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QUATERNARY GEOLOGY OF THE OTTAWA REGION, ONTARIO AND QUEBEC

edited by
R.J. Fulton

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Cover illustration

View up the Ottawa Valley from near Breckenridge. The flat valley floor is underlain by Champlain Sea sediments, locally capped by fluvial sands; the fault-line scarp on the right is the Eardley Escarpment which in this region marks the boundary between the Ottawa Valley (St. Lawrence Lowlands) and the Gatineau Hills (Laurentian Highlands). GSC 200183-D

Vue sur la vallée des Outaouais près de Breckenridge. Le fond de la vallée est constitué de sédiments de la mer Champlain et localement de sables fluviatiles; la ligne de faille à droite est l'escarpement Eardley qui dans cette région se trouve à délimiter la vallée des Outaouais (basses-terres du St-Laurent) et les collines de la Gatineau (hautes-terres des Laurentides). GSC 200183-D

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QUATERNARY GEOLOGY OF THE OTTAWA REGION, ONTARIO AND QUEBEC

Abstract

The Ottawa region is a lowland area, underlain largely by flat-lying bedrock consisting of Paleozoic carbonates and shales, enclosed on the north, west, and south by uplands underlain by Precambrian igneous and metamorphic rocks. The contrasting sedimentary and crystalline bedrock have produced till facies which can be readily characterized by carbonate content. In addition, occurrence of areas of Precambrian rocks which contain distinctive trace element components has produced chemically distinct trains of till. Till, which occurs as a thin and discontinuous cover, is the main Quaternary sediment in the uplands. Quaternary sediments in the lowlands generally consist of fine grained marine sediments and fluvial and offlap marine sands. Eskers, kames, and outwash trains are common features in valleys of upland areas. In the lowlands glaciofluvial deposits do not occur as distinct, easily traced features because they were buried by later sediments and were reworked by marine erosion during regression of the Champlain Sea. Most glaciofluvial deposits in the lowlands contain features which suggest that they were deposited largely as subaqueous outwash. Glacial deposits and most marine sediments are Late Wisconsinan and fluvial deposits are mainly Holocene.

During the single recognized Late Wisconsinan advance, ice moved in a general north to south direction but three major lobes, separated by thin belts of glaciofluvial sediments, developed during deglaciation. Early during the retreat of ice from the lowlands, ice marginal lakes entered the area from the Lake Ontario and Lake Champlain basins. At about 12 ka (according to marine shell dates) marine waters of the Champlain Sea replaced the freshwater in the depressed lowlands.

Sediments deposited in the Champlain Sea consist of clays, silty clays, and sands which can in general be related to deposition in a shoaling basin. The deposits contain a variety of assemblages of macrofossils, foraminifers, and ostracodes which indicate that the Champlain Sea was a salinity stratified water body. By at least 11.3 ka the basin was invaded by water with salinities as high as 34‰ which formed a high salinity wedge. Later, during recession, salinities gradually decreased.

Tundra-woodland vegetation was present around the basin at the time of deglaciation, but a more closed forest developed soon after. Spruce was the first tree species to dominate the forest in southern parts of the area but poplar was the first to colonize areas immediately west and north of the Champlain Sea basin. In the south the spruce dominated forests were replaced by pine dominated forests, whereas in the north the poplar dominated forest was invaded first by spruce and then by birch before pine became the dominant species. Many of the other species that make up the modern forest invaded the area shortly after 8 ka. A chronology based on pollen stratigraphy suggests that the Champlain Sea invaded the area between about 11 and 11.5 ka. This is several hundred years later than the date of invasion indicated by marine shells. One reason suggested for this discrepancy is a dilution of the sea water carbon by old carbon derived from the melting ice.

Limit of marine submergence varies from about 125 m in upper St. Lawrence River valley to 200 m near Ottawa and about 165 m near the western limit of the Champlain Sea. Uplift apparently occurred at a rate as great as 10 m per century immediately following deglaciation but insufficient data are available to produce good uplift curves. By about 10 ka uplift had caused the Champlain Sea to drain from the area and a major fluvial system had developed in the approximate position of the modern Ottawa River.

Résumé

La région d'Ottawa comprend une zone de basses-terres sous laquelle repose un socle rocheux largement horizontal, composé de roches carbonatées et de schistes argileux paléozoïques; cette zone de basses-terres est entourée au nord, à l'ouest et au sud par des hautes-terres qui recouvrent des roches ignées et métamorphiques du Précambrien. Les socles sédimentaire et cristallin ont produit des faciès de till que l'on identifie facilement à leur teneur en carbonate. En outre, la présence de zones de roches précambriennes qui contiennent des oligo-éléments caractéristiques a donné lieu à des traînées de till chimiquement distinctes. Le till, trouvé sous forme de couverture mince et discontinue, représente le principal sédiment quaternaire des hautes-terres. Les sédiments quaternaires des basses-terres se composent généralement de sédiments marins à grain fin, de sables fluviatiles et de sables marins régressifs. Les eskers, les kames et les épandages fluvio-glaciaires sont répandus dans les vallées des hautes-terres. Dans les basses-terres, les sédiments fluvio-glaciaires ne se manifestent pas sous forme de dépôts distincts, facilement repérables, car ils reposent enfouis sous des sédiments plus récents et ont été remaniés pas l'érosion marine au cours de la régression de la mer de Champlain. Dans les basses-terres, certaines caractéristiques propres à la plupart des sédiments fluvio-glaciaires semblent indiquer que ces derniers se sont accumulés en grande partie sous forme d'épandages sousaquatiques. Les dépôts glaciaires et la plupart des sédiments marins datent du Wisconsinien supérieur, et les sédiments fluviatiles, surtout de l'Holocène.

Au cours de la seule avancée glaciaire connue du Wisconsinien supérieur, la glace s'est déplacée suivant une direction généralement nord-sud; trois grands lobes, séparés les uns des autres par de minces zones de sédiments fluvio-glaciaires, se sont formés au cours de la déglaciation. Au début du retrait glaciaire dans la zone des basses-terres, des lacs proglaciaires ont envahi la région à partir des bassins du lac Ontario et du lac Champlain. La datation de coquillages marins révèle que les eaux de la mer de Champlain ont remplacé l'eau douce dans les basses-terres déprimées il y a environ 12 000 ans.

Les sédiments déposés dans la mer de Champlain se composent d'argiles, d'argiles limoneuses et de sables dont l'accumulation s'apparente, de façon générale, au type de sédimentation rencontré dans les bassins caractérisés par la formation de hauts-fonds. Les dépôts renferment divers assemblages de macrofossiles, de foraminifères et d'ostracodes dont la présence indique que la mer de Champlain contenait des zones de salinité distincte. Peu de temps après sa formation, il y a près de 11 300 ans, le bassin a été envahi par des eaux dont le degré de salinité pouvait atteindre 35 ‰; leur intrusion a mené à la formation d'un coin à degré de salinité élevé. Plus tard, au cours de la régression, la salinité a baissé progressivement.

La végétation de type toundra arboréenne croissait dans le bassin à l'époque de la déglaciation mais une forêt de type fermée l'a rapidement remplacée. L'épinette a été la première espèce dominante dans le sud de la région, tandis que le peuplier a été la première espèce à coloniser les régions immédiatement à l'ouest et au nord du bassin de la mer de Champlain. Au sud, les pessières ont été remplacées par des forêts de pins, tandis qu'au nord, la forêt de peupliers a été envahie d'abord par des épinettes, puis par des bouleaux et enfin par des pins. Une bonne partie des autres espèce qui forment la forêt actuelle ont envahi la région il y a un peu moins de 8 000 ans. La stratigraphie palynologique semble indiquer que la mer de Champlain a envahi la région entre 11 et 11.5 ka, soit plusieurs centaines d'années après la date établie à partir de l'analyse des coquillages marins. Une dilution de l'eau de mer par du vieux carbone provenant de la glace en fonte pourrait expliquer cet écart.

La limite de la submersion marine varie de près de 125 m dans la partie amont de la vallée du Saint-Laurent à 200 m près d'Ottawa et à environ 165 m près de la limite ouest de la mer de Champlain. Le soulèvement aurait atteint un rythme maximum de 10 m par siècle immédiatement suite à la déglaciation, mais on dispose de trop peu de données pour produire de bonnes courbes du soulèvement. Il y a environ 10 000 ans, le soulèvement a provoqué le drainage de la mer de Champlain et un important réseau fluviatile s'est formé presqu'au même endroit que la rivière des Outaouais actuelle.

INTRODUCTION

The best documented Quaternary event in the Ottawa region is submergence of the area by the Champlain Sea, a body of water that occupied St. Lawrence and parts of the Ottawa River valleys between 12 and 10 ka. Literally hundreds of ^{14}C dates related to this event are available and deposits and features formed during this episode are widespread. Little is known, however, about Quaternary events prior to retreat of Late Wisconsinan ice (prior to 13 ka) and even though post-Champlain Sea episode river terraces are abundant, their chronology and genesis are virtually unknown.

This report consists of several segments. The first, by N.R. Gadd, covers the geological setting of the Ottawa region, discusses Late Wisconsinan glacier flow patterns, and provides a general description of Quaternary deposits. The next report by I.M. Kettles and W.W. Shilts describes tills with particular emphasis on geochemistry, the origin of unique till geochemical signatures, and examples of how these have been tied to bedrock source areas and glacier flow patterns. The next segment, by B.R. Rust, is a brief review of what is known about the subaqueous outwash deposits in the immediate vicinity of Ottawa. Studies of these deposits led to extensive use of the term subaqueous outwash and provided descriptions of sedimentary structures that could be used to recognize this significant facies of outwash. A report by C.G. Rodrigues describes the invertebrate macrofaunal and microfaunal assemblages and associations that have been recognized in the area, uses the patterns of successions in the sediments to outline the general evolution of the western basin of the Champlain Sea, and supplies considerable data related to Champlain Sea chronology. The segment prepared by R.J. Fulton and S.H. Richard uses the available radiocarbon dates to establish a chronology of late Pleistocene and early Holocene events. The final paper by T.W. Anderson looks at pollen stratigraphy and development of Holocene vegetation patterns and uses pollen stratigraphy of marine sediments as a tie between early vegetation migration and development of the Champlain Sea.

An attempt has been made to present a complete report on the Quaternary history of the area. In referencing information in this report however credit should be given to the author(s) of the individual segments. Figure 1 shows most of the localities mentioned in this text and the location of NTS map areas specifically referred to in the contribution by Rodrigues. A synthesis of this report is published as Part I of Fulton (1987).

GEOLOGICAL SETTING AND QUATERNARY DEPOSITS OF THE OTTAWA REGION

N.R. Gadd[1]

Physiography and bedrock geology are the dominant factors that control both the patterns of glaciation and the nature and distribution of Quaternary sediments of the Ottawa region. The distribution of highlands and lowlands determined the patterns of ice flow during glacier advance, and the pattern of flow and location of residual ice blocks during retreat. Texture and composition of glacial sediments are controlled by the composition of bedrock from which they were derived, and the distribution of fine grained waterlain sediments is controlled by location of topographic basins.

PHYSIOGRAPHY AND BEDROCK GEOLOGY

Laurentian Highlands (Bostock, 1970; Fig. 2) rise abruptly to elevations of 400 m along the Eardley Fault escarpment on the north side of Ottawa River valley (Fig. 3) and from there northward consist of rolling hills that gradually rise to elevations of at least 600 m in the vicinity of Maniwaki and Mont-Laurier (Fig. 1). The difference in elevation between the lowland and the area immediately north of the Eardley escarpment is almost 300 m and relief within the highlands in places approaches 200 m.

