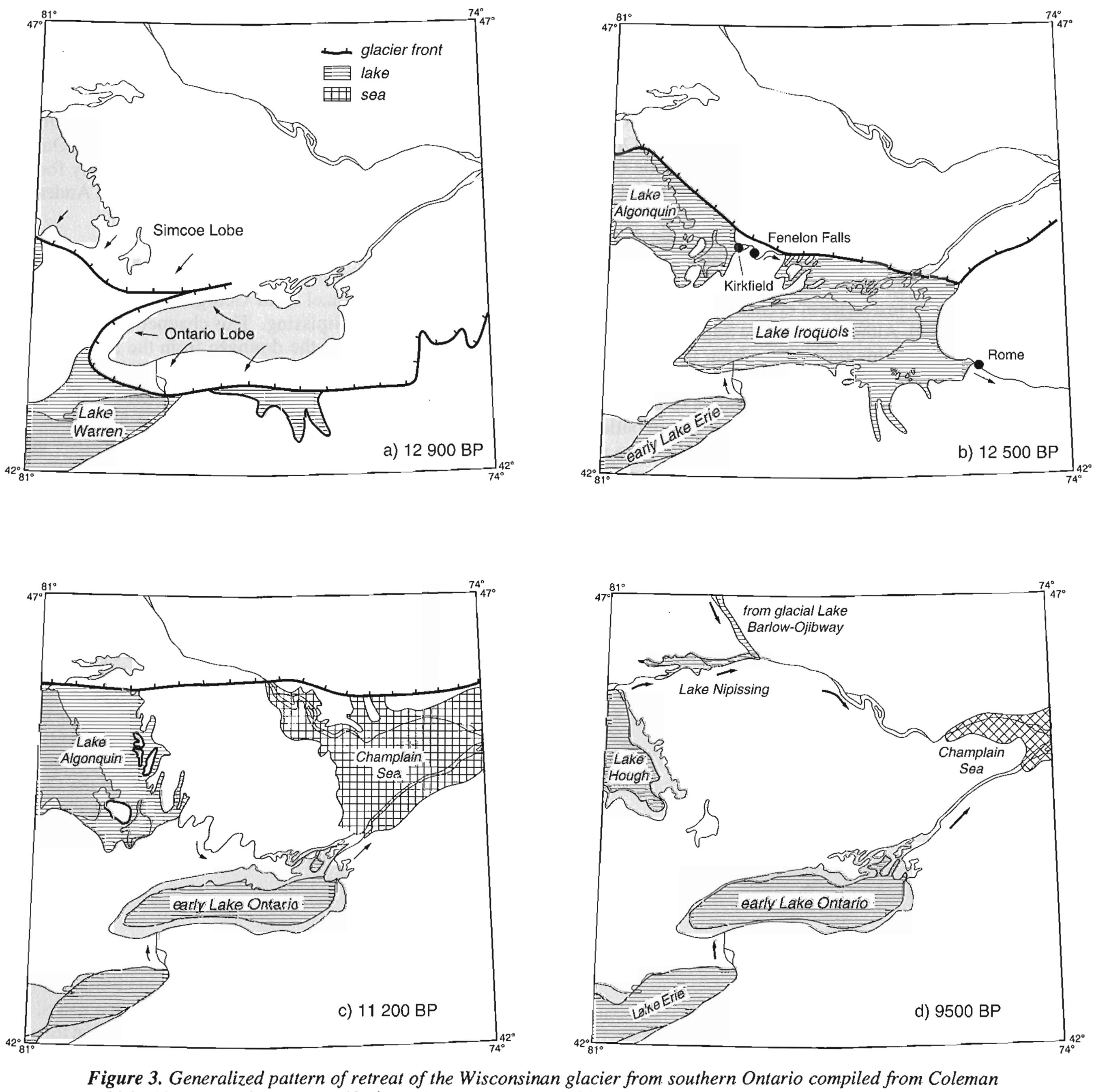


Figure 3. Generalized pattern of retreat of the Wisconsinan glacier from southern Ontario compiled from Coleman (1936), Chapman (1975), Prest (1976), Anderson and Lewis (1985), Finamore (1985), Muller and Prest (1985), Dyke and Prest (1987), Lewis and Anderson (1989), and Gilbert and Shaw (1992).

New York and thence to the Hudson River. Except when local ice advances closed the channel, at least part of the drainage from the newly forming upper Great Lakes entered the northern embayment of Lake Iroquois through outlets at Fenelon Falls and Kirkfield (Fig. 3b).

From the time of the maximum extent of Lake Iroquois about 12 400 BP, the damming glacier withdrew to the northeast and the water level fell in stages marked by beaches mapped in the Trenton area (Muller and Prest, 1985). A glimpse of the physical environment of the lake is seen in the processes described at Stop 4, and of conditions near the front of the retreating ice sheet at stops 5 and 20 as well as in accounts by Kaszycki (1987) and Henderson (1988). When the dam disappeared completely, the level of Lake Ontario fell below the present level (Fig. 3c). Interpretation of this environment and of the subsequent history of water levels in eastern Lake Ontario is the subject of stops 13 and 14. With the opening of the St. Lawrence valley about 11 800 BP (earlier according to Gadd, 1980), not only was Lake Ontario free to drain that way, but the waters of the Atlantic Ocean spread into the isostatically depressed Ottawa and Upper St. Lawrence valleys (Fig. 3c). This embayment is referred to as the Champlain Sea. Its retreat as the land rebounded, and the drainage into it from the upper Great Lakes and lakes to the north, dominated the environment of eastern Ontario for more than 3000 years in a series of events described by Lewis and Anderson (1989). Although the Lake Superior basin was still glacier covered at this time, water from the Huron and Michigan basins continued to flow into early Lake Ontario through the Kirkfield outlet (Fig. 3c). Isostatic tilting toward the northeast was sufficiently great that the southern outlet of Lake Huron to Lake Erie was dry. The Champlain Sea may have reached the Ontario basin for a brief period, but outflowing fresh water probably prevented marine conditions from becoming established (Sharpe, 1979; Pair et al., 1988). As the ice continued to retreat to the north in the period following 11 000 BP, the Champlain Sea extended up the Ottawa valley to its maximum northwesterly extent between Pembroke and Mattawa (Barnett, 1988). Reports on the glaciomarine deposits in the Ottawa valley associated with the retreating ice, and the understanding from chemical and biological evidence of the environment of the Champlain Sea are presented in the volume edited by Gadd (1988) and are the subject of Stop 16. The engineering importance of the marine clays which are subject to catastrophic slope failure is summarized by Locat and Chagnon (1989).

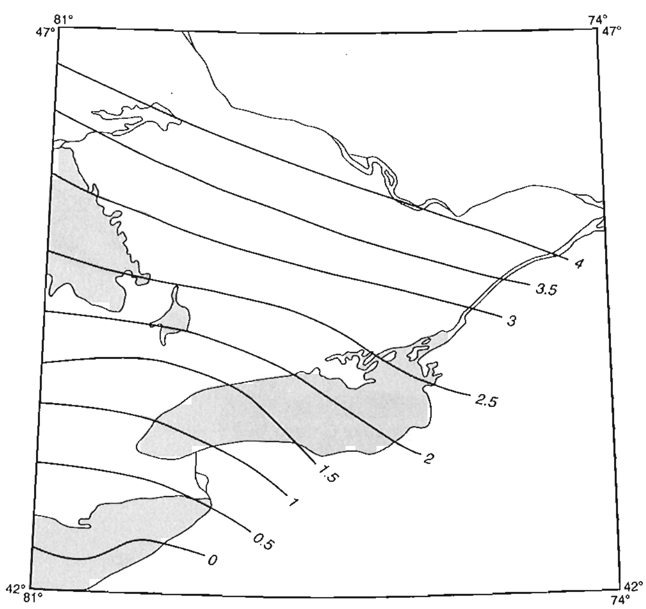


Figure 4. Present rates of isostatic rebound in mm/a based on water level records interpolated from Clark and Persoage (1970). See also Tushingham (1992)

At this time a passage was opened from the upper Great Lakes along the ice front to the Ottawa valley. Glacial Lake Agassiz, which covered large areas of Manitoba, northwestern Ontario, and the adjoining United States, drained through the newly forming Lake Superior. This water was added to the flow from the Great Lakes which ran into the Ottawa valley through this new passage, as well as to Lake Ontario through the Kirkfield outlet, and through the newly formed outlet from Lake Huron to Lake Erie (Lewis and Anderson, 1991).

By 10 100 BP the Lake Agassiz water source had been cut off by ice advancing in the Superior basin. However, ice retreat opened a channel from Georgian Bay to the Ottawa River through Lake Nipissing. This channel was sufficiently low that it captured all the drainage from the Michigan and Huron basins. At the same time, flow in the upper Ottawa River from glacial lakes in northern Ontario and Quebec joined the flow from the west. With the reopening of the link to Lake Agassiz between about 9600 and 8300 BP, flow again increased down the Ottawa valley to the Champlain Sea which was retreating as the land rebounded (Fig. 3d). The Ottawa valley was the only outlet for this flow estimated by Lewis and Anderson (1989) to have been about 200 000 m^3/s. By way of comparison, this is almost 200 times the present average discharge of the Ottawa River and about 20 times that of the St. Lawrence River below Montreal. The fluvial landforms created as a result in the Ottawa valley are described by Chapman and Putnam (1966, p. 48-50) and by French and Hanley (1975).

By 8700 BP the Champlain Sea had withdrawn to the Montreal area, and within another 300 years to the Quebec City region (Prest, 1976). Flows continued through the Nipissing outlet from the upper Great Lakes to the Ottawa River until about 4700 BP when the outlet was closed by isostatic uplift and flow from the upper Great Lakes followed the present route by way of southern Lake Huron into Lake Erie.

The postglacial period

Forests spread into southeastern Ontario soon after the glaciers and glacial lakes and seas retreated. Pollen zones, that is, pollen assemblages that typify specific vegetation associations, conform to pollen sequences and radiocarbon chronologies applying to most of southern Ontario that have been formulated by Terasmae (1981), Webb (1982), and Anderson (1989).

The general vegetation history outlined in Table 1 comprises four zones, the third of which has three subzones. The open spruce and open pine forests (pollen zones 1 and 2) represent successional forests, i.e., transitory phases that reflect, in the first instance, the colonization by plants of newly exposed glacial sediments, devoid of a soil mantle. As the climate changed from cold to cool and dry around 10 600 BP, the cold-adapted species migrated northward as the glaciers retreated. At the same time, other species immigrated, each at its own pace and in accordance with evolving edaphic conditions. These initial forests colonized a rapidly changing environment and cannot be directly equated with the spruce and pine forests that occur today in more northern regions; these have developed during thousands of years under relatively constant environmental conditions.

A mixed coniferous-deciduous forest known as the Great Lakes Forest became established about 7500 BP in response to warmer conditions during the Hypsithermal period. A decline of hemlock pollen, about 4700 BP, is generally attributed to a disease that almost eradicated this species and that had run its course by 3000 BP. More recently, chestnut blight and Dutch elm disease had similar effects.

Although there has been human habitation in the region almost since deglaciation (Ellis and Ferris, 1990), it was not until the nineteenth century that European settlement greatly altered the environment (Warwick, 1980). Widespread deforestation and the introduction of weed and cultivated plant species have drastically changed pollen signals (zone 4) from recent sediments. These changes, occurring at different times in different places, reflect changes in land use driven by political and economic forces. A case study of human influence on a portion of Lake Ontario is presented as Stop 3.

The most active geomorphical sites, other than areas of human activity, are those around the large lakes where changing levels throughout the Holocene have influenced coastal landforms and changed base levels of inflowing streams. Stops 2 and 14 deal with these environments. Smaller lakes and wetlands in the region contain physical, chemical, and biological records that provide a wealth of paleoenvironmental information. Paleoecological sites include those at stops 9 and 14.

Table 1. *Postglacial vegetation history of southeastern Ontario*

Time (ka BP)	Pollen zone	Features
	1	**Open spruce forest in dwarf-shrub tundra**. Spruce (*Picea*) dominant, with willow (*Salix*) and pine (*Pinus*); weeds: wormwood (*Artemisia*) and ragweed (*Ambrosia*)
10.6		— from cold to cool and dry —
	2	**Open pine forest**. Pine (*Pinus*) dominant, declining spruce (*Picea*), modestly increasing oak (*Quercus*).
7.5		— from cool and dry to warmer and wetter —
	3	**Mixed coniferous-deciduous forest**
	3a	**Hemlock (*Tsuga*) dominance**, with declining pine (*Pinus*), rise of basswood (*Tilia*) and hickory (*Carya*).
4.7		— Hemlock decline —
	3b	**Hemlock minimum**, rise of birch (*Betula*).
3.0		— Hemlock recovery —
	3c	**Hemlock dominance**, increasing beech (*Fagus*), elm (*Ulmus*) and birch (*Betula*); declining pine (*Pinus*), and oak (*Quercus*).
0.15		— deforestation —
	4	**Post-settlement vegetation.** Increasing non-arboreal (not from trees) pollen, e.g. ragweed (*Ambrosia*) denoting time-transgressive onset of impacts of lumbering, mining, and agriculture. Major time markers are: 1880 AD: Chestnut decline 1930 AD: Elm (*Ulmus*) decline

☑ STOP 1 Large-scale cross-stratified gravel sets, Brighton

John Shaw, George Gorrell and David Sharpe

NTS 31C/4, UTM 745785

Follow Highway 30 5.3 km south from Highway 401 to the stoplights in Brighton, and Highway 2 west for 6.2 km to the gravel stockpiles of the Trent Valley Sand and Gravel Company. Turn north (right) up the steep wave-cut bluff, taking the left hand fork of the road. The main exposures are in the lower pit to the west of the road about 0.75 km from the turn-off from Highway 2.

The large gravel pits at this site expose crossbedded gravels associated with both glaciofluvial deposits and lacustrine shoreline features formed by waves (beaches and spits). It is important to our understanding of the history of this area to know whether the crossbeds were deposited directly by meltwater from the glacier, or whether they were reworked by waves during high lake stands.

The large-scale cross-stratified gravel sets are exposed in two main sections cut into a broad bench (wave-cut terrace) that terminates on the lake side at a steep, wave-cut bluff (Fig. 5). There are old beach lines on the bench related to a former high stand of Lake Ontario. A lower bench is more heavily dissected by streams.

The benches or wave-cut terraces are related to the falling stages of glacial Lake Iroquois. Since the bench at A (Fig. 5) was eroded by waves, it is not immediately obvious whether the crossbeds were also formed by wave action, or whether they relate to the preceding glacial environment. Of course there was a period during deglaciation when the site at Brighton would have been influenced by both glacial and lacustrine processes. At that time, subaqueous fans were deposited where subglacial meltwater tunnels discharged sediment-laden meltwater into the lake (cf. Rust and Romanelli, 1975). Sediment of one such fan is exposed beneath the wave-cut bench at A on Figure 5. There is no morphological expression of the feeding esker to this fan and it is probable that it, too, was levelled by wave action.

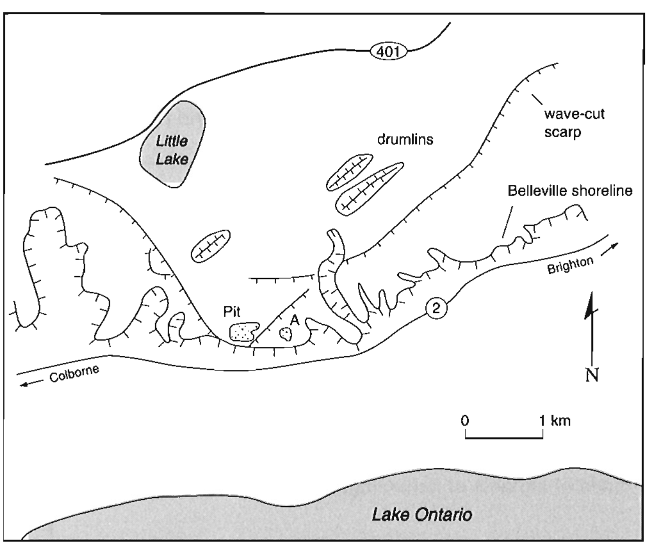


Figure 5. Location and local geomorphology of the large-scale cross-stratification site near Brighton.

The fan sediment, mainly ripple crosslaminated sand, is most likely an ice-marginal or subice shelf deposit; it is at about the same elevation as the large-scale crossbeds. Because the crossbedded gravels lie to the west of the fan and the paleoflow was westward, it is probable that the gravel bedforms were deposited first and the fan sediment was later banked against them. If the subaqueous fan is an ice-marginal deposit connected to an esker, then the earlier deposited gravel bedforms were probably formed subglacially, when the ice margin lay further to the southwest.

The crossbeds are found at two levels within the pit (Fig. 5) and there appear to be only two sets (Fig. 6). The lower set is approximately 5 m high and the upper set is about 3 m high. The lower set (Fig. 6, section b) has thicker strata and its coarsest units are coarser than those in the upper set (Fig. 6, section a). Sand beds and thick prisms of sand interrupt the gravel crossbeds and are banked against them, marking a reduction in the depositional slope (Fig. 7a, b).

The large-scale crossbeds overlie horizontal gravel and sand beds in the southern exposure (Fig. 6, section c), suggesting that the crossbeds represent gravel dune bedforms rather than large deltas (Shaw and Gorrell, 1991). A consistent and unidirectional paleocurrent of 200° is estimated from three measured foresets from section b. Similar measurement for section a gives a paleocurrent of 210°. Trough crossbedded sand in the horizontal gravel and sand beds (section c) indicates a flow direction of about 275° (eight measurements), which is oblique to the flow direction for the crossbeds. Trough crossbedded sand within the sand prism of the upper set of crossbedded gravel, also shows a flow direction oblique to the direction of migration of the larger-scale bedform. A complex flow is implied with local currents sweeping obliquely across the fronts of large gravel dunes. The trough crossbedded sands were deposited on small-scale dunes climbing across the face of the large-scale gravel dune which generated the crossbedded gravel.

Crossbedded sand in the subaqueous fan indicates a paleocurrent of about 285° (ten measurements). This is close to the flow direction for gravel dunes associated with the crossbedded gravel facies exposed in the pits to the west. But the

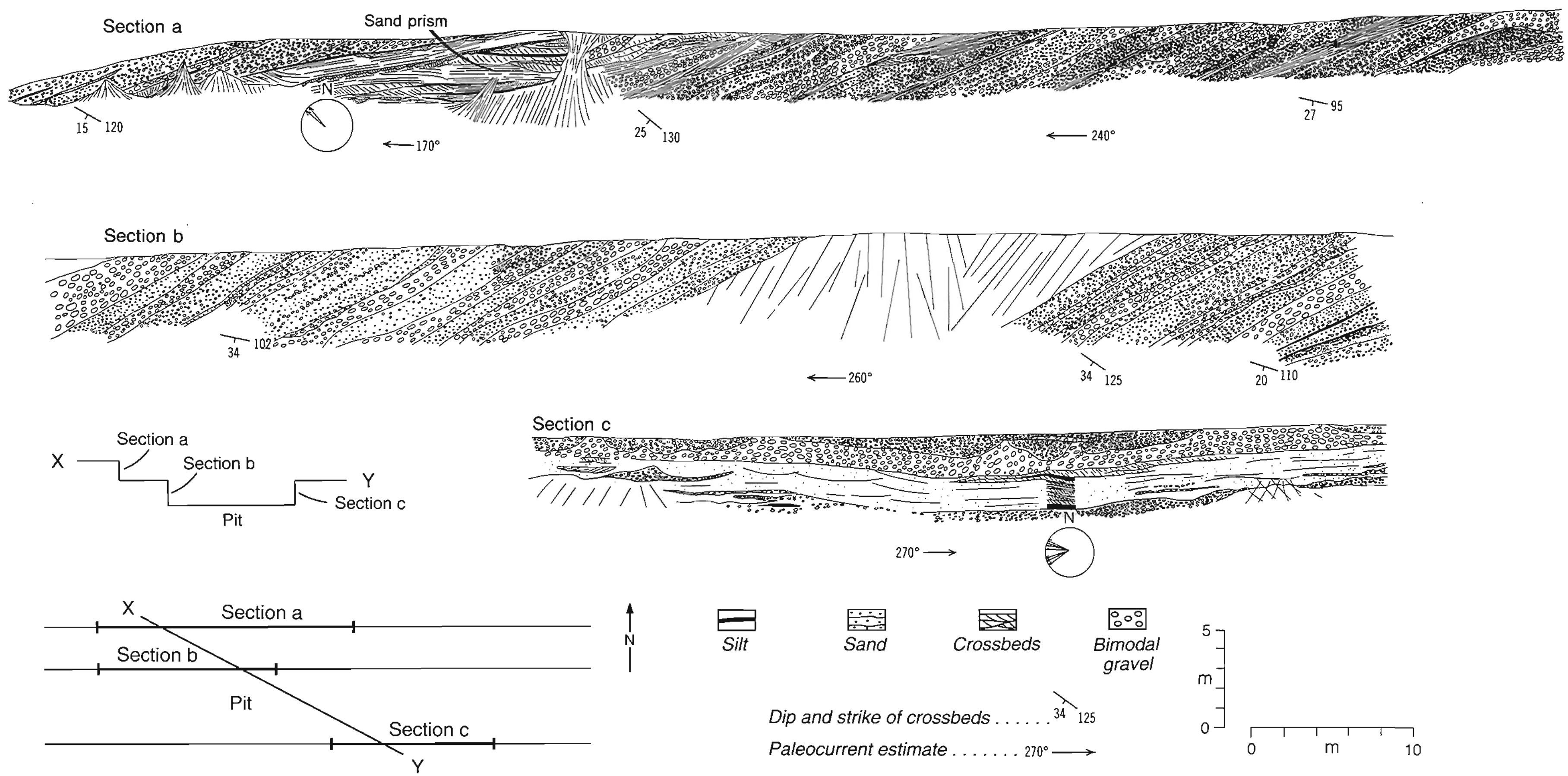


Figure 6. *Sedimentary cross-sections in bimodal gravel and sand near Brighton; section a, 3 m high crossbeds; section b, 5 m high crossbeds; section c, bimodal gravel and sand.*

gravel in the crossbeds represents higher energy of deposition than the sandy fan deposits. It is unlikely that these sediments were deposited at the same time, which would require dramatic distal increase in energy; therefore it is more probable that the fan (A, Fig. 5) was deposited after the gravel crossbeds in large-scale dunes in the pit (Fig. 5).

The details of the bedding and sorting of the crossbedded gravels are highly significant to their interpretation. Almost all of the gravel is strongly bimodal with boulders, cobbles, or pebbles making up the framework. The clasts themselves are extremely well sorted and show remarkable, usually normal grading (Fig. 8). In some cases the gravel is clast supported; the majority of clasts are in contact. In others, the gravel is supported by the matrix. The matrix is usually well-sorted, fine to medium sand and is commonly laminated. Both the primary lamination in the matrix and the nature of support of the matrix discount emplacement of the matrix as a result of sand filtering down through the framework.

Shaw and Gorrell (1991) explained such bimodal, graded crossbeds partly by longitudinal sorting, with bed shear stresses well above critical; larger clasts outrun smaller ones because of the effects of relative bed roughness trapping small clasts (Iseya and Ikeda, 1987). The matrix and gravel are interpreted to have been deposited simultaneously on the foreset slopes of gravel dunes by suspension sedimentation from the return currents of the separation eddies, and avalanching, respectively.

This explanation is not entirely satisfactory for the Brighton site because horizontally bedded gravel is also strongly bimodal (Fig. 6, section c). A possible variation on the flow separation theme is that the gravel and its sand matrix were deposited simultaneously from two sources. The gravel may have been deposited directly from a main current which veered away from the deposition zone; Ashmore (1982) described how, in a curved flow, heavy particles are carried by their momentum in relatively straight paths. The sand may have been deposited in secondary currents of local eddies, perhaps related to the large gravel bedforms themselves.

Boulders, cobbles, and pebbles in the crossbeds are well rounded and many bear percussion marks. Clearly, they were transported by vigorous currents and most closely resemble clasts from eskers in the local area. Carbonate clasts predominate, although Shield clasts are also well represented.

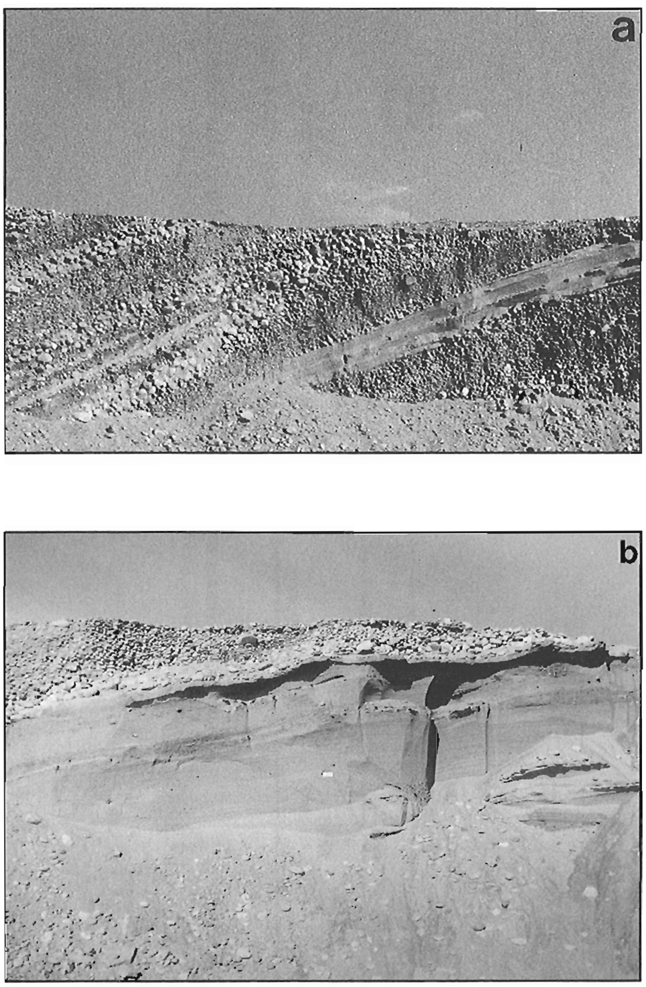


Figure 7. a) *Sand beds within crossbedded gravel, Brighton (section b);* ***b)*** *Sand prism within crossbedded gravel, Brighton (section c). GSC 1993-164A, 164B*

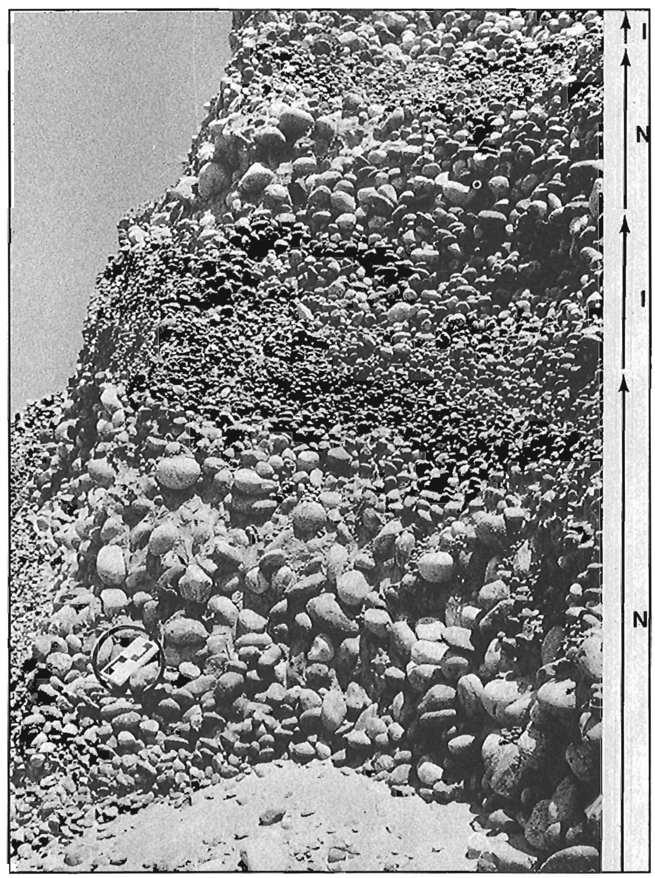


Figure 8. *Normally graded (N) and inversely graded (I), bimodal gravel, Brighton. Notebook is about 20 cm long. GSC 1993-164C*

Unidirectional flow from the east or northeast, vigorous sediment transport, high rates of bedload and suspension load deposition, and strong secondary currents imply glaciofluvial deposition for the large-scale crossbeds. The crossbeds themselves indicate bedforms on a scale that is most commonly associated with outburst floods (cf. Baker, 1978; Shaw and Gorrell, 1991). Shaw and Gorrell (1991) also argued that the gravel dunes were formed subglacially in tunnel channels. Although there is no direct evidence for this environment of deposition for the Brighton sediments, the inferred age relationship for the crossbeds relative to the subaqueous fan and supposed associated esker suggest that the crossbeds preceded the esker and might well have been associated with a phase of tunnel channel flow or with an earlier esker phase. The extraordinary characteristics of these crossbeds are certainly consistent with formation by catastrophic flows.

☑ STOP 2 The beach and dunes of Presqu'ile Provincial Park

Jane Law and Cheryl McKenna Neuman

NTS 30N/13, 31C/4

To reach Presqu'ile Provincial Park travel south 5.3 km on Highway 30 from Highway 401 to the stop lights in Brighton; turn right (west) 0.9 km, then south 4.2 km following signs to the Park. There is a fee for entering the Park ($6.00 per vehicle in 1992) and visitors should be cautious of poison ivy while walking on the dunes.

The Park occupies most of an L-shaped peninsula which extends into Lake Ontario (Fig. 9). Presqu'ile Peninsula was created during the mid- to late Holocene, when abundant glaciofluvial and glaciolacustrine sediments were driven onshore by prevailing west winds over the rising waters of early Lake Ontario. This created optimum conditions for the development of a sandy tombolo formed between two offshore limestone cuestas and the north shore of Lake Ontario. The area contains unique and diverse habitats including Carolinian forest remnants on the pre-tombolo spits, a marsh, dunes, an alkaline interdune panne (basin), a prograding beach, offshore limestone islands, and a variety of forest and meadow communities. This stop describes the development, evolution, and contemporary processes that are affecting the tombolo, its prograding beach, and superimposed dunes.

The limestone foot of the Presqu'ile Peninsula and adjacent High Bluff Island, together with the various limestone headlands along the western shoreline of Prince Edward County represent the most westerly, above water, extensions of a series of cuestas with adjacent bedrock valleys that occur in eastern Lake Ontario. The intervening bedrock valleys are controlled by a well-developed joint system (Mirynech, 1962), which was exploited by interglacial rivers and eroded by glacial and subglacial meltwater erosion (Shaw and Gilbert, 1990) in the Pleistocene. One of these bedrock valleys extends southwest from the Bay of Quinte beneath Presqu'ile tombolo into Lake Ontario (Fig. 10).

Sources of sand for the beach and dunes of Presqu'ile are glaciolacustrine sands along the Lake Ontario shore (Leyland and Mihychuk, 1980) and sands in Popham Bay to the west of the tombolo. The latter deposits are 14.5-19 m thick near the tombolo but they thin toward the lake (Fig. 11). Comparison of the 1815 Admiralty chart and the 1989 Canadian Hydrographic Survey reveals that the depth of Popham Bay has decreased from 1.5-11.7 m (lake level 74.9 m) to 0.45-9 m (lake level 75 m) along the same transect over the 174 year period. Since little sediment is supplied by inflowing streams (Law, 1989b), almost all of the sediment contributing to this shallowing has come from offshore. Popham Bay acts as a trap for eastward moving sediment (Brebner and LeMehaute, 1961).

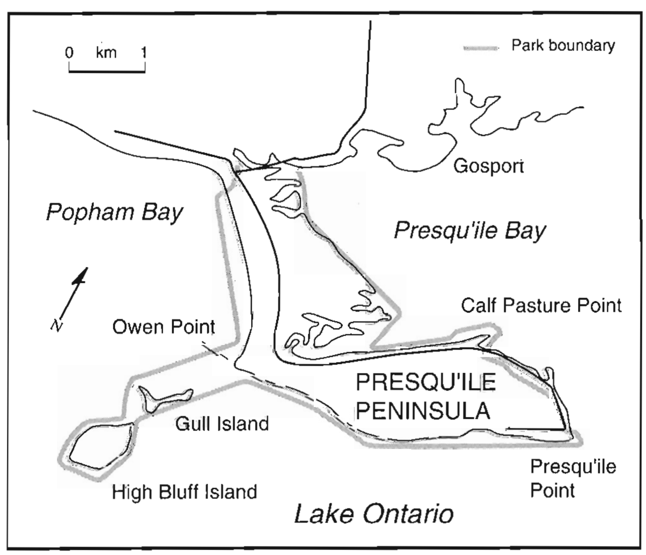


Figure 9. The location of Presqu'ile Peninsula, Presqu'ile Provincial Park.

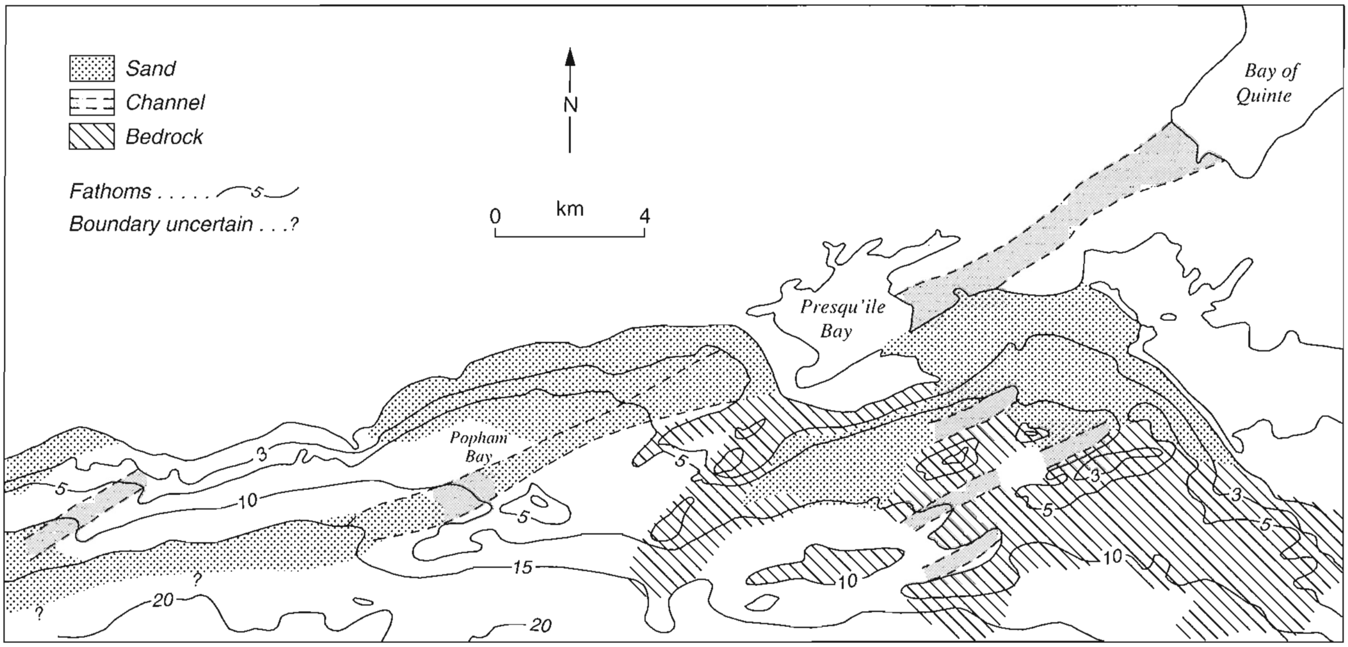


Figure 10. *Nearshore sand deposits around Presqu'ile Peninsula and buried channel that underlies the tombolo (modified after Martini and Kwong, 1985).*

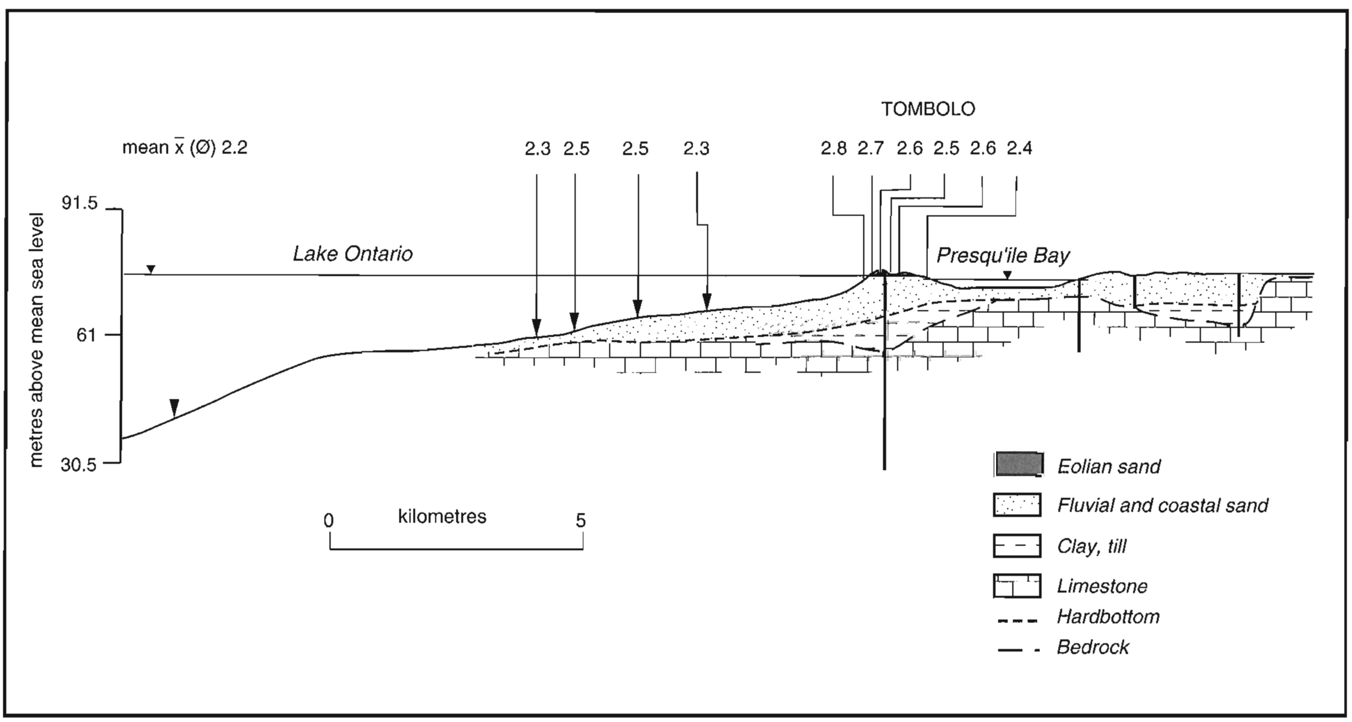


Figure 11. *Cross-section through the Presqu'ile tombolo showing the thickness of the sediments and mean grain size in Ø units (after Martini and Kwong, 1985).*

A model for the development of the tombolo (Fig. 12), presented by Ernsting (1976) shows recurved spits built simultaneously from the mainland and from Presqu'ile Island. After the spits joined to form the tombolo, wave-borne sediment became trapped to the west and a broad beach of well-sorted, very fine sand developed at the head of the new Popham Bay.