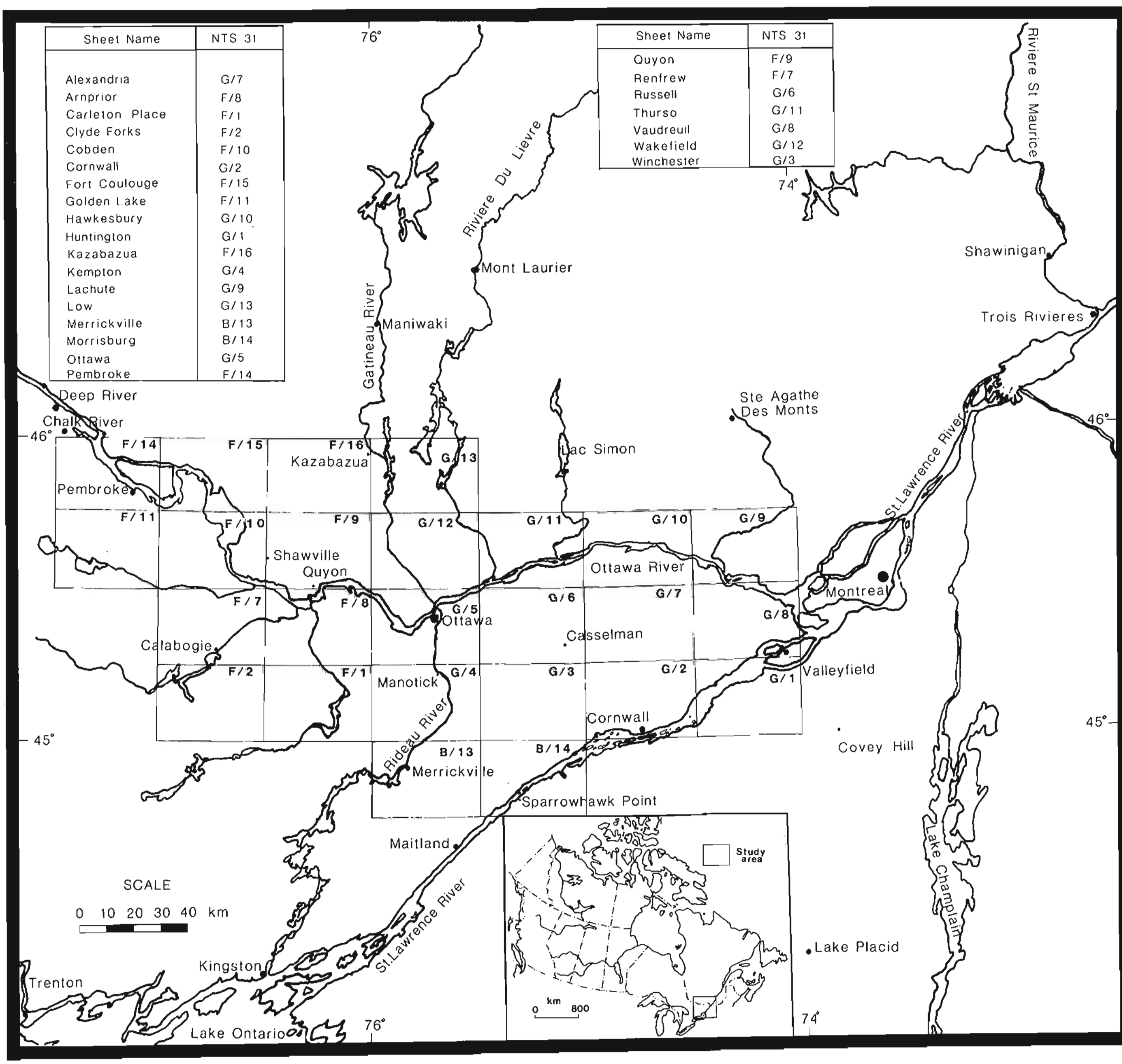

Figure 1. Location map showing the boundaries and names of NTS 1:50 000 scale map areas.

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

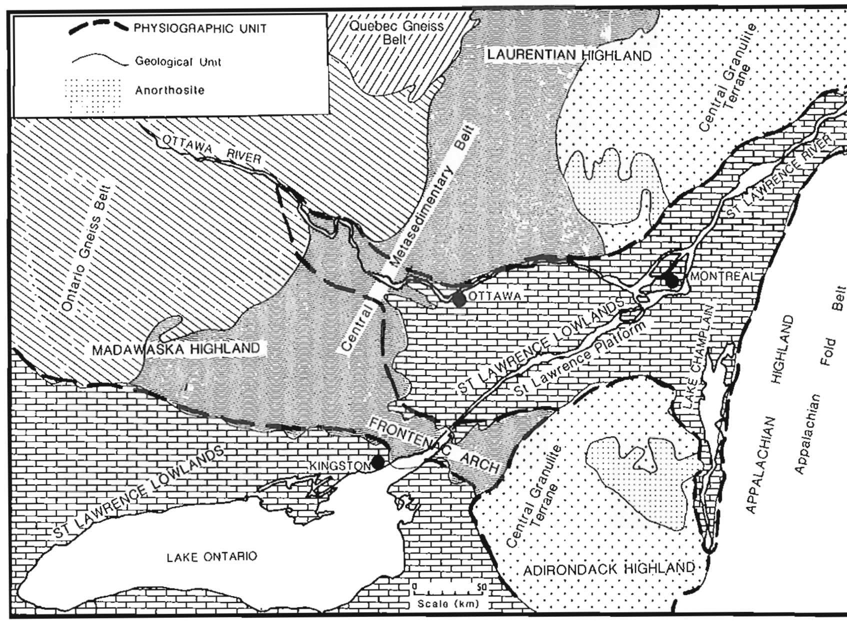

Figure 2

Physiographic and bedrock map of the Ottawa region. Physiographic subdivisions after Bostock (1970); bedrock geology after Sanford et al. (1979).

In the western sector of the area another part of the Laurentian Highlands, known as the Madawaska Highland, is separated from the lowland by major fault systems trending northwesterly (St. Patrick Fault scarp) and southerly (Fernleigh-Clyde Fault) from the vicinity of Calabogie (Fig. 1, 3). Elevations attain 300 m (in some places 450 m) along the St. Patrick Fault scarp and decline southward towards the St. Lawrence Lowlands where Precambrian rocks are overlain, in a minor scarp, by Paleozoic rocks dipping gently towards Lake Ontario (Kay, 1942; Fig. 2). South of St. Lawrence River in New York State, the Adirondack Mountains rise in graduated steps of moderate relief to elevations in excess of 1200 m in the High Peaks area, near Lake Placid (Fig. 1). Bridging the area between Madawaska Highland and the Adirondacks is a low arch of Precambrian rocks, the Frontenac Arch (Fig. 2). This low ridge constitutes the sill at the outlet of Lake Ontario where it is represented by the Thousand Islands in St. Lawrence River. All other areas of the region are underlain by relatively flat lying sedimentary rocks of the St. Lawrence Lowlands characterized by surface elevations which do not exceed 150 m a.s.l. and relief which is generally less than 30 m.

The main structural feature of the region is the Ottawa-Bonnechere Graben system (Kay, 1942; Kumarapeli and Saull, 1966; Kumarapeli, 1985). Upstream from Ottawa, the graben is defined by the Eardley Fault (north) and the St. Patrick Fault (south) that parallel the portion of Ottawa River between Ottawa and Pembroke (Fig. 3). This graben system and related less prominent faults control the location of a preglacial valley in the Ottawa-St. Lawrence system which connects with the "Laurentian Channel" of the Gulf of St. Lawrence. Other related features in the vicinity of Ottawa are the prominent cuesta on the right bank of Ottawa River just east of Ottawa and waterfalls on the Rideau and Ottawa rivers.

Precambrian bedrock in the Gatineau Valley region of the Laurentian Highlands, in the eastern part of the Madawaska Highland, and along the Frontenac Arch are described by Baer et al. (1977) as belonging to the Central Metasedimentary Belt (Fig. 2) which consists of "marble, quartzite, aluminous gneiss and metavolcanics" whose metamorphic grade "ranges from upper greenschist to granulite but is mainly amphibolite". West of Gatineau Valley and in the western Madawaska Highland, highland rocks belong to the Ontario Gneiss Belt (Fig. 2) which includes gneisses that "are dominantly quartzo-feldspathic, pink or grey, well-layered, and mainly at amphibolite, rarely granulite grade of metamorphism". East of the Central Metasedimentary Belt and in the Adirondacks the rocks are described (Baer et al., 1977) as Central Granulite Terrane (Fig. 2) whose rocks are "leucocratic to melanocratic para- and orthogneisses of unknown age". Metasedimentary rocks in these terranes are early Proterozoic (1600-2500 Ma) and Archean (older than 2500 Ma) and the latest phase of metamorphism and mountain building was related to the Grenville Orogeny ca. 1000 Ma.

The St. Lawrence Lowlands, in most of this region, are underlain by sedimentary rocks of the St. Lawrence Platform that lie unconformably on the Precambrian crystalline basement. Baer et al. (1977) described the St. Lawrence Platform rocks as: "Basal, northwesterly transgressive Cambro-Ordovician orthoquartzite....grades upward into Lower Ordovician carbonates. Disconformably overlying Middle Ordovician carbonates are succeeded by southerly thickening black shale and siltstone and the Upper Ordovician Queenston deltaic redbeds....". No younger pre-Quaternary rocks are present in the area. The sedimentary rocks of the St. Lawrence Platform are cut by northwest and northeast trending faults of the Bonnechere graben system which supposedly formed during opening of the Atlantic Ocean in Cretaceous time but which follows a similar, late Precambrian system which is associated with minor intrusives, such as carbonatites.

QUATERNARY DEPOSITS AND HISTORY

Pleistocene deposits consist almost exclusively of glacial and related deposits. One exception is at Pointe-Fortune, on Ottawa River 100 km east of Ottawa, where 10 m of nonglacial sediments occur between two tills (Veillette and Nixon, 1984). These deposits could relate to an Early Wisconsinan Interstade or might even be older.

Wisconsinan glaciation

Wisconsinan glaciation began with early ice accumulation in the highlands of Nouveau-Quebec (Occhietti, 1982). The ice sheet expanded progressively westward and southward, occupied the central St. Lawrence Lowland and Lake Champlain area, and extended into the Ottawa region. In the Laurentian Highlands (Gatineau Hills) north of Ottawa River, a major ice stream in the Gatineau and adjacent subparallel valleys produced an ice lobe that

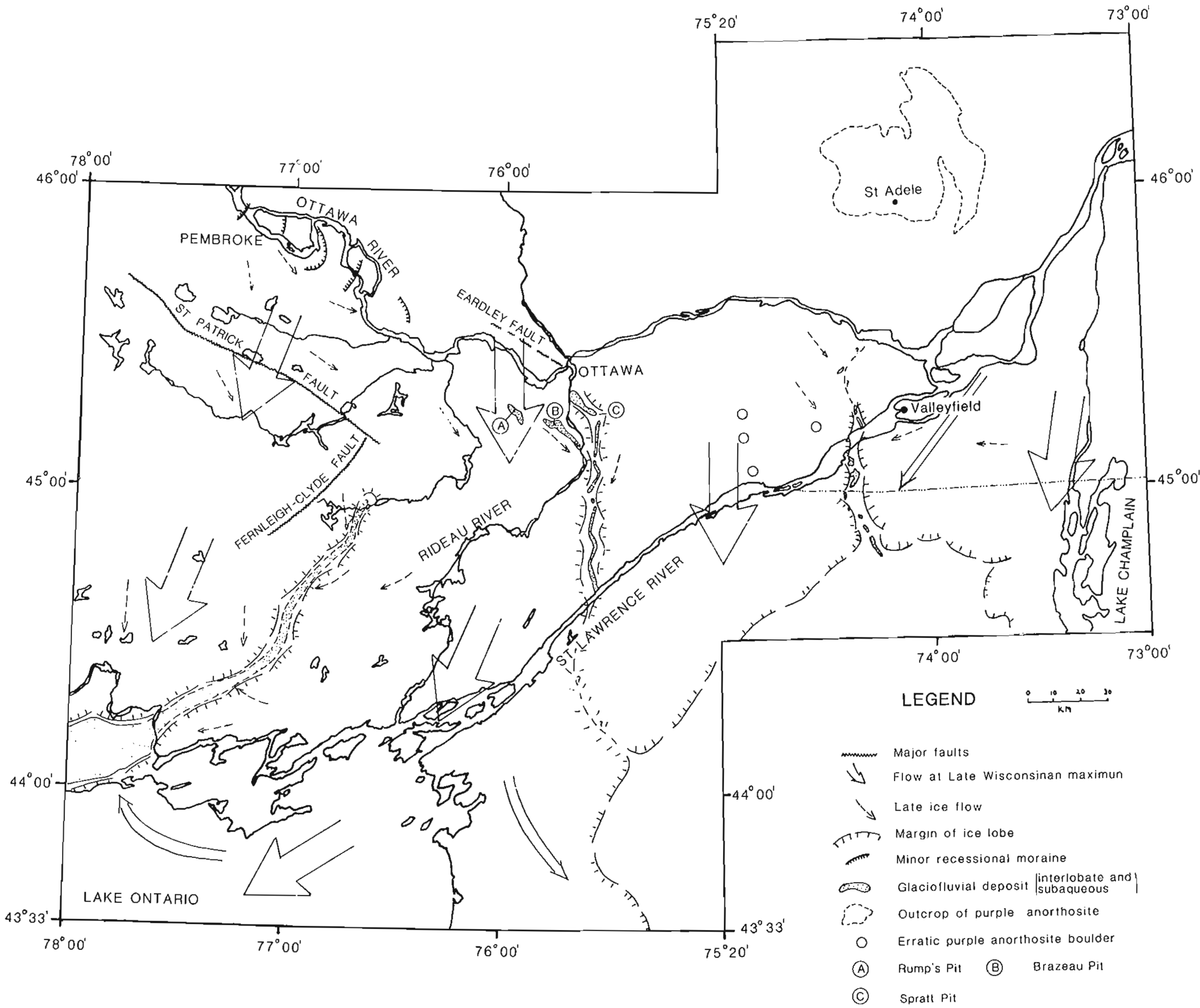

Figure 3. Main pattern of Late Wisconsinan ice lobation and flow

flowed south across Ottawa River valley and swung southwestwards and westward across the Frontenac Arch to flow into Lake Ontario basin. Most authors refer to the southern part of this lobe as the Ontario or Lake Ontario Lobe and Gadd (1980a) referred to it as the Ottawa-Lake Ontario Lobe in order to emphasize its Laurentide origin. Ice flow in Ottawa Valley upstream from Ottawa was initially southeastward, but as the ice sheet thickened it overtopped the fault scarp barrier on the south side of Ottawa Valley. At that time ice flow was southward across the valley and the Madawaska Highland. This movement produced the Madawaska Highland Lobe that extended south towards the Ontario basin.