A transect across the dunes, the panne, and the beach

A walk across the Park traces the development and modern processes of the tombolo, the beach, and the dunes. Start on the Park access road at the entrance gate to Beach 1 (Fig. 13). The first beach west of the new tombolo runs beneath the Park access road. The first foredune ridge occurs to the east of that beach. The present beach now lies 240 m to the west and the Park access road is located on a panne (a flat, alkaline, interdune sand plain deflated to the water table). The accumulation of sand in Popham Bay caused the beach to grow westward, and the dominant westerly winds transported sand from the beach to create foredunes to the lee of the advancing beach face (Fig. 12c). The oldest, easterly foredunes subsequently became devegetated and deflated to the water table. Sand from these dunes moved eastward creating a panne and an irregular, 4-5 m high back dune ridge (Fig. 13).

Walking east from the Park access road we traverse a section of panne and cross a deflation ridge (<0.5 m) that marks the former western edge of the 4-5 m back dune. Three metres east of the deflation ridge, and approximately parallel to it, there is a line of grey limestone slabs that probably marks the site of one of the original recurved spits that extended south from the shore of Lake Ontario. The back dune ridge is located 4 m further east. This dune became relict when the foredunes to the west were revegetated and the beach had moved so far west that the sediment supply was cut off. At that time the back dune may have been fully vegetated because there is a buried soil two-thirds of the way up the west face of this dune. The windward slopes of this dune are now sparsely vegetated as a result of human activity, including commercial sand removal from three locations during the past 40 years. These slopes are deflating. Aerial photographs show that the dune has moved eastward into the Presqu'ile Bay marsh over the past 50 years (Law, 1989a), causing the panne to widen. Cedar stumps now exposed on the windward slopes of this dune probably germinated on the lower eastern slopes of the dune adjacent to the marsh. These trees were overwhelmed by sand and now the stumps have emerged from the windward slopes.

The lower lee (eastern) slopes of the back dune are covered by cedar forest that slows the eastward movement of sand. The nonforested slopes are covered by dense herbaceous vegetation that can survive in this arid dune environment because the lee slopes trap snow and sand in the winter. Snow buried by sand provides moisture into early summer. The crest of the back dune ridge, opposite the Beach 1 gate, overlooks the "Fingers" marsh to the east. This marsh is the remnant of the recurved spit attached to the shore of Lake Ontario. Higher, dryer spit remnants are now covered by

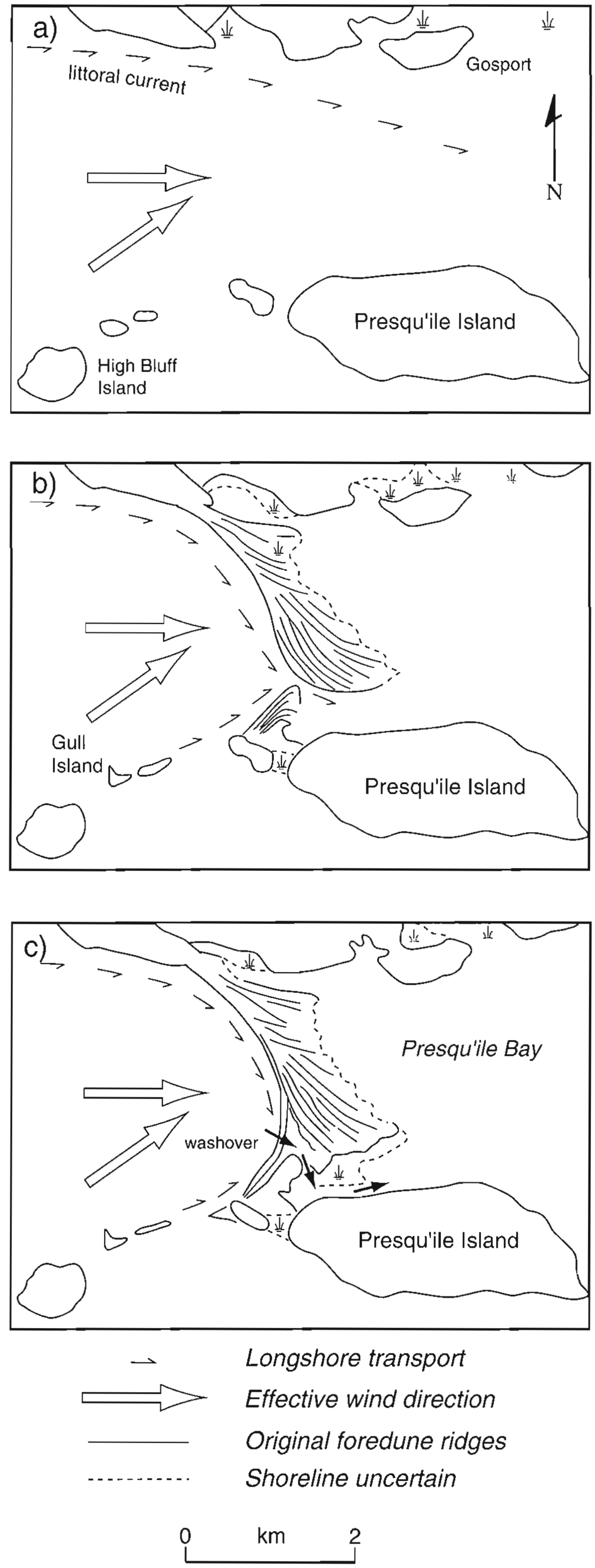


Figure 12. *The first three stages in Ernsting's (1976) model of the development of the Presqu'ile tombolo:* ***a)*** *initial bedrock formation,* ***b)*** *spit development,* ***c)*** *formation of the tombolo.*

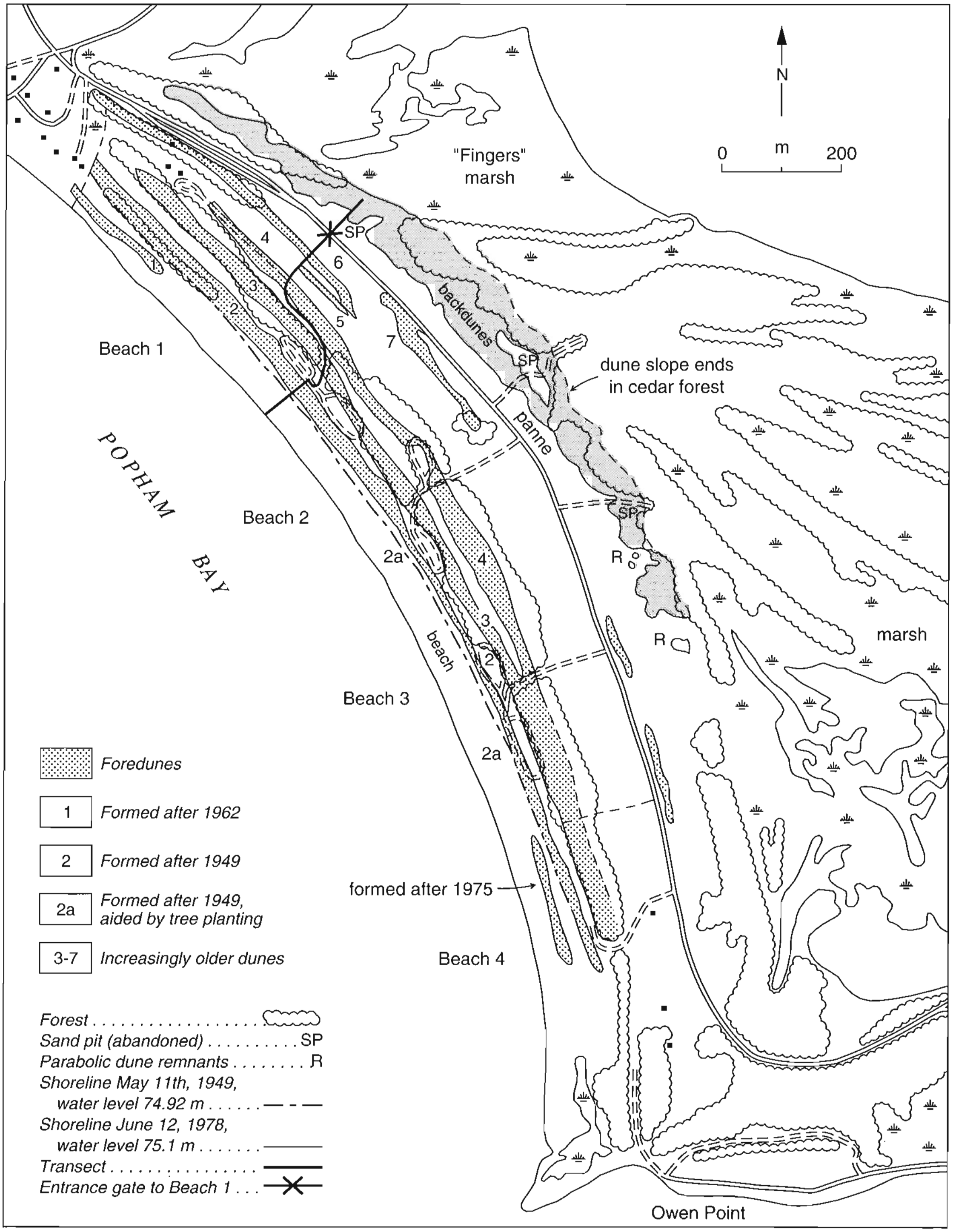


Figure 13. *Details of the beach and dune formation at the main beach at Presqu'ile.*

cedar forest; intervening shallow channels are infilling with sediment. Bathymetric charts indicate that in 1815 Presqu'ile Bay was 3.6-4.8 m deep; in 1989 it was only 1.8-2.7 m deep. This infilling is related to European settlement, land clearing, and agriculture (Law, 1989b).

To complete this field transect, return to the Beach 1 gate, and walk west along the road toward the beach, crossing the panne and six foredune ridges (Fig. 13) that increase in elevation and decrease in age toward the beach. The oldest ridges are marked by an irregular line of cedar trees growing on slightly elevated (<0.5 m) areas above the floor of the panne that becomes flooded in spring. The foredune ridge located between the parking lot and the beach has developed since 1949. The newest foredune that lies to the north of Beach 1 but ends in the middle of the beach, has developed since 1962.

The present Popham Bay beach is 2.8 km long, about 125 m wide in the centre, and 35 m wide at each end. It is composed of very fine sand, although at the extreme southern end, around Owen Point, pebbles accumulate from the adjacent rocky islands, and a pebble lag develops on the northern part of Beach 1 during the winter months. The beach is accreting. Beach 1 prograded 80 m into Lake Ontario between 1949 and 1986 (Law, 1989a; Fig. 13). The southern end of Beach 4 shows similar accretion (Van Heyningen, 1989) and a marina built on Owen Point in the early 1960s silted up completely by 1979. Only the extreme ends of the beach are natural (unraked). In these locations the beach is narrow and foredunes are growing. The centre of the beach is groomed regularly to remove litter throughout the summer. This practice disrupts the establishment of micro- and macrophytic plants, important in surface stabilization and dune development. Consequently the beach is widest here. Human traffic has locally disturbed the development of a foredune and sand accumulates in the parking lots, to the east of the beach, during the fall and winter. Snow fencing, installed along the back of the beach in September, traps this sand, but until recently it was removed each spring. Since September 1991 efforts have been made to establish a new foredune. The fencing has been relocated to maximize the amount of sand trapped and to restrict access to the embryo dune throughout the year. Dune grass has been planted. Regular ground surveys and anemometry, began in autumn 1991 and spring 1992, evaluate sediment transport, morphometric, and airflow relationships as the dune develops.

There is a distinct seasonal variation in the factors that affect sand transport across the beach and dunes, i.e., the sand supply, wind speed, ground frost, and lag development, as well as beach raking and vegetation growth (Law, 1990). In summer and early autumn the beach face progrades rapidly, as much as 17 m in 35 days (June/July 1989). In winter the ice foot stops the sand supply to the beach. Wind speeds in summer rarely exceed threshold velocity but they are much stronger in autumn and as much as 20 cm of deflation takes place within a few storm events. In 1991, major transport events occurred in early November when the wind shear velocity averaged 45 cm/s and in December when the surface was frozen and shear velocity varied between 40 and 50 cm/s. Transport is predominantly northwest to southeast. By December, the north end of Beach 1 has a shingle lag and the beach surface becomes concave. Scouring of frozen sediment occurs during the winter except where an ice sheet develops in the beach hollow, then further winter transport is negligible. During the spring melt the surface is saturated and denivation forms predominate.

Direct measurement of the eolian sediment flux using traps has been difficult due to rapid deflation of the surface and blocking of the traps by sand and snow. Predictions of sediment transport from models based on wind tunnel experiments overestimate transport by an order of magnitude or more. Approximately 30 cubic metres of sand per metre width of beach was actually transported inland during the autumn of 1992. The discrepancies between the actual and predicted values may be attributable to 1) differences in scale and therefore in shear velocity calculated from wind profiles in the field and laboratory, and 2) failure to account for flux reduction resulting from the presence of lags, and damp or frozen surfaces. These results illustrate complex surface eolian transport in cold, humid, coastal settings.

☑ STOP 3 Moira River and Bay Of Quinte: natural and human factors in an enclosed embayment of Lake Ontario

Adele Crowder

NTS 31C/3, 4

Follow Highway 37 south 0.5 km from Highway 401 to a small park on the Moira River located on the left (west) side of the road on the northern outskirts of Belleville for the first part of this stop. Then proceed south 2.3 km through Belleville to Highway 2. Turn right and travel 0.9 km to Highway 62. Turn left and travel south over Zwick Island and the Bay of Quinte high level bridge from which the bay may be viewed, although stopping is not permitted. An overview of the Bay of Quinte may be had from a private road opposite the Mountainview Airport 14.0 km from the turn. To reach another overview, travel 1.7 km further along to Crofton, then turn left (east) on the road to Demorestville, and travel 5.0 km.

The Moira River

The Moira River (catchment area 2840 km^2, mean annual flow 30 m^3/s) is the second largest of five rivers flowing into the Bay of Quinte (Fig. 14). Flooding is frequent during spring runoff, when bridges at Belleville have been damaged by ice jams. In summer, mean flow drops to 8 m^3/s. A plume of calcium carbonate derived from the Ordovician limestone in the basin is carried by Moira River into the bay, where sediment has approximately twice the detrital inorganic carbon found in Lake Ontario sediments.

Weirs and falls along Moira River have been used for water power since settlement began, after 1780. Old mill buildings and weirs can be seen to the right of the road entering Belleville. The peak use of water power was about the 1850s, when there were 40 mills on the river.

Lumbering dominated industrial development on the Quinte rivers during the 18th century. At present, 69% of the watershed is forested and agriculture uses another 27%. Subsistence farming has given way to cereal production, dairying, and a period of intensive corn growing in the 1970s and early 1980s. Both forest clearances and farming have affected sediment carried into the bay. Warwick (1980) developed an erosional index (Fig. 15) based on the ratio of the clay fraction smaller than 10.5 ø, to the clay fraction smaller than 8.5 ø. This showed that more fine sediment was eroded during the initial forest clearances and periods of barley production; the latter leaves fields open longer to erosion.

Deposition of fine sediment on fish spawning beds and the destruction of migration routes by dams have eliminated some populations of fish, including Atlantic salmon which ceased to spawn here in the early nineteenth century, having previously been abundant. To improve fish habitat and water quality, rehabilitation of parts of the Moira basin has begun. Vegetated strips are left beside the creek, cattle are fenced away from it, runoff from manure piles is collected, and incentives are provided for leaving a cover of vegetation on fields all year.

A gold rush occurred in the Moira valley in the 1860s, when shafts were sunk at Eldorado and Deloro (Fig. 14). After a decade, arsenic, produced as a by-product of smelting, proved more lucrative than the gold, and its sale continued until the 1950s, mainly for pesticide use. Other ores were brought to Deloro for smelting, including silver, cobalt, copper, nickel, chromium, and uranium. The Deloro smelter was shut down in 1961, leaving tailings and ponds high in arsenic, heavy metals, and radioactivity. In 1971 poisoning of cattle downstream alerted the Ontario government to the toxicity of this site and its drainage. The Ontario Ministry of the Environment has now covered tailings with limestone and demolished contaminated buildings. Tailings drainage is pumped through an arsenic treatment plant before the water is released to the Moira River.

Arsenic and heavy metals have been found in the water of the Moira River, in sediment in Moira Lake, 10 km south of the mine, in sediments downstream and at the river mouth, and in the Bay of Quinte (Crowder et al., 1989). Moira River is not used for drinking water although, since arsenic has been treated, guidelines for drinking water quality have not been exceeded; daily loadings decreased from 35 kg in 1979-82 to 6 kg in 1989 (Environment Ontario, 1991a). Drinking water for Belleville is taken from the Bay of Quinte and occasionally has exceeded guidelines for copper, cadmium, and iron (Fig. 16). Sediments in the bay generally exceed guidelines for dreged material both for arsenic and some heavy metals (Beak Consultants Ltd., 1988). Arsenic in a core off Belleville peaked in 1868 and in 1936 (Mudroch and Capobianco, 1980) probably related to high flow or ice jams (Sly, 1986). A comparison of sediment along the shore of the Bay of Quinte found that levels of metals such as chromium and nickel decreased with distance from Moira River, but other metals (such as lead) did not. The latter are generally associated with outlets of sewage treatment plants in the bay (Crowder et al., 1989).

The only metal found to correlate with decreased growth of aquatic plants in the Bay was copper (Dushenko, 1990). Copper also was bioconcentrated by a species of snail (*Stagnicola elodes*) in the marshes, without detrimental effects (Greig, 1989). Muskrats did not show detectible levels of arsenic, nickel, and lead in their kidneys, but copper averaged 8 ppm (Greig et al., 1989).

Zwick Island

The route continues through Belleville, the largest centre on the bay, population about 36 000. From the Highway 62 bridge, a park on Zwick Island is visible on the right. It covers a former landfill site, used from 1949 to 1971. Leachate occurs in surface drains and is thought to enter the bay in groundwater. Phenol concentrations are up to 2 mg/L and loading to the bay is being monitored (Environment Ontario, 1991b).

In four years between 1973 and 1987, phenol concentrations in Moira River exceeded the drinking water guideline of 2 μg/L. Generally, phenolic compounds in the bay have been derived from the industries on the Trent River, but at Belleville they were contained in discharges from a factory which has now closed. Temporal trends in contaminant loads are in general downwards (Fig. 16).

Bay of Quinte

The Z-shaped Bay of Quinte stretches 64 km east-west and has an area of 250 km^2. It is 4-8 m deep where seen from the viewpoints; it is up to 17 m deep where Highway 49 crosses the narrower middle bay, and is over 50 m deep in the lower bay as it opens into Lake Ontario. Flushing occurs 2-3 times during high inflow in spring. Generally flow is west to east, but cold Lake Ontario water enters the bay as a bottom layer in the extreme east. The upper bay is generally well mixed, but a thermocline develops in the lower bay during summer (Minns et al., 1986; Remedial Action Plan Coordinating Committee, 1990).

About 12 500 BP, glaciolacustrine clays were laid down in Lake Iroquois which covered the area. The water level dropped below the present level of Lake Ontario about 11 400 BP and the bay was gradually flooded from the east in the rising waters of Lake Ontario. Middle bay was flooded

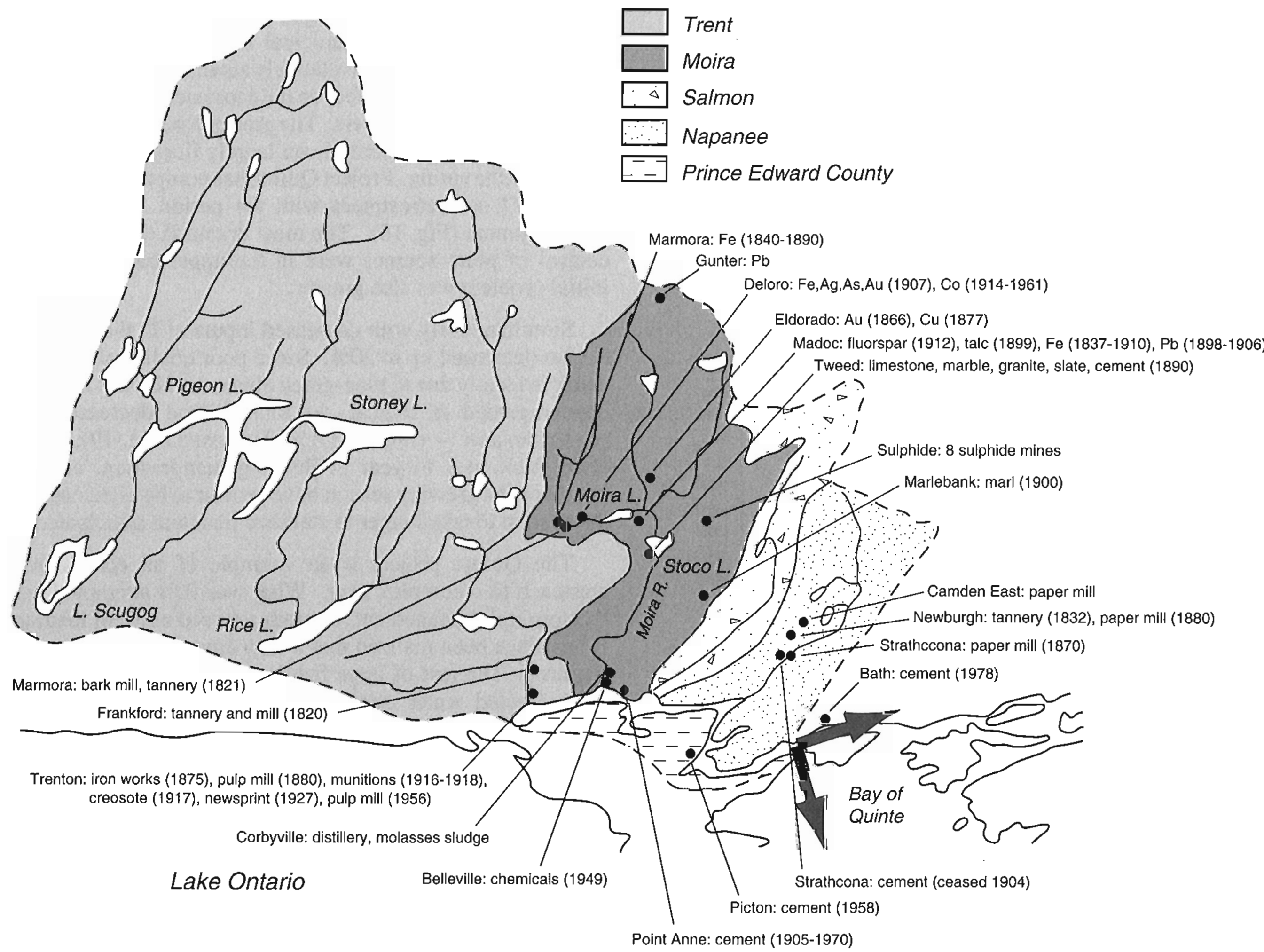


Figure 14. Minerals, mining, and quarrying in the Bay of Quinte watershed (after Sly, 1986).

about 2500 BP. Most of the creek mouths have coarse sediment covered by finer deposits (Sly, 1986; Damiani and Thomas, 1974) reflecting the rising base level.

The region has been inhabited since about 5000 BP. Between 1668-1682 French missionaries settled with the Huron Indians who were farming in the area, but by the time American settlers arrived in the 1780s, only Mississauga Indians were using the area for hunting. Since then, farming and the timber trade developed as the economic base.

The effects of this development are reflected in the sediment of the bay. Warwick (1980) showed that the bay was oligotrophic to mesotrophic (low to moderate in plant nutrients) until the beginning of this century (Fig. 17), although total phosphorus in sediments peaked sharply during initial forest clearances at the end of the eighteenth and beginning of the nineteenth century. After clearance, the proportion of finer sediment being eroded increased, as did the rate of sediment accumulation; clay minerals confirm the terrestrial origin of fines.

During the 1950s total phosphorus (P) began to increase rapidly in the sediment from sewage and detergents, and the bay became moderately eutrophic. In 1972, Project Quinte was formed to study the status of the bay and the effects of decreasing point source inputs of P. This group of scientists and managers from federal and provincial ministries and from Queen's and Trent universities published an account of the effects of point source control of P (Minns et al., 1986) and a Remedial Action Plan (RAP) for the bay was initiated.

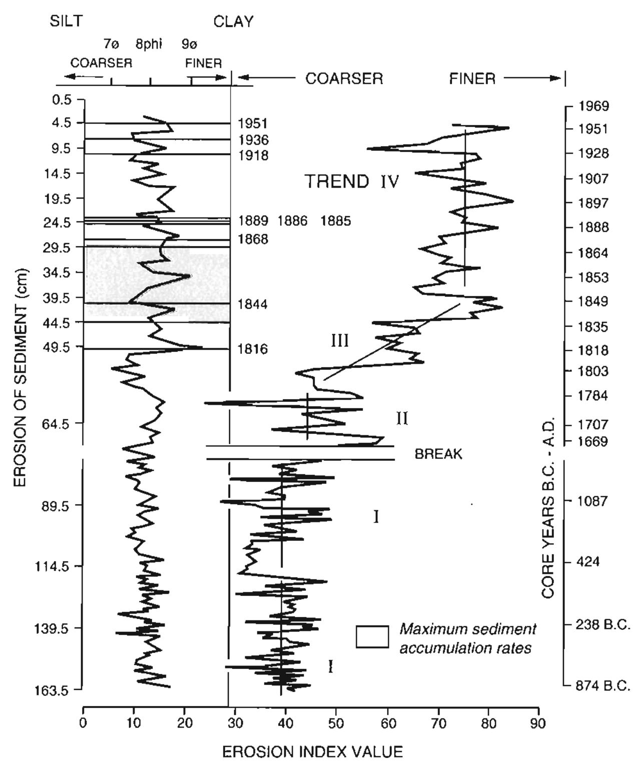


Figure 15. *Mean particle size of Quinte sediment showing the effect of European colonization. The data are from Warwick (1980) as presented by Sly (1986).*

Economic incentives for studying and treating eutrophication in Quinte include (a) quality of drinking water, for example at Belleville where eutrophication led to blooms of blue-green algae which make the water unpalatable even after treatment, and (b) deterioration of fishing, a major component of the regional economy. More recently the Remedial Action Plan has defined other objectives, such as reduction of toxic contaminants in fish (notably mercury, mirex, and dioxin), control of bacteria that cause closures of beaches, and maintenance of wildlife habitat.

In the period 1956-1983 the annual commercial harvest of fish from the bay was 390 Mg. Some species became unsaleable because of contamination (e.g., American eel and carp). By the 1970s a very large sports fishery had developed with a total annual harvest of all species being 217 Mg in 1984. The combined effects of exploitation, eutrophication, introduction of non-native species, and climatic shifts have resulted in dramatic reduction in certain fish stocks in the bay (Hurley, 1986 in Minns et al., 1986).

Control of point sources of P through improvement of sewage treatment plants during 1972-1982 demanded a target of 1 mg/L P or less in municipal inputs to the bay. This component of the P load is relatively small, but is controllable and also has a maximal effect on the ecosystem during periods of low water in hot summers. The greater loads in river water, which are not easy to control, are largely flushed out to Lake Ontario in the spring. Project Quinte has compared the period 1972-1977 as pretreatment with the period 1978-1981 as post-treatment (Fig. 18). The most dramatic decreases after control of point sources were in the upper bay, where the initial problem was also greater.

Simultaneously with decreased inputs of P, the inputs of N also decreased up to 20%. Since poor quality of drinking water is locally due to blue-green algae, the N loads have not been regarded as critical. As total P load decreased, the phytoplankton (= chlorophyll a) decreased (Fig. 19). Variation from year to year in flushing, temperature, and the length of the growing season have proven to be high, causing the system to take longer to stabilize than was anticipated.

The Quinte project is an example of an ecosystemic approach to a complex area. What was first attempted was 'bottom up' management, through reduced nutrient loading. It has since been realized that a 'top down' approach is also required. The loss of large fish is seen as one of the factors that affected water quality, because indirectly their loss increased the quantity of algae present, as predation on zooplankton was decreased. Loss of aquatic habitat has now also been recognized as having contributed to instability. The value of research at Quinte is not only due to the large scale of this experiment in managing water quality and the fishery, but also to the way in which government agencies and the public have co-operated.

Large areas of cattail marsh can be seen from the viewpoints. Such marshes are now characteristic of the bay, whereas there are few remaining areas of swamp forest; in total it is estimated that two-thirds of the wetlands present in the early nineteenth century have been converted for agricultural or residential use. Terasmae and Mirynech (1964)

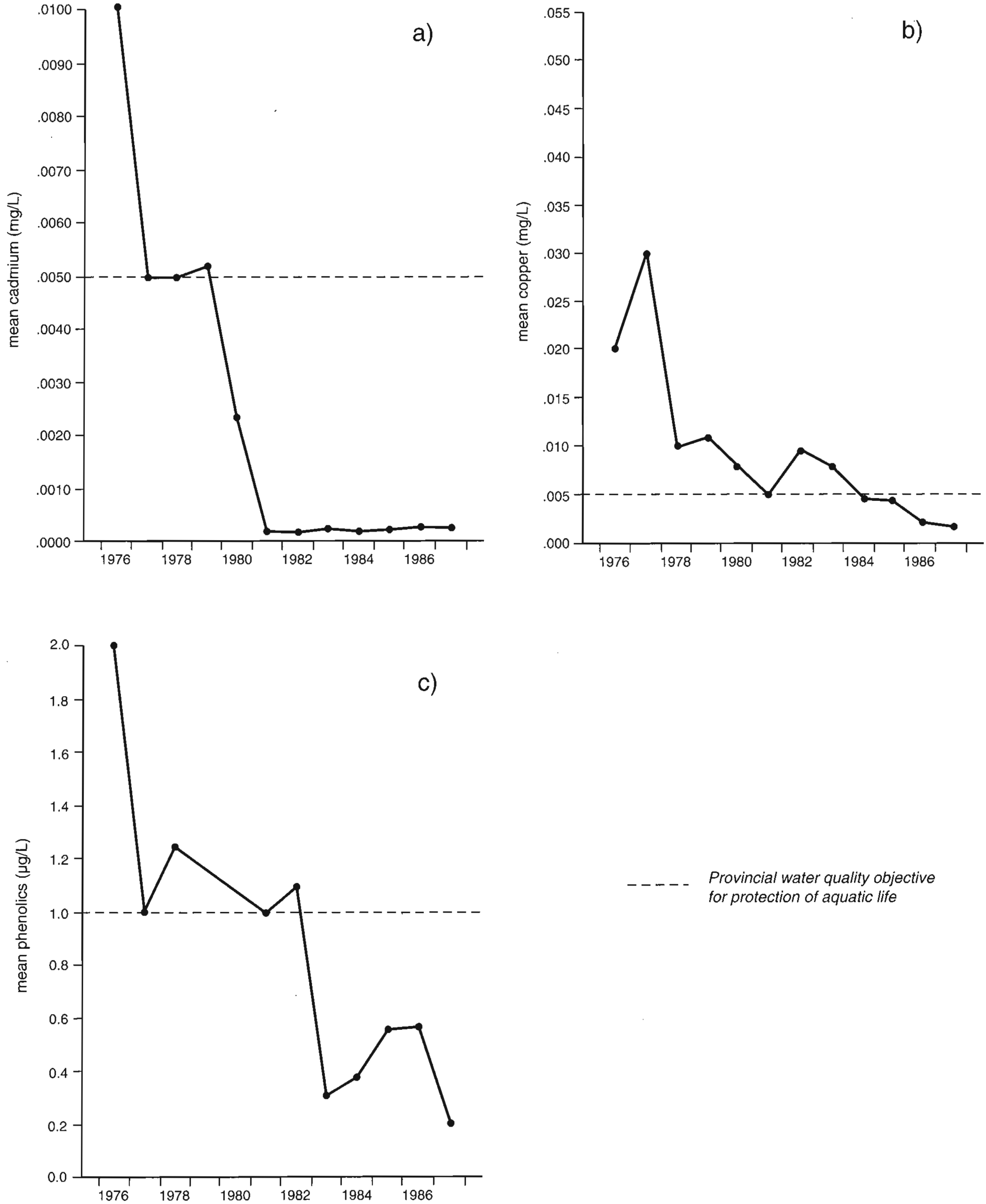


Figure 16. *Mean annual concentrations in the Moira River of* ***a)*** *total cadmium,* ***b)*** *copper,* ***c)*** *phenolics (after Beak Consultants Ltd., 1988).*

considered that the marshes are of relatively recent origin, as they can only have formed after submergence of the upper part of the bay. Cattails (*Typha* spp.) are favoured by a stabilization of water levels, so may have increased as the Lake Ontario level has been controlled for shipping and hydro power (Busch et al., 1989).

Marshes at Muscote Bay and Hay Bay are excellent habitat for waterfowl, muskrats, amphibians, and reptiles. Local opinion is that most of the animal populations have recently declined (Crowder et al., 1989; Greig, 1989). In the hope of increasing habitat diversity and therefore the diversity and number of animals, the cattail marshes near North Port have had channels dug in 1990 and were partially burned in 1992. Some of the marshes were formerly used for wild rice cultivation. They are currently used for trapping muskrats and beaver when the price of pelts is high, and for hunting and fishing.

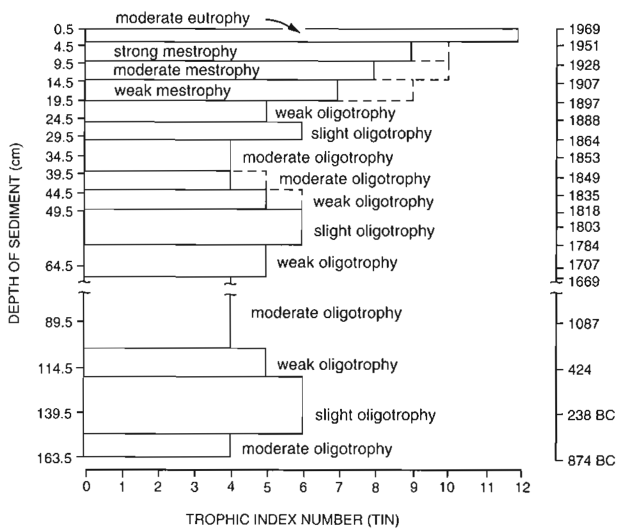


***Figure 17.** Tropic (nutrient) status of the Bay of Quinte interpreted from core data (Warwick, 1980)*

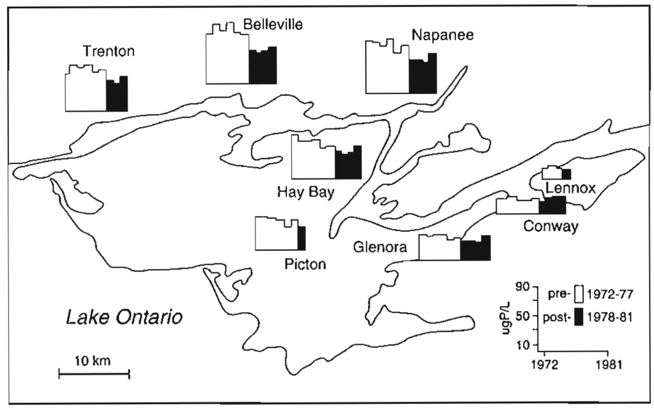


***Figure 18.** Mean total phosphorous concentrations in the euphotic zone in the Bay of Quinte for the period 1972-1981 (after Robinson, 1986).*

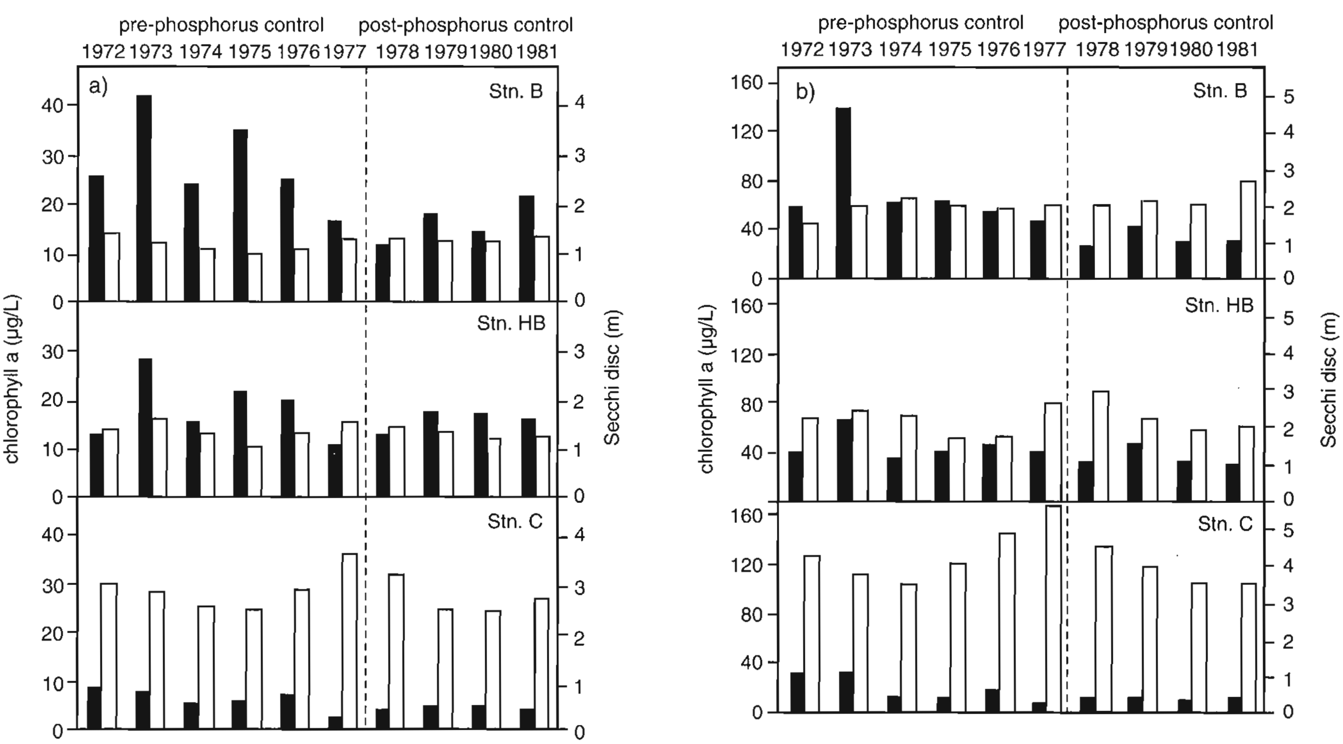


***Figure 19.** Chlorophyll a concentrations (solid bars) in euphotic zone and Secchi disc (measures water clarity) depths (open bars) at stations in the upper (B), middle (HB), and lower (C) bays for the pre- and post-phosphorus control periods: **a)** seasonal means, **b)** seasonal maxima (after Robinson, 1986).*

☑ *STOP 4* Ice scours in glacial Lake Iroquois

Robert Gilbert

NTS 31C/3, UTM 308851

These features described by Gilbert et al. (1992) may be seen along Highway 49 north of Picton, but the best example is the location marked by the arrow on Figure 20. If approaching from the east, turn west on the Woodville sideroad 9.4 km north of Picton, or 15 km south on Highway 49/2 from junction at Marysville on Highway 401. Bear right (north) at 3.1 km and turn right at 4.3 km where the road turns left (west). Proceed 0.5 km along the farm lane to a left turn to view the large scour in a field to the northeast. If approaching from Stop 3, travel east on the road from Crofton through Demorestville. Bear right at a T-junction 14.2 km from Croften, and left at a second T-junction 1.0 km further along. Travel 2.4 km and turn right at the Bethesda Community Hall. Travel 1.9 km to the farm lane referred to above. The best time to examine scours is in spring or fall before or after crops are growing in the fields. The landowner, Mr. Henry MacHill, lives in the white house nearby.

On the gently undulating top of the limestone escarpment in eastern Prince Edward County, a concentration of long, shallow furrows is found in the thin glaciolacustrine sediments of Lake Iroquois. Most occur at elevations of 110-125 m a.s.l. on the south of an area of higher ground, although a few occur to 135 m and below 100 m a.s.l. The mean length and width are 810 m and 73 m respectively. Most are straight (Fig. 20), but about one quarter show curvature of greater than 3°. Orientation is closely clustered around a mean azimuth of 261° (Fig. 21).

The furrows are incised a maximum of about 1 m. Ridges rising less than 1 m above the ground surface extend irregularly along the sides and there is considerable variation in detail of cross-sectional shape. Part of this may be due to subsequent erosion and to agricultural practice which appear to have smoothed and muted the profiles, but only subtle features have been obliterated completely. Some still have a small ridge or ridges in the troughs in the orientation of the furrow. The scours have formed in a thin cover of glaciolacustrine silts and clays on a nearly flat-lying bedrock surface. In some cases, the sediment has been removed to expose bedrock in the trough, but there appears to be no control of plan form by patterns in the bedrock. At the west end of some furrows, the lateral ridges wrap around the end.

The furrows are interpreted to be ice scours formed in Lake Iroquois either by icebergs calved from the damming glacier, or by seasonal, wind-blown lake ice. It is unlikely that the scours formed under an ice shelf of the damming glacier because of their crosscutting relationship (Fig. 20). The glaciolacustrine sediments at the study site are fine grained with only a few pebbles and cobbles (probably ice rafted) and so were more likely deposited in a distal lake environment, rather than near the margin of a large, active ice sheet. If they formed under a readvancing ice shelf, more evidence of a high-energy, proximal environment should be found.

At the time of highest Lake Iroquois the water level would have stood about 205 m a.s.l. in the study area (Gilbert et al., 1992). As the damming glaciers retreated to the northeast, water levels stood for short periods at 182 m (Frontenac phase), 124 m (Sydney phase), and 103 m (Belleville-Sandy Creek phase). It is possible that some of the scours formed at the highest Iroquois and Frontenac phases when the water depth would have been 60-90 m in the region. However, concentration of scours in a narrow range of elevation, their similar orientation, and the absence of large amounts of ice-rafted debris (especially dumped mounds of coarse sediment) suggest that bergs are less likely to have created the scours than lake ice.

During the Sydney phase, the site of most of the scours would have been in less than about 9 m water depth on the south side of a low island about 20 km offshore from the north

Figure 20. *Part of air photograph A23662-55 showing ice scours in the sediments of glacial Lake Iroquois. Arrow marks sites referred to in the text where scours are clearly visible on the ground.*

shore of the lake. At this time the ice dam was sufficiently withdrawn to have created a maximum fetch to the east-northeast of at least 200 km. Rapid lake level lowering from water depths of about 60 m at the end of the Frontenac phase brought recently deposited, moderately distal, undisturbed glaciolacustrine sediment very near to the water surface. These soft sediments could have been scoured by wind-driven lake ice in relatively shallow water (less than about 10 m for the most part) around an offshore island where ice ridging and pile-up would have been promoted.

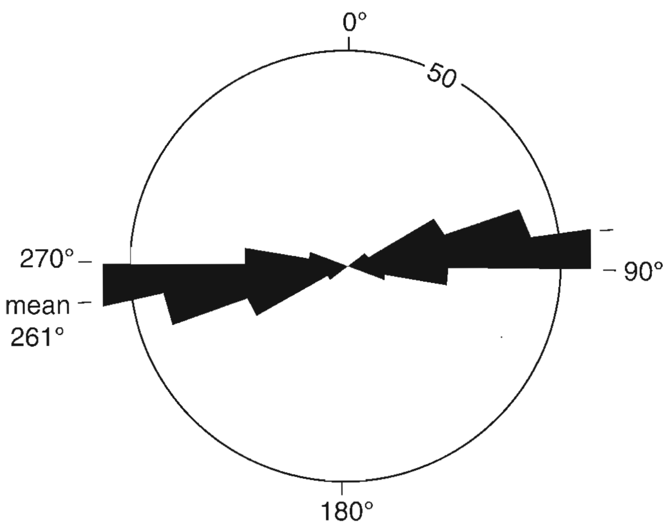


Figure 21. *Orientation of 164 ice scours in the study area plotted in 10° increments. Data are mirrored across the origin.*

Scours are formed instantaneously, and so their orientation does not necessarily correspond to the prevailing wind direction. Nevertheless, the highest probability is that they will be created by the dominant wind. From the sedimentary record in sand dunes, David (1988) records a strong, prevailing wind direction from the northeast and east (orientation toward 230-270°) in southern Quebec during the immediate postglacial period. These were katabatic winds draining cold air from the glacier to the northeast. It is probable that the situation in southern Ontario was very similar and that the orientation of the scours (mean 261°, Fig. 21) reflects the effect of this wind. The ridges that wrap around the western end of some scours supports the interpretation of push from the east to a point where the ice jammed sufficiently to stop.

Large lake ice scours have been reported from water depths from 13-25 m in Lake Erie (Grass, 1983) and in very shallow water (less than about 1 m) in Great Slave Lake (Weber, 1958). This range encompasses the estimated depth of about 9 m at the scour sites in our study area during the Sydney phase. The island at the higher elevation may have acted as a centre for ice jamming, encouraging the pile-up of the moving lake ice sheet and causing a greater number of scours here (see also Grass, 1983). Some overtopping of the island above the water level by ice pile-up may have occurred, creating the smaller scours on its upper surface. During the open-water season, the island may also have protected the scours from erosion by waves created by the prevailing northeast winds. Grass (1983) reports that scours in Lake Erie are preserved through the summer only in protected areas.

Ice scouring may have lasted for only one year or a few years before the water level again fell rapidly leaving the land surface well above water level and the possibility of eradication of the scours by waves.

☑ STOP 5 Ridged glaciolacustrine clay terrain near Kingston, Frontenac County

Willem J. Vreeken

NTS 31C/7, UTM 800050

Follow Division Street 1.2 km north from Highway 401. An overview of the features may be had from the roadside as it descends into Little Cataraqui valley. To walk among them, turn left (west) into Little Cataraqui Conservation Area where parking is available. A fee is charged for admission to the Area and the interpretive centre.

Distribution, external and internal form, depositional environment, and age.

Limestone uplands between Belleville and Kingston, known as the Napanee plains, have a thin (<1.5 m) mantle of rhythmically laminated, upward-fining clay and silt, attributed to glacial Lake Iroquois, which existed between 12 400 and 11 800 BP. In the southwesterly oriented intervening valleys, these sediments are up to 30 m thick and commonly have inclusions of glaciofluvial sands and gravels with evidence of southwesterly paleocurrents. These thick-clay bottomlands are commonly marked by distinct subparallel, sinuous, flat-crested ridges (Fig. 22 and 23), many of which bifurcate or join in a pattern resembling linear current-ripple bedforms. Essentially symmetrical in transverse profile and with a wavy crestline, they are up to 6 m (normally 2 to 3 m) high, 300 m long, and 30 to 100 m apart. The ridges are oriented northwest-southeast, perpendicular to the trend of the bedrock valleys, glacial striae, and subglacial meltwater erosion marks (S-forms - Kor et al., 1991; Fig. 23).

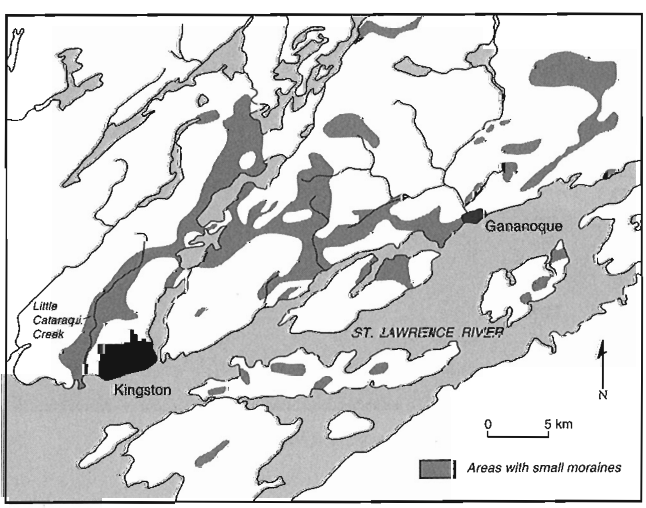


Figure 22. Distribution of ridged glaciolacustrine clay terrain in the vicinity of Kingston, Ontario (from Løken and Leahy, 1964).

Patches of ridged-clay terrain occupy a 50-m elevation range above the level of Lake Ontario and the St. Lawrence River, up to 120 m a.s.l. The continuous ridged terrain of the Little Cataraqui valley rises 25 m to the northeast, with an average gradient of 0.2%, over a distance of 12 km. The weakly undulating floor of swales between the ridges is occupied by small seasonal or perennial ponds in which through-flowing runoff is rare. Valley thalwegs are occupied by vastly underfit meandering creeks without evidence of significant postglacial downcutting or lateral migration. Dissolved load accounts for more than 90% of sediment transport in the streams (Ongley, 1973). In summary, these clay landscapes have experienced little postglacial geomorphic change and their landforms mainly reflect late glacial and deglacial processes.

Observations on internal form pertain to the composition and deformation of the sediments. The bedrock pavement beneath has striae, flutes, and S-forms and is locally separated by diamicton or glaciofluvial sands and gravels from rhythmically laminated silts and clays with dropstones. Rhythmic beds are inferred to be annual: the summer layers commonly are pale, silt-dominated, multiply-graded units with several, simple depositional cycles, each containing up to ten graded laminae; these beds thin upwards. Winter layers are of constant thickness and have dark clay-dominated material which is less calcareous than the summer deposits (Naldrett, 1991). Bedding planes and partings in summer layers have small, faint locomotion or feeding traces resembling those created by oligochaetes or similar organisms. Gastropod shells from the uppermost deformed clays include *Succinea avara* (Say), a freshwater species living along streams and lake shores, currently distributed from the Gulf of Mexico to James Bay and climatically undiagnostic (Løken and Leahy, 1964). Rare pollen grains include *Pinus, Ambrosia, Picea, Larix, Dryopteris*, and *Quercus* (Naldrett, 1991), species that are also found in the first post-Iroquois pollen zone and, consequently, are of limited climatic significance.

Distortion features similar to those reported by Henderson (1967) and Sangrey (1970) are illustrated in Figure 24. A broadly arched sand zone occurs 2.5 m beneath the surface of the ridge and conformable with it. Superposed are many unidirectional high-angle thrust faults, with displacements from 1 to 13 cm, that suggest thrusting to the northeastward, perpendicular to the ridge. A borehole transect across a second ridge revealed a sand bed at similar depth deformed to conform to the ground surface. Miniature faults in a third ridge showed horst-and-graben structure without a unidirectional lateral thrust. Dense laminated clays less than 1.5 m deep, in a fourth ridge, displayed diapiric deformation and low-dip thrust faults (overthrusts). Geotechnical data reported by Sangrey (1970) from the Great Cataraqui and

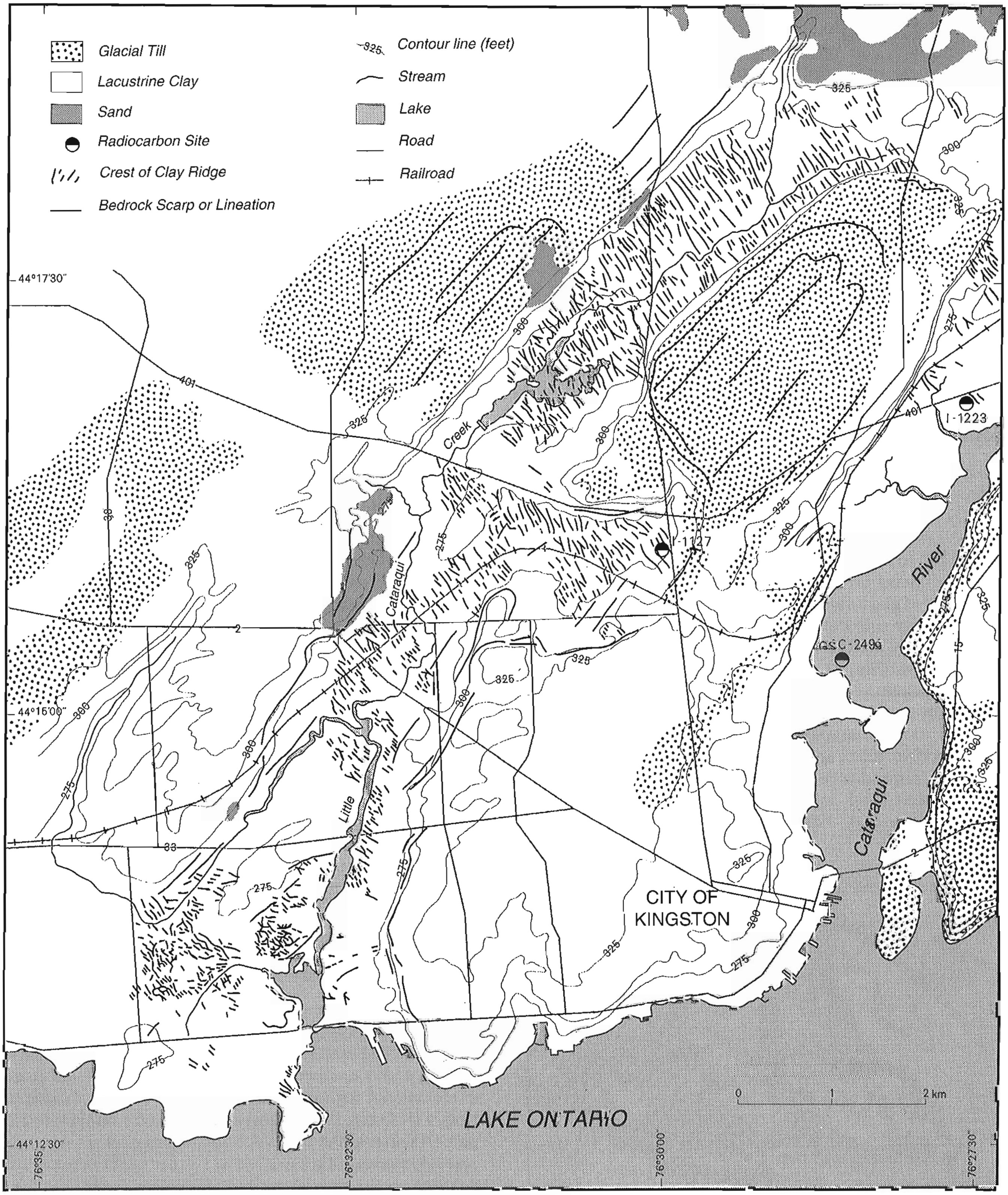


Figure 23. Distribution of clay ridges in the Little Cataraqui Creek valley near Kingston, Ontario (W.J. Vreeken, unpub. data, 1992).

Collins Creek valleys indicate high preconsolidation stresses for the stiff distorted clays. This implies that an overlying glacier lay on the ridges, producing the observed deformation structures.

Radiocarbon dates from shells in the distorted clays do not correspond to the age of Lake Iroquois sediments. One date (I-1127) is older than 18 000 BP and the other 10 050 ± 390 BP (GSC-314), is younger than many basal dates from postglacial sediment in the region (Vreeken, 1981). More detailed dating is necessary to explain this anomaly.

Little Cataraqui Creek Conservation Area

The valley of Little Cataraqui Creek (Fig. 23) has linear scarp-forming flanks cut into Ordovician limestone. The bedrock surface is marked by northeast-southwest oriented striae and S-forms that include a 5 m deep, 10 m wide channel at the foot of the northern flank. The bottomland is underlain by about 22 m of glaciolacustrine clays and silts on thin glaciofluvial sands and gravels that form an artesian aquifer above bedrock. Sands and gravels with southwest-dipping foreset laminae occupied the minor bedrock channel mentioned above and also are exposed in pits elsewhere along the northern valley flank. The sand body wedges out into the clay body, indicating that they are coeval deposits. The bottomland is marked by prominent regular clay ridges. The artificial lake which dates to the early 1970s is clearly visible from the overview on Division Street or from within the Conservation Area. Its irregular shoreline reflects topographic details of the valley bottom. Figure 24b depicts displacements observed when the foundations for a building were being laid, the borehole transect (Fig. 24a) refers to the second ridge (B) northeast of the building.

Origin of clay ridges in the Kingston area

Løken and Leahy (1964) proposed that the ridges were formed as lake bottom clays were squeezed up into crevasses along the bottom of ice that readvanced into glacial Lake

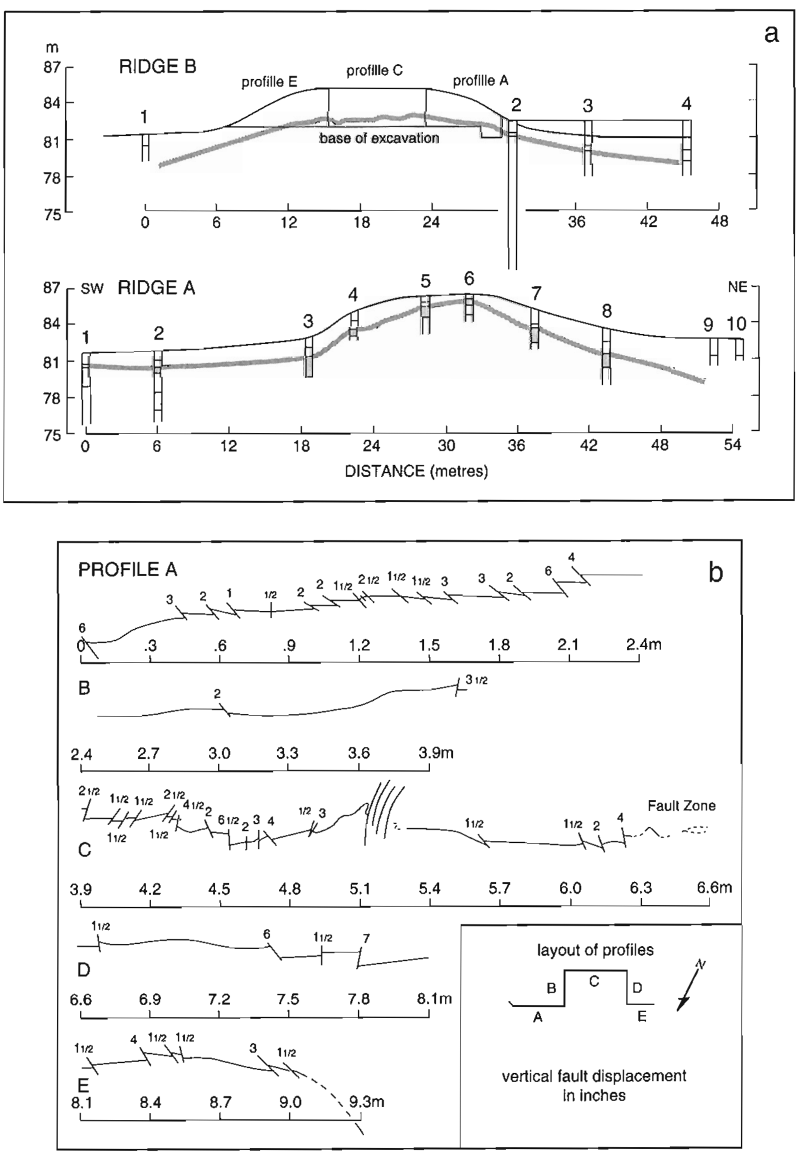


Figure 24.

***a)** Transverse sections across two clay ridges in Little Cataraqui Creek Conservation Area, showing distribution of subsurface sand zone (dotted line). **b)** Faults and folds in sand zone exposed in the foundation pit for the Conservation office (W.J. Vreeken, unpub. data, 1992).*

Iroquois. Henderson (1967) accepted the hypothesis of ice-pressing but not the idea of ice advance because he did not find any till deposits on the ridges or on adjacent unridged lacustrine sediments. He proposed the ridges were formed when crevassed floating ice, unbroken and still to some extent attached to land-based ice to the northeast, was lowered onto the clays, probably during partial drainage of the lake. Sangrey (1970) proposed that overconsolidation of clays from ridged clay terrain reflected a readvance of partly buoyed but not free floating glacial ice.

The regular surface morphology of the ridged clay terrain, the absence of overlying till, and the absence of unidirectional southwest-oriented internal deformations indicate that these landforms did not result from glacial overriding per se. By default, an ice-pressed origin involving stagnant ice must be invoked and Henderson's explanation of ice lowering during drainage of the lake seems plausible. But the regular pattern of subparallel and in places bifurcating ridges need not reflect a pattern of tension crevasses at the base of the ice invoked by Løken and Leahy (1964) and by Henderson (1967). Because the glacier is laterally unconstrained, unlike the situation of alpine glaciers, it is likely that such a pattern would have included many perpendicular and oblique crevasses, whereas ridges of such orientation are uncommon (Fig. 23). Alternatively, the ridge pattern could reflect the basal configuration of noncrevassed ice. Ice ripples, erosional forms sculpted into the base of river ice by subice waterflow, in patterns resembling those of current-rippled fluvial bedforms, have been reported by Ashton and Kennedy (1972). Such forms could have developed on the sole of glacier ice during subglacial meltwater discharge. They would have been pressed onto the soft substrate when the ice was lowered during drainage of the lake. Sangrey's (1970) preconsolidation stress estimates suggest that the thickness of ice that would have pressed the sediments upon drainage ranged from 19 to 35 m, with a differential of 16 m. Considering that the maximum height of the clay ridges (6 m) could represent a minimum estimate for the relief on the base of the ice if the ridges were formed under conditions of confined compression, this would leave a maximum topographic relief of 10 m on the ice surface.

☑ STOP 6 Large-scale bedrock fluting, Elginburg

John Shaw

NTS 31C/7, UTM 740082

Follow Highway 38 1.2 km northwestward from Highway 401, then turn north (right) on Cordukes Road for 2.6 km to the T-junction at the Elginburg Road. Large-scale flutings are seen in the rock cut 150 m west of Cordukes Road. The largest is located just east of the house on the south side of Elginburg Road.

Relatively low, lithologically controlled escarpments form the risers of the broad terraces in the predominantly bedrock landscape between anastomosing channels of the Kingston area (Fig. 25). Northeastward facing escarpments are shaped into prominent noses pointing into the former ice flow. Large-scale fluting of the bedrock extends southwesterly downflow from many of these escarpments. It is easily mapped from aerial photographs and indicates a strongly unidirectional flow over the immediate Kingston area (Fig. 25). The origin of large-scale bedrock fluting at Elginburg is discussed at Stop 6.

Although such fluting is widely regarded as a product of direct glacial scour (Smith, 1948), similar fluting in areas subject to catastrophic flooding, but not directly affected by glacial abrasion, makes this conclusion far from certain (Bretz, 1969; Baker, 1978; Kehew and Lord, 1987). Indeed, careful examination of the large-scale fluting in the Kingston area leads to the conclusion that it, too, is fluvial rather than glacial (Shaw, 1988).

Fluting occurs on relatively flat surfaces as a result of the erosion of furrows into the bedrock. The furrows separate remnant bedrock ridges that commonly taper downflow to the southwest (Fig. 26). Furrow rims are generally sharp and, where they close in the upflow direction, the fluting resembles the open flutes described by Kor et al. (1991). Elsewhere, the flutes wrap around the upstream ends of the escarpment noses to form horseshoe scours which probably relate to horseshoe vortices in the formative flow (Sharpe and Shaw, 1989). Flutings, which may be as much as 10 m wide and 1 m deep, show a smoothly curving cross-section, in some cases symmetrical and elsewhere asymmetrical. There are both types at the Elginburg site. Meltwater flutings of smaller scale show similar cross-sections (Shaw, 1988). Some large-scale furrows show secondary, lateral furrows similar to those alongside classical flutes sculpted by turbidity currents (Allen, 1982).

Such form analogies support a meltwater origin for the bedrock fluting. This interpretation is strengthened by evidence from the distribution of glacial striae relative to fluting. Where fluting is aligned obliquely to striae, the bedrock immediately to the lee of the fluting crest is either unstriated or weakly striated. A good example of this is seen at Stop 7.

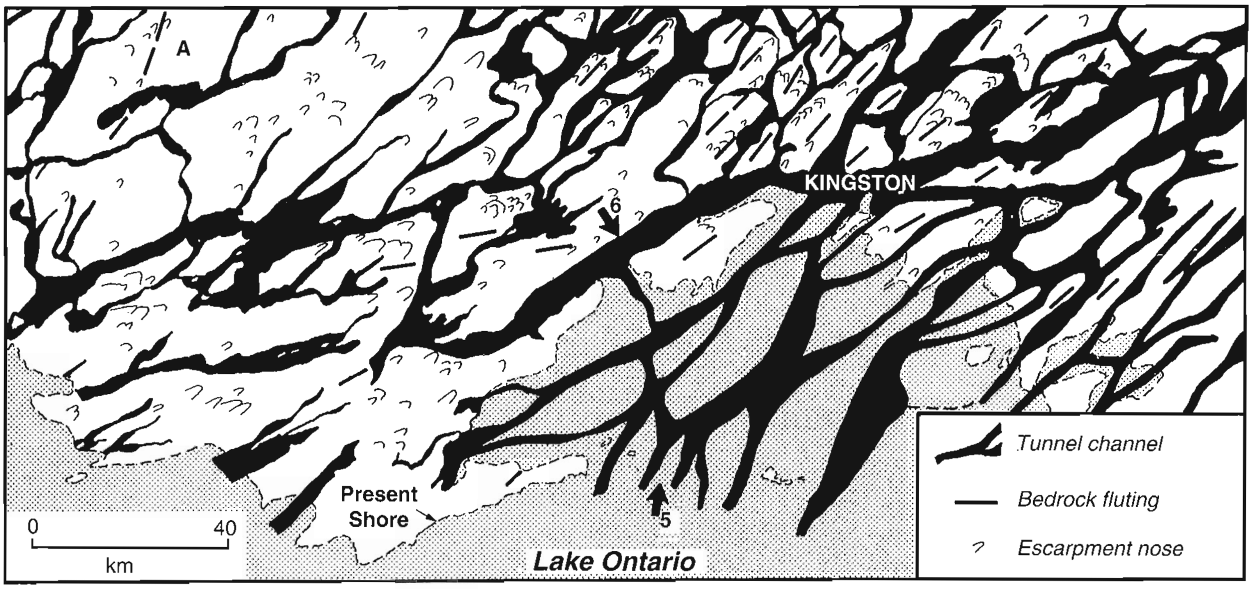


Figure 25. Subglacial channels, bedrock fluting, and escarpment noses around Kingston. Arrow 6 = North Channel; arrow 5 = area with submerged escarpment noses.

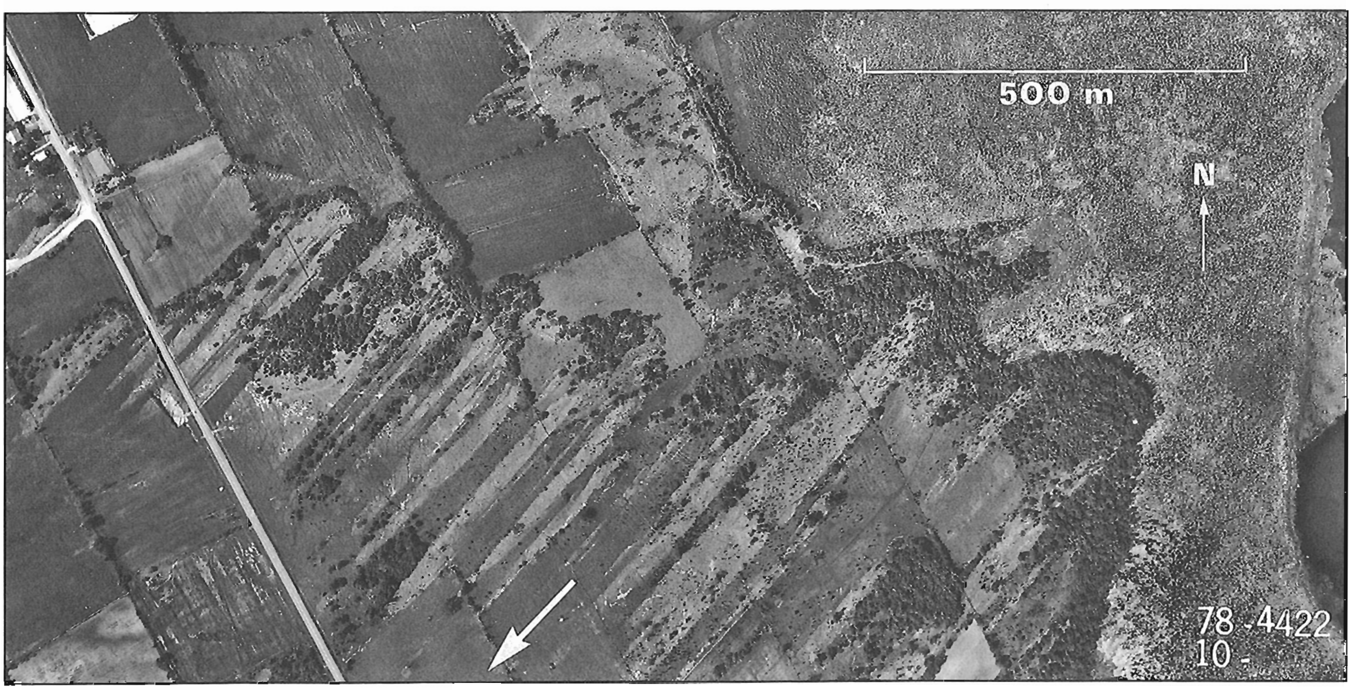


Figure 26. Bedrock fluting downstream from an escarpment with flow noses. Light areas between the flutes are remnant ridges. Ontario Government air photograph 78-4422 379.

It is most unlikely that an unstriated portion of a furrow was eroded by glacial abrasion. By the same token, if glacial abrasion were the primary erosional mechanism, the heavily striated areas of the bedrock should be the most deeply incised. Because this is not the case, it is reasonable to suppose that furrows were eroded by turbulent meltwater; striae simply represent minor surface ornamentation by the glacier as it made contact with parts of the bed following a meltwater drainage event.

The widespread distribution of fluting between the deeply incised tunnel channels (Fig. 25) and its unidirectional trend indicate that the channels were unable to cope with some enormous discharge and consequently a broad interchannel area was also inundated below a vast subglacial flood (Shaw and Gilbert, 1990; Ontarian event Fig. 27). Study of erosional marks around Georgian Bay and drumlin fields formed beneath the Laurentide Ice Sheet also gives evidence for numerous subglacial outburst floods (Shaw and Sharpe, 1987a, b; Shaw et al., 1991; Kor et al., 1991).

The complementary relationship between large-scale fluting, bedrock erosion, and drumlins is discussed in the interpretation of the drumlin at Stop 8.

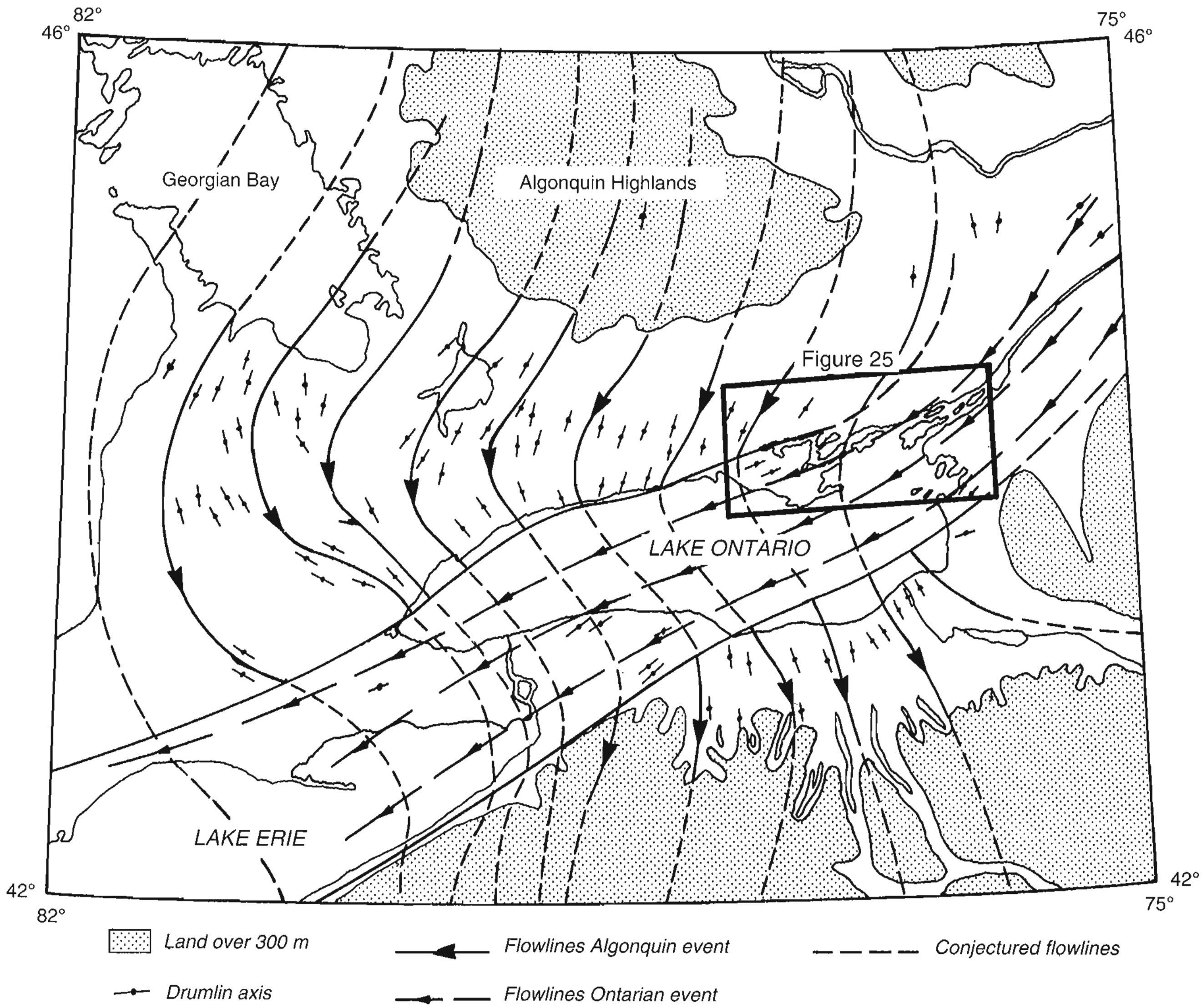


Figure 27. *Regional flow patterns in southern Ontario and northern New York State of subglacial meltwater flood events interpreted from the orientation of drumlin long axes (after Shaw and Gilbert, 1990). Location of Figure 25 is shown.*

☑ *STOP 7* Meltwater erosional marks, Wilton Creek (Thorpe Pit)

John Shaw

NTS 31C/7, UTM 582049

Follow Highway 133 north 0.9 km from Highway 401 across Wilton Creek to the first turn on the right (east) and continue across the creek again. The erosional marks are in an old sand pit southwest of the road. To follow the sequence of this guide, enter through the gate at the top of the rise from the valley floor beyond the Wilton Creek bridge.

The Thorpe site is notable for the variety of erosional marks cut mainly into limestone. The character of the marks changes with location relative to the valley (Fig. 28). Preservation is excellent because the rock surface has only recently been exhumed from beneath glaciofluvial deposits.

Upper benches

The upper benches on the shoulder of the valley carry large-scale flutes with sharp rims (Fig. 29). These flutes are part of a large fluting field (Stop 6, Fig. 26) that includes a particularly well-developed set in the dividing median of Highway 401 where the highway dips into Wilton Creek valley, just east of the Highway 133 junction (Fig. 30; Note that stopping on Highway 401 is dangerous and illegal except in an emergency). At the Thorpe site flutes trend at 056° to 236°, whereas striae were formed by ice flowing toward 248°. In contrast with the Marysville site (Stop 12), there is only one set of striae at the Thorpe site.

The most prominent flute on the broadest bench has rims which close in the upflow (northeast) direction; one rim is better developed and more extensive than the other. Inset, lateral flutes give the extensive rim a broad, scalloped form (Fig. 31). Striae are less well developed in the flutes than on the adjacent rock and the flute surface is unstriated immediately to the lee side of the rims.

In addition to the large-scale flutes, the upper benches show remarkably straight grooves, many of which are paired, and range in width from a few millimetres to centimetres (Fig. 32). The larger grooves may be traced for several metres (Fig. 32a) and the smaller ones are normally less than a metre long. Unlike striae, which have a very rough surface texture, grooves have smooth floors and sharp, unbroken rims. Where paired grooves can be traced to their upstream (northeasterly) limit, they invariably join to form a crescentic scour. Many of the crescentic scours are enlarged to form "thumbprint" pits with relatively sharp proximal rims (Fig. 32b). The straight grooves and striae are parallel.

Large-scale flutes at Elginburg (Stop 6) are interpreted as meltwater forms partly on the basis of evidence at the Thorpe site. A strong incentive for this interpretation is the morphological similarity between large-scale flutes and other forms considered to be products of erosion by turbulent flows: sharp rims closing upstream, smoothly curving cross-sections, occurrence in groups, inset lateral flutes, and broadening and loss of definition downflow (Allen, 1982; Kor et al., 1991). Evidence from striae provides a second argument in favour of this interpretation; it seems most unlikely that glacial

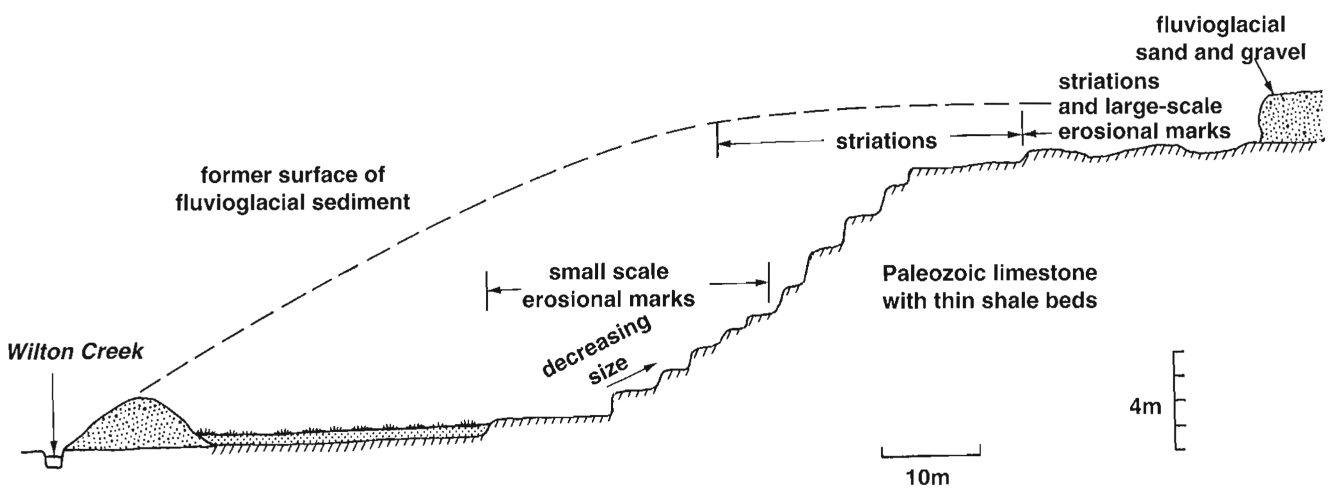


Figure 28. Cross-section of Wilton Creek valley showing the locations of striations and erosional marks and the former extent of glaciofluvial sediment (after Shaw, 1988).

abrasion could have formed the flutings when parts of the flute surfaces are unstriated. Consequently, a meltwater rather than a glacial process is preferred for flute formation.