Ice flowing across Ottawa Valley east of Ottawa moved mainly southward towards the Adirondacks but was deflected southwestward by that highland and by the Lake Champlain Lobe which already occupied the Lake Champlain basin. Some of this southwestward deflected flow apparently carried distinctive mauve to maroon coloured anorthosite from the Sainte-Agathe area north of Montreal to the area north of Cornwall (Fig. 1) and on through the Lake Ontario basin to Buffalo, New York (Gadd, 1981).

Gadd (1980a,b) used north trending interlobate ice contact and glaciofluvial deposits and the convergence towards these linear features of fine striations and other ice flow direction indicators as evidence that the retreating Late Wisconsinan ice margin in these areas had several lobations. The main ice lobe positions are delineated by interlobate deposits trending southerly from Rigaud Mountain to and beyond St. Lawrence River, extending from Ottawa to and beyond St. Lawrence River, and trend southwest from the vicinity of the Fernleigh-Clyde Fault (near Calabogie) to the eastern tip of Oak Ridges Moraine at Trenton, north of Lake Ontario (Fig. 3). The existence of these lobes suggests that the retreating ice margin was not the relatively straight, roughly east-west line extending between the eastern end of Oak Ridges Moraine and Covey Hill as proposed by Prest (1970), Karrow et al. (1961), and Dreimanis (1977).

At the time of deglaciation the St. Lawrence Lowlands were isostatically depressed below sea level. Proglacial lakes developed as ice masses shrank in the Lake Ontario basin to the west and in Lake Champlain basin to the south. As the ice was removed from lower St. Lawrence Valley to the east, marine waters invaded the area and the Champlain Sea developed (Elson, 1969a). There are two different ideas on submergence of the Ottawa area. Those who visualize a straight ice margin (see above) contend that St. Lawrence Valley between Montreal and Lake Ontario opened early so that glacial lakes in the Lake Ontario and Lake Champlain basins joined and extended into the Ottawa area. Retreat of ice near Quebec City then permitted the Champlain Sea to invade the area. This does not, however, explain why radiocarbon dates on marine shells from beach deposits near marine limit in upper St. Lawrence Valley are consistently several hundred radiocarbon years younger than those of the Ottawa and Gatineau valleys. Karrow (1981) and Hillaire-Marcel (1981) suggested that the discrepancy results from the northern shells containing old carbonate originating from glacial meltwater. As an alternative, Gadd (1980a) suggested opening of Ottawa Valley in early Champlain Sea time by a calving bay mechanism. He proposed that a calving bay would have extended along the Laurentian Channel and entered the deeper Ottawa Valley rather than penetrating the shallow upper St. Lawrence Valley. He referred to calculations by Thomas (1977) to support this hypothesis. A corollary of this proposal is that residual ice would have remained in upper St. Lawrence Valley south of the calving bay keeping marine waters from entering that area until some time later. Under this hypothesis, a proglacial lake could not have formed in the Ottawa area prior to arrival of marine waters, and Lake Iroquois in the Lake Ontario basin could have remained behind the residual ice at the time the Champlain Sea occupied the Ottawa area. The calving bay hypothesis provides a means of explaining the age discrepancy between marine shells in the Ottawa area and those in upper St. Lawrence Valley, but several other lines of evidence are in conflict with this proposal (for further discussion see Fulton and Richard, 1987).

Quaternary deposits

Glacial deposits in the area closely reflect the regional bedrock geology. Glacial and glacially derived sediments eroded from Precambrian rocks generally have sandy to gravelly textures and are either noncalcareous or variably calcareous depending on the local abundance of Precambrian marbles and their position in relation to flow from areas of Paleozoic carbonates. In the St. Lawrence Lowlands area, equivalent sediments are highly calcareous. The general distribution of tills of this region is shown in Figure 4 and they are discussed in more detail by Kettles and Shilts (1987).

Glaciofluvial sediments are common deposits in the Ottawa region. The most studied glaciofluvial deposits occupy ridges which rise above or are buried by marine deposits within the Champlain Sea basin. These are interpreted as having been deposited where meltwater conduits debouched from the ice into deep marine waters (Rust, 1977). These features were heavily reworked during regression and are capped by fossiliferous marine sediments (see Rust, 1987). In western parts of the basin, ridges and isolated bodies of glaciofluvial deposits occur in arcuate patterns. These isolated deltas and ice-contact fan complexes were deposited at the margin of an ice tongue retreating westward in Ottawa Valley.

Small eskers, and various types of outwash including kettle and kame terraces occur above marine limit in the Madawaska Highland. Most such deposits are small in extent but they occur in most of the valleys and lake basins of the area. Extensive glaciofluvial deposits lie near and north of marine limit but few details are available on their the nature and distribution; they consist of a variety of stratified sediments ranging from bouldery gravel to fine grained sand. In most places local units are well sorted, washed, and stratified but are deformed, are juxtaposed with contrasting units, are cut by closed depressions, and occur as hummocks and irregular benches. One major area of glaciofluvial sediment lies within the northern limits of the Champlain Sea north of Quyon and Shawville (Fig. 4). Along its southern margin it has been completely buried by Champlain Sea sediments but farther north it occurs as hummocks separated by hollows partly filled by marine sediments. Outwash terraces occupy many valleys north of this area and outwash fans are common at the edge of Ottawa Valley. Apparently this was an area where abundant glaciofluvial sediments were deposited on and at the margin of stagnant ice sitting in the Ottawa Valley.

Isolated bodies of glaciofluvial sediments occur in segments of north shore valleys lying below marine limit. Generally these are attached to one or both valley walls and apparently formed at locations where englacial or supraglacial debris, which was being flushed through these valleys, was trapped. Above marine limit, especially in the area adjacent to Gatineau Valley, glaciofluvial sediments line the bottoms of most valleys; these occur either as ridges and kame terraces, which apparently formed in valleys that were largely filled with ice at the time of sediment deposition, or terraces and kettled terraces, which formed in valleys that were largely ice free at the time of outwash deposition. It may be noted that the outwash terraces and fans in this area appear to have been constructed in short segments, each segment beginning at a lake and ending a few kilometres downvalley. One particularly large area of outwash lies in the northwest corner of the Kazabazua map area (Fig. 1, 4). It has been suggested that this might be related in some way to the glacial activity that to the east constructed the St. Narcisse Moraine (LaSalle and Elson, 1975; Fig. 5).

Preliminary maps showing the limit of the Champlain Sea were presented by Fransham et al. (1976), but the limit on these is based primarily on the elevation and distribution of fine grained sediments. Figure 5 is a summary of current knowledge of marine limit in the western basin of the Champlain Sea. It is taken largely from unpublished data by J-S. Vincent (Geological Survey of Canada) and Occhietti (in press) and is based on location of deltas and beach features in addition to the distribution of fine grained sediments. Of particular note is the area in Gatineau Valley which supposedly was contiguous with the Champlain Sea but was occupied by freshwater. Vincent (in press) explained that the former presence of a large water body in this area is based on the work of Dadswell (1974) and that the interpretation that this was freshwater is based on a lack of evidence that would prove the rhythmites, which are present in this area, are of marine origin.

Late Pleistocene and postglacial fine grained sediments as thick as 100 m occur in the Ottawa basin. The thickest deposits of these materials occur in deeper parts of the bedrock valley and adjacent to the north shore of the basin. Deposits on the south side of the divide between Ottawa and upper St. Lawrence Valley are generally thin and discontinuous. This suggests that the great bulk of sediments entered the basin during deglaciation of the north shore of Ottawa Valley and currents were such that most sediments were trapped in basins adjacent to the present position of Ottawa River. The limited thickness of fine grained marine sediments in upper St. Lawrence Valley might also be used as evidence that this area was occupied by stagnant ice during degaciation of the area to the north so it did not receive as much sediment as the Ottawa basin.

The fine grained marine sediments occur in several lithofacies which are best observed in deep borings. Cores up to 100 m in length are described briefly by Gadd (1977) and in more detail by Gadd (1986b). In stratigraphic order the suite consists of: varve-like rhythmites, massive clays and silty clays, laminated silty clays (red and grey banded) with some

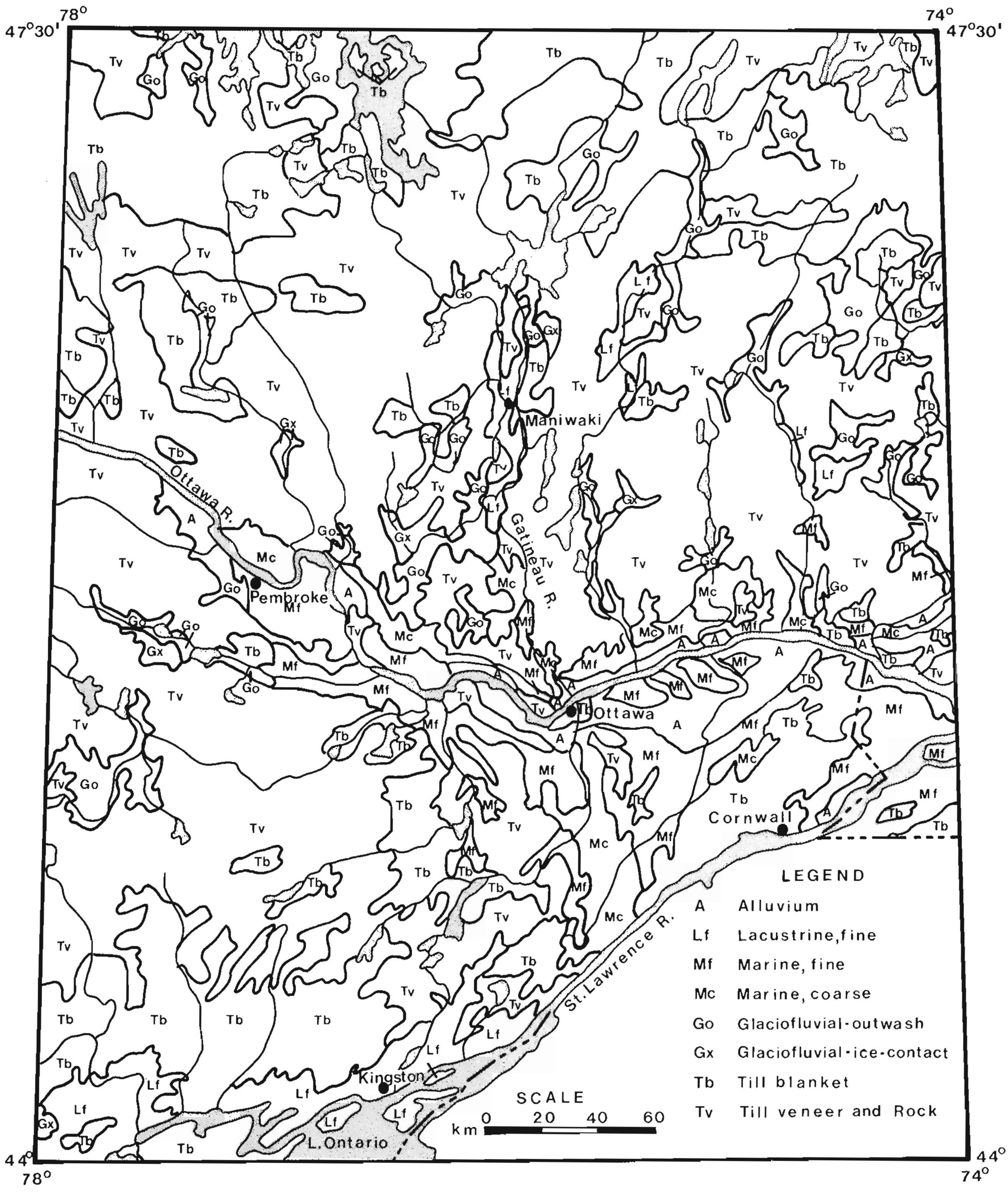

Figure 4. Surficial materials of the Ottawa area; geology compiled at a scale of 1:1 M by N.R. Gadd and J.J. Veillette, Geological Survey of Canada

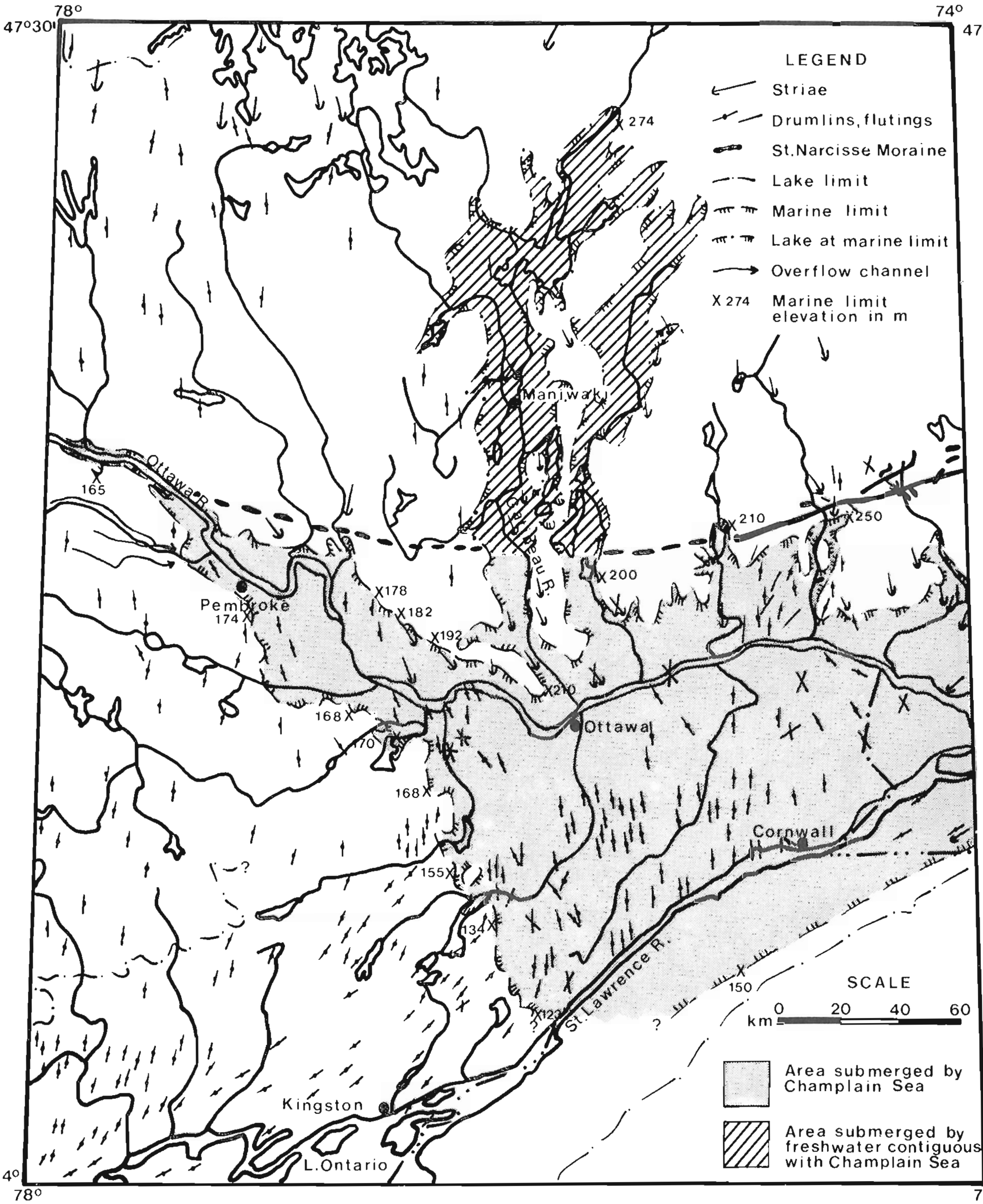

Figure 5. Marine limit in the western basin of the Champlain Sea, glacial flow features, and miscellaneous late glacial features of the Ottawa Region. Limit of Champlain Sea is from an unpublished compilation by J-S. Vincent, Geological Survey of Canada.

sand, and silty sand and sand with abundant channel structures. All contacts within the suite are gradational, suggesting sedimentation without significant erosional breaks. The lower laminated deposits are quiet water sediments derived from a nearby glacial source and deposited in freshwater. The massive clays and silty clays were similarly derived from a glacial source but were deposited in a marine environment. The upper sediments are largely the product of a delta migrating through the area as isostatic uplift drained the Champlain Sea.

The similarity of the older rhythmites to glacial lake varves was noted by earlier workers (Johnston, 1917; Antevs, 1925) and the sediments have been attributed to a glacial lake named glacial Lake Frontenac (Leverett and Taylor, 1915) or glacial Lake St. Lawrence (Goldthwait, 1971). According to Prest (1970) and Clark and Karrow (1984), this lake was part of a single lake that extended via upper St. Lawrence Valley from the Lake Ontario basin to the Lake Champlain Valley. If Gadd (1980a) is correct and a calving bay extended up Ottawa Valley, there would have been no opportunity for a glacial lake to precede the Champlain Sea in the area and the varve-like sediments would have to be related to an area of freshwater in a zone of maximum meltwater influx at the head of the calving bay.

The massive grey clays and silty clays were deposited in parts of the basin subject to high sediment influx at the time of deglaciation and in deeper parts of the basin following deglaciation. The younger laminated silty clays and silty sands can be related to the fluvial system that operated following deglaciation. Gadd (1986b) concluded that a major delta system developed in Ottawa Valley near Chalk River and another in Gatineau Valley near Kazabazua. These deltas prograded seaward as offlap of Champlain Sea took place and thus the several delta facies were stacked or superposed in a more or less continuous series. These delta facies included laminated red and grey silty clays with sand lenses and some turbidite beds, and silty sand and sand displaying channel structures. As the delta tops emerged, the modern drainage system was incised and two new facies of sediment were produced, a sandy facies in channels and a silty clay overbank facies in slack water areas.

The legacy of these proto-fluvial systems is a series of channels and cut terraces that largely parallel the present river channels (included in unit A of Fig. 4). In some parts of the area the terraces and channels are underlain by marine clay, locally with a thin cover of fluvial sand. In places the incising river uncovered buried glaciofluvial deposits or exposed till or rock ridges; bars of sand and gravel were deposited downstream from these sites. Rock areas that were exposed were swept free of sediment and bouldery lag was left on top of exhumed glaciofluvial deposits and till. Many large landslide scars indicate that during channel cutting the saturated clays that made up the channel walls were unstable. In some places landslide deposits lying on the channel floor indicate that failure occurred after the channels were abandoned. In many places, however, landslide debris has been removed from the toe of the scars suggesting that many of the slides occurred at a time when flow in the channels was sufficient to remove the slide debris. Slope failures are still occurring in the area but these are largely induced by human activities or occur in areas where rivers or streams are actively undercutting clay banks.

Eolian processes were active in the area immediately following deglaciation. This is shown by the presence of stabilized dunes in most areas underlain by sand. In addition a cap of loessal sediment several centimetres thick is present in most parts of the region and a blanket of silty cover sand, up to 50 cm thick, occurs in areas of abundant glaciofluvial, marine, and fluvial sand. Peat bogs have developed in poorly drained areas such as abandoned river channels; Mer Bleue on the eastern outskirts of Ottawa is a good example of a large peat bog that has developed in such a feature.

The chronology of late Quaternary events in the Ottawa region is discussed by Fulton and Richard (1987) but is mentioned briefly here for completeness. There is no information on the timing of the beginning or maximum phases of the last glaciation. Ice is known to have begun to retreat from its maximum position in New Jersey by 18.5 ka (Cotter et al., 1985). The Ottawa basin apparently was undergoing deglaciation by 12 ka when the Champlain Sea entered the area.[1] By 10 ka the water remaining in the basin was sufficiently fresh to support **Lampsilis** (Rodrigues, this publication) and by 8 ka many of the early proto-Ottawa River channels were abandoned and had become the sites of peat bogs.

[1] A date of 12 700 ± 100 BP (GSC-2151; Richard, 1978) at Clayton west of Ottawa is controversial. Some consider it to be too far out of line with other dates in the area to be reliable (see Fulton and Richard, 1987); others accept it at face value.

TILLS OF THE OTTAWA REGION

I.M. Kettles[1] and W.W. Shilts[1]

INTRODUCTION

The Geological Survey of Canada carried out a program of systematic till sampling in the Ottawa area from 1980-1984 (Shilts, 1982; Kettles and Shilts, 1983). The project was designed to quantify regional variations in drift composition in order to provide baseline data that might be used in assessing the effects of acid rain. For this purpose two groups of compositional characteristics of till were mapped: (1) texture and carbonate composition (the buffering components) and (2) concentrations of naturally occurring trace and minor elements (potential sources of environmental contamination if released by acid leaching or by exchange reactions with groundwater). Till was observed at and collected from more than 1500 man-made and natural exposures along roads and streams and also from sand and gravel pits. Information obtained from this study pertaining to sedimentary characteristics and mineralogical composition of till are discussed in this report.

Bedrock geology

The till of the Ottawa and surrounding regions has inherited most of its physical and lithological properties from erosion of two contrasting bedrock lithologies. Precambrian metasediments and igneous rocks (Baer et al., 1977), referred to collectively as "crystalline" bedrock, underlie much of the area and are characterized by massive outcrops and sharp but low relief. Relatively flat-lying Paleozoic sedimentary rocks, dominated by carbonate lithologies, form flaggy outcrops in Ottawa Valley and east and west of Thousand Islands in the Central and West St. Lawrence Lowlands (Bostock, 1970; Fig. 2). Most Precambrian rocks are part of the Central Metasedimentary Belt of the Grenville Structural Province. This is characterized by extensive areas of carbonate metasediments (outlined in Fig. 6), felsic (less commonly mafic) plutons, and belts of metavolcanic and noncalcareous metasedimentary rocks. The Central Metasedimentary Belt includes the Frontenac Arch, a prominent geological structure that connects the main body of the Canadian Shield in the north to the Adirondack Mountains south of St. Lawrence River (Fig. 2). The remainder of the crystalline terrain comprises gneisses of the Ontario Gneiss Belt.

Glacial geology

The predominant ice flow direction across the Ottawa region during the Late Wisconsinan was southerly down Gatineau River valley in Quebec and south-southwest over the southwestern part of the area (Fig. 3). During the last stages of glaciation, one lobe of ice flowed southwestward, retreating from Lake Ontario towards St. Lawrence Valley and another flowed south-southeastward and retreated up the Ottawa Valley (Richard, 1975a; Gadd, 1980a).