The surface texture contrast between the straight grooves and striae also suggests that these two erosional features were formed differently. The striae have the characteristics of erosion by individual tools and are confidently ascribed to glacial abrasion. On the other hand, the crescentic scours at the upstream end of paired grooves are best explained as products of horseshoe vortices around small obstacles (Sharpe and Shaw, 1989). A similar explanation is discussed in more detail for erosional remnant ridges (rat tails) at Stop 12.

The small depth, straightness, and extent of the grooves suggest formation under thin sheets of high-velocity meltwater; only a limited water layer thickness is required to accommodate the small diameter vortices that are thought to have produced the grooves, and their downstream extent suggests a high Reynolds number with inertial forces dominating over viscous resistance. For shallow depths, high Reynolds numbers imply high velocity.

The combination of erosional marks on the valley shoulder may be used to speculate about the events that produced them. The large-scale flutes are part of an extensive field, indicating that a regional-scale subglacial flood overspilled the valley sides and submerged the high ground. Such flows were relatively deep, at least deep enough to contain the large vortices that eroded the flutes (maximum diameter about 10 m). In the immediate vicinity of Wilton Creek valley at Thorpe, the large-scale flutes parallel the valley trend.

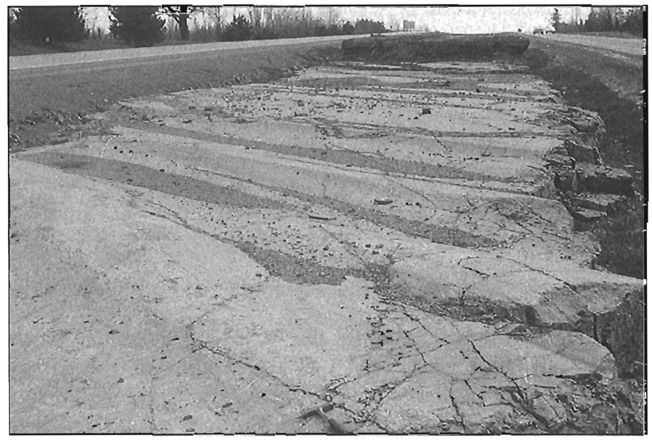


Figure 30. *Flutes within the central median of Highway 401 near Wilton Creek. GSC 1993-164E*

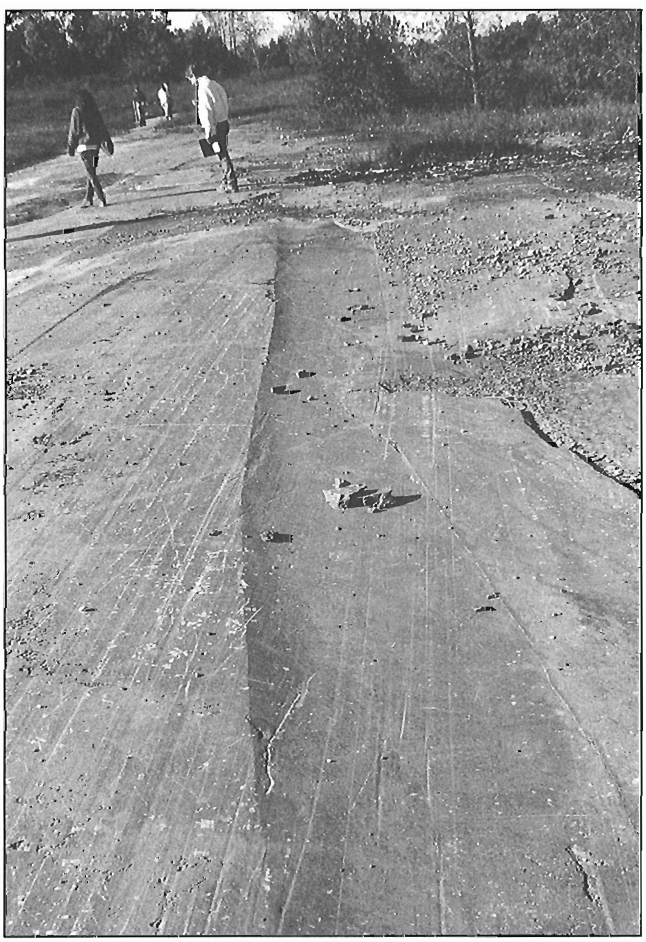


Figure 29. *Large-scale flute on an upper step, Thorpe site, Wilton Creek. GSC 1993-164D*

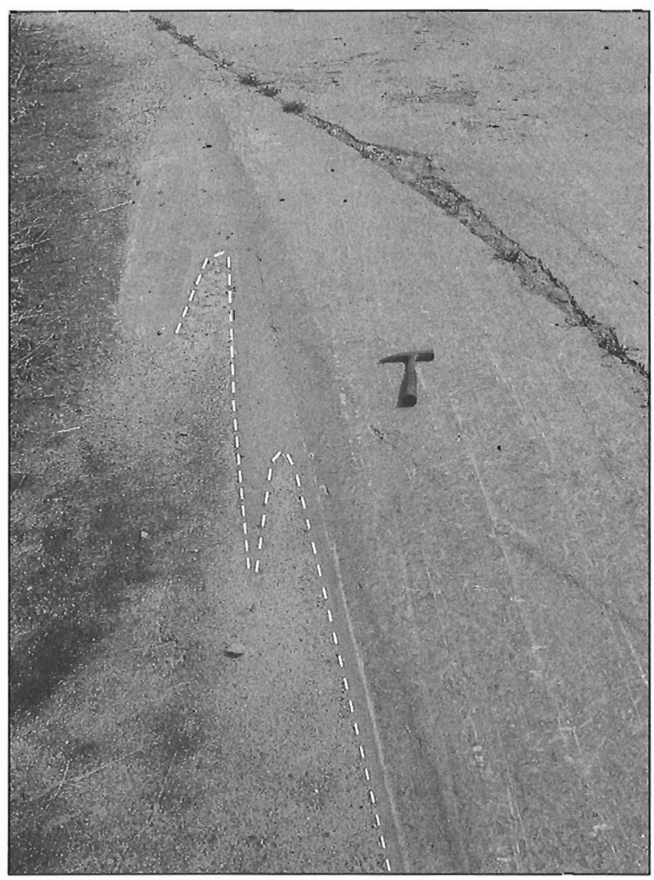


Figure 31. *Inset flutes and scalloped flute margins, Wilton Creek. GSC 1993-164F*

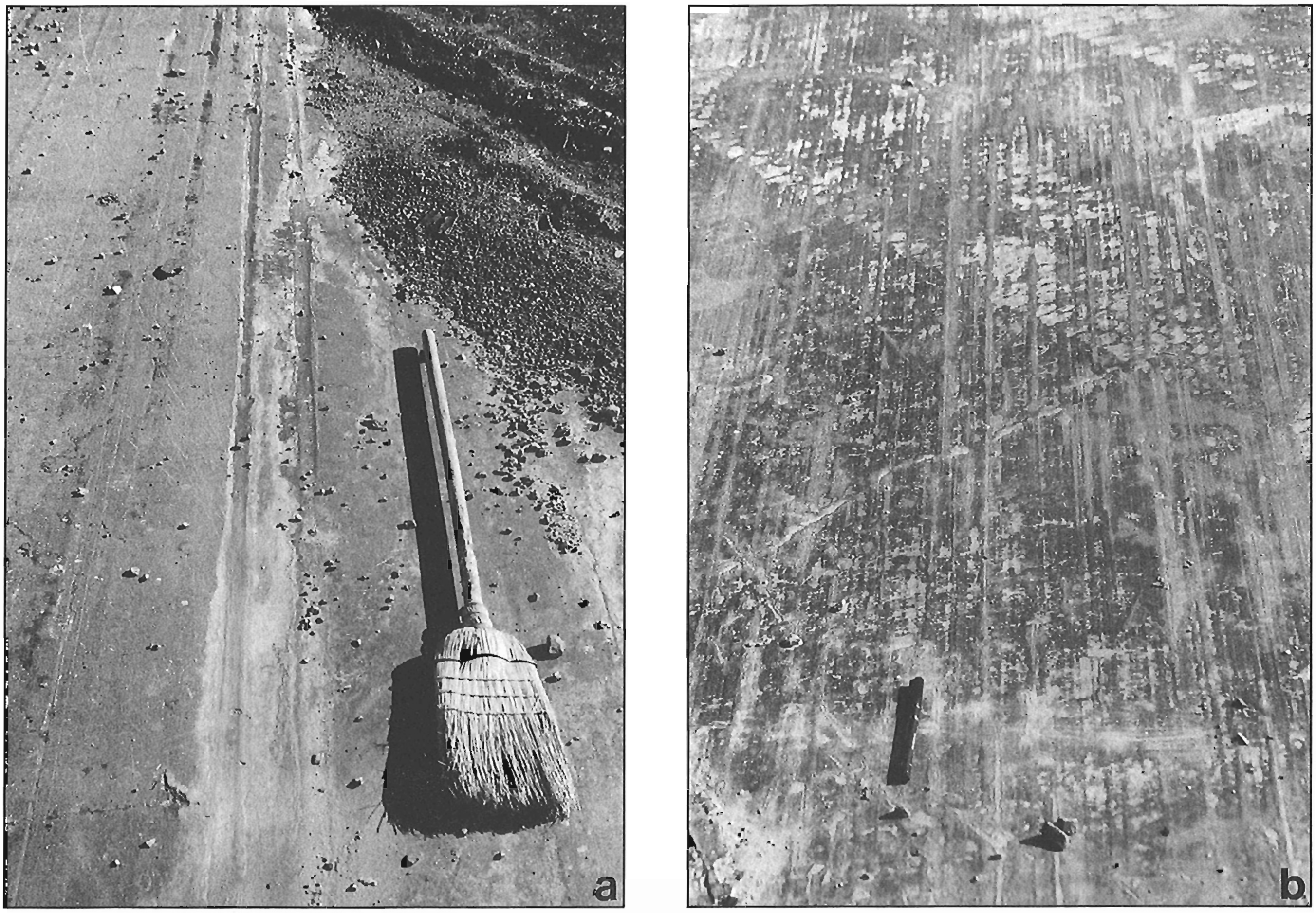


*Figure 32. Parallel furrows, Wilton Creek: **a**) intermediate scale paired furrows; **b**) small-scale furrows with crescentic closure and thumbprint scours at the upstream end. GSC 1993-164G, 164H*

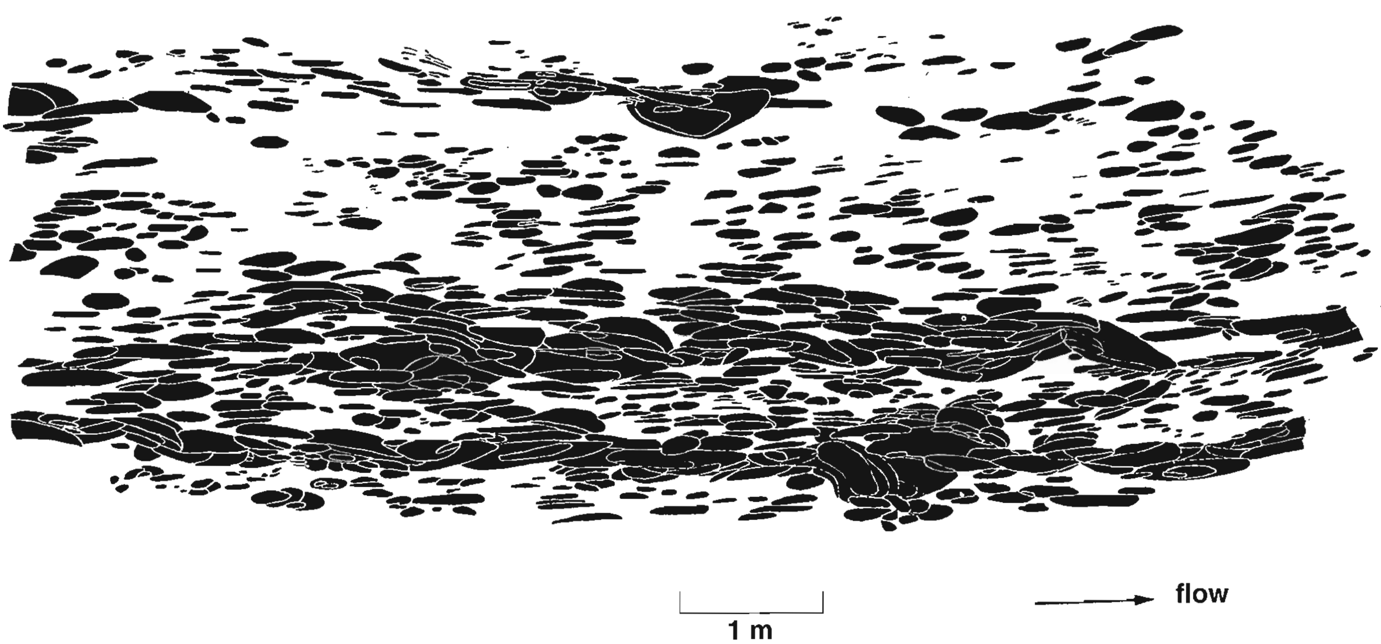


Figure 33. Map of small scale erosional marks on a lower step, Wilton Creek (after Shaw, 1988). See Figure 28 for location.

Subsequent striation and smaller-scale meltwater forms deviate by about 10° from this direction, suggesting that these small-scale features were formed later when subglacial meltwater and ice-flow directions were coincident. It is simplest to assume that the small-scale forms represent a decreasing gap width between the ice and its bed. Finally, with reattachment as meltwater supply decreased, flow was ultimately confined to the tunnel channels during the waning stages of an outburst flood, probably the Ontarian event of Shaw and Gilbert (1990).

Intermediate benches and valley floor

Benches at an intermediate level on the valley side are striated but do not carry meltwater erosional marks. At about 4 m above the lowest exposed rock bench there are ill-defined indentations together with striae on both the risers and treads of rock steps (Fig. 28). The steps themselves result from preferential erosion of shales interbedded with carbonates. The size of the indentations increases toward the valley floor: the uppermost forms are commonly 2-3 cm long; those on the lowest step may be 40-50 cm. The indentations also become more obviously recognizable as erosional marks and cover more of the bed toward the valley floor. There is only a very limited zone of overlap between the erosional marked and striated bed. Below this, there is no striation.

It is difficult to do justice to the beauty and complexity of these erosional marks. The majority are closed spindles completely enclosed by sharp rims (Shaw, 1988; Kor et al., 1991). The spatial complexity of the erosional features is seen in a map compiled from vertical stereo photographs of the rock step with the best developed features (Fig. 33).

The simplest forms have straight longitudinal axes and are usually asymmetric (Fig. 34a); one rim is more curved than the other and the side of the flute below the curved rim is gentler than the side below the straight rim. Besides asymmetry in cross-section and about longitudinal axes, the upstream ends of most flutes are slightly more pointed than the downstream ends. In places, where these straight flutes intersect, they virtually occupy the whole bed. Straight flutes are from 15 to 30 cm long, and have maximum widths and depths of about 7 cm and 3 cm, respectively.

Where the density of erosional marks becomes particularly high in two clearly defined, flow-wise linear bands, spindle axes curve and are usually concave toward the axes of the bands (Fig. 34b, c). The curved spindle forms are generally deeper and longer than straight forms. They also have more pronounced cross-sectional asymmetry, being steeper, even overhanging in places, on the outside of the curve. They are also, like the straight flutes, more pointed at their upstream ends. These forms invariably intersect in a fashion indicating superposition of the formative flow structure of an upstream form over that for the downstream form it intersects. It appears as though flow features responsible for the flutes were responding to relatively local conditions that also caused high intensity erosion.

This intertwining of flutes is most extreme in braided flutes (Fig. 35). These involve curved erosional marks in which successive flutes have an opposite sense of curvature, and it is clear that each flute resulted from a flow structure that originated in the flow structure over the flute downstream. Braided flutes are deeper than other small-scale flutes (less than 20 cm) and their outer sides are invariably

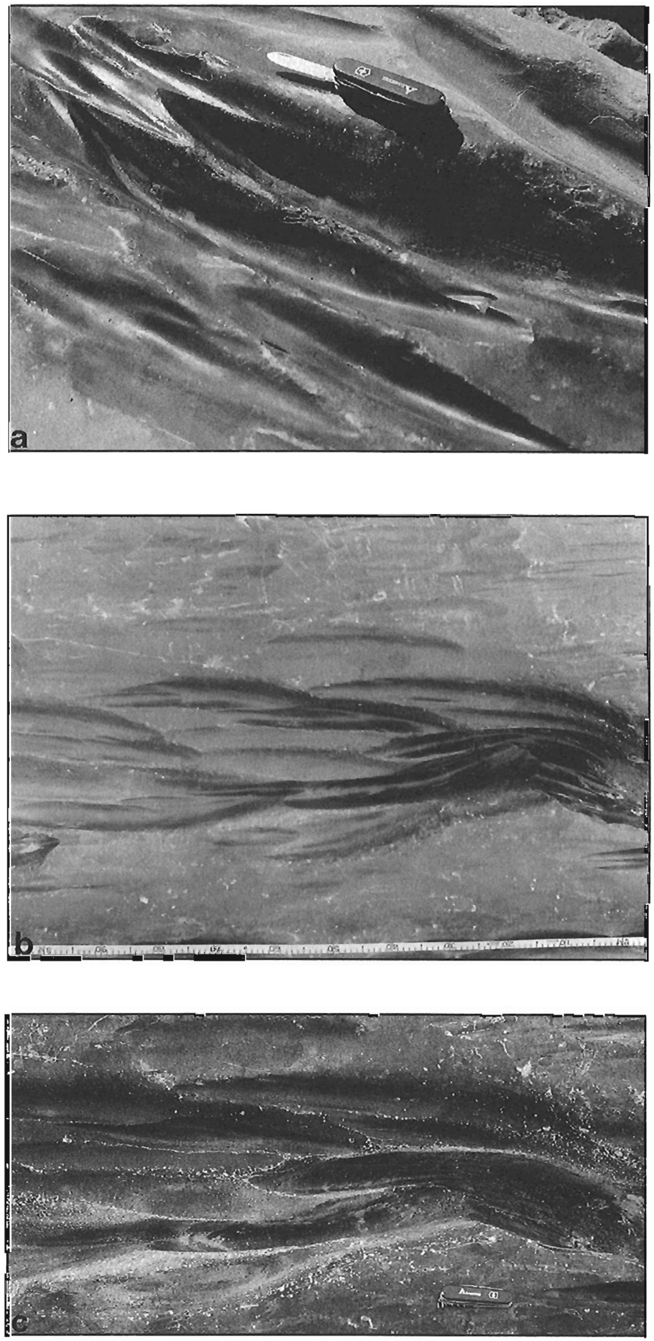


Figure 34. *Spindle flutes:* ***a)*** *straight,* ***b)*** *curved, and* ***c)*** *twined. GSC 1993-164I, 164J, 164K*

overhanging. Individual marks are up to 40 cm long, although they are commonly compound features with smaller flutes inset in the main scours.

Although flutes are the most striking features on the bed, close inspection shows that it is also scalloped into a series of broad, shallow hollows defined by relatively sharp rims. Straight spindle forms are commonly clustered in scallops (Fig. 36; Allen, 1982).

It is already noted that the most intense erosion was concentrated in two flow-parallel bands (Fig. 33). Relatively deep troughs occur along these bands, and individual flutes bordering these troughs appear to converge at nodes beyond which erosion was less intense. A pothole is located at one such node (Fig. 37).

The small-scale flutes evidently owe their existence to streamwise flow structures that cut downward and laterally into the bed. Unlike the horseshoe scours producing the rat tails described at Stop 12, these flow structures cannot be related to obstacles on the bed; they appear to have impinged on the bed, then separated from it. Because the erosion occurred selectively at specific points, the flow structures that created them must have been relatively stationary. As well, the formative structures must have been involved in a complex intertwining to form the braided flutes. These requirements severely limit explanations and the only plausible interpretation seems to be that the spindle flutes were formed where vortices played on the bed (Fig. 38). Straight spindles represent the case of flow-parallel vortices, curved spindles represent the impingement of curved vortices, and braided spindles represent intertwined vortices. The number of vortices involved in twining is strictly limited by the bed geometry and the requirement that adjacent vortices have an opposite sense of rotation; two vortices intertwined would produce flutes with the same sense of curvature; three vortices must involve at least two adjacent vortices with the same sense of rotation, a situation that leads to vortex destruction. Thus, the simplest explanation for the braided flutes involves four vortices.

Figure 36. *Scallops with inset flutes. GSC 1993-164M*

Figure 35. *Braided flutes. GSC 1993-164L*

Figure 37. *Pothole (p) at downstream termination of trough and convergence nodes for flutes (c). GSC 1993-164N*

A number of mechanisms may have generated the vortices producing flutes, all of which are likely in the subglacial environment with its abundance of bed irregularities and potential obstacles in the form of boulders on the bed and projecting from the ice (Hjulström, 1935; Shaw, 1988). Of particular interest is the clustering of spindle forms in scallops (Fig. 39). Taylor-Görtler vortices are known to develop over concavities in the bed as fast-flowing outer fluid is accelerated centrifugally and displaces slower-moving fluid near the bed (Tani, 1962; Floryan and Saric, 1982). The scallops which contain flutes may themselves have been the concavities that generated the vortices. Allen (1971) made a similar suggestion for small-scale flutes superimposed on transverse bedforms.

The glacier appears not to have abraded the valley floor as occurred on the upper benches after the meltwater event. The sand and gravel that were removed from the pit may have been deposited subglacially and the glacier came to rest on it. Alteratively, the glacier may have continued to float over the valley floor in the vicinity of the Thorpe section; at a second site, the Wilton Ball Park, striations indicate that the glacier settled to the valley bed after being separated from it by a meltwater event. Glaciogenic sand and gravel, similar to those at Thorpe, also covered the bedrock floor of Wilton Creek valley at the Ball Park, suggesting that their deposition was not directly related to the flow events producing the erosional marks. It is more likely that the sand and gravel are equivalent in chronological sequence to the deposition of eskers in other tunnel valleys and represent a relatively late stage of deglaciation (Shaw and Gorrell, 1991). This interpretation, although largely speculative, is in keeping with the conclusion that the major meltwater outburst floods, the Algonquin and Ontarian events, happened when the ice front was close to its maximum position (Shaw and Gilbert, 1990).

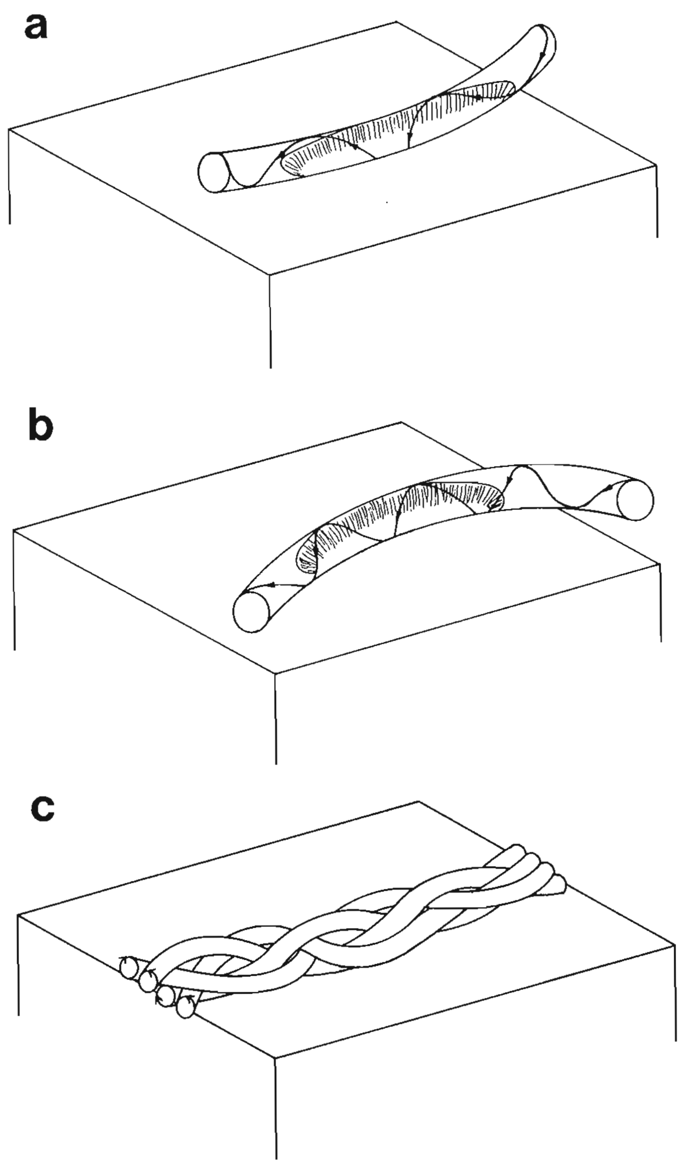


*Figure 38. Vortices impinging on the bed to produce flutes. **a**) left-hand spiral, **b**) right-hand spiral, **c**) intertwined spirals in three dimensional meanders.*

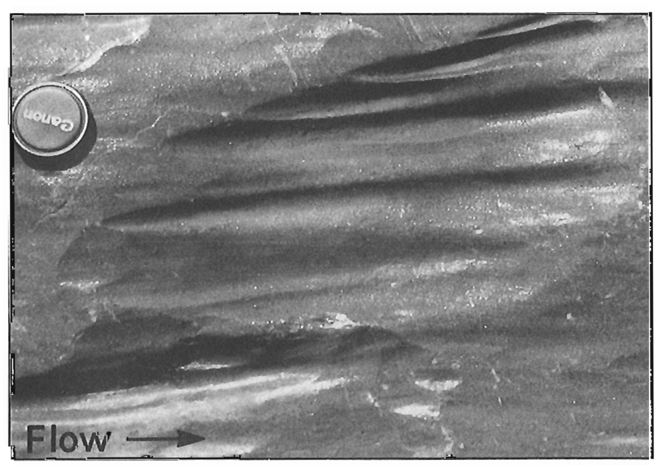


Figure 39. Clustered spindle flutes in a scallop. GSC 1993-164O

☑ STOP 8 The Kimmett Drumlin at Camden East

John Shaw

NTS 31C/7, UTM 555125

Follow Highway 133 8.6 km northwest from Highway 401 through the village of Camden East to the first intersection beyond. Turn right (northeast) and travel 2.5 km to a small pit in a drumlin to the north of the road.

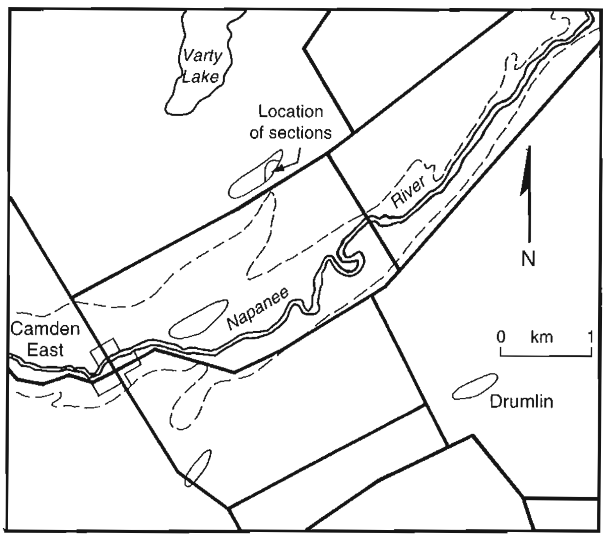


***Figure 40.** Location map, Kimmet drumlin. See Figure 2 (Stop 8) for location (from Shaw and Sharpe, 1987a).*

The Kimmett drumlin (Fig. 40) lies in the extreme northeastern part of the Belleville drumlin field (Fig. 2) which, in turn, is part of a larger system associated with a flow pattern extending along the axis of Lake Ontario (Fig. 27, Shaw and Gilbert, 1990). It stands on a broad, bedrock interfluve with Palaeozoic carbonates at, or close to, the surface. The bedrock is striated and carries meltwater erosional marks associated with regional-scale flooding (see Stop 6; Shaw and Gilbert, 1990).

The drumlin contains interbeds of diamicton and sorted, stratified sediment that are truncated by the landform surface (Fig. 41). The lower parts of the exposed sections are dominated by stratified deposits interbedded with thin diamicton units. Diamicton is dominant in the upper part of the section where the beds are relatively thick and associated stratified beds are thin and discontinuous.

Diamicton beds are clast-rich and the matrix is silty or sandy; sandier matrix is common in the thin diamicton beds intercalated with sand and gravel. Exotic clasts from the Canadian Shield are found in the diamicton together with local carbonate and shale clasts. Some large boulders, left in piles on the pit floor, show faceting and striation; several of these are "bullet-shaped" (Boulton, 1978) and many smaller clasts in the diamicton beds are also striated. Preferred orientations of clast long axes in the diamicton parallel the drumlin axes (Holden, 1988). The lower parts of some diamicton beds contain eroded clasts of underlying stratified sediments.

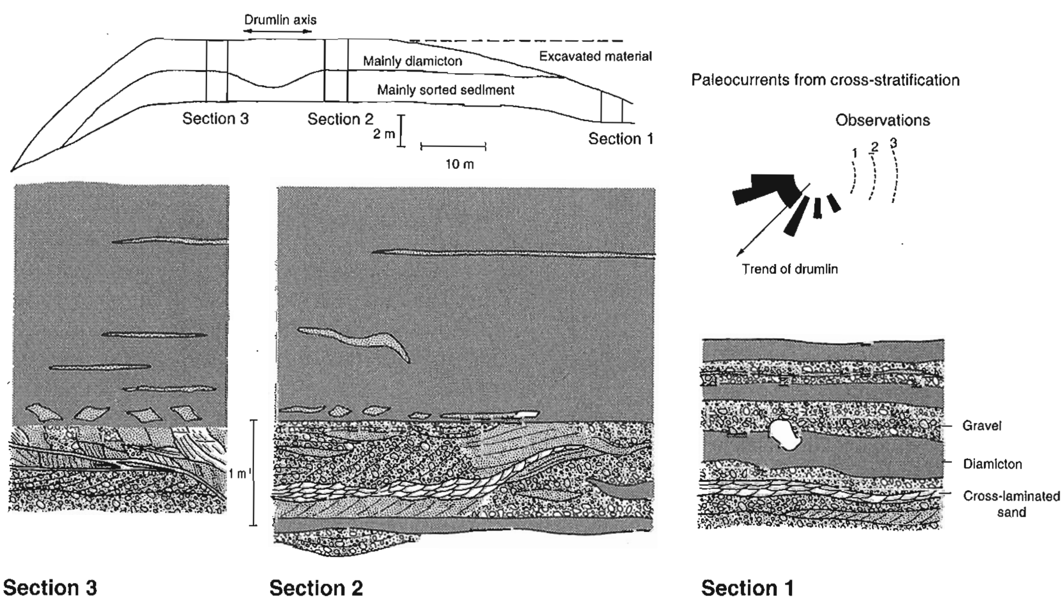


***Figure 41.** Cross-section of the Kimmet drumlin (from Shaw and Sharpe, 1987a).*

The sorted and stratified sediment ranges from silt to gravel. Gravel units commonly rest on an erosional surface. The silts and sands are parallel or crosslaminated. Gravel is either horizontally bedded with clast imbrication or cross-bedded. Flow directions estimated from crossbeds and cross-lamination are toward the southwest, dispersed about the axial trend of the drumlin.

Some stratified units beneath diamicton beds are faulted, folded, or attenuated. However, the broad stratification of the drumlin is intact and has not been subject to wholesale deformation.

The broad expanse of exposed bedrock around this drumlin and others in the vicinity and the truncated bedding suggest drumlin formation by erosion. The other possibility, that the drumlin was created by deformation of previously deposited material (Boyce and Eyles, 1991), is unlikely given the extent of exposed inter-drumlin bedrock. Such deformation cannot sweep bedrock clean of sediment. Undeformed, intratill stratified sediment less than a metre below the landform surface also speaks against deformation on the scale proposed for drumlin formation (Boulton, 1987; Boyce and Eyles, 1991).

Deposition of sediment in the drumlin appears to have involved running water to produce the sorted stratified sediment; thin diamictons with sandy matrix interbedded with sand and gravel are probably debris-flow deposits; bullet-shaped boulders suggest some deposition by lodgement, and clasts of sorted sediment within thick diamicton units with parallel preferred clast orientations relative to ice flow imply melt-out. Such complex depositional sequences are as expected for a subglacial setting in which the glacier is periodically decoupled from the bed as subglacial meltwater cavities are formed and deposition of cavity fills by running water and debris flows interrupts the deposition of till by lodgement, minor deformation, and meltout (Dreimanis et al., 1987; Shaw, 1987).

These processes in themselves could not have formed the drumlin; the cavity fills are much smaller than the landform and, besides, the truncated strata point to formation by erosion. The simplest inference is to presume that the process that produced the fluted bedrock also eroded the surficial sediments (Shaw and Sharpe, 1987b). By this explanation, drumlin fields and bedrock scour are complementary forms in southeastern Ontario; the exposed bedrock areas either representing areas of thinner primary sediment cover, or areas of more intense meltwater erosion.

☑ STOP 9 Holocene lacustrine marl and peat sediments at Dry Lake, Hastings County

Willem J. Vreeken

NTS 31C/6, UTM 322204

Travel 19.3 km north on Highway 41 from Highway 401 through Roblin. Turn left (west) on the Marlbank sideroad for 4.8 km. Turn left (west) on Esker Road which passes between Dry and Lime lakes and proceed to the pull-off on the right hand shoulder of the road at the west end of Dry Lake (1.5 km from the intersection). The remains of the marl processing plant can be seen in the woods on the south side of the road that passes to the north of Marl Lake.

Marl is an unconsolidated, highly calcareous lacustrine sediment, usually containing shells, variable amounts of organic matter, sometimes with clastic admixtures, and with pale colours, such as white, cream, and light greenish grey (Guillet, 1969). It is normally found with peat in calcareous, hard-water lake basins, fens, and marshes, in formerly glaciated areas with calcareous rocks nearby. Marl may be seen in the shallow waters of a number of lakes in the region, especially Camden Lake, Varty Lake, Loughborough Lake, Knowlton Lake, and Stoco Lake, however, one of the best examples is Dry Lake.

Dry Lake marl deposits

Dry Lake, in the headwaters of Parks Creek (Fig. 42), is one of several marl lakes (including Inglesby Lake, White Lake, and Lime Lake) that flank the Codrington esker and are sited in a tunnel valley. Removal of the peat cover of this formerly infilled lake during commercial marl dredging between 1891 and 1914 and during World War II, reactivated marl accumulation. The artificial lake has an area of 50 hectares, a maximum depth of 7.3 m (Fig. 43), an average depth of 3.5 m, and a volume of 1650 m^3. The marl is 6 to 10 m thick. The view across the lake, from the top of the esker, presents an impression of landscapes that, only 6000 years ago, characterized many wetlands in the region (Fig. 42).

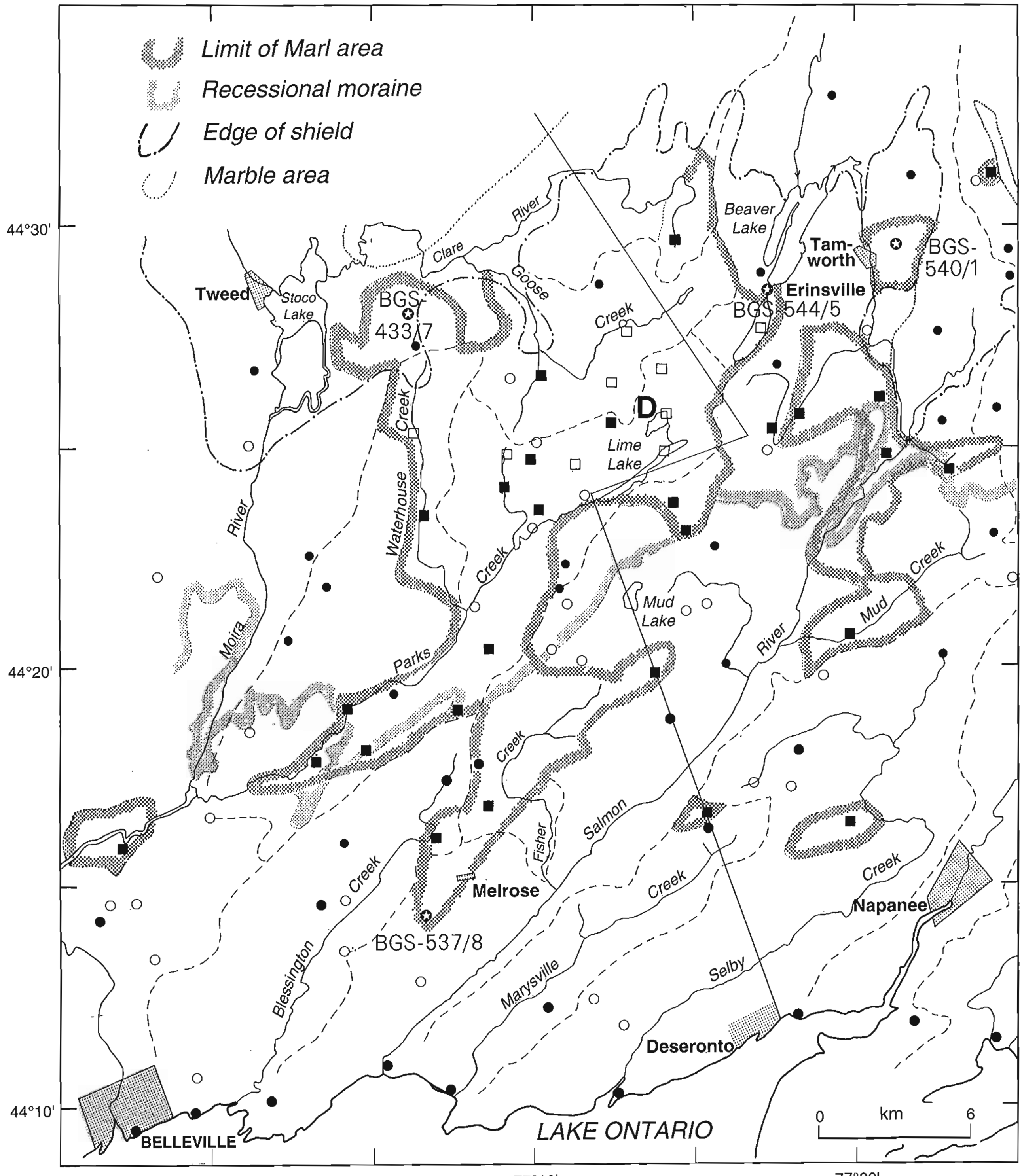


Figure 42. Distribution of reconnaissance sites, radiocarbon sites, and marl-producing drainage basins in the area between Kingston and Belleville, Ontario (from Vreeken, 1981). D = Dry Lake.