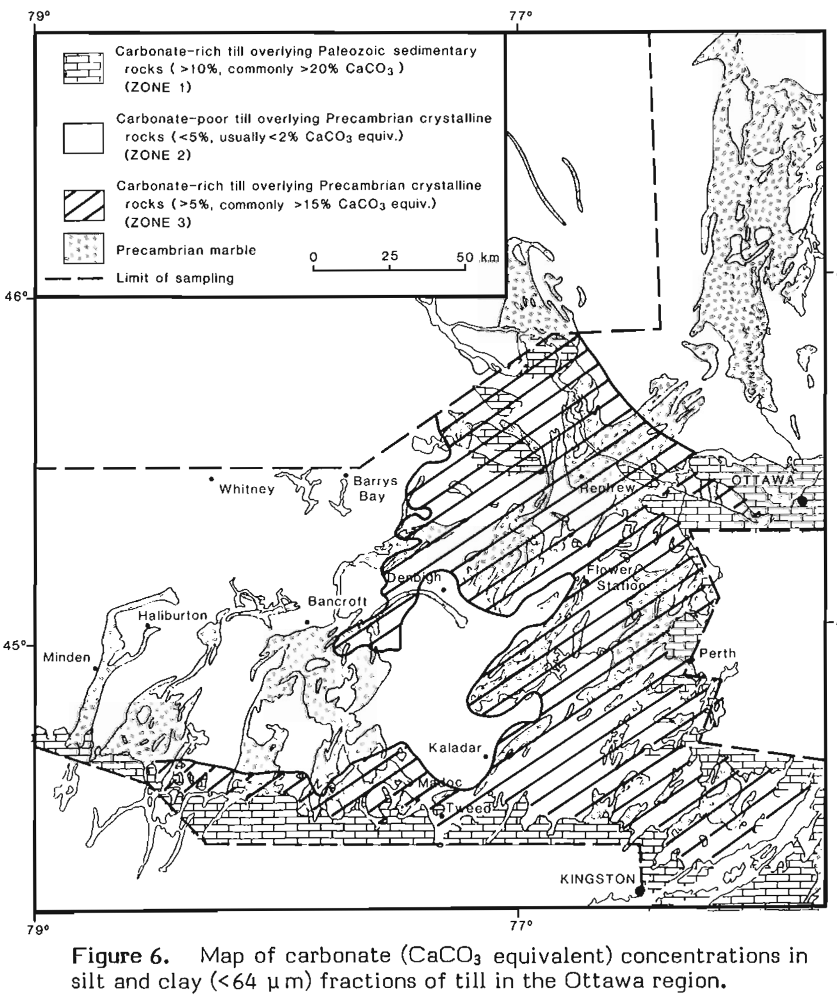

Figure 6. Map of carbonate ($CaCO_3$ equivalent) concentrations in silt and clay (<64 μm) fractions of till in the Ottawa region.

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Stratigraphically, only one unit of till, which is associated with the last (Late Wisconsinan) glacial expansion of the Laurentide Ice Sheet, has been identified in the areas sampled (Gadd, 1963; Richard, 1975a). This till was referred to as Fort Covington till in the Merrickville map area at the eastern edge of the area covered by this work (Sharpe, 1979). It forms a persistent, but thin cover (<5 m) over both Precambrian and Paleozoic bedrock of the region but can be thicker in depressions, along valley walls, and on the up-ice side of bedrock ridges. Till is the most common surface deposit, except below 200 m elevation along Ottawa and St. Lawrence valleys where marine and lacustrine sediments predominate.

TILL SEDIMENTOLOGY

Sedimentary characteristics of till

The till of the Ottawa region is stony and sandy, similar to Canadian Shield tills in many other areas. The average texture of the <2 mm fraction of more than 800 samples is 67% sand (2 mm-64 µm), 26% silt (64-4 µm), and 7% clay (<4 µm) with ranges of 34-97%, 2-55%, and 1-29%, respectively (Fig. 7). In shallow exposures till is typically not very compact and fractures into small (<1 cm) angular

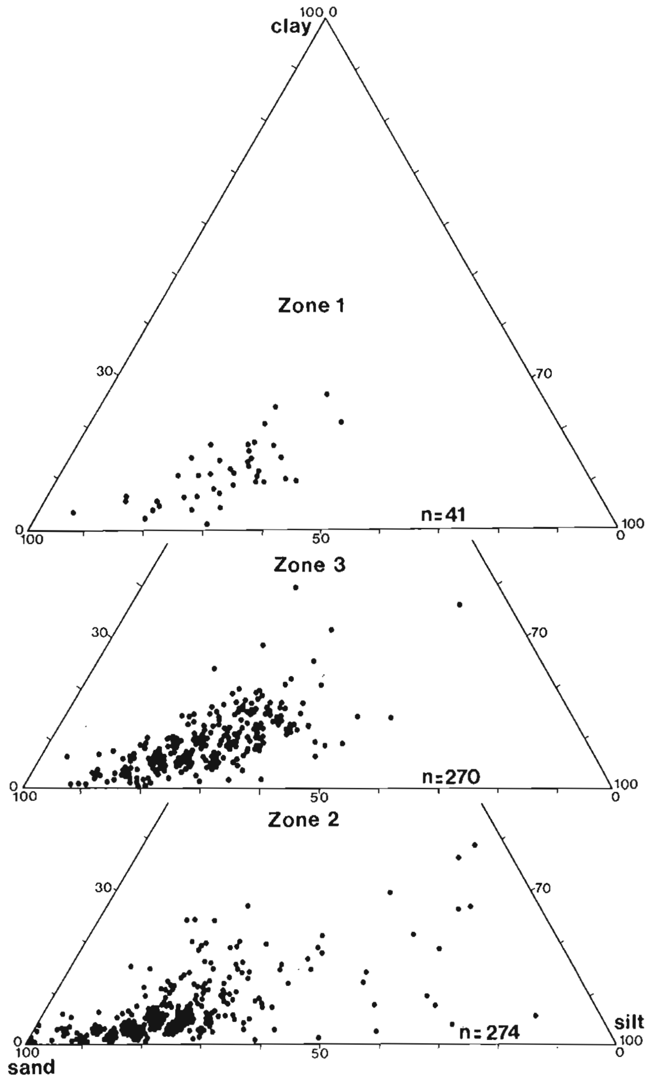

Figure 7. Ternary diagrams showing the texture of the <2 mm fraction of tills in the Ottawa region. Zones refer to areas outlined in Fig. 6.

"chunks" composed of sand and granules with a cohesive silt-clay matrix. In deeper exposures, till of similar texture can be fairly compact, but the origin of its compactness cannot be ascribed confidently to its mode of deposition. Increasing compactness with depth may simply reflect compaction by the overlying sediment combined with desiccation on exposure to the air.

In good near-surface exposures, till usually can be seen to contain stringers or clasts of water-sorted sediment, suggesting that it was emplaced by slumping or melting from a supraglacial position (Boulton and Deynoux, 1981). In deeper sections, the compactness, faint evidence of subhorizontal discontinuities, interbedded waterlaid sediments, and sediment-filled reticulate fractures may indicate that till melted out from a subglacial or englacial position, possibly from stagnant ice.

In many pits the surface of ice contact gravels and sands is mantled by a till-like diamicton which contains structures and has a distribution that indicates that the material flowed from ice onto or into the deposit. Because this "flow till" is present in or on so many glaciofluvial deposits, it probably also is present at the surface of much of the rest of the area. This diamicton, however, can only be recognized as a distinct deposit where it overlies sorted sediments.

Till composition

Carbonate minerals. The distribution pattern of carbonate minerals in till (Fig. 6) is determined from carbonate content of the silt plus clay-sized (<64 µm) fraction. Carbonate content was determined using a Leco carbon analyzer to measure carbon concentrations which were converted to %$CaCO_3$ equivalent. The distribution patterns of carbonate in the fine sand fraction (64-250 µm) and weight per cent Paleozoic limestone and dolomite erratics in the total granule (2-6 mm) fraction produce patterns that have similar configurations.

The geographic distribution of till characteristics may be divided into three zones (Fig. 6) based on carbonate content of till matrix and on the structure and lithology of underlying bedrock: 1) Low lying areas east and south of the Frontenac Arch which are underlain by flat-lying Paleozoic rocks characterized by a basal quartz sandstone overlain by carbonates with minor shales. The carbonate content of the silt-clay fraction of till in this region is high (generally >10%). 2) The Gatineau River valley region of Quebec and an area which extends southwestward across the western part of the Frontenac Arch from Barry's Bay to Minden (Gatineau-Minden area). This zone is underlain by crystalline bedrock of the Central Metasedimentary Belt and the Ontario Gneiss Belt. Till in this region has low carbonate content (<5% and commmonly <2%) even where it overlies areas underlain by marble. 3) The eastern and southern parts of the Frontenac Arch, which are also underlain by lithologies similar to those underlying Zone 2 but in this area the carbonate content of till is high (>5% and commonly >15%).

Texture. Textural analyses of samples collected in each of the three zones have been plotted on ternary diagrams (Fig. 7). The low carbonate till of the Gatineau and Minden areas (Zone 2) generally has a smaller component of silt and clay-sized material than carbonate-rich till from the other two areas. There appears to be little difference in sand-silt-clay ratios between carbonate-rich till overlying the Paleozoic sedimentary rocks (Zone 1) and carbonate-rich tills overlying Precambrian crystalline rocks (Zone 3).

Trace elements. The clay-sized (<4 μm) fraction of till has been analyzed for 14 trace and minor elements (Kettles and Shilts, 1983). With the possible exception of U (not shown here), the geographic distribution of trace element characteristics in till appears related to underlying bedrock and does not correspond to the distribution patterns of texture and carbonate concentrations. The lowest concentrations of U appear to be associated with carbonate-rich till of the first and second zones.

On a regional scale, high levels of various elements are preferentially associated with the Frontenac Arch. High levels of Zn, As, Fe, Pb, Mn, Cd, Mo, Hg, and sometimes Co are related to the northeast-southwest striking belts of metasedimentary and metavolcanic rocks, whereas others such as Co, Cr, and Ni are commonly associated with some large basic plutons which are also part of the Arch.

The dispersal of arsenic in till will be discussed here because its provenance can be related easily to geological features within the region. Major areas of arsenic-enriched till occur within a triangle, the apices of which are the towns of Flower Station, Coe Hill, and Madoc (Fig. 8). This triangle encompasses the original Ontario gold mining belt that was the site of active mining of arsenical gold in the late 1800s and early 1900s. The areas of As enrichment outline metasedimentary and metavolcanic belts which within shear zones and fractures contain abundant deposits of arsenic sulphides, some of which are host to gold-bearing veins (Carter et al., 1980).

Weathering and postdepositional modification. The postglacial soil profile is commonly about 1 m thick. Below this, till usually shows some evidence of minor weathering such as a tan (presumably oxidized) colour, staining from iron and manganese oxide precipitates, and the common presence of disaggregated, coarsely crystalline clasts of marble and/or granitoid rocks. Although signs of oxidation can be found to considerable depths, carbonate leaching is rarely deeper than the base of the solum. Also, secondary carbonate is commonly observed around root casts and along joints in till less than 1 m from the surface.

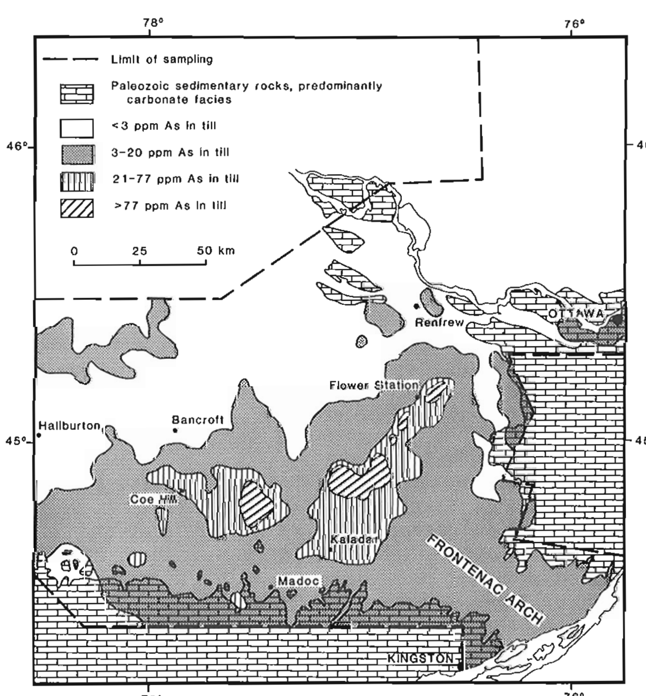

Figure 8. Arsenic concentrations in 1500 samples of till from the Ottawa area.

In the most sandy, permeable tills there has been a considerable amount of downward translocation of silt- and clay-sized particles by percolating ground water. Undisturbed exposures commonly reveal <1 mm-thick coatings of silt and clay on the tops of clasts, apparently washed down from the higher parts of the section. Below the local water table, chemical weathering does not seem to accompany these phenomena.

Glacial erosion insights from carbonate and trace element data

There is a high frequency of Paleozoic erratics in carbonate-rich till overlying crystalline rocks of the eastern and southern part of the Frontenac Arch, and a low concentration of carbonate minerals in till overlying the marble belts of the Gatineau-Minden area. This suggests that carbonate components were more readily glacially eroded from the flaggy, flat-lying unmetamorphosed rocks of the Paleozoic basins than from the massive, little fractured Precambrian marble. The carbonate derived from Paleozoic rocks south of Ottawa has been transported more than 70 km southwestward over the relatively low eastern part of the Frontenac Arch. In the western part of the area carbonate debris has been moved up the larger river valleys (the Madawaska and Mississippi and their tributaries) into areas underlain by Precambrian crystalline rocks. Also, along the southern edge of the Frontenac Arch the carbonate content of till abruptly increases from 5% to 10% and is commonly >30% directly down ice from numerous Paleozoic inliers that occur in this area. In contrast, in all areas where carbonate-poor material was carried from the Shield onto Paleozoic rocks, it was quickly diluted by large quantities of Paleozoic debris.

The strong influence of the Paleozoic lithologies on till composition over a wide area can be attributed to physical characteristics of these strata. These sedimentary rocks, composed largely of fine grained "soft" minerals, calcite and dolomite, are jointed and thinly bedded and consequently highly susceptible to plucking, the most effective form of glacial erosion. Once dense concentrations of these blocks were entrained in the basal parts of the glacier, they were broken up and abraded by collisions with each other and the underlying bedrock. The breakdown process would be further enhanced as the blocks moved over the coarse grained, structurally massive crystalline rock of the Frontenac Arch, composed largely of more erosion resistant minerals such as quartz and felspar. Although the marbles of the Frontenac Arch are also composed largely of calcite and dolomite, their components were not so readily entrained because they occur in massive outcrops, which were not as susceptible to erosion by plucking as were the Paleozoic outcrops. Erosion of marble outcrops has produced characteristic moulded forms, but relatively little debris coarser than sand size. Thus, the secondary comminution that occurred through clast-to-clast contact during transport of high concentrations of Paleozoic erratics near the base of the ice was not an important process down ice from marble outcrops. The presence of fine carbonate detritus most likely accounts for the larger component of silt- and clay-sized material (Fig. 7) in carbonate-rich till (zones 1 and 2) because the terminal grain size of calcite and dolomite is silt- and fine sand-size (Dreimanis and Vagners, 1971).

A broad arsenic anomaly is clearly outlined over the Frontenac Arch (Fig. 8). One known source of arsenic is small sulphide deposits, some of which have hosted gold, that are located in fractures and shear zones of metasedimentary and metavolcanic rocks of this area. Other possible sources are the metasedimentary and metavolcanic rocks themselves which may have high background concentrations of As in the Frontenac Arch area. In either case, the area covered by

As-rich till is large compared to the area underlain by either type of source rock and illustrates how high concentrations of As might have been homogenized and dispersed over a wide area through glacial transport.

CONCLUSIONS AND SUMMARY

The ubiquitous presence of clasts, stringers, and lenses of waterlaid material in till, subhorizontal discontinuities in the deeper parts of sections, and till-like diamictons in ice contact sand and gravel sediments, all suggest that much of the till of this region was emplaced during deglaciation when debris melted out of or slumped off retreating or stagnating ice at the glacier margin. A similar conclusion was reached by Kaszycki (1983) for the Haliburton area at the western edge of this region.

The effects of glacial erosion and transport on the geographic distribution of till characteristics are readily apparent. Over the eastern and southern parts of the Frontenac Arch the composition of till is strikingly different from that of underlying bedrock. Large concentrations of carbonate minerals derived from the extensive Paleozoic terranes of the Ottawa and St. Lawrence valleys have been dispersed over a variety of noncalcareous rocks which comprise part of the Frontenac Arch.

On a smaller scale, arsenic-rich debris, derived from small source areas of arsenic sulphides or possibly from high background concentrations of arsenic in metasedimentary and metavolcanic rocks of the Frontenac Arch, is dispersed over much larger areas by glacial processes of homogenization through transport of debris.

SUBAQUEOUS OUTWASH DEPOSITS OF THE OTTAWA REGION

B.R. Rust[1]

The sediments discussed here are late Quaternary sands and gravels exposed in pits along a series of ridges to the south of Ottawa (Fig. 3). They overlie Late Wisconsinan till and are largely unfossiliferous. Locally, however, they contain marine fossils and are commonly overlain unconformably by a thin fossiliferous sheet of littoral sand and gravel (Rust and Romanelli, 1975, Fig. 3) formed as isostatic rebound caused the ridges to emerge from the Champlain Sea. Because this area was submerged by the Champlain Sea (possibly also by a short lived lake which may have preceded it; Naldrett and Rust, 1984; Anderson et al., 1985) and progressively drained by isostatic uplift, the ridge sediments accumulated well below the level of the Champlain Sea. The sand and gravel can therefore be regarded as ice proximal equivalents of the fine grained Champlain Sea deposits and older lacustrine clays.

The ridge deposits are commonly deformed and include coarse boulder gravels, which fine laterally to sand facies within short distances. These features indicate deposition in an ice-contact environment, in which unidirectional paleocurrents point to a dominance by meltwater flow from the glacier. The sediments show stratification typical of glaciofluvial deposits and were mapped as such by Gadd (1962) and Richard (1974, 1975a). Because these deposits were formed well below the surface of a standing body of water, however, Rust and Romanelli (1975) proposed the term subaqueous outwash. The ridges are interpreted as a series of longitudinally overlapping subaqueous fans formed at the mouths of subglacial conduits as the ice front retreated (Rust, 1977; Fig. 9). In effect the ridges are submerged eskers that were modified by wave action during isostatic rebound. The locations of the subglacial conduits were probably controlled by longitudinal crevasse systems in the ice, or in some cases the conduits may have issued from the suture zone between ice lobes. Changes in the morphology of the ice margin as it retreated gave rise to facies variation within the ridges (Cheel, 1982; Cheel and Rust, 1982).

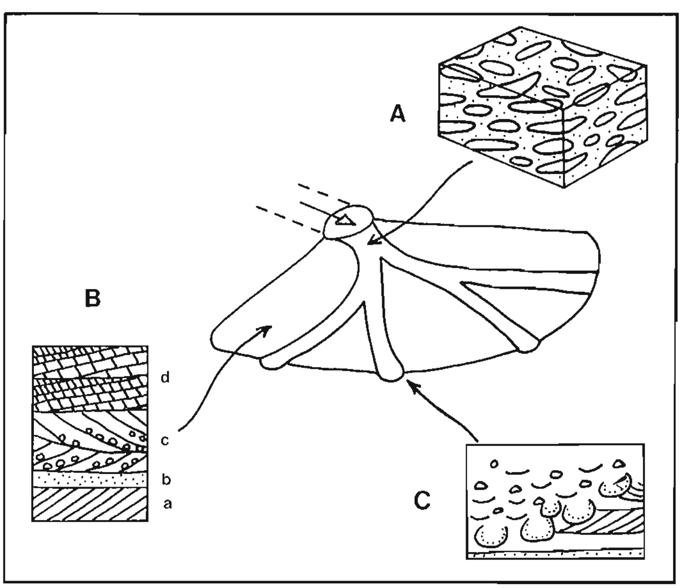

Figure 9. Depositional model for subaqueous fans south of Ottawa. Dashed line represents margins of meltwater conduit ice; for simplicity, ice contact features are omitted. A: gravel facies. B: stratified sand facies; a: planar cross-stratified sand; b: structureless sand; c: trough cross-stratified, pebbly sand; d: graded ripple-drift units. C: massive sand facies in a channel locally with ball and pillow structures due to upward escape of water (from Rust, 1977).

The facies exposed at any time are dependent on the activities of the pit operators, who generally prefer to extract well washed sand. Gravel facies include imbricate boulder gravel (Rust and Romanelli, 1975, Fig. 4), horizontally stratified gravel, and cross-stratified gravel, formed within or adjacent to the mouth of the glacial conduit. Boulder-bearing diamictons with sandy mud matrix are interpreted as till flows, formed as supraglacial melt-out till slid from the glacier surface and was redeposited on the seafloor. Sand facies include large-scale cross-strata of coarse pebbly sand, medium grained massive sand and ripple-drift units that fine upwards from fine sand to silt (Rust and Romanelli, 1975, Fig. 8). Massive sand commonly occurs in steep-sided channels (Rust and Romanelli, 1975, Fig. 6; Rust, 1977, Fig. 3) and in places contains scattered boulders. The massive sand is attributed to mass-flow deposition within channels on the subaqueous fan surface, and the cross-stratified facies to deposition on interchannel areas of such fans (Rust, 1977; Fig. 9). Massive sand also occurs as thin, normally or inversely graded sheets, attributed to overbank spillage of mass flows from channels, and also as fills of large slumps on the fan surface that resulted from melting of buried ice masses (Cheel and Rust, 1982, Fig. 7).

Deformation is common in the subaqueous outwash deposits near Ottawa. Faults resulted from compaction as buried ice melted (Rust and Romanelli, 1975, Fig. 12). The water released by ice melt commonly was confined by overlying impermeable strata, so that pore pressure increased and a series of soft-sediment deformation structures resulted. The typical upward sequence is from convolute stratification to ball and pillow structure (the scale of the pillows decreasing upwards) to dish structures. This sequence is attributed to foundering of initial layers as fluidized sediment moved upwards in progressively smaller streams, eventually forming dishes by general upward percolation (Cheel and Rust, 1980).

The phenomena described above can be viewed at three pits in the Ottawa area. Rump's Pit (site S-6 of Cheel, 1982) is located on the southwest side of the Stittsville-Huntley ridge, a short distance northwest of the intersection between Highway 417 and Regional Road 5 (A of Fig. 3). Exposures show relatively undeformed stratified sand and silt (mostly climbing ripple facies) sharply transitional into deformed strata of similar lithology. Deformation is attributed to rising water, in some cases derived from buried ice blocks. The deformed sequence contains ball and pillow structures transitional upward into dish structures in massive sand. The sand contains scattered larger clasts ranging from pebbles to boulders.

Brazeau Pit is located in the Twin Elm Ridge about 5 km west of Manotick (B of Fig. 3). In it there are extensive exposures of subaqueous outwash which locally contain marine fossils, in places highly concentrated. Horizontally stratified imbricate boulder to cobble gravel is interpreted as proximal subaqueous outwash, formed within or close to the mouth of a subglacial tunnel. This is transitional distally to cross-stratified sand and pebbly sand, and thence to ripple cross-laminated sand and silty sand. The rippled units commonly occur as fining upward sequences, attributed to waning flows, such as a storm-induced discharge through the glacier. Internally massive sheets of diamicton are interpreted as till flow deposits, generated by debris flow

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initiated by sliding of supraglacial till either subaerially or subaqueously, followed by deposition on the subaqueous fan surface (Rust, 1982). Most of the deformation structures in the Brazeau Pit are faults, which are attributed to foundering of sediments above melting blocks of ice. Fossiliferous littoral deposits of the Champlain Sea were formerly present as a thin sheet over the Brazeau Pit locality, but have been largely removed by pit operations.

Spratt Pit at South Glouchester exposes an extensive section of subaqueous outwash in the ridge extending southward from Ottawa Airport (C of Fig. 3). The section forms a continuous, mildly faulted, fining upward succession that was deformed by melt-out of buried ice. Facies include climbing ripple units of fine sand and silt and massive coarse sand in graded sheets (both inverse and normally graded) and channel fills. The uppermost unit of the succession contains various lithologies, including silt, and abundant **Hiatella** with some **Balanus** plates. **Hiatella arctica** from gravel which possibly was deposited in a kettle has been dated at 11 100 ± 130 BP (GSC-4166). It is overlain unconformably by a sheet of fossiliferous littoral gravel, which contains stunted **Hiatella.** **Hiatella arctica** from the bottom of the littoral gravel has a radiocarbon age of 10 500 ± 120 BP (GSC-4173).

Shells associated with subaqueous outwash at Brazeau and Rump's pits have also been dated. As discussed by Fulton and Richard (1987) they are younger than would be anticipated for ice marginal sediments at these locations. In all cases, however, they are found in down-faulted or loaded strata and probably reached their present location by downward movement due to melting of underlying ice masses. The fossils therefore come from a higher stratigraphic location and do not date the time at which the location was under the influence of an active ice margin. Rather, they indicate the time lapse required for the melting of an ice mass covered by a thick insulating succession of sediments.