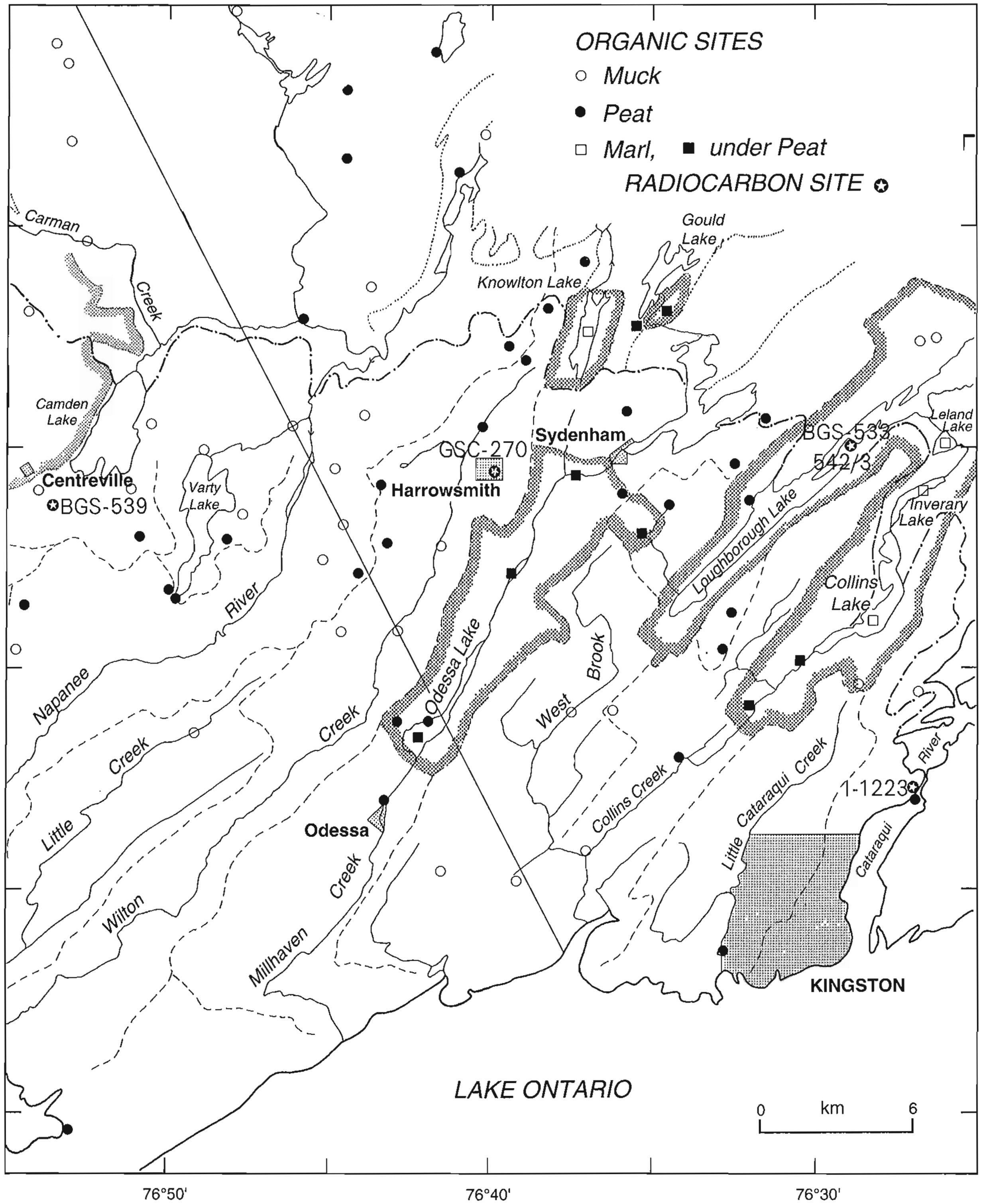


Figure 42. (cont.)

Marl was used in the manufacture of Portland cement and in industrial fillers, and in the production of paper, paint, plastic, rubber, and linoleum. Marl met most important requirements for fillers, namely, whiteness or brightness and particle sizes in the micrometre range. Disadvantages were that it had to be dried and that peat interlayers detracted from quality. In time, limestone became a preferred substitute for cement production and English Chalk became a preferred paper filler. Trans-Atlantic shipping of marl substitutes was stopped during World War II, in favour of strategically more important cargo, which led to temporary resumption of marl exploitation.

Ontogeny of hard-water calcareous lakes and lake marl

As discussed by Wetzel (1983), lakes tend to pass through a series of stages that mark the succession (ontogeny) of their ecosystems until they have filled and are incorporated into the terrestrial landscape. The general development is from low to higher productivity (from oligotrophic to eutrophic) but rates of filling with clastic or organic materials vary greatly. Low organic productivity rates in oligotrophic lakes, largely maintained by low inputs of inorganic nutrients from external sources, concomitant low rates of decomposition, and oxidizing conditions in the hypolimnion (the basal water stratum) result in low rates of nutrient release in a cyclical causal system. Low quantities of dissolved organic compounds also limit the availability of organic micronutrients. Under eutrophic conditions, the supply rates of inorganic nutrients, especially of phosphorus and combined nitrogen, are relatively high. As rates of photosynthetic productivity and organic loading to lower lake strata increase, nutrients are released from sediments into the hypolimnion, increasing nutrient recycling rates.

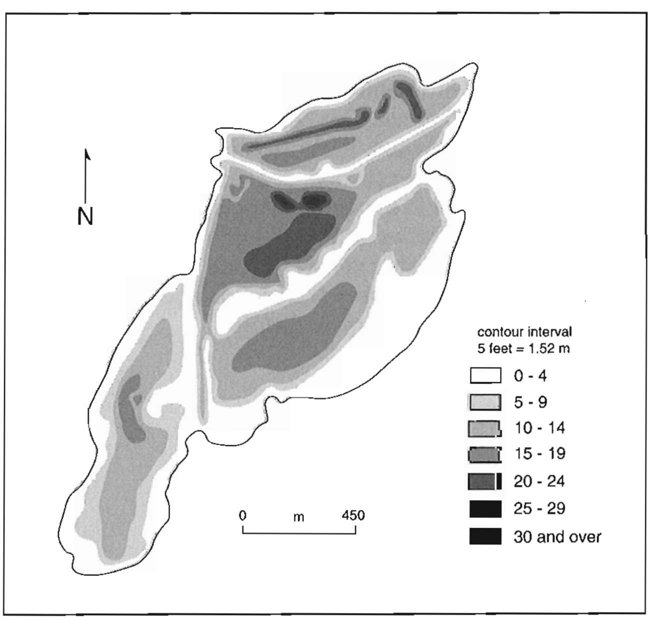


Figure 43. *Bathymetry of Dry Lake, Ontario (from Roddick, 1970).*

Calcareous hard-water lakes receive high natural inputs of carbonates and associated cations. Yet they are oligotrophic because excessively buffered alkaline conditions reduce the availability to plants of inorganic nutrients (phosphorus and metallic micronutrients) and organic micronutrients (vitamins), by inactivation or sedimentation with inorganic particulates (coprecipitation with $CaCO_3$). Still, loadings of organic and inorganic nitrogen compounds tend to be high. Once the buffering capacity is reduced, owing to diminished carbonate supply or increased content of dissolved organic matter, inhibitions on nutrient availability are reduced and eutrophication can proceed rapidly. Transitions to a very acidic, organic-rich state are known to have occurred within a millennium.

Marl is a product of biochemical and physicochemical precipitation and mechanical (re)sedimentation. Precipitation results mainly from CO_2 withdrawal from the lake waters during photosynthesis by hard-water adapted plants, such as *Chara, Potamogeton,* and *Elodea.* To compensate for this CO_2 withdrawal, the dissolved $Ca(HCO_3)_2$ decomposes into CO_2, H_2O, and precipitating $CaCO_3$. Also, CO_2 may be released to the atmosphere when cool groundwater enters the warmer lake environment, or when wave action enhances degassing of high-pCO_2 water. Redistribution of marl due to waves, currents, and ice can lead to formation of submersed bars and shoreline accretions (Gilbert and Leask, 1981). Layering in marls may reflect this redistribution as well as seasonal variation in precipitation.

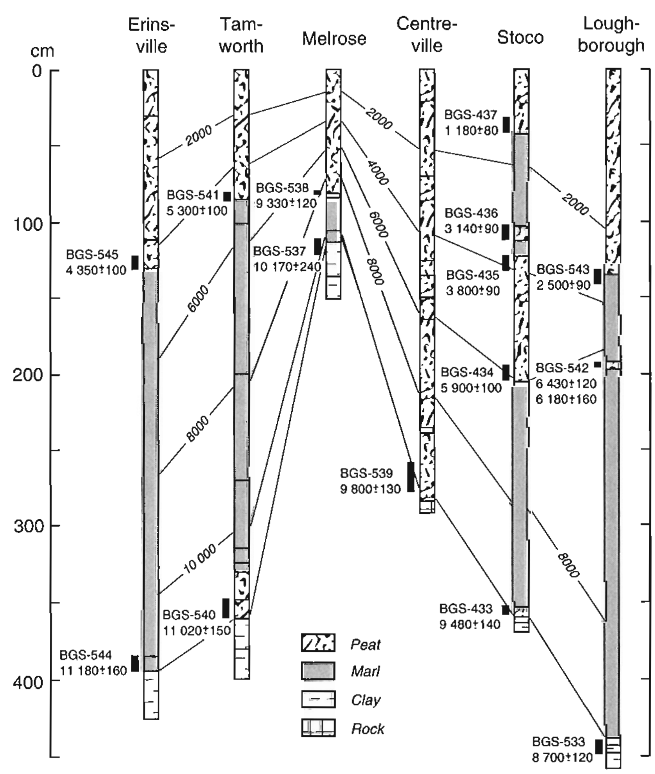


Figure 44. *Radiocarbon-dated organochemical sequences in the survey area (from Vreeken, 1981).*

Holocene lake marl deposits between Kingston and Belleville

A survey of 184 organic-terrain sites in this area revealed that 34% of them were underlain by muck (organic-enriched mineral material), 39% by peat, 20% by peat-mantled marl, and 7% by marl only (Fig. 42; Vreeken, 1981). The complete absence of clastic interbeds from the postglacial sequences suggests that erosion and deposition by hillslope wash and channel action were of minor consequence in this area. Marl deposits, up to 10 m thick, were typically sited in calcareous till terrain with highly jointed and fractured limestone bedrock, but they were invariably underlain by a layer of clay. Thus, the typical marl lake basin received bicarbonate-rich groundwater from pervious surroundings and its impervious clay base permitted maintenance of the perennially wet conditions needed by its specialized life forms.

Marl accumulation began between 11 200 and 8700 BP (Fig. 44), soon after glacial Lake Iroquois drained. At places with minimal scour or redeposition, accumulation averaged from 0.30 to 0.47 mm/a. Marl lake extinction was time-transgressive (Fig. 45), beginning between 10 000 and 8000 BP on internally drained uplands where it was complete by 4000 BP. Extinction in bottomlands began between 6000 and 4000 BP and was pronounced between 4000 and 2000 BP. In interconnected bottomlands, it progressed upbasin, but marl still accumulates in several basin headwaters and in isolated lakes. Peat, up to 3.5 m thick, accumulated at average rates from 0.09 to 0.31 mm/a. The lowest rates (0.09 and 0.15 mm/a) occur at sites prone to oxidation owing to seasonally low water tables (Fig. 43). Average rates from perennially wet bottomlands were between 0.29 to 0.31 mm/a. These rates were used to estimate the time elapsed since the shift from marl to peat accumulation.

Explanations for the demise of marl accumulation, especially between 6000 and 2000 BP, the Main Hypsithermal climate interval (Fig. 46), remain speculative given the range of site conditions and factors involved. Prerequisites for accumulation include a continued carbonate supply to a perennial lake and maintenance of the hard-water condition within it. Progressively deeper pedogenic leaching of the regolith would have diminished carbonate supplied via shallow groundwater systems to shallow upland basins. Bottomland basins with deeper groundwater systems would have been affected much later. Maintenance of the hard-water condition in bottomland drainage lakes could have been jeopardized by shortening of the mean residency time of their water. That would have been promoted by any increased efficiency of the fluvial network aided by a cumulative isostatic rebound differential between headwaters and outlet, currently about 0.5 mm/a. This could help explain an upbasin demise of marl formation. Early Holocene climate warming and a trend toward more pronounced summer rainfall deficits could soon have led to seasonal drying up of shallow upland basins. Lowland basins whose lakes became shallower, owing to marl accumulation itself, would have been affected later. Impacts of progressive organic loading of lake waters should be added to geomorphic factors.

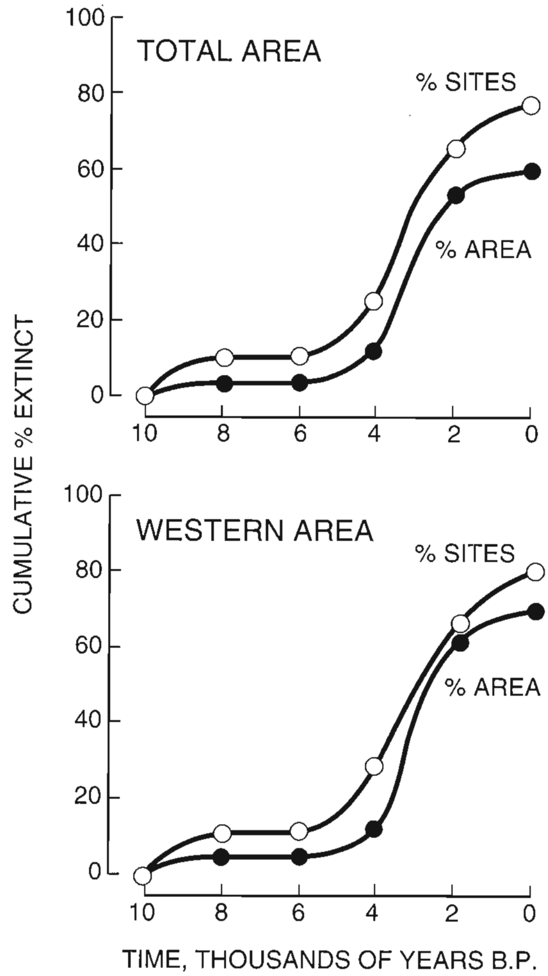


Figure 45. The demise of marl accumulation through geological time, using percentage frequency of sites (open circles) and of associated catchment areas (dark circles) (from Vreeken, 1981).

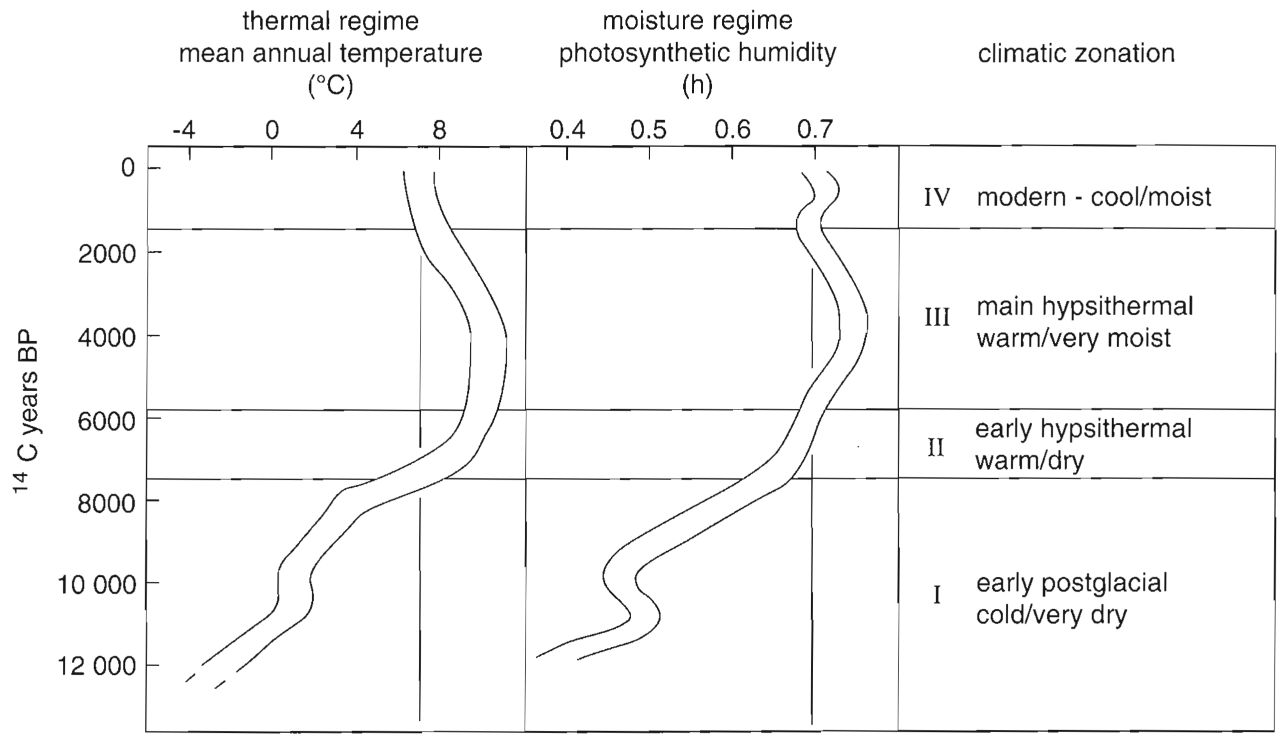


Figure 46. *Composite Holocene paleoclimate reconstruction for southern Ontario from stable-isotope studies. The curves are based on fossil wood-cellulose ^{18}O and ^{2}H data from a site at Brampton, supplemented by marl ^{18}O data from Inglesby Lake and Little Lake (from Edwards and McAndrews, 1989).*

☑ *STOP 10* Landscape of the Dummer Moraine

John Shaw

NTS 31C/6, UTM 232175

Proceed west 10 km from Stop 9 to the Coulters Hill farm.

The Dummer Moraine is located downflow (the direction of ice flow) from the junction of Palaeozoic carbonate and Precambrian Shield rocks (Fig. 2). A prominent upflow facing escarpment in carbonates occurs at this junction. The moraine is hummocky and consists mainly of angular blocks of carbonate bedrock in a silty matrix. Between the hummocks, bedrock lies very close to the surface.

Eskers and tunnel channels extend across the Dummer Moraine and are not covered or infilled with morainic deposits. This indicates that the moraine was formed before these subglacial meltwater forms and is also likely to be of subglacial origin, not a recessional moraine as indicated by Gravenor (1957).

Meltwater flowed in tunnels and subglacial channels after the large-scale regional floods had occurred (Fig. 27). Consequently, the Dummer Moraine may relate to these floods. Clearly, some form of plucking was involved to produce the limestone blocks (Shulmeister, 1989). This may have been a result of a meltwater flow prying ice with frozen blocks of bedrock attached from the bed. Alternatively, low pressures generated in the flow where it accelerated over the escarpment may have caused hydraulic plucking of the heavily jointed bedrock. In either case, the blocks would have been piled subglacially in hummocks.

The moraine area was cleared in the early days of settlement; fences, stone piles, and rock walls around abandoned farmsteads attest to the backbreaking efforts to cultivate this stony wasteland. Since the 1940s much of the cleared land has reverted to scrub woodland and recreational properties have replaced the farmsteads.

☑ *STOP 11* Tweed esker

Tracy Brennand and John Shaw

NTS 31C/6, UTM 121153

Follow Highway 37 21.2 km north from Highway 401 through Roslin to the Moira side road. Turn west toward Moira for 1.4 km to where the road crosses the esker. The stop is in a gravel pit to the south of the road. If coming from Stop 10, continue 11 km west from Coulters Hill to Highway 37 at Roslin (turn right and cross the Moira River at Chisholm on this road). At Roslin turn right (north) on Highway 37 for 2.8 km to the Moira side road.

The Tweed esker extends for approximately 75 km southwest from Actinolite and crosses both the Dummer Moraine and parts of the Belleville-Trenton drumlin fields (Fig. 47). Like other eskers in the drumlin field, the Tweed esker lies in a broad tunnel channel (Shaw and Gorrell, 1991). Transverse ridges are observed adjacent to the esker and within the tunnel channel, toward the southern end of this system.

For the most part, the esker is a single ridge, but in places it divides into multiple ridges (Fig. 47). The ridges reach a maximum height of about 15 m and are up to 50 m across

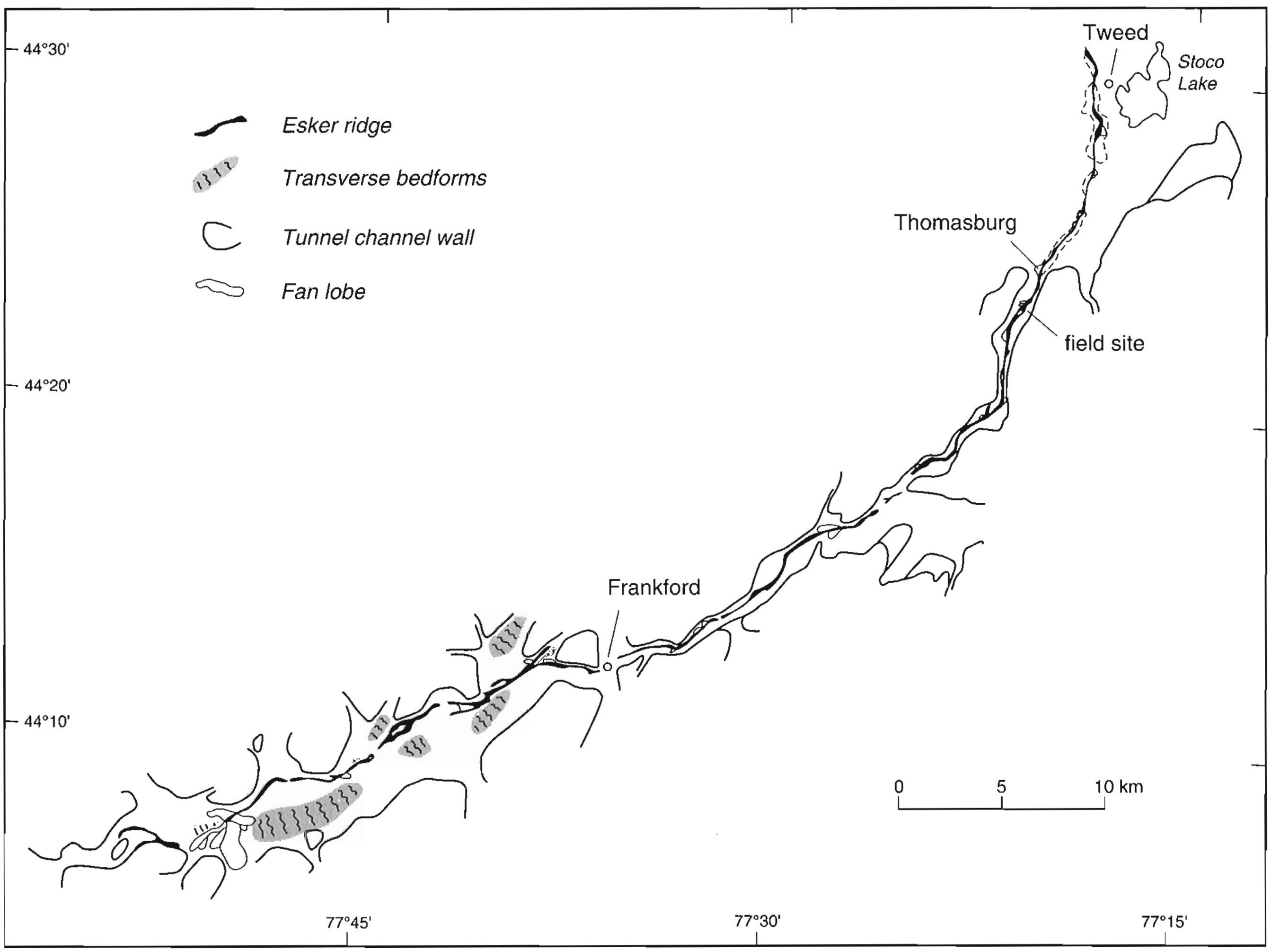


***Figure 47.** Tweed esker and associated tunnel channel.*

(Shulmeister, 1989). Esker deposits are up to 25 m thick. The surface of the esker rises and falls in a series of swells, perhaps suggesting deposition of the esker sediment in large-scale bedforms or macroforms; the latter probably corresponded to changes in height of the esker conduit. Small distributary ridges, terminating in lateral fans, extend from the main esker ridge (Fig. 47).

The sedimentary sequences in the esker core are dominated by thick couplets of gravel and sand. Much of the gravel in the esker core is disorganized and with no apparent lateral continuity to the internal bedding. There is an abrupt transition from gravel to sand or pebbly sand. The fine grained member of the couplet may be plane bedded, trough crossbedded, crosslaminated, or massive. Only two or three couplets are exposed at a section.

Sediments in the core of the esker are dramatically folded in the northern part of the section (Fig. 48). The core of the fold is a poorly sorted, almost diamictic, boulder gravel with a sand matrix. Gravel and sand couplets are arched over this core. The sand is discontinuous and the gravels amalgamated.

Clast fabric on imbricate gravels in the southeast side of the ridge indicates a down-ridge, crest-convergent paleoflow (toward 277°). This is interpreted as indicating a secondary flow vortex within the conduit. The esker sediment fines toward the sides of the deposit, becoming predominantly sand. Faulting is prevalent in these lateral deposits. Faults are mainly high-angled and normal.

Figure 48. *Folded core of the Tweed esker. GSC 1993-164P*

The esker is a product of subglacial drainage and marks a late stage in a complex drainage history. As discussed at previous stops (6-8), earlier drainage events, which produced tunnel channels, were probably associated with catastrophic outburst floods of subglacially stored meltwater (Murray, 1988; Shaw, 1988; Shaw and Gorrell, 1991; Shaw and Gilbert, 1990). The drumlins through which the esker wends its way, are also products of major meltwater flows when the tunnel channels were overtopped and the area was inundated by regional sheet floods (Shaw and Gilbert, 1990; Stop 6). The eskers themselves appear to record more normal glacial drainage, with much of the meltwater being derived from the glacier surface. Since melt rates vary seasonally, the gravel-sand couplets of the core sediments may represent annual deposits. The diamicton beds in the esker then indicate winter periods when meltwater flow in the conduit had nearly ceased and debris flows were emplaced without winnowing (Gorrell and Shaw, 1991). The folding at this site may also indicate that there may have been a relative increase in the rate of conduit closure during periods of reduced flow, compared to the rate of melting back of the conduit walls by fast-flowing water. This closure would have exerted a pincer-like compression, folding the sediment between the ice walls (Fig. 48). Alternatively, low pressure within the conduit during a high discharge event (Röthlisberger, 1972) may have caused basal sediment to be sucked into the conduit.

In the final stages, the esker tunnel appears to have widened appreciably and finer grained deposits were laid down alongside the core deposits. These ice-contact deposits failed by faulting when lateral ice support was removed. Such widening may also have accompanied uplift of the ice sheet and the emplacement of the marginal fans (Gorrell and Shaw, 1991).

It is difficult to determine exactly when the transverse bedforms were formed in the tunnel valley. They may have preceded esker deposition and have formed at a late stage in the last tunnel-channel forming event. Indeed some ridges appear to be overlain by eskers as would be expected in this interpretation. Otherwise, they may have formed as the ice sheet was lifted from its bed in a late stage of glaciation. In this case, they would postdate the main phase of esker formation and possibly relate to the formation of the marginal fans.

☑ STOP 12 Meltwater erosional marks, Marysville

John Shaw

NTS 31C/3, UTM 313002

The erosional marks at Marysville are best seen on the bed of the drainage channel just east of the entrance and exit ramps to the west-bound lanes of Highway 401 at the Marysville turn-off (Fig. 49). Park to the side of the ramp and follow the channel eastward. At high stream discharge, the bed is only exposed about 50 m east of the bridge on the ramp where the channel turns westward. This is also where the most spectacular marks occur.

The erosional marks are found in a complex morphological setting. Large drumlins with a southwesterly orientation correspond to those of the Ontarian event (Shaw and Gilbert, 1990). Large-scale bedrock fluting and striae (Fig. 49 sets 1 and 4) on exposed bedrock at a site cleared for quarrying about 500 m west of Stop 7, show the same trend. However, other, smaller drumlins, meltwater scours, and large-scale bedrock fluting in this area trend at about 80-260°. The small-scale erosional marks at this site also show this trend.

Small-scale erosional marks (Fig. 50) are mainly classical rat tails with tapering remnant ridges extending downflow from resistant chert inclusions in the Ordovician carbonate. The remnant ridges are defined by crescentic scours wrapped around the proximal nose of the ridge and extending downstream in parallel furrows. The ridges taper and become lower downflow relative to the bottom of the furrows because the furrows broaden and become shallower. In some cases, classical comma forms result where only one furrow is well developed (Kor et al., 1991). Other rat tails are arranged in an offset pattern with upstream furrows bifurcating around the noses of downstream ridges.

The rat tails at Marysville are best developed on slightly raised parts of the bedrock surface and commonly extend downflow from upflow-facing steps. Some tails are heavily fractured and appear to have been quarried (Fig. 51). Besides this quarrying, the bedrock was also striated subsequent to formation of the rat tails.

The erosional marks at Marysville are unlike those at Wilton Creek (Stop 7) in that the dominant forms are positive ridges. Nevertheless, ridges must be explained in terms of the erosional processes causing the scour marks around them. The form and pattern of these scours are best explained by the action of horseshoe vortices generated by the boundary-layer pressure distribution upstream of the obstacle (Fig. 52). The residual ridges are considered analogous to desert yardangs and some erosional drumlins, which also have crescentic scour around their proximal ends (Shaw and Sharpe, 1987b).

Other details of the ridges allow a fuller interpretation of the specific conditions as erosion progressed. The distribution of the most prominent rat tails on high points and downstream from upstream-facing steps is analogous to that of large-scale fluting in the Kingston area (Stop 6, Fig. 26). This is probably not coincidental but can be explained in terms of confined subglacial meltwater flow (Kor et al., 1991). At a step, the confined flow is forced to accelerate and, in doing so, stretches longitudinal vortices within it (Fig. 52a, b). Stretched vortices have a reduced radius and, in order to

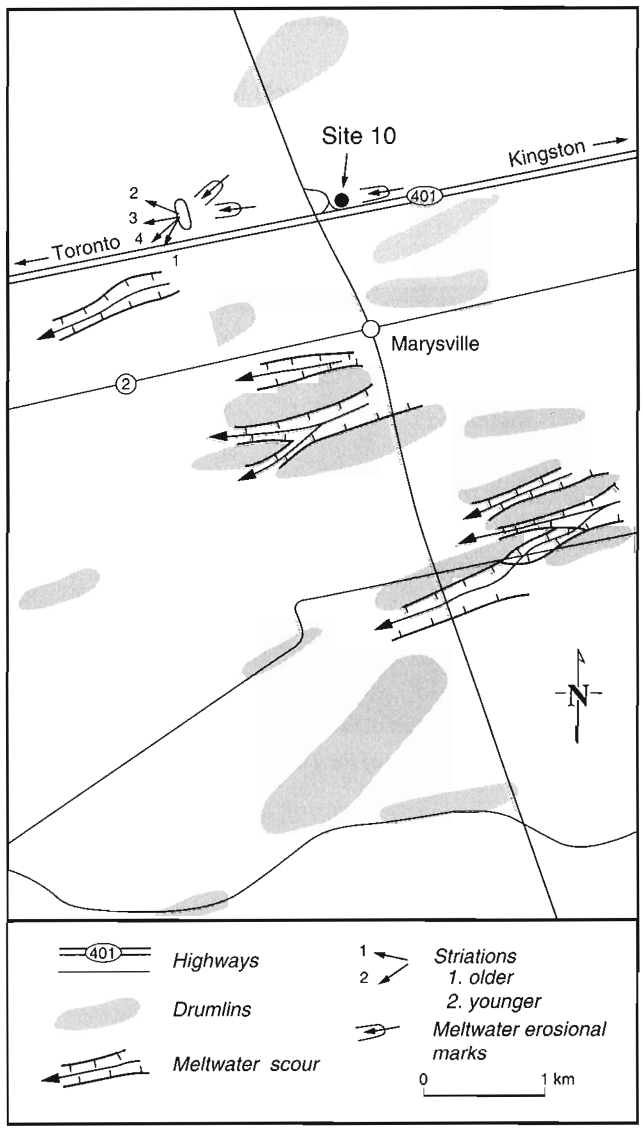


Figure 49. Location map of Marysville erosional marks.

conserve angular momentum, they rotate vigorously and become more erosive. This explains the intensity of erosion by longitudinal vortices downstream from steps on raised parts of the bedrock (Kor et al., 1991).

Longitudinal vortices also break down, expanding rapidly (van Dyke, 1982, p. 75). Expansion and turbulent mixing of the vortex with surrounding fluid causes a reduction in erosional intensity; expansion and shallowing of furrows (Fig. 50 and 52b, c) are thus related to vortex breakdown.

As meltwater flow decreased and the glacier was lowered to the bed, it would have made contact with the highest points first. The rat tails would, therefore, have carried extremely high normal stresses and ice would have deformed rapidly around them. But should a protruding boulder come into contact with a rat tail, the ice would deform downwards around it, causing downward drag on the boulder and extremely high normal stresses between the rat tail and the boulder (Hallet, 1981). Highly fractured and plucked rat tails (Fig. 51) are explained by this process, striation representing more normal processes of sliding and abrasion after a meltwater drainage event.

Figure 50. Rat tails downstream from resistant chert nodules, Marysville. GSC 1993-164Q

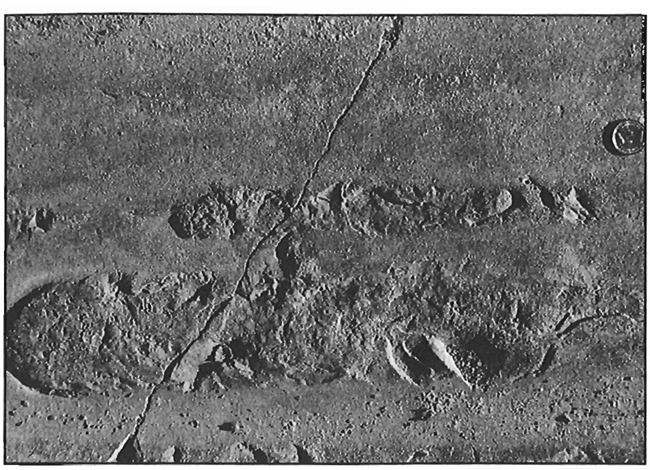


Figure 51. Fractured rat tails, Marysville. GSC 1993-164R

The drainage event that produced the Marysville rat tails appears to have been relatively late; the flow directions correspond closely with those of eskers in tunnel channels close to this site (Shaw and Gorrell, 1991). Esker formation may correlate with the formation of the Oak Ridges Moraine and was later than the Algonquin and Ontarian events of Shaw and Gilbert (1990).

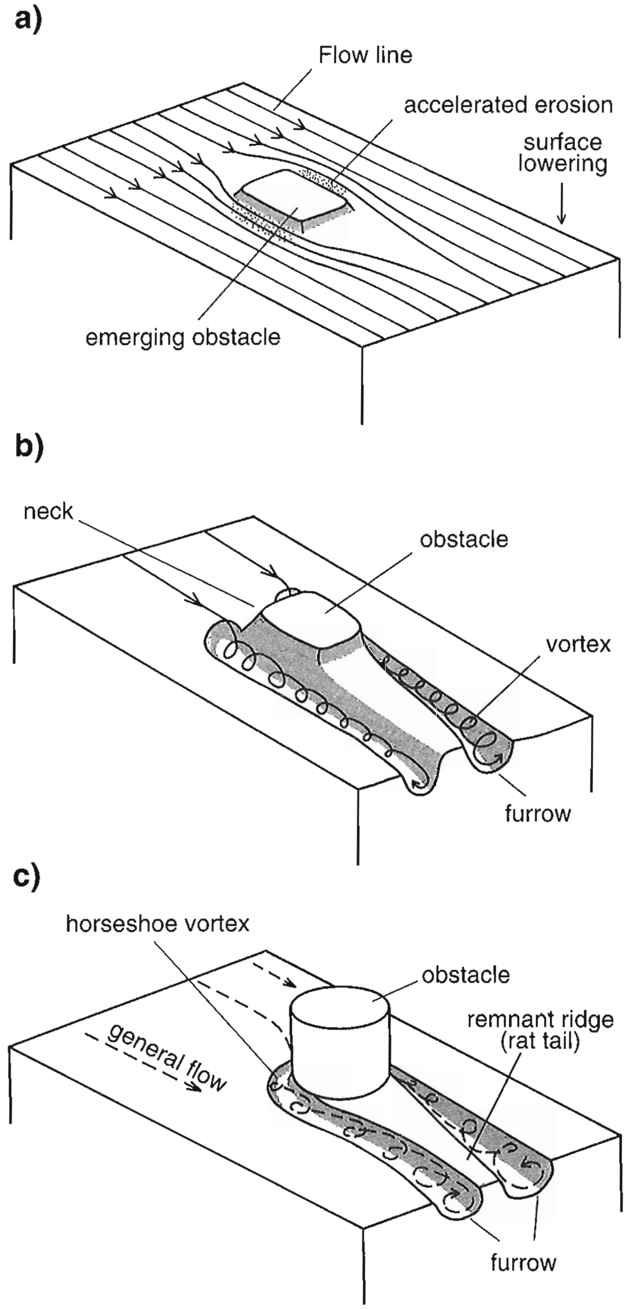


*Figure 52. Horseshoe vortices producing rat tails (after Sharpe and Shaw, 1989) **a)** flow around emerging obstacle, **b)** vortices with enhanced erosion adjacent to and down flow of obstacle, **c)** horseshoe vortex producing crescentic furrow with areas which become shallower and broader downflow.*

☑ STOP 13 Large subglacial fluvial landforms and the postglacial lacustrine environment of part of northeastern Lake Ontario and the upper St. Lawrence River

Robert Gilbert

NTS 31C/1, 2

There is no single best place to view the region containing the features described in this stop. It may be seen along Highway 33 between Picton and Kingston, from the foot paths in the Lemoine Point Conservation Area (south off Highway 33, 2.0 km west of the stop lights at Days Road), from the parks along the Kingston waterfront, or from the ferries between the mainland and Amherst and Wolfe islands.

The northeastern arm of Lake Ontario (mean depth 23 m) and the upper St. Lawrence River are shallow compared to the main body of the lake, except where a number of deeply incised channels occur especially among the islands of the region. Shaw and Gilbert (1990) have proposed that, during glaciation, the Kingston area was part of a major conduit for the flow of subglacial water from the northeast that created this extensive channel network (Fig. 25). This account summarizes detailed reports on several of the interisland channel networks (Gilbert, 1990; Gilbert and Shaw, 1992).