LATE PLEISTOCENE INVERTEBRATE MACROFOSSILS, MICROFOSSILS AND DEPOSITIONAL ENVIRONMENTS OF THE WESTERN BASIN OF THE CHAMPLAIN SEA

Cyril G. Rodrigues[1]

Late Pleistocene glaciation isostatically depressed lower Ottawa and St. Lawrence valleys below sea level; consequently these areas were submerged during deglaciation. A sequence of depositional environments from freshwater to marine (Champlain Sea) to freshwater is recognized in the deposits related to this episode. The main water body, the Champlain Sea, was salinity stratified throughout most of its brief history. Cold subarctic high salinity water occupied the deeper parts of the basin and was overlain by lower salinity water. Warmer boreal and cold subarctic low salinity waters were present at shallow depths in the western basin of th^ Champlain Sea (lower Ottawa and upper St. Lawrence valleys) during the latter part of the marine episode. These conclusions are based on the temporal and spatial distribution of eight marine and one freshwater macrofaunal associations and fifteen foraminiferal and nine ostracode assemblages from Champlain Sea and related deposits.

MACROFOSSILS

Marine assemblages consist of bryozoans, cirripeds, gastropods, pelecypods, and sponges, whereas the much less common freshwater assemblages are characterized by pelecypods and gastropods. Pelecypods are the most abundant invertebrate macrofossils. The macrofaunal assemblages, which are commonly characterized by one dominant species, are arranged into groups termed associations and are named after the dominant species of the assemblage (Rodrigues and Richard, 1983).

Eight marine and one freshwater macrofaunal associations are recognized: **Balanus hameri, Hiatella arctica, Macoma balthica, Macoma calcarea, Mya arenaria, Mya truncata, Mytilus edulis, Portlandia arctica,** and **Lampsilis.** The temporal distribution of seven of the associations is shown in Figure 10. The ranges are based on ^{14}C age determinations on the dominant species of the associations, except for the range of the **Mytilus edulis** association which is inferred from the presence of the association at sites at which other species have been dated. Only one date has been published for the **Macoma calcarea** association, 10 600 ± 100 BP (GSC-3614; Blake, 1983), and one for the **Mya truncata** association, 10 300 ± 100 BP (GSC-2261; Lowdon and Blake, 1979).

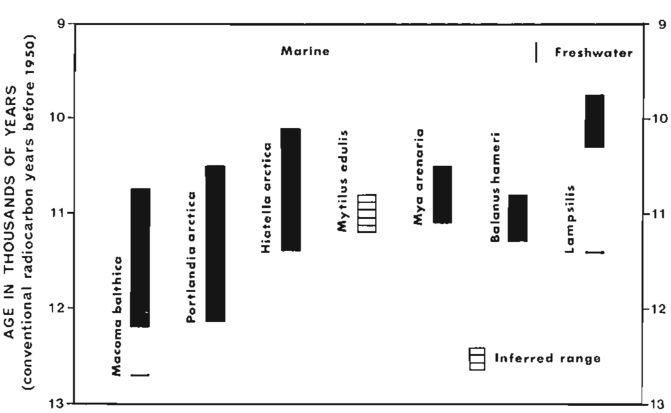

Figure 10. Temporal distribution of some macrofaunal associations in the western basin of the Champlain Sea; ranges are based on available radiocarbon dates and may not represent the total ranges for the associations.

The **Macoma balthica** and **Portlandia arctica** associations are widely distributed in the western basin of the Champlain Sea. The **Hiatella arctica** association penetrated as far west as the Arnprior map area in Ottawa Valley and the Morrisburg map area in St. Lawrence Valley (Fig. 1). The **Balanus hameri** association appears to be restricted to the Lachute map area on the north side of Ottawa River and was not observed west of the Kemptville map area on the south side of Ottawa River. The western limit of **Mya arenaria** association is in the Lachute map area on the north side of Ottawa River and in the Cornwall and Alexandria map areas on the south side of Ottawa River. The **Macoma calcarea** association is present in the Vaudreuil map area and the **Mya truncata** association is found in the Huntingdon and Vaudreuil map areas. The **Lampsilis** association occurs in the Russell, Hawkesbury, Cornwall and Huntingdon map areas, and at one site in the Cobden map area near the western margin of the Champlain Sea basin.

MICROFOSSILS

Foraminifers, ostracodes, and the cartwheel holothurian sclerite **Myriotrochus vitreus** (Sars) are present in the late glacial and postglacial deposits; foraminifers and ostracodes generally accompany macrofossils in Champlain Sea deposits. **Myriotrochus vitreus** was observed in samples of Champlain Sea sediments from a borehole in Lac Deschênes. The foraminiferal fauna consists of calcareous benthonic species. Ostracodes are not as abundant as foraminifers in the marine deposits. Approximately 210 000 foraminiferal tests and 10 000 ostracode valves were counted in 148 samples of Champlain Sea sediments from gravel and sand pits, drainage ditches, roadcuts, riversides, and boreholes. The dominant foraminiferal and ostracode species are listed in Table 1; they are illustrated in Rodrigues and Richard (1986). Fifteen groups or foraminiferal assemblages and nine groups or ostracode assemblages are recognized (Table 2). The freshwater ostracode **Candona** occurs in low numbers in Champlain Sea deposits and is commonly the only invertebrate microfossil in the rhythmically laminated sediments underlying marine deposits in lower Ottawa and upper St. Lawrence valleys.

FAUNAL SUCCESSIONS

In some exposures a succession of macrofaunal associations occurs. One of **Balanus hameri** or **Portlandia arctica** or **Hiatella arctica** association is found at the base of the successions and the **Macoma balthica** or **Mya arenaria** association commonly occurs at the top of the successions (Fig. 11). Marine pelecypod associations are replaced by the freshwater **Lampsilis** association in the Huntingdon map area. Successions of foraminiferal assemblages and at some sites successions of ostracode assemblages accompany the successions of macrofaunal associations; an example from a section in the Russell map area is given in Rodrigues and Richard (1986). The most common succession of microfaunal assemblages is that in which **Cassidulina reniforme**-dominant assemblages are replaced by assemblages characterized by **Elphidium** spp. and **Haynesina orbicularis.** The sequences of macrofaunal associations and microfaunal assemblages represent short-term successions that are noncyclical at the study sites in the western basin of the Champlain Sea.

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Table 1. Dominant species of foraminiferal and ostracode assemblages from late Pleistocene marine sediments of the western basin of the Champlain Sea

Foraminiferal Species	Ostracode Species
Astrononion gallowayi Loeblich and Tappan	**Cytheromorpha macchesneyi** (Brady and Crosskey)
Cassidulina reniforme Nφrvang	**Cytheropteron arcuatum** Brady, Crosskey and Robertson
Elphidium clavatum Cushman	**Cytheropteron inflatum** Brady, Crosskey and Robertson
Elphidium incertum/asklundi (Williamson)/Brotzen	**Cytheropteron latissimum** (Norman)
Elphidium sp.	**Cytheropteron nodosum** Brady
Eoeponidella pulchella (Parker)	**Cytheropteron paralatissimum** Swain
Haynesina orbicularis (Brady)	**Cytheropteron pseudomontrosiense** Whatley and Masson
Islandiella helenae Feyling-Hanssen and Buzas	**Heterocyprideis sorbyana** (Jones)
Islandiella norcrossi (Cushman)	**Palmenella limicola** (Norman)
Pateoris hauerinoides (Rhumbler)	**Sarsicytheridea punctillata** (Brady)

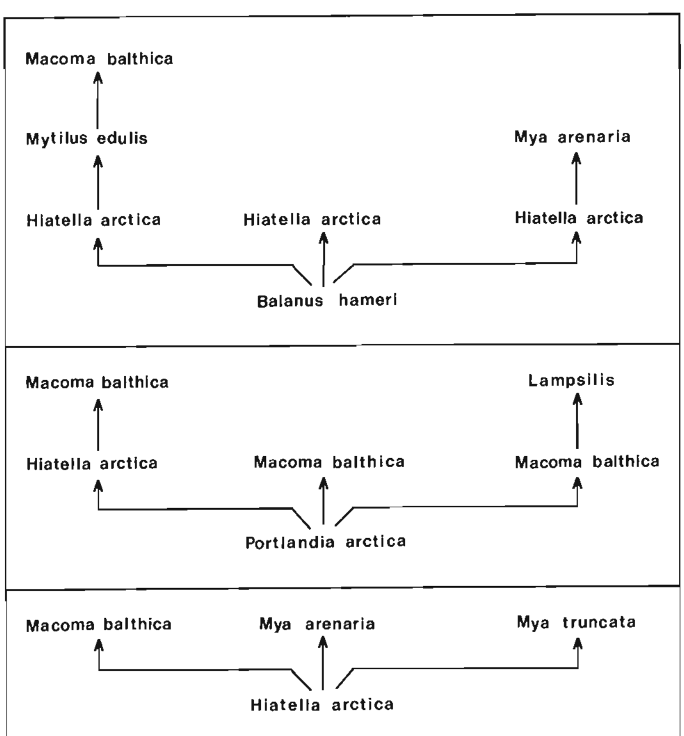

Figure 11. Some successions of macrofaunal associations in the late glacial and postglacial sediments of the western basin of the Champlain Sea.

MARINE ENVIRONMENTS

Salinity, temperature, and substrate are generally the primary factors controlling the distribution of marine benthonic invertebrate organisms. The bottom water in the western basin of the Champlain Sea was essentially cold throughout the marine episode and hence in most cases water temperature was not a significant controlling factor. However, warmer boreal water did migrate into the eastern part of the study area during the latter part of the marine episode and was characterized by the **Mya arenaria** association in the Cornwall, Alexandria, Huntingdon, Vaudreuil, and Lachute map areas. The dominant species of the macrofaunal associations and microfaunal assemblages occur in large numbers in both coarse and fine grained marine sediments. Hence, substrate was generally not a controlling factor. Exceptions to this are **Portlandia arctica** (Gray) and **Cytheropteron pseudomontrosiense** Whatley and Masson which are found mainly in clay and **Balanus hameri** (Ascanius) and **Elphidium incertum/asklundi** (Williamson)/Brotzen which occur almost exclusively in pebbly sandy clay and pebbly sand. Because water temperature and nature of substrate were not major controlling factors, the assemblage occurrence must have been controlled primarily by salinity. The successions of assemblages indicate that the salinity of the bottom waters of the Champlain Sea decreased with time.

Modern distribution data for some of the dominant foraminiferal species and the successions of foraminiferal assemblages are used to reconstruct the paleosalinity of marine water in the study area. Based on modern salinity preferences, three general salinity zones can be recognized. The highest salinity bottom water (30-34‰) was characterized by assemblages in which **Cassidulina reniforme** Nφrvang, **Islandiella helenae** Feyling-Hanssen and Buzas, **Islandiella norcrossi** (Cushman), and **Astrononion gallowayi** Loeblich and Tappan were abundant. **Elphidium clavatum** Cushman and **Haynesina orbicularis** (Brady) were the most abundant species in bottom water with salinity between 15 and 30‰. The lowest salinity bottom water (<15‰) was characterized by **Elphidiuim** sp. – dominant assemblages. The macrofaunal associations and ostracode assemblages

Table 2. Paleosalinity for microfaunal assemblages and macrofaunal associations in the western basin of the Champlain Sea (see Fig. 12).