A subbottom acoustic survey system was used to map the topography of the lake floor and to penetrate the sediment that partially fills the channels (Fig. 53). Striking features lead to the interpretation that they originated by subglacial fluvial processes: (1) The channels are deeply incised in a landscape that is otherwise remarkably flat. The trenches in Bateau and North channels reach more than 100 m below the level of the surrounding landscape. Others in the region which have yet to be studied, including the channel of the St. Lawrence River beneath the Ivy Lea bridge, appear to be similar. (2) The channels are narrow (in most cases from 0.5 to 1 km) with respect to their depth which enhances the impression of their deep incision. (3) Although some appear associated with structural features in the underlying Precambrian or Palaeozoic bedrock, suggesting that they may have originated along zones of weakness or in pre-existing depressions, the slopes of the channels and the interconnected network that they form (Fig. 25) indicate that they are not created by the present streams in the region. They could only have formed in a closed hydraulic system as exists beneath a large glacier covering the land surface. (4) The shape of individual elements on the channel floors and sides is similar especially to the comma-shaped features described at Stop 7, although the scale is up

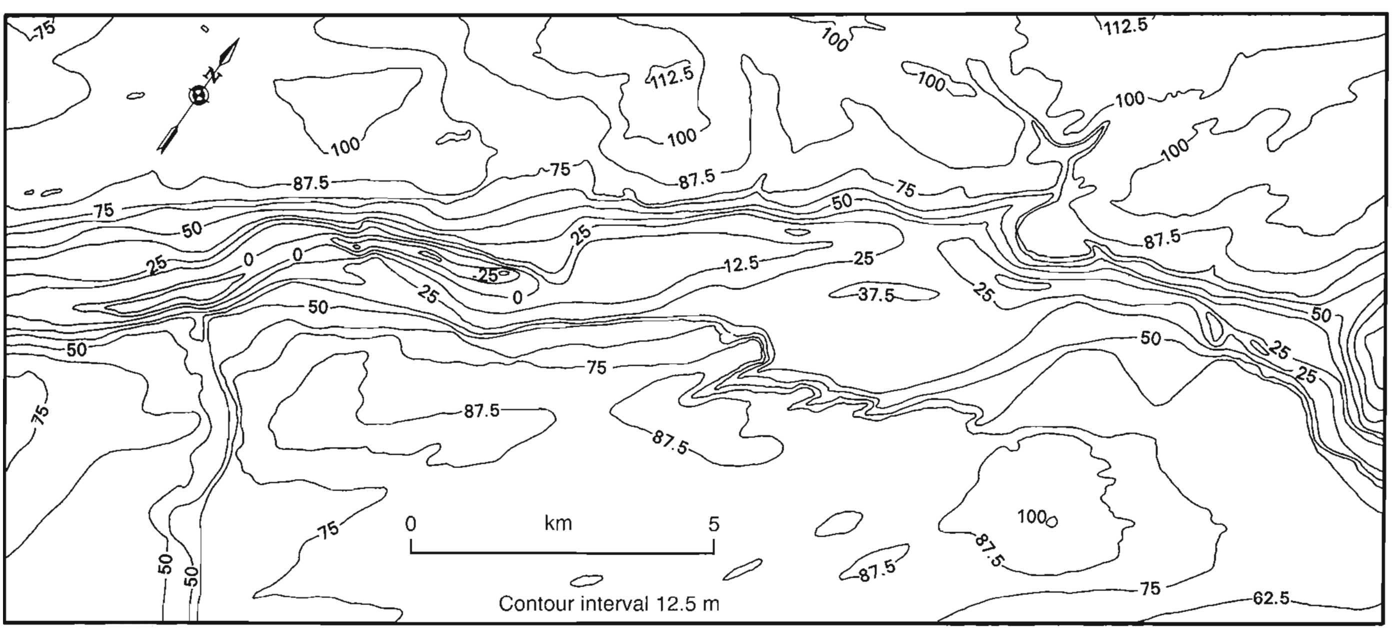


Figure 53. Surface of bedrock in and near North Channel (<75 m) interpreted from NTS topography map 31C/2 and acoustic subbottom records. Elevations refer to sea level; the contour interval is 12.5 m. The 75 m contour approximately outlines the present shoreline. Bateau channel (<75 m). See Figure 2, Stop 13 for location (after Gilbert and Shaw, 1992).

to four orders of magnitude greater. The flow direction indicated by their form is from the northeast to east, in accordance with the hydraulic gradient beneath the continental glacier, but opposite to the present flow direction from Lake Ontario.

From the subbottom survey the character of the glacial and postglacial sediments that partially fill these channels has been determined and the environment shortly after deglaciation interpreted. Four distinct acoustic facies are recognised (Fig. 54): (1) an acoustically transparent sediment which forms the top layer of sediment, (2) a partially opaque, massive sediment in some places underlying facies 1, (3) a pervasive acoustically well-stratified sediment in some places beneath facies 1 and 2, in other places exposed at the lake bottom, and (4) a nearly completely opaque sediment underlying facies 3 in a few places. Three of these are shown in Figure 54.

Facies (4) is the oldest and represents sediment deposited in close proximity to the retreating continental glacier. The sediments of facies 3 were deposited in ice-dammed Lake Iroquois (Fig. 3b). Those of facies 2 are the remnants of an environment during the low level phase of early Lake Ontario (Fig. 3c) described below, while those of facies 1 are the modern sediments of Lake Ontario. From the patterns of these four, and the presence of completely opaque surfaces below which there is no acoustic definition, eight distinct acoustic assemblages are mapped in Figure 55 and are described below.

On exposed shores in water depths less than 5-10 m, the lake floor is swept clear of fine sediments and bedrock (Br) or coarse, hard sediment is exposed. In deeper water and in areas protected from wave action, a patchy or thin veneer of acoustically transparent sediment (facies 1) is found over the opaque surface (BrV). A sloping zone mantled with sediment (MS) found only in the western portion of the study area (Fig. 55) is transitional from the Br and BrV zones above to the channel bottom zones. Throughout North Channel, nearly flat-lying shelves are found at 18-36 m depth (depths increasing westward). Those east of Kerr Point occur as a single platform along the sides of a well-defined channel (Fig. 55), while those to the west have a different acoustic character and rise in several steps comprising upper shelves (US) and lower shelves (LS) flanking the sides of a deeper basin occupying North Channel at Upper Gap. A drowned river channel occupies a portion of the present waterway from Adolphus Reach through North Channel to Kingston (Fig. 55). Three facies assemblages recognized from their acoustic character occupy part of the channel: acoustically stratified sediments mantling the deep channel floor west of Kerr Point (Gl); a mixture of massive and stratified materials partly filling the channel between Kerr Point and Cataraqui Bay (C); and a zone of flat-lying, acoustically opaque sediments in Kingston harbour which have the character of a submerged floodplain (F) with a series of anastomosing river channels (R) on the surface.

The paleogeography of the northern part of Kingston basin (Fig. 2) from about 11 400 BP when the last effect of ice damming ended until about 8000 BP has been assessed in detail by Gilbert (1990) and Gilbert and Shaw (1992). Although the history of each channel is different, the conceptual model shown in Figure 56 summarizes the general development of these channel systems from their formation by subglacial fluvial discharge to their filling with postglacial lacustrine sediment. In brief, isostatic depression which was greater toward the northeast (Fig. 4) created an early stage of

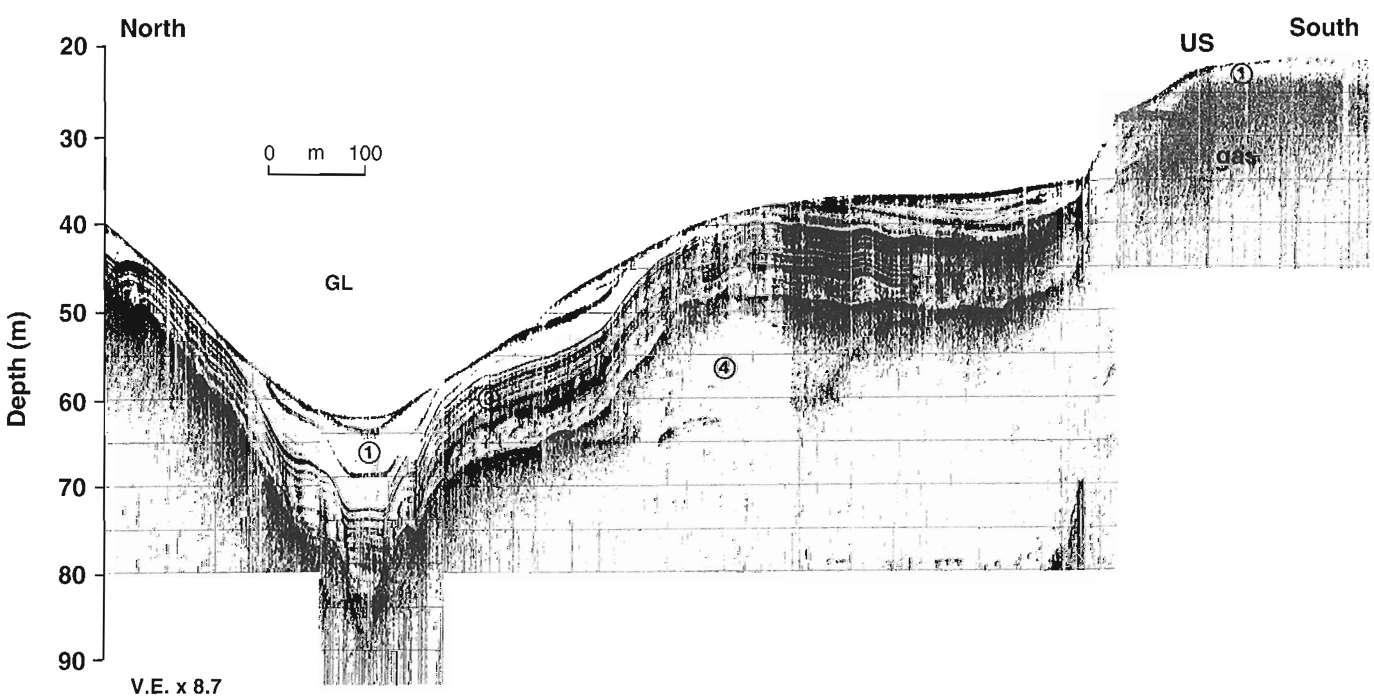


Figure 54. Subbottom record (accentuated track, Fig. 55) showing conformably lying glaciolacustrine sediments (3) in the deep channel (GL), glacial sediments beneath (4), and the upper shelf (US) with its gaseous modern sediment (1) (after Gilbert and Shaw, 1992). Vertical exaggeration is 8.7.

Lake Ontario which was much lower than the present level (Fig. 3c, d). The entrance to the early St. Lawrence River was located in the southern part of the present Kingston basin and, as a result, an isolated drainage system of lakes and rivers developed in the northern part of the basin. This drainage system flowed eastward from the present Bay of Quinte, through a small deep lake in North Channel. Tributaries east and west of Amherst Island drained a larger, shallow lake in the Kingston basin to the south. A broad floodplain between the city of Kingston and Wolfe Island acted as a local base level controlling water levels in this drainage system. For a short period, large flows from the upper Great Lakes and possibly Lake Agassiz to the west (Fig. 3c, 56e) may have come through this system by way of the Kirkfield-Fenelon Falls spillway to the Trent River system. However, through much of this period a moderate-sized river occupied North Channel-Kingston harbour with extensive shallow shelves along the sides (Fig. 56f), probably very much as wetlands flank nearby Cataraqui River today (Stop 14). After about 8000 BP isostatic rebound had raised the St. Lawrence River sufficiently that the sills in the Kingston area were back-flooded and the water levels rose sufficiently to connect the main body of Lake Ontario with the lakes in the Kingston basin. Gradually the connecting rivers in North Channel and upper St. Lawrence River were flooded to the present level of Lake Ontario by about 4000 BP.

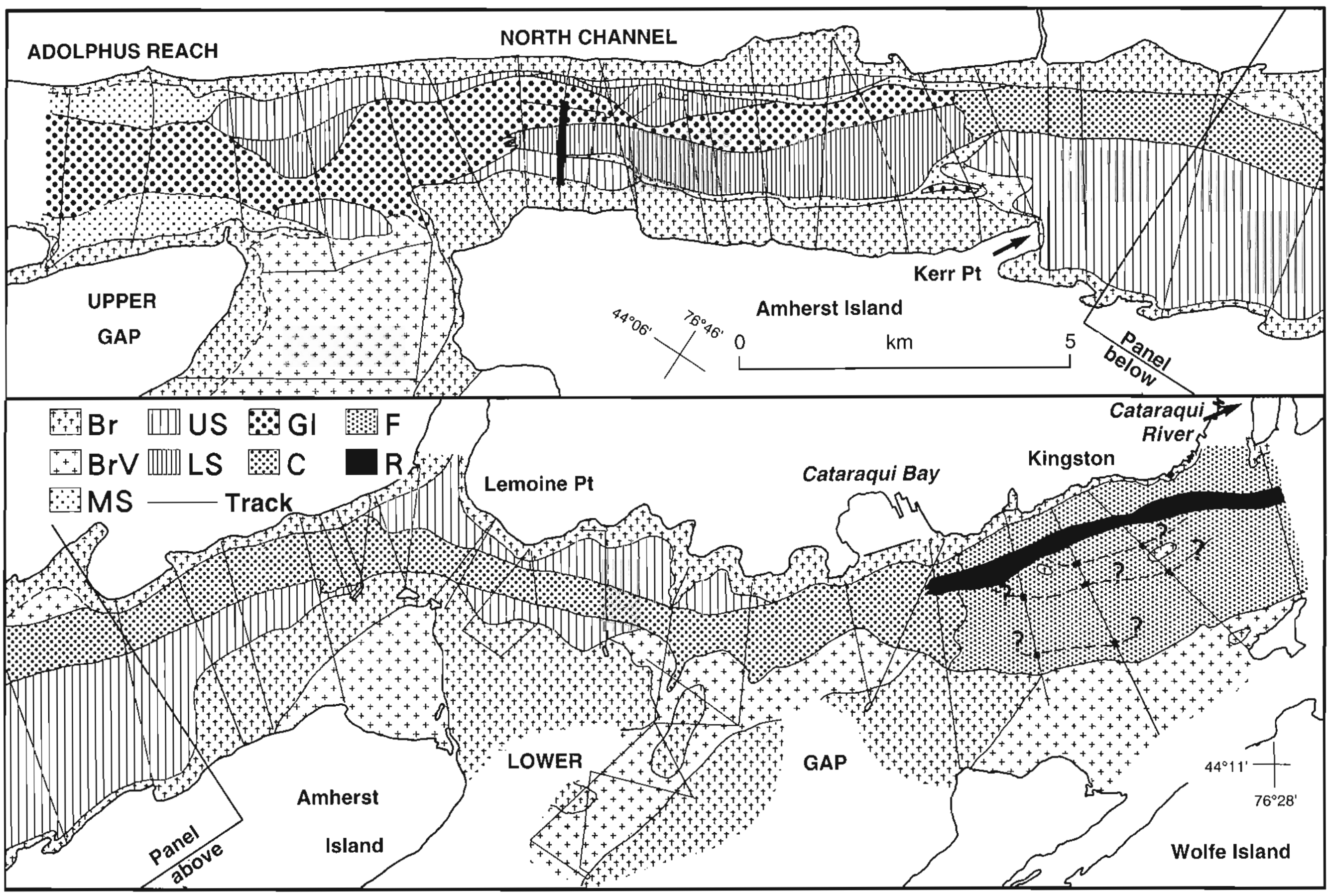


Figure 55. Acoustic facies assemblages of North Channel and Kingston harbour mapped from subbottom acoustic data: bedrock at the surface (Br); bedrock with a thin or discontinuous veneer of sediment (BrV), slopes mantled with undifferentiated sediment (MS), upper shelves (US), lower shelves (LS), glaciolacustrine sediments exposed at or near surface (Gl), channel deposits, (C) and floodplain deposits (F) with river courses (R) (after Gilbert and Shaw, 1992).

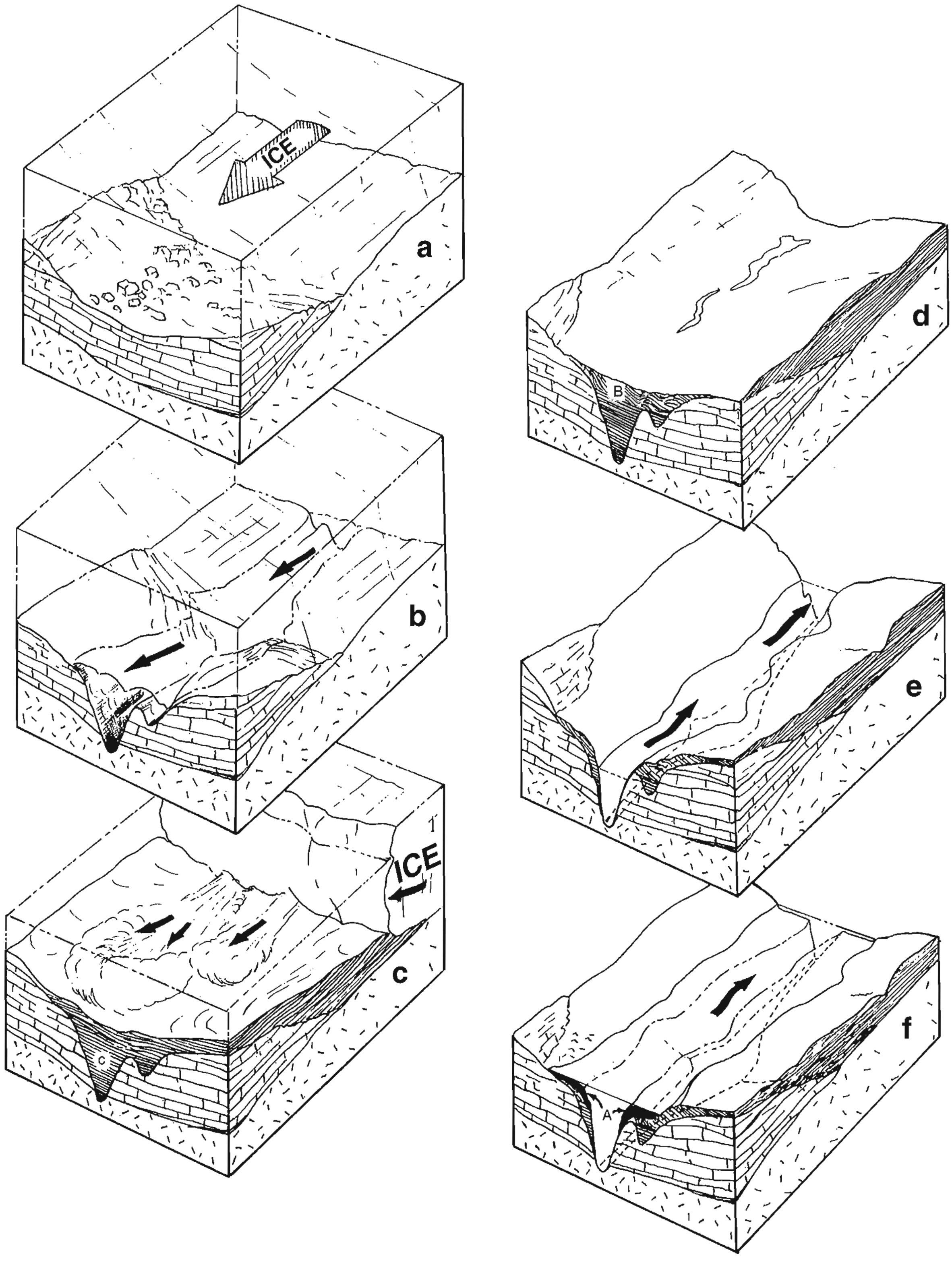


Figure 56. *Sketch of the proposed sequence of events in the formation of Bateau Channel:* ***a)*** *during glaciation,* ***b)*** *during subglacial fluvial erosion by meltwater,* ***c)*** *during sedimentation in Lake Iroquois,* ***d)*** *as erosion occurred from surrounding land during the low-level phase of early Lake Ontario,* ***e)*** *during reinundation and exhumation of the trench, and* ***f)*** *as the marl-rich sediment of the flanking shelves were deposited (after Gilbert, 1990).*

☑ STOP 14 Evolution and hydrology of wetlands in the Cataraqui Lagoon, Kingston

Robert W. Dalrymple and Jonathan S. Price

NTS 31C/1, 8, UTM 825025

The Cataraqui Lagoon and its associated wetland occupies the mouth of the Cataraqui River immediately east of Kingston. An overview of the area may be obtained from east-bound Highway 401, 0.5-1 km east of the Montreal Street exit (only emergency parking is permitted on the shoulders of Highway 401), and from the top of "Mount Cataraqui", an old landfill site on the north side of the Bell Park Golf Course, which is located immediately east of Montreal Street, 3.4 km south of Highway 401. (Bear left after entering the Golf Course grounds, and park at the tennis courts. Proceed on foot to the top of the hill.)

Wetlands form an important interface between the terrestrial and lacustrine environments. They are valuable wildlife habitats and form an important part of the hydrological cycle. They contain in their sedimentary record the history of changes from one to the other and so of climatic and other global changes.

The open-water portion of the Cataraqui Lagoon has a length of 12 km, and a maximum width of 1.3 km. It is bordered on the west by an extensive area of marsh up to 1 km in width (Fig. 57). Near the mouth, water depths exceed 2 m, but most of the inner lagoon above Bells Island is less than 1-1.5 m deep. The vegetation in the outer, lower and wetter parts of the marsh is dominated by *Typha* (cattails), while meadows of blue joint grass, reed canary grass, and sedges occur in the higher and drier areas (Catling, 1985). The shallow portions of the lagoon are also heavily vegetated, the most abundant species being waterweed (*Elodea canadensis*), pondweed (*Potamogeton crispus*), coontail (*Ceratophyllum demersum*), and mud plantain (*Heteranthera dubia*). The density of the subaquatic vegetation decreases as depth increases, and the areas deeper than about 1.7 m are largely unvegetated.

This lagoon has played an important role in the history of the area, and is responsible for the location of Kingston. The good harbour that it provided was one of the major reasons for the siting of Fort Frontenac at its mouth by the French in 1673 (Osborne and Swainson, 1988). This location subsequently attained strategic importance for the English when the Rideau Canal, which joins Lake Ontario with the Ottawa River, was constructed between 1826 and 1832, followed by the building of Fort Henry near its mouth in 1826-1832 (Fig. 57). Periodic dredging of the inner Kingston harbour, and construction of the Lasalle Causeway in 1916 have altered the outer part of the lagoon. Dredging in the inner lagoon is restricted to the channel of the Rideau Canal, and to the vicinity of the Cataraqui landfill site which lies on the northern side of the neck of land which joins Bells Island to the western shore (Fig. 57). Leachate from this landfill site, which was decommissioned in 1975, is entering the lagoon (Frape, 1979; Creasy, 1981), but routine water quality measurements at the LaSalle Causeway reveal that only phosphorus values periodically exceed the Ontario Water Quality Objectives. Organic content, conductivity, pH, suspended solids, and

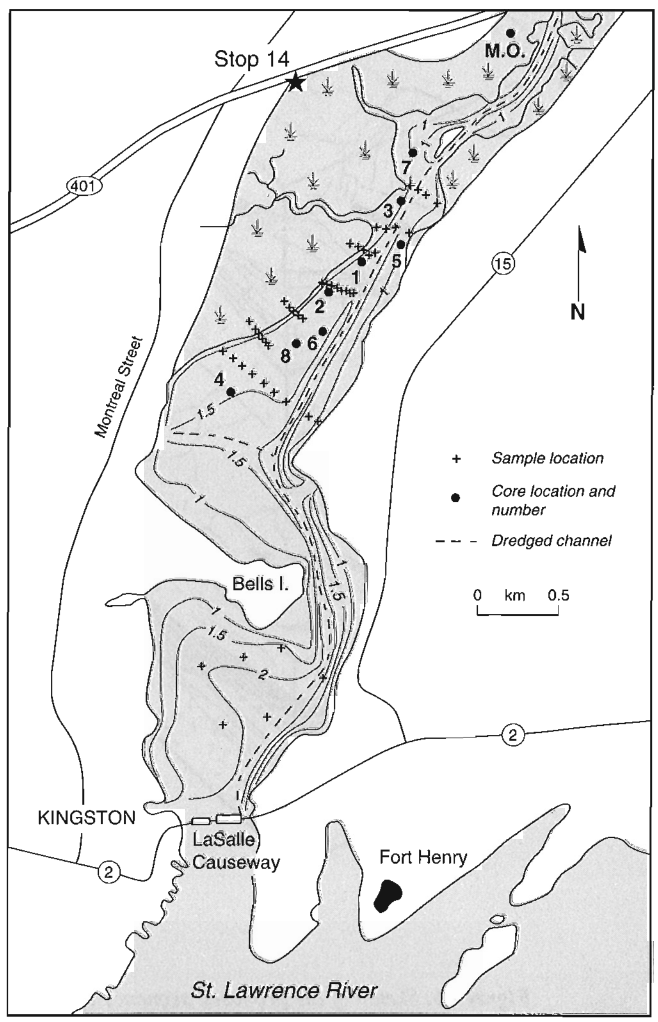


Figure 57. Map of the Cataraqui Lagoon and marsh showing the stop location (star), bathymetry based on Canadian Hydrographic Chart 1513H, and the distribution of surfical grab samples and cores.

various metals are all well within the acceptable range. The inner lagoon and marsh area now face the threat of disturbance as demand for a second road link between Kingston and the communities to the east increases.

Evolution of the lagoon and lake-level fluctuations

The Cataraqui River and lagoon occupy one of the subglacial channels (Fig. 25) that cut the area (Gilbert, 1990; Shaw and Gilbert, 1990). The lagoon itself is underlain by Ordovician limestone, which outcrops in cliffs along the margins of the valley. The edge of the Precambrian Shield lies less than a kilometre upriver of Highway 401, and most of the 478 km^2 drainage area of the Cataraqui River is underlain by gneiss and granite. Throughout the immediate area, tills are thin or absent (Henderson, 1966), presumably because of the subglacial floods (Shaw and Gilbert, 1990).

The oldest known sediments beneath the lagoon consist of thinly-laminated, grey clays which are interpreted to have accumulated in glacial Lake Iroquois (Dalrymple and Carey, 1990). These sediments are probably equivalent to Gilbert and Shaw's (1992) acoustic facies 3 (Stop 13). In cores, the upper surface of these clays is sharp and lacks evidence of soil formation, perhaps due either to fluvial erosion during the time when water levels in Lake Ontario were below present, and/or to the presence of wet, boggy conditions in the bottom of the valley. The lack of coarse fluvial sediments above the Lake Iroquois clays (Dalrymple and Carey, 1990) suggests that the many lakes within the upper part of the Cataraqui River basin have acted as settling basins and therefore restricted the sediment input to fine grained suspended material.

Isostatically-rising water levels, coupled with the limited sediment input, have allowed Lake Ontario to flood up the valley, with the establishment of lagoon and marsh conditions above Bells Island about 4000 years ago (Dalrymple and Carey, 1990). Since then, up to 2.5 m of organic-rich clayey silts (gyttja) and peat have accumulated (Fig. 58).

At the present time, the surficial sediments display a strong inverse correlation between the amount of organic matter and water depth (Fig. 59). Sediment in water >1.5 m deep (standardized to the mean, midsummer water level) contain approximately 30% (dry weight) organic material, whereas sediment in water <0.75 m deep, including the modern marsh, contain 60-90% organic material. Based on this correlation, the layers of gyttja in the cores are interpreted to have accumulated in an open-water, lagoon setting, whereas the peats (organic content >70-75%) represent the deposits of a very shallow lagoon or marsh environment. The alternation of these environments through time (Fig. 58) implies that water levels have not risen smoothly.

The relationship between organic content and water depth (Fig. 59) allows us to obtain an estimate of past water depths, based on the variation in organic content within cores (Dalrymple and Carey, 1990). This can in turn be converted into a plot of lake level through time by "hanging" the inferred

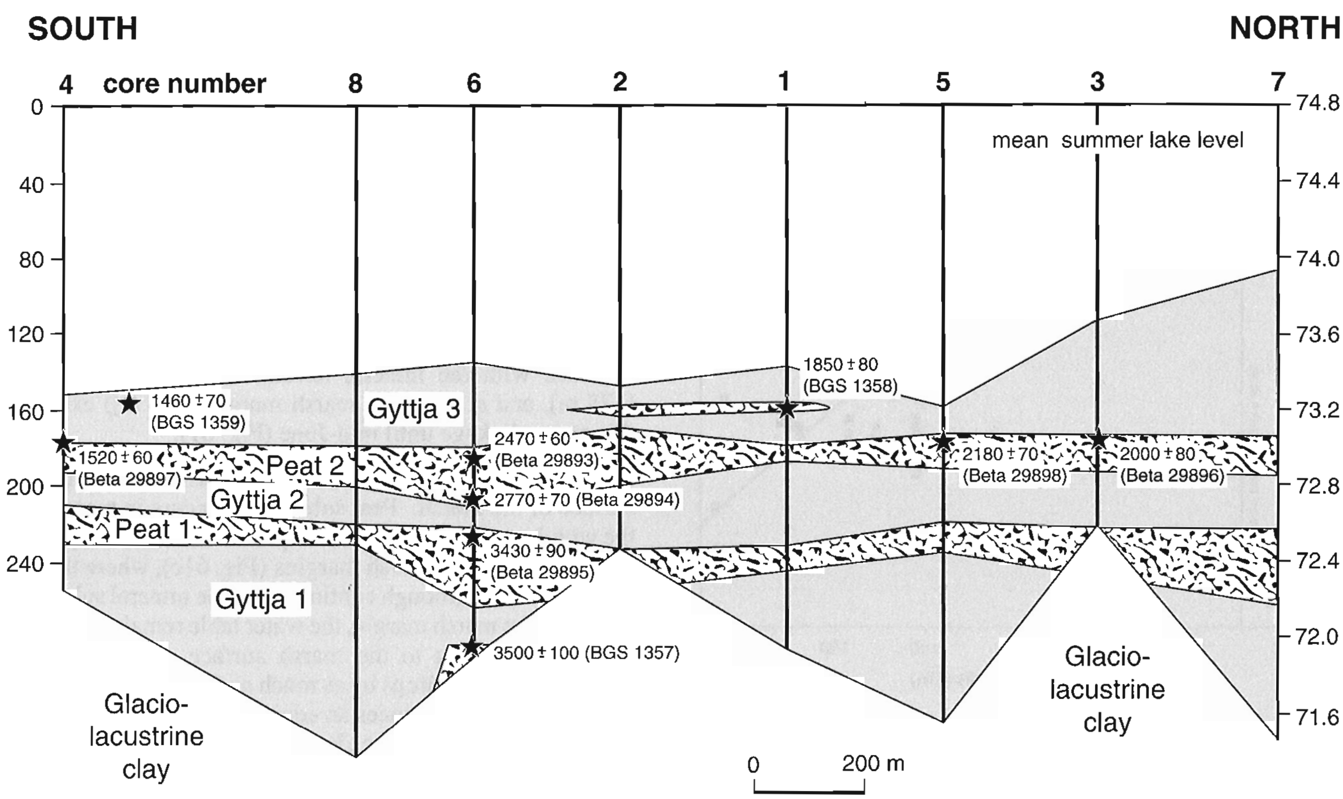


Figure 58. Stratigraphy and chronology of Holocene sediments beneath the Cataraqui Lagoon. See Figure 57 for core locations (from Dalrymple and Carey, 1990).

water depths on a plot of sediment elevation against time (Fig. 60). This diagram shows that the water surface was below the level of the sediment (i.e., the area was a marsh) from about 4000-3200 BP and again from 2400-2000 BP. Before, between, and after these periods of marsh development, an open-water lagoon existed.

As discussed by Dalrymple and Carey (1990), the water level oscillations deduced from the lagoon stratigraphy correspond closely in direction, amplitude, and time with the lake-level fluctuations reported by Flint et al. (1988) and McCarthy and McAndrews (1988) from lagoons at the western end of Lake Ontario. Therefore, these water-level fluctuations are interpreted to reflect climatic changes in the Great Lakes drainage basin, with wetter periods coinciding with lake-level rises and high stands. An analysis of independent climatic indicators (Flint et al., 1988; McCarthy and McAndrews, 1988; Dalrymple and Carey, 1990) supports this correlation. Less agreement exists, however, on the relationship between temperature and lake levels. Using both paleoclimatic and instrumental data, Flint et al. (1988) and Dalrymple and Carey (1990) suggested that wet periods are associated with warmer intervals in the eastern portion of the Great Lakes basin. McCarthy and McAndrews (1988) concluded that the opposite correlation existed (wet periods are synchronous with

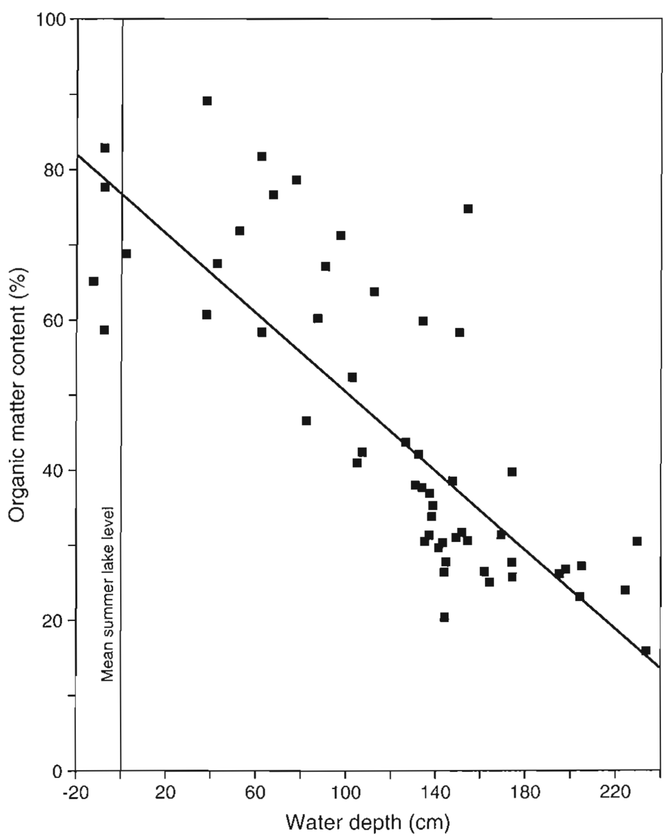


Figure 59. Variation of organic matter content (weight loss on combustion at 550°C) with water depth in the surficial sediments of the Cataraqui Lagoon and marsh. The regression line has a slope of -0.28%/cm and a y-intercept of 77.1%. The relationship is statistically significant at greater than the 99% level of confidence (from Dalrymple and Carey, 1990).

cool temperatures), based on paleoclimatic data which was derived largely from the western Great Lakes area. Most predictions of the potential impact of global warming also suggest that precipitation and lake levels will drop as the temperature rises (Cohen, 1986).

Hydrological aspects of the Cataraqui Lagoon wetland

The hydrology of the *Typha* marsh can be inferred from measurements made from May-August 1991 on a similar marsh in Bayfield Bay, Wolfe Island, 10 km southeast of the Cataraqui Lagoon (J.S. Price, unpub. data, 1992). Both marshes are subject to similar climatic stresses, lake-level fluctuations, and have similar vegetation. The deposits of the Cataraqui Lagoon marsh extend from just above the average summer lake water level to a depth of 2.4-3.3 m (Dalrymple and Carey, 1990). Peat in the upper 1 m is poorly to moderately decomposed, except at the outer marsh margin, where decomposition is moderate to good. The upper 20-30 cm contains abundant *Typha* roots and rhizomes.

The hydrological regime of the marsh is closely related to the water level of Lake Ontario, which peaks in spring following snowmelt, and experiences a general decline over the summer (Fig. 61a). Superimposed on this general decline are periodic peaks due to rainfall events, and frequent variations associated with seiche activity in the lake. In spring the marsh hydrology is dominated by high lake levels which, in 1991, elevated the lagoon water level up to 30 cm above the general surface of the marsh (Fig. 61b). During this period, fluctuations in water level in the lagoon are rapidly transmitted across the surface of the marsh as water seeps through the tangled *Typha* mat left standing over winter. The fluctuations generated by seiche activity in 1991 caused rapid water-level variations of up to 15 cm in the lake, producing equivalent water-level fluctuations at the marsh margin (Fig. 61c). On June 16 (Julian day 169), for example, a 5.7 cm change was recorded over a 15 minute period at the margin of the Wolfe Island marsh. In the marsh interior, 300 m inland from the margin, delayed water-level fluctuations of subdued intensity were recorded. Water levels measured at three locations across the Wolfe Island marsh in 1991 (10 m from the interface with the mineral terrain, near the marsh centre [175 m], and at the outer marsh margin [375 m]) exhibited this strong linkage until mid-June (Fig. 61).

By mid- to late June the lake level falls below the general surface of the marsh. Peat subsidence occurs in response to the withdrawal of water from the peat matrix. This is particularly marked at the marsh margins (Fig. 61c), where the peat is more buoyant (though continuous to the mineral substrate). In fact, at the marsh margin, the water table remains relatively fixed with respect to the marsh surface, even though its absolute elevation drops by as much as 50 cm (Fig. 61c). The marsh interior experiences an equivalent water level decline, but peat subsidence is much less marked because of the denser peat there (Fig. 61b). During this period of lower lake levels, there is a reversal of the water flow direction, and water generally drains away from the marsh to the lake.

A high water table in the surrounding land area during spring causes drainage toward the marsh. However, in summer, the water table drop is amplified in the mineral sediments by their low specific yield, resulting in deep recession of the water table in the areas adjacent to the marsh. By late June the lower water levels in the mineral sediments induce drainage away from the marsh. For the remainder of the summer, therefore, the marsh is normally isolated from lateral water influxes, and its water table regime is dominated by rainfall and evaporation.

During the spring, evaporation from the marsh dominates over transpiration since there is a free water surface, and plant development is limited. As shoots develop, the water surface becomes more shaded, and transpiration increases. By the time the water table drops below the marsh surface, the canopy provides total shade and high atmospheric resistance, which decrease the turbulent flux of vapour, causing evaporation to approach zero. As a result, evapotranspiration is almost entirely due to transpiration. Since the cattails have essentially unlimited access to the water table, which is never far below the surface, the overall evapotranspiration rate is high. At the Wolfe Island marsh, energy balance measurements indicate that evapotranspiration in June and July 1991 ranged from 0.6-7 mm/d, and averaged 4.6 mm/d. These evapotranspiration rates produce small daily changes in the water level on the marsh (Fig. 61). However, during the summer of 1991, cumulative evapotranspiration losses exceed inputs by rainfall by over 100 mm. The deficit accounts for the low water table position in the marsh by the end of summer.

Electrical conductivity during the spring is generally low, reflecting the lake water values, and has little variation because mixing at this time of high water levels is relatively good. In summer, electrical conductivity increases, and is greatest 10-20 m from the margin of the marsh (>300 μS), decreasing toward the centre (<100 μS). The low values in the centre reflect the low electrical conductivity of precipitation. The elevated values at the margin result from limited penetration of lake water (≈ 230 μS), which is subsequently evaporated, raising the concentration of dissolved solids. The limited penetration of lake water during the summer indicates that most of the marsh is uncoupled with the lake for most of the year. This is important because it suggests that the ability of the marsh to accept and release pollutants and nutrients from and to the lake is limited to the springtime, except along the marsh margin.

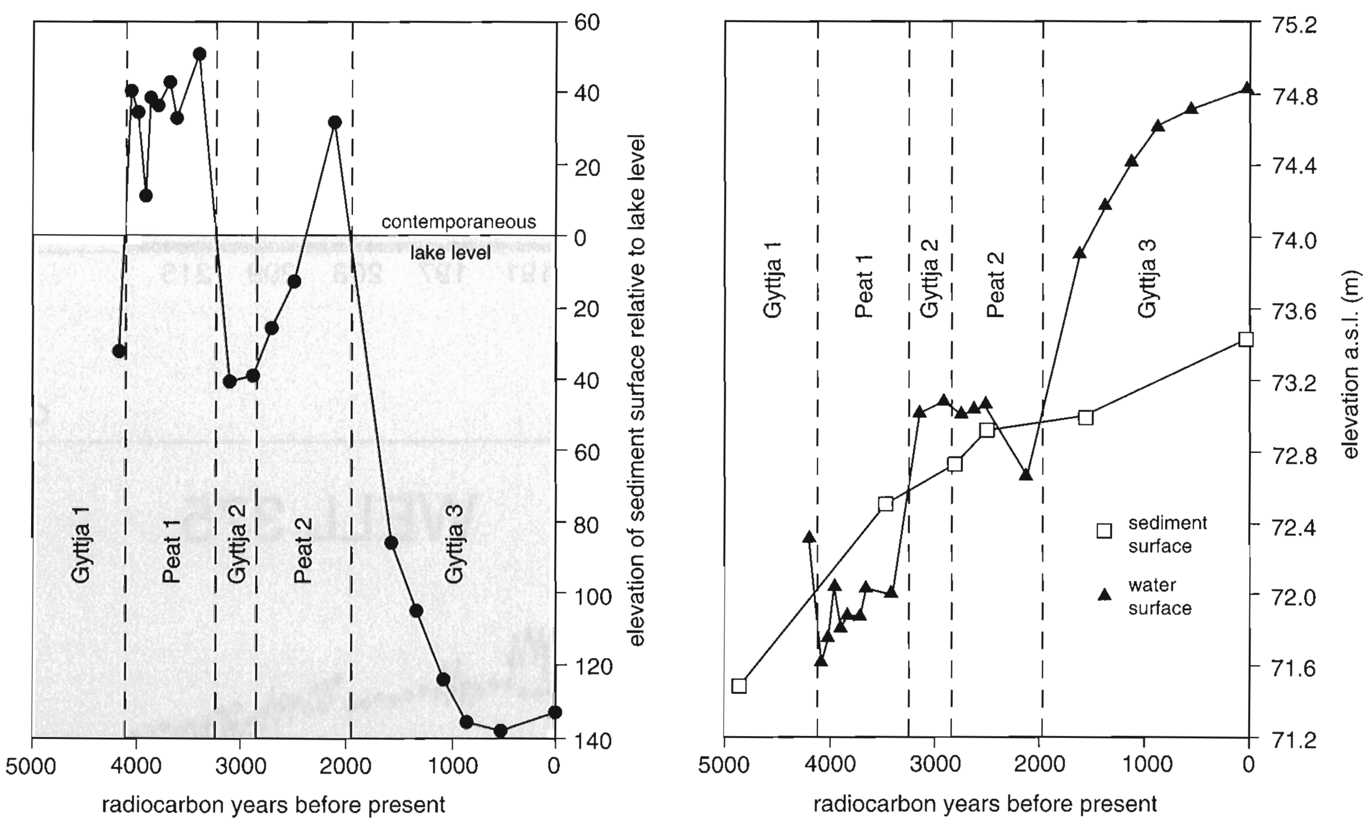


Figure 60. Lake-level curve (triangles) as reconstructed from the stratigraphy of the Cataraqui lagoon, using the organic content-water depth relationship of Figure 59 (from Dalrymple and Carey, 1990).

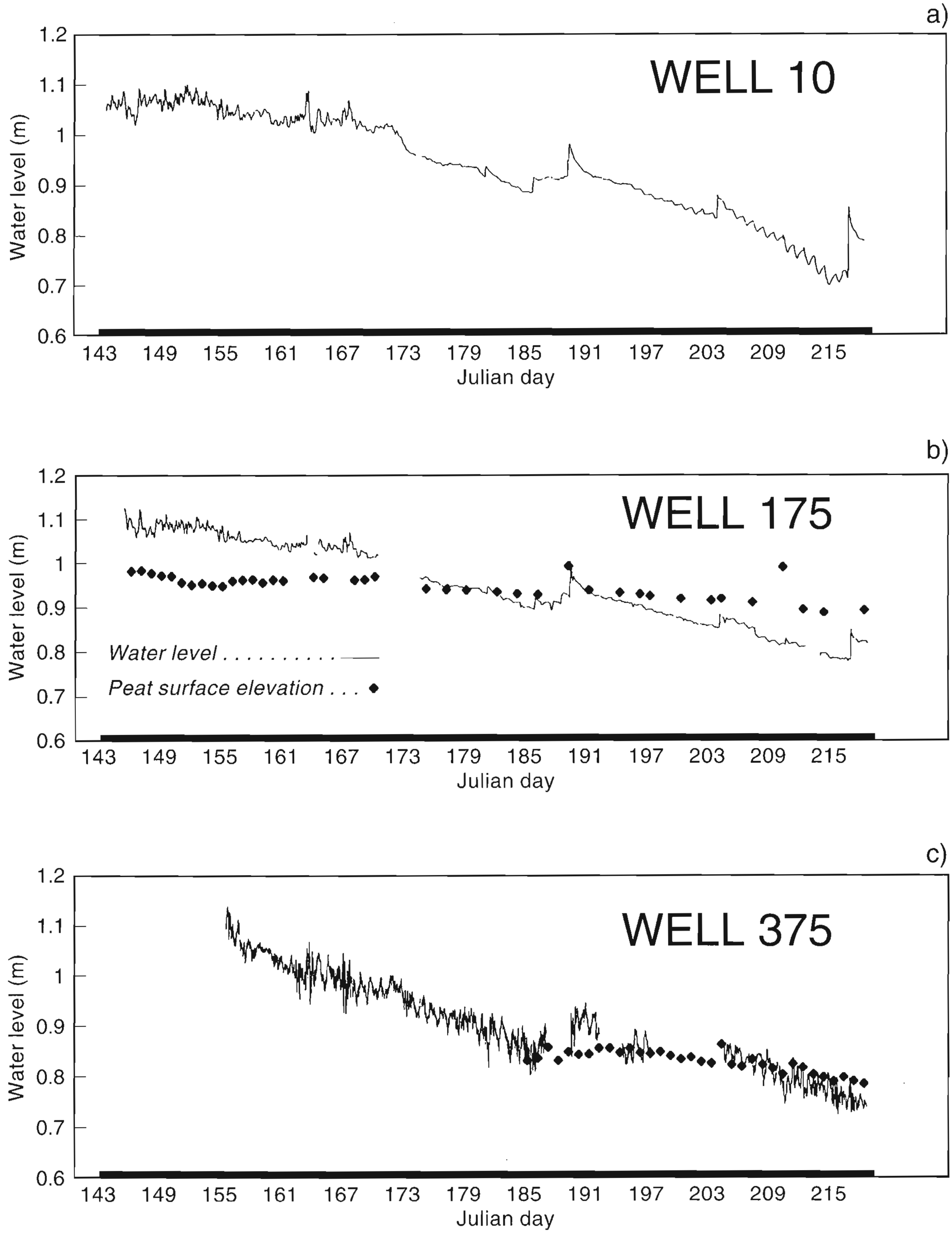


Figure 61. *Temporal variation of water surface (solid line) and marsh surface elevation (diamonds) at three sites on the Bayfield Bay marsh, Wolfe Island:* ***a)*** *10 m from the adjacent land;* ***b)*** *175 m from land; and* ***c)*** *375 m from land, at the marsh margin (J.S. Price, unpub. data, 1992). The measurement period extends from late May to mid-August, 1991.*

☑ *STOP 15* Hallville subaqueous fan

George Gorrell

NTS 31G/4, UTM 570005

The Hallville esker and fan complex is located halfway between Ottawa and the St. Lawrence River. To arrive at the site, turn east on Highway 43 from the junction of highways 16 and 43. Travel east on Highway 43 for approximately 5.5 km and turn left (north) on Boundary Road. Continue north on Boundary Road for approximately 1 km and turn right (east) onto the paved Township Road. Travel east on this road for 2.5 km and turn into the Redmond/Permanent Concrete Pit (after checking with the pit manager) on the north side of the road.

Glaciofluvial ridges composed of sorted sediment are located throughout eastern Ontario (Fig. 2). Many of the ridges are buried or confined by thick sequences of clay. The extent of these sedimentary features was determined by geophysical studies, drilling, and examining sections in pits. Several authors have described portions of these deposits and related the information to depositional models (cf., Gorrell and Shaw, 1991; Rust, 1977; Rust and Romanelli, 1975; Richard, 1982).

The Redmond pit is found on the eastern side of one of these glaciofluvial ridges. The ridge is one of the highest areas between Ottawa and the St. Lawrence River (approximately 116 m a.s.l.). Bedrock in this pit lies at a depth of about 56 m below the surface and rises to the south (Fig. 62).

Description

An esker ridge trends northeast from the exposed face. Proximal gravels exposed near the ridge contain a variety of sediment types which suggest considerable fluctuation in the current velocity at the time of deposition: (1) fining upward cycles of open-work gravel, (2) matrix-supported gravel, (3) beds of massive coarse sand and fine gravel, and (4) massive gravel fining upward into massive coarse to medium sand. Deformation structures (e.g., convolutions), probably formed due to failure as ice melted or as flow scouring occurred, are also present.

A few steep-walled scours in sediments immediately downcurrent of the gravels are filled with one of two sequences: (1) fining-upward cycles, from medium to fine gravel to medium to coarse sand, and (2) diffusely bedded sand which appears massive on close examination. However, faint grading is visible when viewed from a distance. Both sequences contain "rip-up" blocks or clasts of adjacent sediment. Two possible explanations for these sequences are scour and fill by mass flows or scouring and filling by

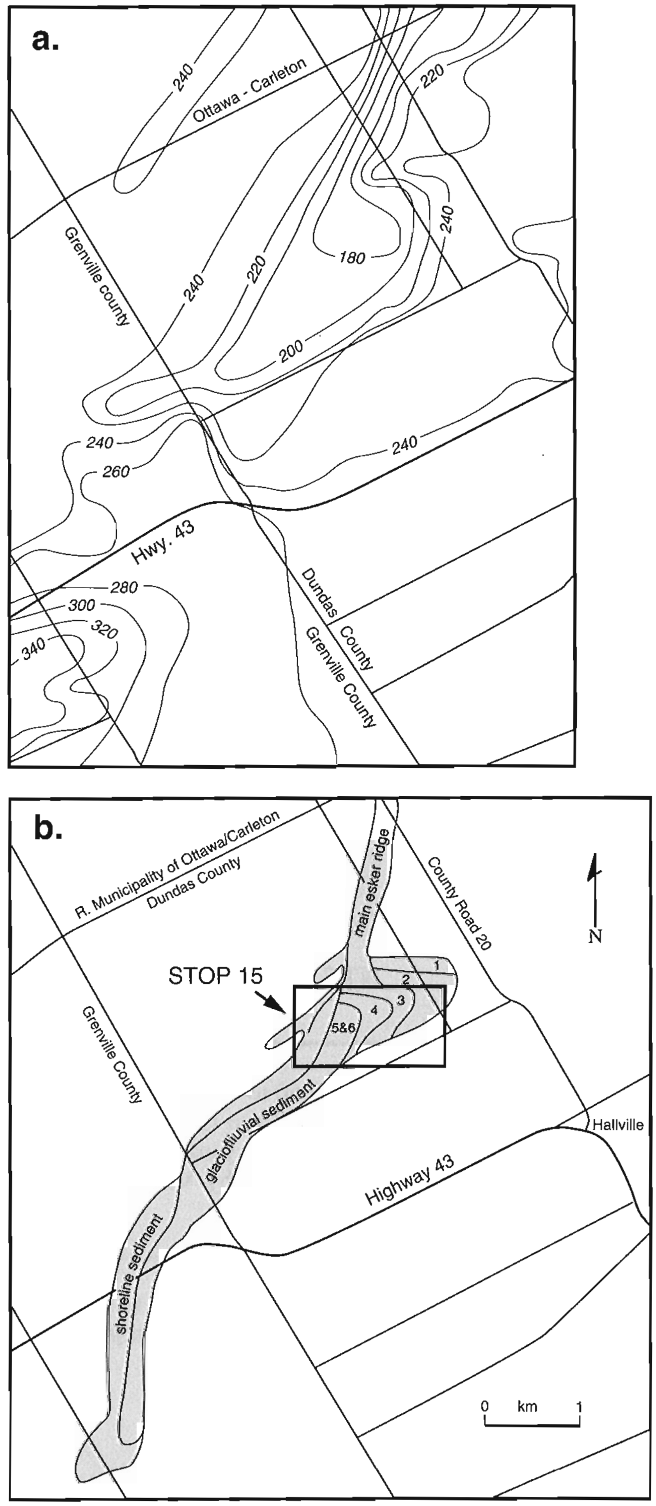


Figure 62. *Location of Stop 15.* ***a)*** *Contour map (m a.s.l.) showing eroded depression beneath,* ***b)*** *Hallville glaciofluvial ridge. Numbers represent lobes.*

high-concentration turbidity currents (Gorrell and Shaw, 1991). The second of these explanations is considered more likely. Other features of interest include:

1) three dimensional wave forms, possibly in phase with waves in the depositing flow;
2) distinct gravel ridges separated from the fan by a clay-rich diamicton;
3) regressive ripples on steep-sided scours;
4) mid-fan fining-upward sequences of massive and/or diffusely bedded sands, through crossbeds, plane beds, tabular crossbeds, and crosslamination (in some sections abrupt paleocurrent changes may be due to channel migration or coalescence);
5) many deformation structures resulting from shocks or rapid sedimentation and associated high pore water pressures; and
6) a marine unconformity truncating the glaciofluvial sequence (3-5 m of sediment have been removed).

Interpretation and discussion

Beneath the esker portion of most of these ridges, channels 4 to 7 m deep have been cut into the underlying bedrock (Fig. 62a). These channels, which also extend beneath the central portions of the fans, are interpreted as Nye (N)-channels (Nye, 1973) cut beneath the glacier bed. Bedrock elevation in the area (derived from 147 well logs) is up to 12 m lower in the central portion of the landform than on its periphery (Fig. 62a).

When the glacier stagnated, the bedrock channel funnelled meltwater and a tunnel (R-channel) was cut in overlying ice. Within a glacier, discharge peaks seasonally and high basal water pressures occur within the subice N-channel. The high water pressure causes the glacier to lift from the bed at channel bends, with the result that smaller channels were cut nearby. This changed the subglacial channel system from one large tunnel to one composed of many secondary or minor channels. Deposition in this channel system produced a central esker ridge and smaller gravel ridges with at least six distinct fans which can be traced to the ridges (Fig. 62b).

The central conduit initially carried large quantities of sediment and water which debouched from the conduit as prolonged or quasicontinuous high-concentration turbidity currents. Rapid deposition resulted from flow expansion and flow deceleration. As the main conduit filled with sediment, meltwater was diverted to the secondary conduits. Prolonged turbidity currents flowed beyond these secondary conduits, cutting the large scours and depositing fill sequences (points 3-5 above) within them. Large volumes of water were released periodically as the system developed. These large channels/scours were cut when meltwater was suddenly released (see Fig. 69).

☑ *STOP 16* Cantley erosional forms and glaciomarine sediments

David Sharpe and John Shaw

NTS 31G/12, UTM 438049

Proceed east on Highway 50 to Gatineau and take Highway 307 (Archambault) to Cantley village; continue north for 5 km to the quarry on the west side of the road. The quarry is owned by Vern Pageau; access is at your own risk.

This stop describes glacial erosional forms cut by powerful subglacial meltwater flow (Sharpe and Shaw, 1989) within a tributary valley of the Gatineau River (Fig. 2).

Description of features

The features described here are eroded into Precambrian marble with included resistant granitic or volcanic clasts. Regional ice flow was approximately north to south. The area was deglaciated about 12 000 BP but until recently, the forms were covered by gravel, sand, and mud deposited rapidly on subaqueous fans formed at the margin of the Champlain Sea (Sharpe, 1988). Similar erosional forms occur on other outcrops in the valley but only 10% of the area has exposed outcrop. Sculpted forms completely cover the streamlined outcrops in this quarry but they diminish and disappear with elevation on outcrops adjacent to and approximately 20-30 m above site level. This pattern of erosion and the position of the streamlined outcrop in the valley bottom suggests flow confined to a valley width of several kilometres.

The studied outcrop is streamlined parallel to regional ice flow. Much of the rock surface is striated and planed-off, such that the granitic and volcanic inclusions lie flush with the surface of the surrounding marble. The erosional forms appear as obstacle marks (Allen, 1982) or ridges in the lee of obstacles (Fig. 63), channels or cavettos, and sichelwannen or crescentic scours (Dahl, 1965), now termed S-forms

(Kor et al., 1991). They consist of depressions and channels (Fig. 64) that are defined by sharp rims, smooth inner surfaces, divergent flow features, and remnant ridges. These forms are found on lee, lateral, and overhung rock surfaces. Carbonate precipitate is also found on polished surfaces in places.

The most common erosional forms are obstacle marks (Fig. 52c; Allen, 1982). These consist of a proximal crescentic furrow wrapped around an upstanding obstacle of resistant bedrock. The proximal furrow commonly has a sharp leading margin. The arms of the crescentic furrow extend leewards in a pair of furrows that become shallower and wider downflow (Fig. 63). The furrows are often smooth or less striated than adjacent surfaces outside the furrow. The remnant ridge, referred to as a rat tail (Prest, 1983), occurs behind the obstacle and between the furrows. The furrows may contain small rat tails.

Interpretation

From the evidence of striations on the surfaces of some marks, it is concluded that the erosional forms at Cantley were created subglacially. Planed-off inclusions indicate glacial abrasion. Obstacle marks are thought to result from glaciofluvial processes (Sharpe and Shaw, 1989). The marks are closely related to features formed by flow separation and horseshoe vortices resulting from a vertical pressure gradient generated at the upstream face of the obstacle. This gradient sets up secondary flow and a pair of oppositely rotating vortices (Fig. 52c). Such secondary flows reattached to the bed where high velocity fluid approached at a high angle and caused maximum erosion (Allen, 1971, 1982). The crescentic furrow cut around the leading side of the obstacle and the paired furrows extending downflow can be easily seen as products of a horseshoe vortex.

As the furrows deepened, flow separation would have occurred together with the powerful vortices. These vortices expanded rapidly and thus, would have been reduced in intensity and erosional power downstream, with the consequence that the furrows would have become broader and shallower downflow (Fig. 63). As a result, rat tails, the remnant (rock) ridges between the furrows, would have become narrower and lower.

The smaller scale forms are identical to some of the sculpted fluvial forms in terrain subject to flooding in Australia (e.g., Baker and Pickup, 1986), which supports the interpretation of formation by water erosion.

Ice abrasion forms

The outcrops show ice abrasion forms, striations, and plucked forms such as gouges and crescentic fractures. The occurrence of abrasion, pitting, polishing, and carbonate precipitate with meltwater forms suggests that the meltwater flows were subglacial. Lifting of ice from its bed by fast-flowing meltwater temporarily suspended glacial abrasion. When ice settled back on the bed as the meltwater flow subsided, abrasion resumed, rounding sharp edges and striating rock faces.

The association of forms produced both by glaciofluvial erosion and ice abrasion suggests that the glacier was lifted from and let down on the bed during each subglacial flood. Depositional sequences related to these catastrophic meltwater outbursts were probably laid down in the adjacent Champlain Sea basin (Sharpe, 1988).

The assemblage of sculpted features at Cantley is best explained by differential erosion produced by strong vortices. Rapid, sediment-laden, turbulent, subglacial meltwater flows likely produced the forms by corrasion and cavitation (Sharpe and Shaw, 1989).

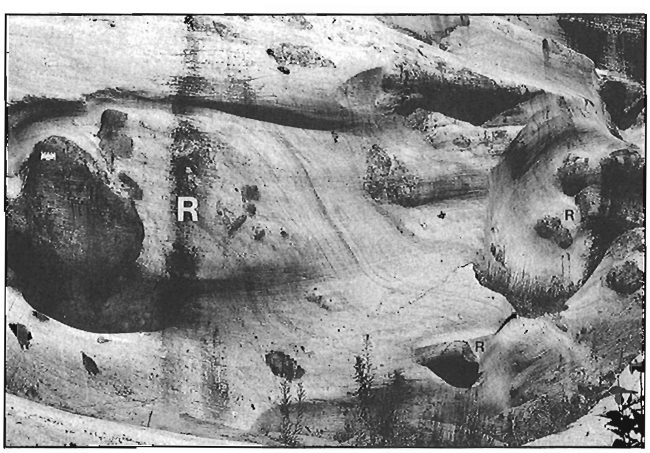


Figure 63. Crescentic obstacle marks – sichelwannen with remnants ridges (R) – in lee of obstacles. Flow was left to right. Card on dark rock is 8 cm long. GSC 1993-165A

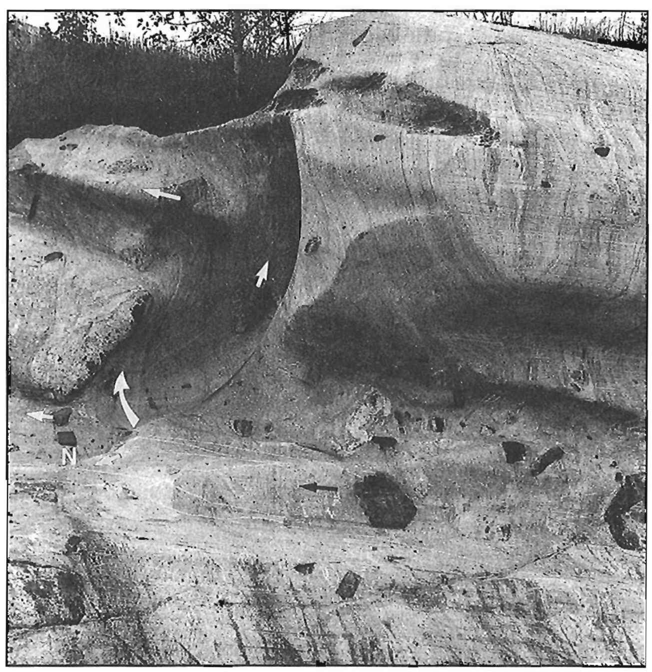


Figure 64. Channel pattern formed by complex flow (arrows) defined by remnant ridges in lee of obstacles. Notebook (N) is 20 cm long (from Sharpe and Shaw, 1989). GSC 1993-165B

Glaciomarine sediment

At the north end of the site a 15 m thick sequence of glaciomarine sediments overlies the eroded rock surface. The beds consist of a fining upwards sequence of sandy gravel and bedded sand overlain by silt and laminated clay. Marine fossils such as *Portlandia arctica, Macoma balthica,* and *Mytilus edulis* may be found at the site.

These sediments were deposited into the Champlain Sea as subaqueous fans at an elevation of about 170 m a.s.l., whereas the local marine limit is about 200 m a.s.l.

☑ *STOP 17* Lac Deschênes acoustic profiles and cores

William W. Shilts

NTS 31G/5, UTM 275299

Lac Deschênes is conveniently accessed from marinas in Aylmer, Quebec; Britannia in west Ottawa; and Port-o-Call Marina at Baskins Beach. The postglacial and recent history of the lake and its implications for early Great Lakes drainage history can be discussed conveniently at Pinhey Point, a municipal park in Kanata. To reach Pinhey Point, follow old Highway 17 from Kanata northwest through South March to the Dunrobin turnoff 2 km northwest of South March. After turning right, make a second right turn and continue across the Constance Lake depression northeast to the Ottawa River where the road makes a sharp turn to the northwest. The entrance to Pinhey Point, on the right (northeast) side of the road is about 4 km from this turn and is clearly marked.

Lac Deschênes is a lacustrine reach of the Ottawa River extending upstream from its bedrock threshold, Deschênes Rapids, northwestward for about 30 km, ending west of Quyon, Quebec. Its surface lies well below the Champlain Sea limit and was once an arm of this sea. Subbottom acoustic profile and side-scan sonar records indicate a complex seismostratigraphic history. Two deep, continuously cored holes were drilled to bedrock from a winter ice platform. Several shorter cores were also collected by scuba diving and drilling from ice. The two long cores were X-rayed and sampled in detail. Micropaleontological and geochemical analyses were done to assess fresh water and marine environments of deposition and provenance signals.

The lake can be divided into three basins of similar length (Fig. 65). The uppermost, northerly basin is relatively shallow (<15 m) with sand covering Precambrian outcrop. The sand carries megaripples which range in size from 0.25 m to sand waves over 2 m high.

The middle section, Constance Bay to Aylmer Island, has a shallow (<10 m), gullied shelf on the northeast and a deep (>50 m), sharp-sided channel on the southwest. The shelf is underlain by two series of acoustically parallel, laminated silty clay sequences separated by a ubiquitous unconformity. Where visible, the sharp-sided channel cuts through both laminated sequences (Fig. 66a).

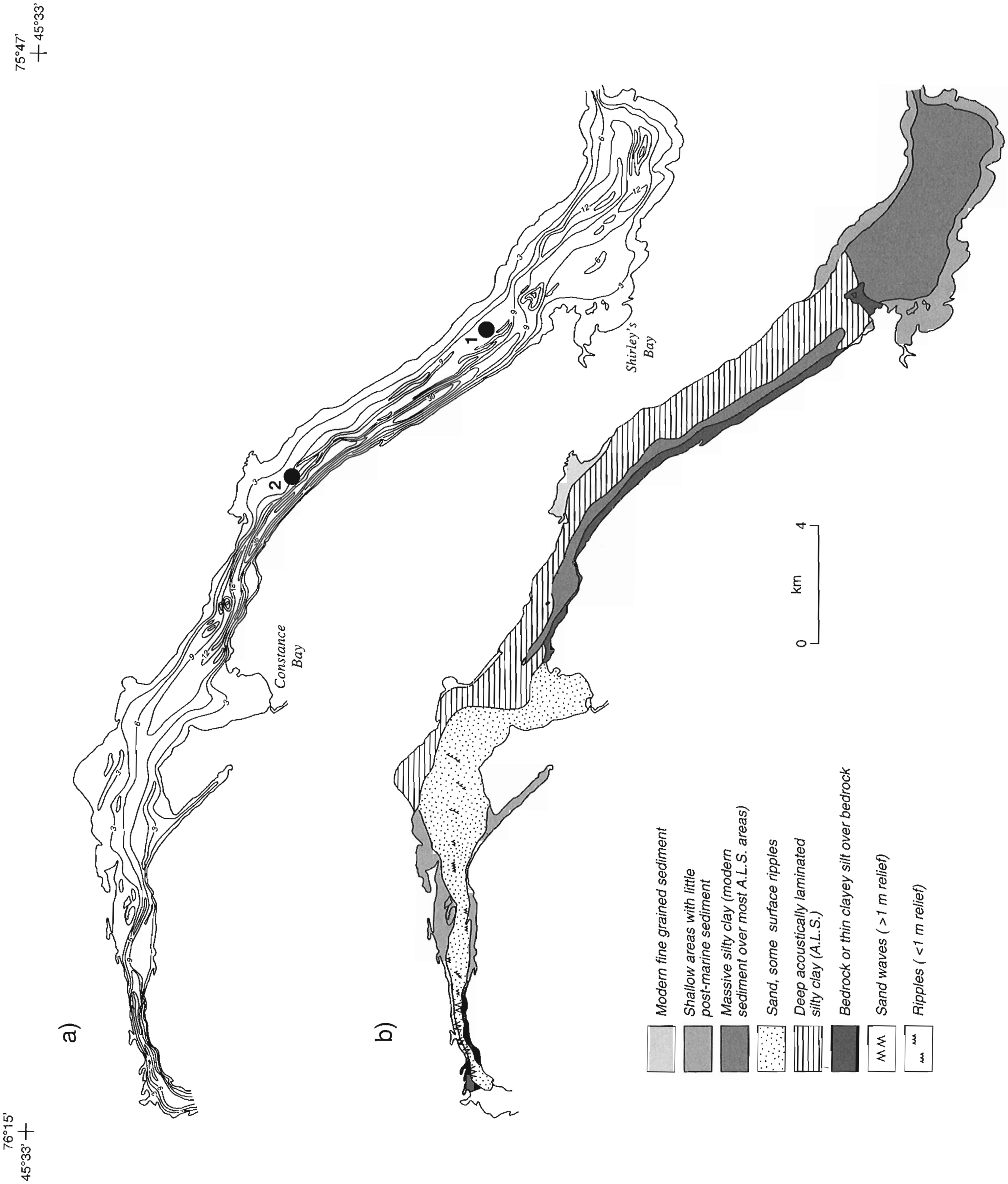
The downstream end of the lake consists of a gently rolling bottom of organic-rich, acoustically massive silt and clay, overlying a low-relief bedrock surface that protrudes through the clay blanket but its shoreward side is largely bedrock (Fig. 65).

Cores collected on the shelf (2) and in the middle of the buried bedrock valley (1, Fig. 65), reveal a sedimentary sequence consisting of a chaotically-bedded basal unit, thick near the valley axis and thinner on the flanks, overlain by a massive, nearly structureless marine unit with shell fragments and a compacted, sandy, worm-burrowed upper boundary (Fig. 66b). In both cores, the hard upper boundary corresponds with the widespread acoustic unconformity noted in the basin. The compacted zone is overlain by massive to faintly laminated silty clay, in turn overlain by massive silty clay with some organic fragments.

Faunal zones in the cores correspond with both the sedimentary stratigraphy and seismostratigraphy. The lowermost, turbiditic zone contains a sparse microfauna and low salinity indicators. The massive unit overlying the turbidites has a compacted upper boundary and has high concentrations of foraminiferal tests from fauna tolerant of high salinity (24-34 ppt). Fauna in the acoustically laminated sediments overlying the compacted zone revert to low salinity forms. The uppermost silty clay is devoid of foraminifera.

Chlorine increases from 200 to 400 ppm in the lower turbidite sequence to 800-1000 ppm in the high-salinity sequence. In the parallel laminated sequence it declines from

Figure 65. a) Bathymetry and (b) sediment types of Lac Deschênes. 1 = core 1; 2 = core 2. ALS = acoustically laminated sediment.


75°47'
45°33'
76°15'
45°33'
a)
b)
Constance Bay
Shirley's Bay
1
2
0
km
4
Modern fine grained sediment
Shallow areas with little post-marine sediment
Massive silty clay (modern sediment over most A.L.S. areas)
Sand, some surface ripples
Deep acoustically laminated silty clay (A.L.S.)
Bedrock or thin clayey silt over bedrock
Sand waves (>1 m relief)
Ripples (<1 m relief)

bottom to top from 600-700 ppm to 300 ppm and drops to <200 ppm in the uppermost silty clay layer. Noncarbonate carbon is not detectable in the lower two units, increases to just over detection limit to 0.3-0.4% in the upper parallel laminated unit, and surpasses 2.0% in the upper silty clay unit. Carbonate is enriched in the basal turbidite zone and in the transition zone at the unconformity.

Element pairs with geochemical affinities in bedrock, Cr-Ni and Zn-Pb, have virtually identical concentration curves in the cores. This suggests that resedimentation of glacially crushed bedrock detritus is primarily responsible for the geochemical signature of the various sedimentary packages. Local geochemical effect was minor.

Interpretation

The Deschênes cores and seismic records appear to provide data that constrain paleogeographic interpretations for the upper Ottawa valley (e.g., Lewis and Anderson, 1989; Teller, 1985). Several authors have proposed that meltwater discharge from glacial Lake Agassiz in addition to normal outflow from the upper Great Lakes passed through the Ottawa valley into the Champlain Sea.

Given this general model for drainage evolution in the Ottawa valley, the following sequence of events is inferred from the various geological, geophysical, geochemical, and micropaleontological data from Lac Deschênes:

1) The lowermost chaotic beds were deposited by density underflow from conduits at the base of the retreating glacier front, standing in 100 m of water (cf. Fig. 56c). Sediments pooled in the deeper depressions, thus accounting for their greater thickness (core 2) near the axis of the preglacial bedrock valley, compared to the its flank (core 1, Fig. 65). Rapid sedimentation and proximal glacial conditions account for distorted bedding, high sand content, low chlorine, foraminiferal species tolerant of low salinities, and low density of foraminiferal tests observed in the cores.
2) With retreat of the ice from the basin, more distal marine conditions prevailed. Little isostatic recovery and weak currents meant that fine grained sediments were deposited. Increased chlorine concentrations, a fauna tolerant of high salinity, high concentrations of foraminiferal tests, low sand content, and low noncarbonate carbon concentrations all confirm open marine conditions.
3) The acoustic unconformity at the top of the marine sequence in Lac Deschênes represents erosion by enhanced flow of fresh water draining from the upper Great Lakes and from glacial Lake Agassiz into the deep northwest arm of Champlain Sea. This unconformity forms a sharp transition between high and low salinity faunal assemblages. The unconformity cuts across acoustic laminae of the earlier high-salinity sedimentary sequence. Organic (noncarbonate) carbon increases and chlorine concentration begins to decrease above this unconformity.
4) The acoustically parallel laminated sequence marks the return of normal estuarine conditions in the northwestern area of Champlain Sea. The reason for the return to estuarine conditions was the temporary deflection into the Mississippi River system of Lake Agassiz and upper Great Lakes drainage by the Marquette readvance into the Lake Superior basin. The low salinity foraminiferal assemblage in the parallel laminated unit indicates more fresh water. Isostatic uplift forced more fresh water into the Lac Deschênes basin compared to the earlier marine phase when shorelines and river mouths were further

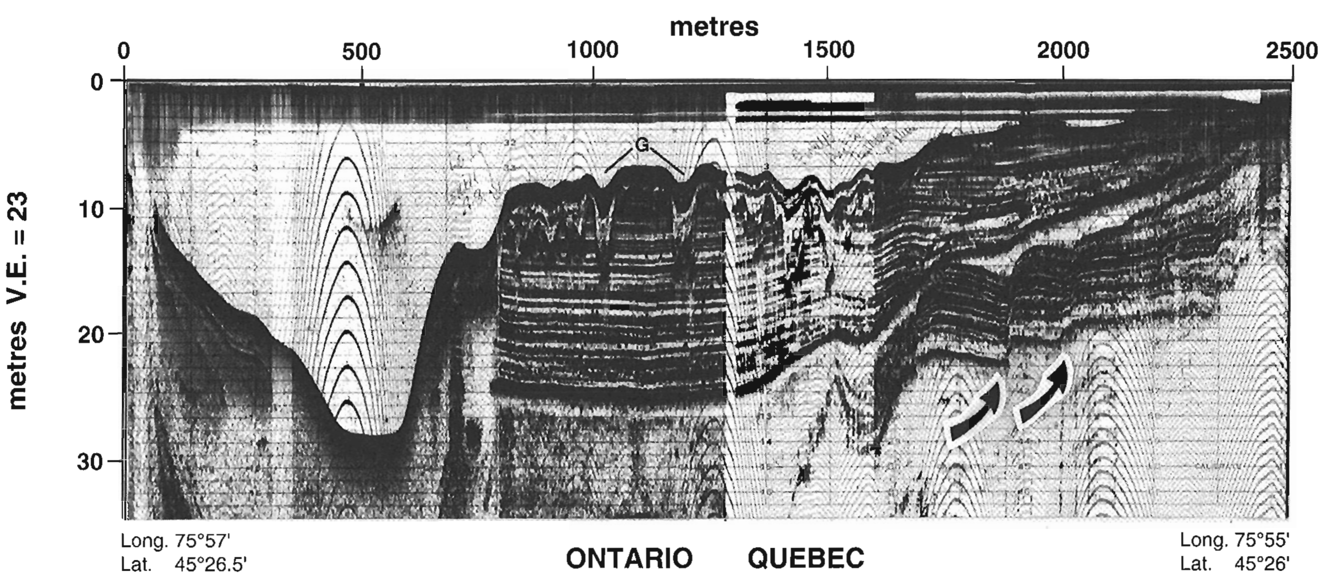


Figure 66. a) Typical profile across central basin of Lac Deschênes. Arrows indicate sharp sediment flexures over Paleozoic limestone outcrops. Note angular unconformity between upper parallel laminated sequence and lower massive to laminated sequence; G = gullies.

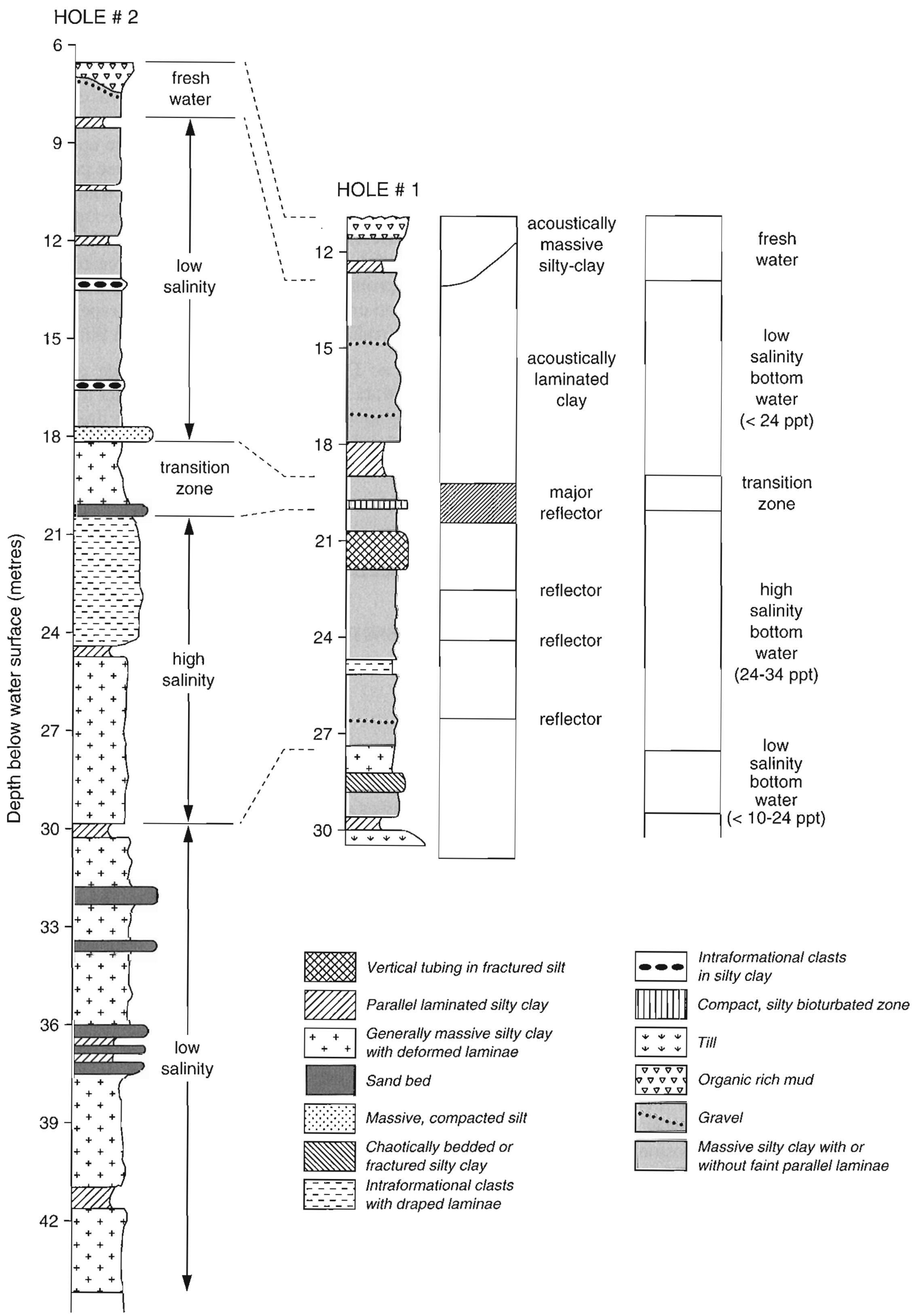


Figure 66. b) *Sediment logs, cores 1 and 2.*

away. Increasing noncarbonate carbon and decreasing chlorine concentrations also indicate increasing fresh water input.

5) Estuarine deposition was modified by renewed flow from Lake Agassiz and the upper Great Lakes, following retreat of the ice after the Marquette event. The results of this erosional event in shallower Lac Deschênes were greater than the gentle truncation of the first overflow event. Fine grained sediment was stripped from the Ottawa channel, only being preserved in pockets where protected by nearby bedrock highs. In the main channel of Lac Deschênes, an erosional trench, over 50 m deep at its upper end and shallowing to less than 3 m at Aylmer Island, was cut into marine sediment at its contact with the steep bedrock. Though crenulated by gullies cut by the strong paleocurrents (Fig. 66), the sediment in the main channel between Constance and Shirley's bays largely survived to form the bench that now flanks the Deschênes trench along its northeast side.

6) Eventually the Nipissing overflow was stopped by isostatic uplift. The Champlain Sea was far east of Ottawa, and bedrock of Deschênes Rapids ponded water in Lake Deschênes to its present level and configuration. Fine grained sediment deposited in this reach was likely contributed by tributaries eroding the thick, subaerially exposed, marine sediments. The observations that those sediments are draped over the scoured surfaces and the reverse gradient on the main channel in central Lac Deschênes suggest that modern currents cannot be invoked to account for the scoured surface. Even the sand waves that are present in the upper part of the lake are thought to be relict from these periods of enhanced flow. Their size is incompatible with the gentle currents presently depositing sediments throughout the rest of Lac Deschênes. The lowermost third of Lac Deschênes, east of Aylmer Island, was swept free of marine sediment, probably by the last overflow event, and now consists of an undulating bedrock surface draped by post-overflow fresh water silty clay from 0-5 m thick.

Lac Deschênes presents a unique opportunity to link disparate types of geoscience data that lends itself to paleogeographic reconstruction of the early drainage history of the Great Lakes.

☑ *STOP 18* Lanark esker, bead, and fan assemblage

George Gorrell

NTS 31F/1, UTM 300500

The Lanark site is located northeast of Perth. The site is reached most easily by taking Highway 7 to Perth. At the traffic lights on the west side of Perth turn right (north) on Highway 511 to Lanark. At the T-intersection turn right and travel for about 3 km. The pit is the second active pit on the north (left) side of the road.

Three kilometres northeast of Lanark, a complex consisting of a single esker ridge, multiple ridges and beads, and a series of ridges and fans (Fig. 67) is found. The complex is exposed in an active sand and gravel deposit revealing spectacular sedimentary sections. The sedimentary structures, facies, stratigraphic relationships, bedrock erosion, and paleocurrent variability indicate that the complex formed due to a combination of meltwater hydraulics and the position of the ice margin (Gorrell and Shaw, 1991).

The esker is a steep-sided, northerly-trending ridge 1.85 km long and about 8 m high. Its width ranges from 70-130 m with the narrowest segment to the north. The ridge broadens to the south where it bends abruptly southwest and then southeast.

The esker ridge and bead complex trends northeast and occupies an area of approximately 900 by 300 m. The main esker, which winds between four beads, is up to 30 m high and 60 m wide. Seismic surveys show that the esker is centred over a bedrock channel. The beads are separated by depressions and are connected to the esker by minor ridges. All the landforms in this complex have steep sides.

The esker/fan complex trends northeast. It is approximately 1.5 km long and varies in width from 0.75 km in the northeast to 2 km in the southwest. The fans, like the beads, flank the main esker ridge. The surface relief varies considerably but is generally more subdued than the esker/bead complex.

The esker

To aid discussion a model of formation of the esker system is shown (Fig. 68).

When glacier ice covered the area and melt started to develop, meltwater flow was restricted to small conduits incised into the underlying bedrock. These Nye-channels (Nye, 1973) are common on the pit floor. They are up to 7 m deep, but because they are filled principally with over-sized boulders and because they are narrow and therefore difficult to excavate, inspection of channels is limited.

Where exposed, the central portion of the channel is covered with sculpted (S-) forms (Kor et al., 1991). These forms become progressively smaller suggesting that, at least in the earlier period of development, when the capacity of the meltwater to keep open a channel within the ice itself (Röthlisberger (R)-channel) was limited, meltwater flow was restricted to channels cut in the substrata (N-channels, cf. Nye, 1973; Walder and Hallet, 1979). As the volume of meltwater at the base of the glacier increased sufficiently to exceed the overburden pressure and the deformation of the ice, R-channels could develop over N-channels. Deposition on the bed of these R-channels initially reduced their cross-sectional area, giving rise to increased flow velocity at constant discharge. Increased meltwater velocity in turn increased melting. As a result, the conduit expanded upwards from the N-channel in a process resembling cave expansion.

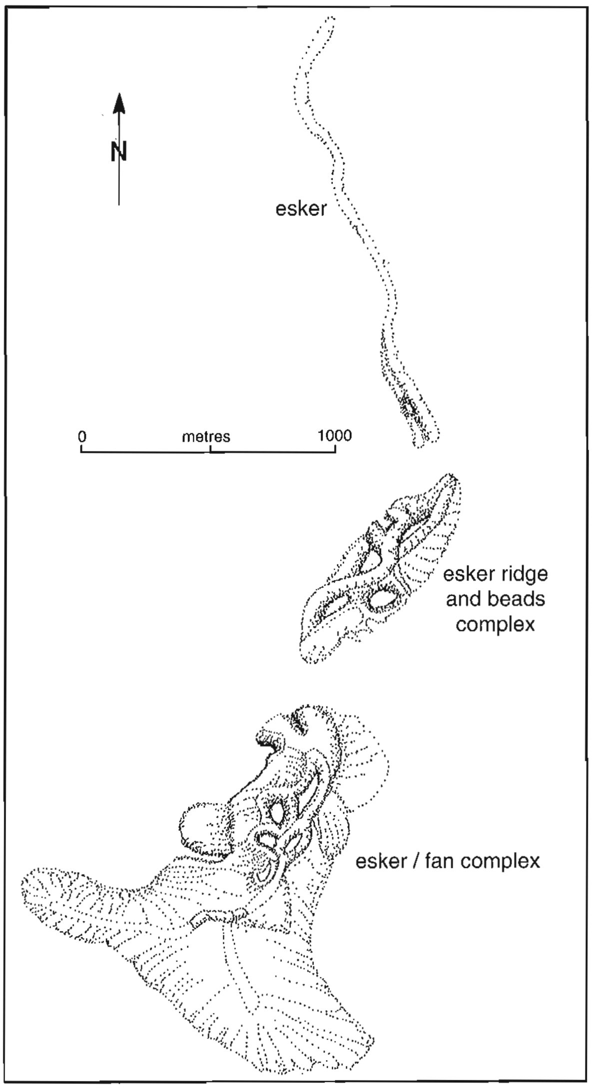


Figure 67. Landform and sediment complexes at Lanark, Stop 18 (from Gorrell and Shaw, 1991).

Three gravel units were identified in the esker. The lower unit within the N-channel consists of boulders with a very fine matrix. Overlying this unit is a multimodal gravel with a-axes transverse and b-axes dipping upflow. This gravel fabric indicates that the clasts were transported as bed load by rolling (Rust, 1972; Johansson, 1976).

The upper unit, where present, consists of multimodal gravel with a random to very slight tendency to a-axes parallel and dipping upstream. The sediment arrangement resembles the disorganized gravels from turbidites (Walker, 1975) and the "sliding bed" facies of Saunderson (1977). It was probably deposited from hyperconcentrated flows.

The esker/beads

The beads are small elongate hillocks, 15-25 m high and 120-150 m wide. They lie close to bends in the esker and are connected to them by short ridges (Fig. 67). Steep-sided depressions separate the beads. Massive and crudely stratified pebble and cobble gravel make up the proximal portion of the beads, which fine downflow to successions including graded and massive coarse sand, crudely stratified medium sand, crossbedded medium sand, and crosslaminated fine sand. These beds are stacked with no preferred vertical arrangement. Faults are common throughout the succession and, in a few cases, large-scale over-folding may be observed.

The faulting and large over-folds suggest that the deposit was not only laterally supported but formed subglacially and in cavities as shown diagramatically in Figure 69. The abrupt reduction in grain size from the proximal to distal portion of the bead may indicate that there was a rapid reduction in competence through the cavity or alternatively, there was a differential transport of coarse and fine sediment caused by rapid aggradation.

The esker/fan complex

The morphology of the esker/fan complex is more complicated and more subtle than either the esker or beads. An anabranching esker extends through the fan complex with small ridges converging on it and diverging from it (Fig. 68). These ridges contain sedimentary sequences fining upward from massive coarse to medium sands to crosslaminated fine sand (Fig. 69). Many fining-upwards sequences exhibit rip-up clasts and load and flame structures near their bases. These sedimentary sequences suggest that the flow events started abruptly and gradually decreased in power.

Fans are located adjacent to the main esker (Fig. 67, 68). Most fan sediment is much finer than that within the esker. Grain size and the sedimentary structure change dramatically downfan. Sediment in the proximal portion of the fans consists of thick sequences of normally and inversely graded coarse grained sand and medium gravel. In many places open-work, fining-upward sets of cobbles and pebbles may

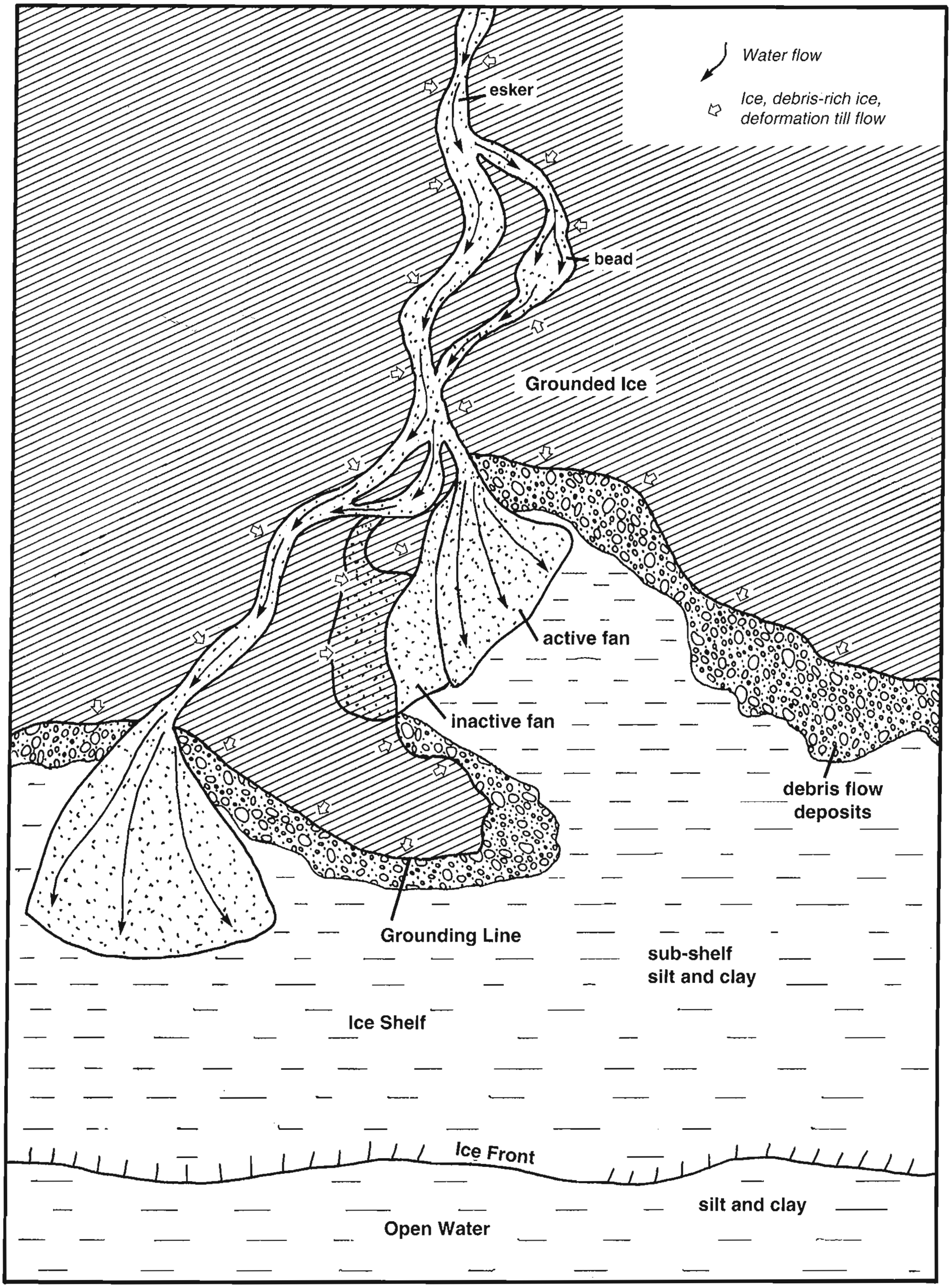


Figure 68. Reconstruction of the depositional environment for the Lanark complex (from Gorrell and Shaw, 1991).

be seen. These are succeeded downfan by large scours with near vertical sides and filled with massive or diffusely bedded medium sand. These scours are succeeded downflow by dunes with preserved stoss-side bedding which are overlain by crosslaminated fine sand, plane, massive and graded beds of medium sand in the midfan zone. Thick cosets of climbing ripple crosslamination with whole bedforms (supercritical) are preserved in the distal fan zone.

The sedimentary sequences indicate that a large number of flow events built the fans. These events or fluctuations within them, gave rise to many distinct fining-upward sedimentary sequences. In the proximal part of the fan, the large steep-walled scours filled with massive to diffusely bedded sand are interpreted to have been eroded and filled in the vicinity of hydraulic jumps in hyperconcentrated turbidity currents. These turbidity currents originated as submerged jets emerging from a submerged glacial tunnel (Fig. 69) found at the grounding line of a small ice marginal shelf (Fig. 68).

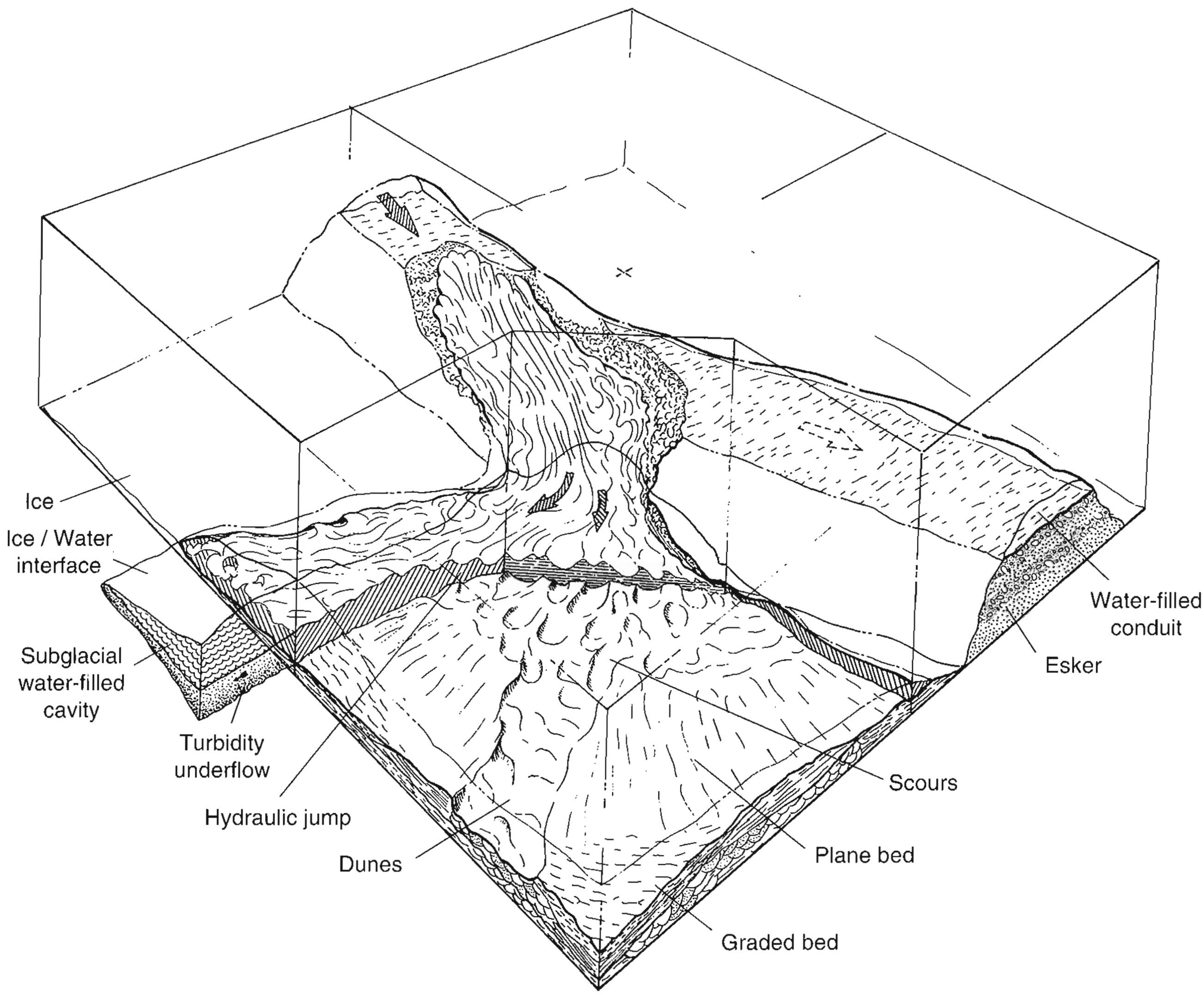


Figure 69. *Schematic interpretation of the formation of fans at Lanark (from Gorrell and Shaw, 1991).*

☑ STOP 19 Peterborough drumlin field

David Sharpe

NTS 31D/8, UTM 205095

Location A: Travel east on Highway 7 to Highway 34, turn south to Keene (about 15 km). Travel east and then south on Lakeview Road to a gravel pit about 5 km east of the village. Check with Mr. Loucks, Highlandview Farms south of pit.

Location B: Return to Highway 7 and travel east 2 km to a pit owned by Mr. Hoekstra on the south side of the highway. Check with the pit operator.

The Peterborough drumlin field (Gravenor, 1957) comprises about 10 000 drumlins covering 900 km^2 on Paleozoic limestone (Fig. 2). The field, oriented northeasterly, mainly lies north of the Oak Ridges Moraine, but in the east appears to underlie the moraine (Fig. 70), a ridge composed of thick sequences of glaciofluvial and glaciolacustrine sediment (Duckworth, 1979). The field may have transitional elements to the Oak Ridges Moraine (tunnel valleys and eskers) but the drumlins themselves are clearly older than the moraine (see Stop 20).

Some Peterborough drumlins have "carved limestone cores" covered by thin diamictons; others are composed of pre-existing sorted and unsorted sediment, in places "carved" from lake sediment (Gravenor, 1957). Sand and gravel ridges north of the Oak Ridges Moraine were also shaped into drumlins (Gravenor, 1957), however there is no sediment deformation within these drumlins (Sharpe, 1987). Although a majority of exposures in drumlins reveal till (Gravenor, 1957; Shaw and Sharpe, 1987a), the drumlin forms themselves may be the product of glaciofluvial erosion (Shaw and Sharpe, 1987b) or erosion by subice deformation according to Boyce and Eyles (1991).

The Peterborough drumlin field consists of small subfields of elongate spindles and less common parabolic forms that occur mainly in groups or aggregates. Long narrow flutings are present on the tops of drumlins and on uplands between drumlins (Shaw and Sharpe, 1987b), commonly as a tail on upflow rises (f, Fig. 71). In places, crescentic scours, or depressions occur prominently on the upflow side of these rises (s, Fig. 71).

Rice Lake drumlin (location A, Fig. 70)

Sections in an abandoned gravel pit expose 10 m of sediment in a 20 m high drumlin (Shaw and Sharpe, 1987a). Exposed sediments show no signs of deformation, folding, or thrusting and they consist entirely of sorted sand and gravel with no diamicton. The upper 5 m consist of large tabular cross-sets of pebbly sand. This is truncated by a channel filled with ripple crosslaminated sand indicating southwesterly paleoflow. In places, silty beds cap the rippled strata.

Interpretation

Gravenor (1957) mapped the Peterborough drumlin field and identified drumlin ridges at this site as kame moraine, emphasizing their sorted character rather than their drumlin form. It was suggested that some drumlins in the field resulted from glaciofluvial erosion at the sole of the glacier and subsequent fill as sediment was trapped in these areas due to flow expansion (Sharpe, 1987). However, it is not possible to demonstrate a conformable relationship between sediment fill and drumlin form. The ridge may result from subglacial floods which scoured the landscape, producing erosional drumlins (Shaw and Sharpe, 1987b), or alternatively, these ridges may be bedforms developed in the large tunnel valleys in the area (e.g., Shaw and Gilbert, 1990; Shaw and Gorrell, 1991).

Hoekstra pit (location B, Fig. 70)

Measured sections in the Hoekstra pit show upward gradation from gravel to sand and to fine grained sediments; they are capped by and interbedded with thinly bedded to massive diamicton (Shaw and Sharpe, 1987a; Sharpe, 1987). The main sediments consist of gravel, sand, rhythmic silt-clay, and diamictons; a brecciated facies cuts across the beds above the gravel sequence. The lower gravel beds commonly have a matrix filling consisting of clayey silt. Contacts between the drumlin sediments are gradational and the beds are generally undeformed and possibly conformable with the drumlin form. Some steeply dipping beds, however, are not conformable with drumlin form.

A large saddle structure occurs in the centre of the exposed beds, revealing diamicton-rich sediments on its flank and rhythmic sediments in the depression (Sharpe, 1987). The rhythmic beds consist of normally graded silt-clay units (2-20 mm) within about 14 thicker graded cycles, 0.2 to 0.5 m thick. The thick cycles have sand at the base with rare ripples (southwest paleoflow) and they are capped by 2 cm of clay. Diamictons overlie and are interbedded with the rhythmic beds (Sharpe, 1987).

Rip-ups, clasts of diamicton, and clay-silt balls occur within the rhythmites. Some folding and minor normal faulting is present, particularly near clastic dykes and diapirstructures found just above the gravel beds. These deformation structures are bounded by undisturbed beds and thus are syndepositional.

At the south end of the pit, new exposure reveals several 1-2 m sets of graded sand and gravelly sand with low-angle inclined stratification or cross-stratification showing polymodal and bimodal textures. Some sets show gravel with mud. The sequence forms a broad arched deposit or apparent macroform. Finer sediments (discussed above) drape the trough beside the macroform. These beds may be similar to polymodal and bimodal deposits (Stops 1 and 11) and indicative of migrating bedforms in a subglacial or near-marginal high-energy flow.

Interpretation

Drumlin sediment appears to record quasicontinuous sedimentation of gravel, sand, and rhythmites containing interbeds of diamicton and sorted sediment. Coarse gravel beds and massive sand indicate high-energy flow, and the mud matrix in gravel beds indicates a subaqueous environment of deposition. Massive sand was deposited by suspension in high-energy turbidity flows or high-concentration mass flows. The rhythmites were deposited in depressions on the sand and gravel surface, and they represent an annual or shorter cyclicity.

The drumlin sediments in the Hoekstra pit show no evidence of moulding or deformation by overriding ice. Above the interbedded diamicton/sorted bed sequence, some deformation may have occurred but no large bed stress was transmitted to lower sediments. Thus passive erosion of the drumlin sediments is required to explain the lack of deformation (Shaw and Sharpe, 1987b).

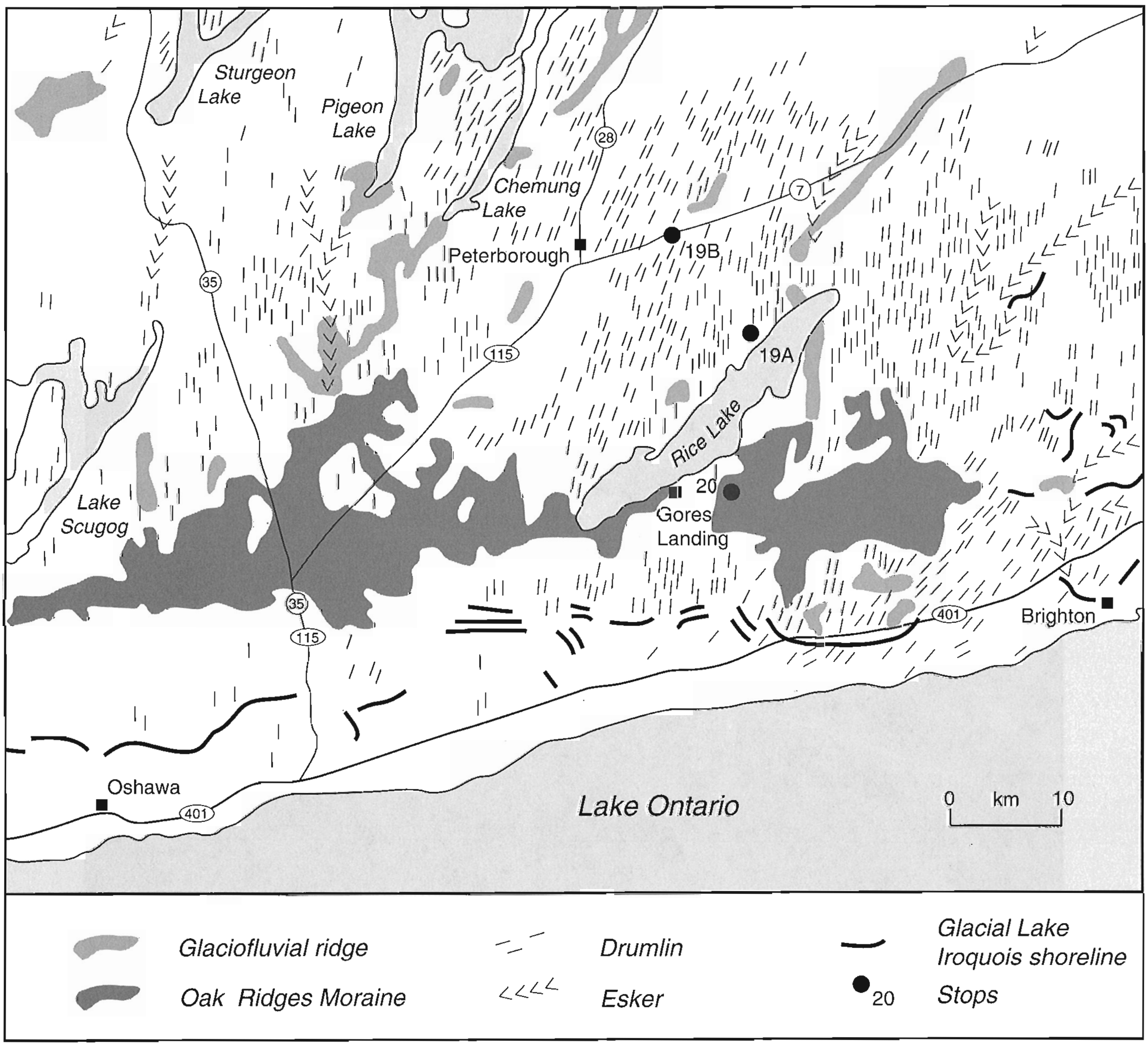


Figure 70. Map of eastern portion of Oak Ridges Moraine showing landform assemblages near stops 19 and 20.

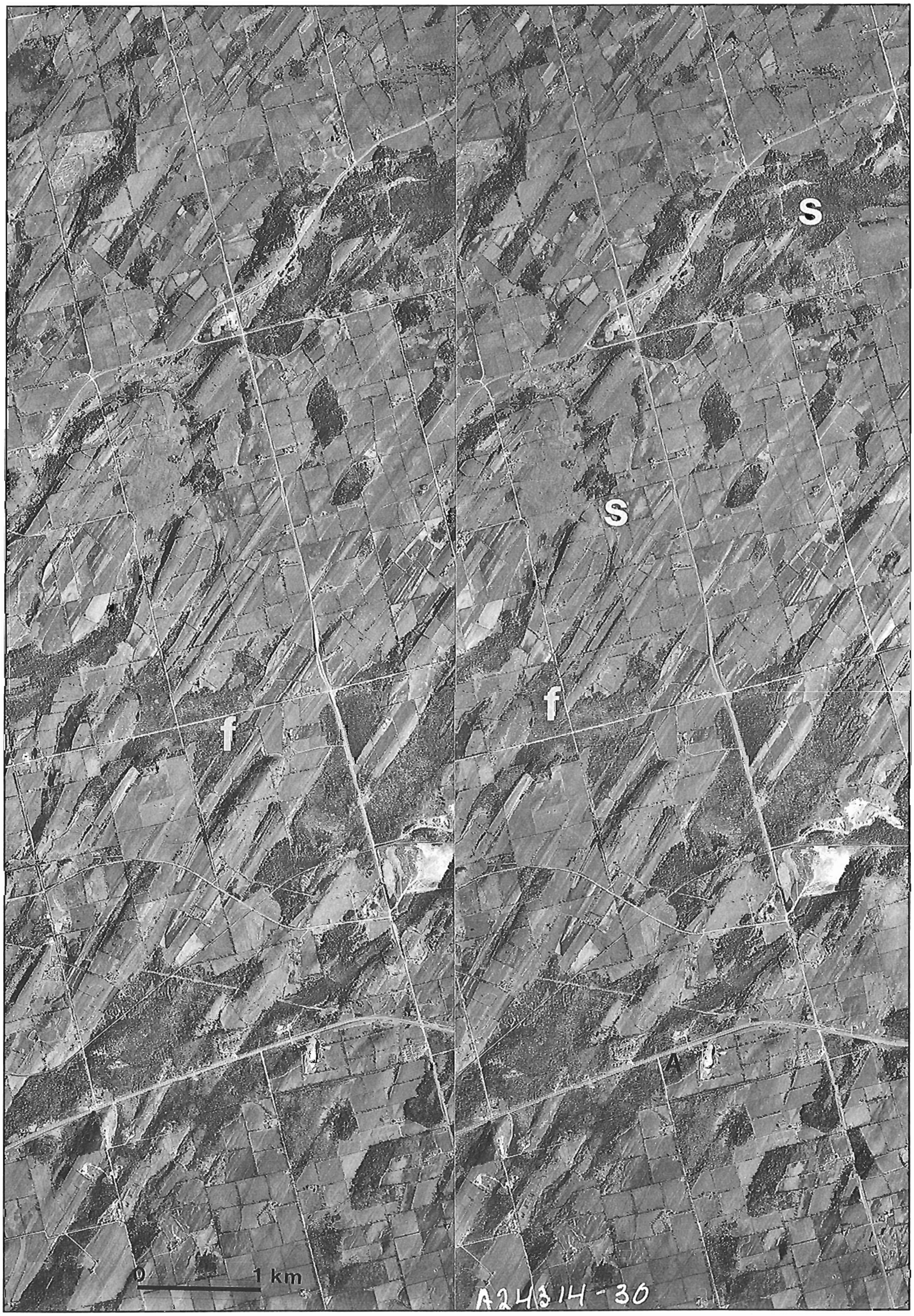


Figure 71. Photo stereopair showing Peterborough drumlins near Stop 19B (A). Note crescentic scours (s) and long narrow flutings (f).

Discussion

Removal of thick interdrumlin sediment is required by erosional models of drumlin formation whether "carved" by subice deforming diamicton (Boyce and Eyles, 1991) or eroded by subglacial meltwater flow (Shaw and Sharpe, 1987b). The proposed ice margin for a till carving event lies south of the Oak Ridges Moraine and north of Lake Ontario (Boyce and Eyles, 1991). Unfortunately, no thick sediment sequence is found in this location or in adjacent basins (Lewis et al., 1992). Subglacial floods with turbulent flow, however, explain the crescentic scours, drumlin forms, and superimposed flutings or rat tails (Shaw and Sharpe, 1987b; Shaw and Gilbert, 1990). Sheetfloods are thought to have removed sediment well beyond the ice margin to the south of the Great Lakes, but as sheetflood evolved to tunnel channels, sediment may have been deposited within the Oak Ridges Moraine.

☑ *STOP 20* Oak Ridges Moraine

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NTS 31D/1, UTM 280905

At Bewedley, located on Highway 28 southeast of Rice Lake, turn east along the County Road through Gores Landing and Harwood 2 km east to the Harden and King property (Wimpy Minerals). The stop is in the eastern portion of the Oak Ridges Moraine, south of Rice Lake, in Haldiman Township, Concession 8 and 9, Lots 30-33 (Fig. 71). See the pit manager for property access.

The Oak Ridges Moraine is a large ridge of mainly glaciofluvial sediment running east-west for 150 km north of Lake Ontario. The moraine straddles the Peterborough drumlin field as indicated by drumlins that are almost buried by moraine sediments. Several large tunnel valleys (e.g. lakes Scugog, Pigeon, Chemung, Rice, etc. Fig. 70) cut across the drumlins and lead into the north side of the moraine. The character and origin of the Oak Ridges Moraine are being investigated and these aspects are briefly discussed for the eastern segment of the moraine.

Near this stop the moraine consists of three wedge-shaped forms (Fig. 72). Each becomes broader to the south and has its apex at the mouth of a tunnel valley.

A series of small tunnel valleys leading into the moraine contains a transitional sequence of landforms (Fig. 72). More than 5 km north of the moraine the tunnel valleys contain eskers and fans. Closer to the moraine, eskers broaden into valley fills similar to outwash fans. The valley fills grade upwards from extensive sheets of gravel to thick sets of climbing ripple crosslamination. Valley fills grade southward into a series of large streamlined landforms about 3 km north of the moraine.

At the stop, streamlined forms are set in tunnel valleys as they merge with the Oak Ridges Moraine. The streamlined forms are at least 20 m high and 50 to 200 m long, forming broad tear-drop shaped ridges pointing to the south. These ridges consist of sand and gravel partly concealed by sheets of finer sediments, which may be the distal equivalent of coarser valley fill sediments to the north. Alternatively, this sediment may have been deposited in quiescent periods after formation of the moraine.

Downflow where the tunnel valleys meet the moraine, there are dunes of medium to coarse sand and gravel capped in many places by thin diamicton. The dunes have stoss side laminae perserved, suggesting high rates of sedimentation and very high suspension deposition. Based on borehole logs, buried dunes are entirely composed of these sorted

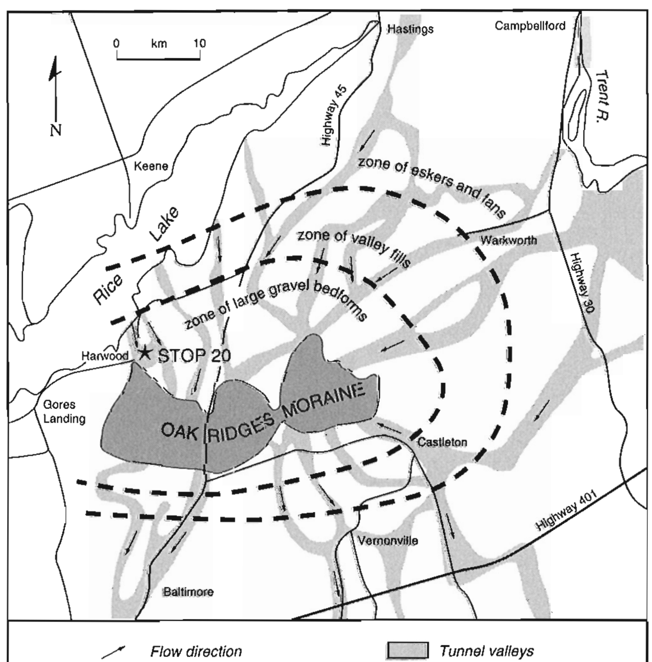

Figure 72. Schematic diagram showing tunnel valleys north and south of the Oak Ridges Moraine and the inset transitional landforms.

sediments and are at least 60 m thick. Where these forms lose definition in the moraine, deformation stuctures such as load structures and fluidized dykes may be seen in the sorted sediments within the forms.

On the south side of the moraine there is a series of smaller valleys. The distribution of bedforms in the southern valleys mirrors that in tunnel valleys north of the moraine. Streamlined forms are found closest to the moraine. These grade downflow to valley fills which in turn grade to eskers and fans.

Interpretation

As discussed at Stop 19, the drumlins, which underlie the moraine ridge and the transitional landforms found in the tunnel valley, may represent the erosion of existing sheetflow deposits (Shaw and Sharpe, 1987b; Shaw and Gilbert, 1990). An alternative explanation of drumlin origin suggests that deforming sediment may have eroded the drumlin forms (Boyce and Eyles, 1991), but this is at odds with the evidence from the extensive tunnel channels and their associated bedforms.

The tunnel valleys represent erosional remnants of catastropic subglacial flows (Barnett, 1990; Shaw and Gorrell, 1991). Much of the eroded sediment passed through the channels; some passed through the area of the present moraine to channels on the south side. Waning flow led to deposition in the tunnel valleys apparently producing the streamlined forms, analogous to the bedforms at Trout Creek (see Stop 1; Shaw and Gorrell, 1991). A series of lobes 4-5 km long and up to 6 km wide (Fig. 72) formed where the tunnel valleys merge with subaqueous fan sediment of the Oak Ridges Moraine.

Apparently, sedimentation within the main tunnel valleys blocked flow but in other places, flow may have been deflected around the fans or later erosion produced valleys which cut through the moraine. On a larger scale (Fig. 71), the Oak Ridges Moraine may have originated as a series of larger fans formed at points where tunnel valleys converge to produce flow expansion (e.g., Sharpe and Cowan, 1990).

Most interpretations of the Oak Ridges Moraine recognize ice support north and south of the ridge; however, the above observation related to mirroring of bedforms in tunnel channels north and south of the moraine leads to interpretations of a subglacial origin for both tunnel valleys and channel deposits in this location. Some tunnel channels may predate the moraine but the similarity of bedforms in tunnel channels north and south of the moraine suggests sedimentation contemporaneous with moraine formation.

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