Foraminiferal Assemblage		Macrofaunal Association		Ostracode Assemblage		Salinity of Bottom Water
No.	Dominant Species	No.	Dominant Species	No.	Dominant Species	
15	Elphidium sp. Haynesina orbicularis	8	Mya arenaria			Low (<15°/oo)
		3	Macoma balthica			
14	Elphidium sp. Haynesina orbicularis Elphidium clavatum	3	Macoma balthica	9	Cytheromorpha macchesneyi	
				8	Cytheromorpha macchesneyi Heterocyprideis sorbyana Sarsicytheridea punctillata	
13	Elphidium clavatum Haynesina orbicularis Elphidium sp.	4	Hiatella arctica	6	Sarsicytheridea punctillata Heterocyprideis sorbyana	Intermediate (15-30°/oo)
		3	Macoma balthica	8	Cytheromorpha macchesneyi Heterocyprideis sorbyana Sarsicytheridea punctillata	
12	Elphidium clavatum	4	Hiatella arctica	7	Cytheropteron latissimum Sarsicytheridea punctillata	
		2	Portlandia arctica	2	Cytheropteron pseudomontrosiense	
11	Elphidium clavatum Haynesina orbicularis	4	Hiatella arctica	9	Cytheromorpha macchesneyi	
		3	Macoma balthica			
		2	Portlandia arctica			
		4	Hiatella arctica	7	Cytheropteron latissimum Sarsicytheridea punctillata	
		5	Mytilus edulis			
		2	Portlandia arctica	2	Cytheropteron pseudomontrosiense	
10	Pateoris hauerinoides Elphidium clavatum	2	Portlandia arctica	2	Cytheropteron pseudomontrosiense	
9	Haynesina orbicularis Elphidium clavatum	7	Mya truncata			
		5	Mytilus edulis			
		3	Macoma balthica			
		4	Hiatella arctica	7	Cytheropteron latissimum Sarsicytheridea punctillata	
8	Haynesina orbicularis Elphidium clavatum Eoeponidella pulchella	6	Macoma calcarea			
		4	Hiatella arctica	6	Sarsicytheridea punctillata Heterocyprideis sorbyana	
7	Elphidium clavatum Elphidium incertum/ asklundi Haynesina orbicularis	4	Hiatella arctica	5	Sarsicytheridea punctillata Cytheropteron nodosum Heterocyprideis sorbyana	
6	Haynesina orbicularis Cassidulina reniforme Elphidium clavatum	4	Hiatella arctica			
		3	Macoma balthica			
5	Elphidium clavatum Cassidulina reniforme Haynesina orbicularis	4	Hiatella arctica	4	Cytheropteron nodosum Cytheropteron inflatum	
		2	Portlandia arctica	3	Cytheropteron arcuatum Palmenella limicola	
				2	Cytheropteron pseudomontrosiense	
4	Cassidulina reniforme	2	Portlandia arctica	2	Cytheropteron pseudomontrosiense	High (30-34°/oo)
3	Cassidulina reniforme Islandiella helenae Haynesina orbicularis	4	Hiatella arctica			
		3	Macoma balthica			
		1	Balanus hameri			
2	Cassidulina reniforme Islandiella helenae	2	Portlandia arctica	1	Cytheropteron paralatissimum Cytheropteron arcuatum Cytheropteron pseudomontrosiense	
1	Cassidulina reniforme Astrononion gallowayi Islandiella norcrossi	1	Balanus hameri			

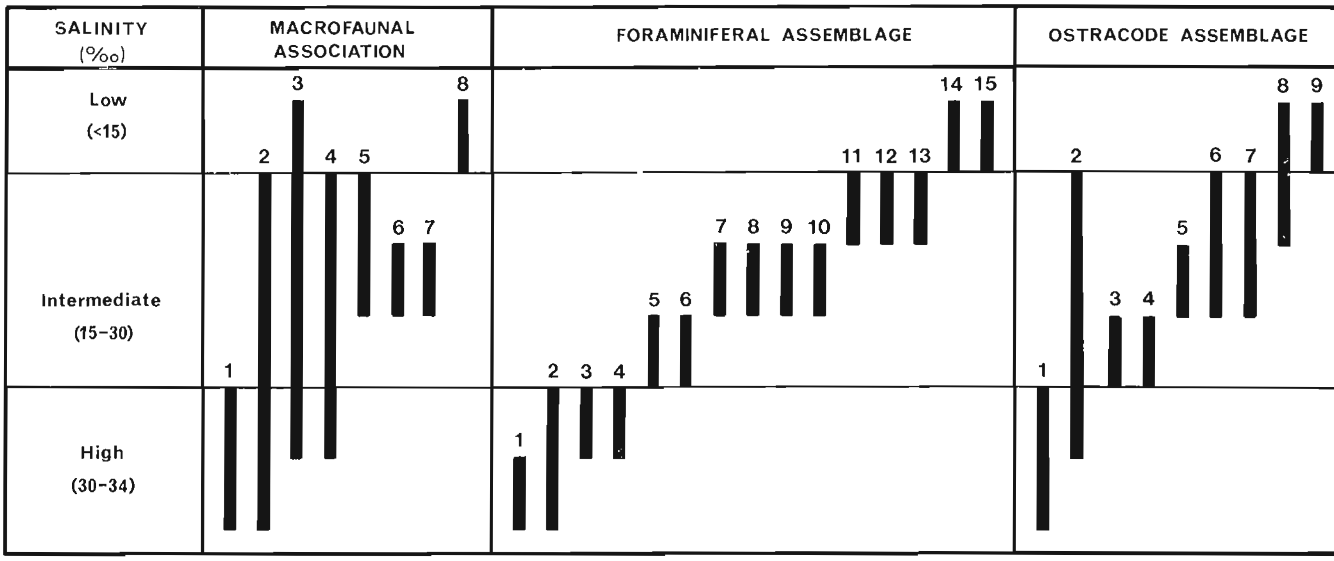

Figure 12. Distribution of late glacial and postglacial macrofaunal associations and microfaunal assemblages with respect to salinity in the western basin of the Champlain Sea. Numbers relate to associations and assemblages listed in Table 2.

Table 3. Dimensions of valves of **Hiatella arctica** (Linné) from sites in the Russell map area, Ontario

Site	No. of Valves	Length (mm) Range	Length (mm) Mean	Width (mm) Range	Width (mm) Mean	Stratigraphic Position
	337	4.1-31.7	17.5	2.1-17.1	8.5	Upper part of marine sequence
Navan	770	4.2-31.2	17.2	2.2-16.4	8.3	Upper part of marine sequence
	86	6.9-22.0	14.4	3.8-10.1	7.2	Lower part of marine sequence
	373	5.9-26.0	14.7	3.5-10.1	7.3	Shell bed at top of marine sequence
Bearbrook	96	12.8-34.5	25.2	6.0-16.5	12.8	Sand below shell bed
	310	5.6-19.9	13.7	2.5-10.9	7.2	Lower part of marine sequence

accompanying the foraminiferal assemblages in the high, intermediate, and low salinity bottom water of the Champlain Sea in the western part of the basin are listed in Table 2; their distribution with respect to the salinity zones is shown in Figure 12. Foraminifer and ostracode species diversity based on the Shannon-Wiener Information Function [H(S)] ranges from 0.00 to 1.98 and 0.00 to 2.23, respectively. The overlap of foraminifer diversity in high salinity bottom water (0.05-1.98) in intermediate salinity bottom water (0.00-1.89) and in low salinity bottom water (0.38-1.57) shows that foraminifer diversity and paleosalinity are not directly related in the western basin of the Champlain Sea.

Size of pelecypod shells has been used as a paleosalinity indicator in the Champlain Sea basin (Goldring, 1922). The size of the pelecypod shells is rather variable in the western basin of the Champlain Sea. The maximum dimensions of valves of **Macoma balthica** (Linné) range from 17.6 to 24.3 mm in length and 14.1 to 21.0 mm in width; those of **Hiatella arctica** (Linné) range from 18.0 to 36.0 mm in length and 10.1 to 20.0 mm in width. A summary of measurements on valves of **Hiatella arctica** from sections of Champlain Sea sediments in the Russell map area is given in Table 3. At the Navan site (No. 8 of Rodrigues and Richard, 1986), the valves from the upper part of the section are larger than those from the lower part of the section. Conversely, the valves from the lower and upper parts of the section are smaller than those from the middle part of the section at the Bearbrook site (No. 10 of Rodrigues and Richard, 1986). The foraminiferal assemblages indicate decreasing salinity from the base to the top of the marine strata at both sites. Thus, there does not appear to be a direct relationship between the size of **Hiatella arctica** shells and paleosalinity at the Navan and Bearbrook sites.

LITHOLOGICAL SUCCESSIONS

The glacial and postglacial deposits of the western basin of the Champlain Sea are grouped into five stratigraphic units. Five successions of these units that have

Table 4. Successions of stratigraphic units in the western basin of the Champlain Sea

Stratigraphic Unit	Succession 1	2	3	4	5
*4. Postmarine sediments	x	x	x	x	x
3. Fossiliferous marine sediments (Champlain Sea sediments)	x	x	x	x	x
2b. Varved sediments					x
2a. Nonfossiliferous stratified sediments	x	x			
1. Till	x		x		x

*The unit may be absent at some sites

been observed in the field are noted in Table 4. Unit 1 (till) and Unit 2a are commonly overlain unconformably by fossiliferous marine sediments (Champlain Sea sediments, Unit 3) which contain about 200 to 600 years of marine record as based on radiocarbon dates for pelecypod shells and cirriped plates from the base and for pelecypod shells from the top of the unit. The successions of macrofaunal associations and microfaunal assemblages within this unit indicate decreasing salinity of the bottom water during deposition.

The most complete succession (No. 5 of Table 4) is described from boreholes in the central parts of lower Ottawa Valley. The portion of the succession overlying till consists of the following lithofacies: rhythmically laminated clay, blue-grey clay and silty clay, banded grey silty clay and red clay, and interbedded sand, silt and clay (Gadd, 1977; 1986). Rodrigues (1984) reported a preliminary interpretation of the foraminiferal and ostracode assemblages from some of these cores (Table 5). The succession of faunal assemblages confirms the sequence of depositional environments recognized by Gadd (1986b), i.e., a freshwater environment, followed by marine and freshwater environments. The foraminiferal assemblages indicate that high salinity bottom water (30-34‰) migrated into the deeper parts of Ottawa Valley shortly after an initial freshwater episode. The maximum salinity conditions were followed by decreasing salinity of the bottom water and finally by freshwater

Table 5. Microfauna and macrofauna found in the lithofacies recognized by Gadd (1977) from boreholes in lower Ottawa Valley

Lithofacies (Depositional Environment)	[1]Foraminiferal Assemblage (Number)	Remarks
Interbedded sand, silt, and clay (upper delta facies of a prograding delta)		Post-Champlain Sea deposits; freshwater conditions; reworked foraminifers and ostracodes present
Regularly banded, grey silty clay and red clay (bottomset facies of a prograding delta)		
	Elphidium clavatum (12)	Champlain Sea deposits; salinity 30-34‰ in lower part of deepwater marine facies decreasing upwards to lower part of regularly banded fine grained facies; ostracodes rare, **Portlandia arctica** and **Yoldiella** sp. present in deepwater marine facies
Vaguely stratified to massive, dark blue-grey clay and silty clay (deepwater marine)	**Haynesina orbicularis** **Elphidium clavatum** (9)	
	Cassidulina reniforme **Islandiella helenae** (2)	
Varved clay (conditions similar to those of a freshwater glacial lake)		**Candona** sp. present; freshwater conditions

[1]See Table 2 and Figure 12

conditions during deposition of the regularly banded, grey silty clay and red clay lithofacies. Because Richard (1975a) has suggested that a mid-Champlain Sea readvance occurred, it is worth noting that the Champlain Sea succession (as seen in cores) was continuous without any significant hiatuses.

The dark blue-grey clay and silty clay (lowest marine unit of Gadd, 1986b) is commonly found in the low areas between ridges in lower Ottawa Valley and is characterized by the **Portlandia arctica** association. The regularly banded, grey silty clay and red clay facies is present in the Arnprior, Ottawa, Russell, Alexandria, Hawkesbury, Vaudreuil, and Lachute map areas. Articulated valves of **Macoma balthica** and **Portlandia arctica** are found in these sediments in the Russell and Alexandria map areas and the **Elphidium clavatum**-dominant assemblage (No. 12, Table 2 and Fig. 12) accompany **P. arctica** in the Alexandria map area. Disarticulated values of **Hiatella arctica, Macoma balthica, Mya arenaria** Linné, and **Mytilus edulis** Linné, some broken and others convex upwards, were observed in the banded sediment in the Vaudreuil map area. The absence of foraminifers and articulated pelecypod valves, however, suggests that the red and grey banded fine grained sediment at the Vaudreuil site is nonmarine. Therefore, the simple presence of marine fossils in the banded sediment does not necessarily prove that it is a marine deposit.

The boreholes discussed by Gadd (1977, 1986b) were located in the deeper parts of the Champlain Sea basin. Therefore, the lithofacies recognized from studies of the cores do not include littoral sediments or fine grained sediments which were deposited on high areas within the basin. The littoral sediments consist mainly of gravel and sand. The oldest, furthest west, and furthest north littoral deposits are characterized by the **Macoma balthica** association whereas younger littoral deposits contain the **Hiatella arctica, Macoma balthica,** and **Mytilus edulis** associations. **Mya arenaria** and **Mya truncata** associations are also present in the younger shallow water deposits; however, they are restricted to the eastern part of the study area. The main, topographically higher areas that have been studied within the western basin of the Champlain Sea are the ridges. The cores of the ridges contain glaciofluvial sediments which are overlain at some sites by a succession from deeper water (high salinity) fossiliferous sand, gravel and/or pebbly mud at the base to fossiliferous littoral (low salinity) sediments at the top. The deeper water deposits are characterized by the **Balanus hameri** or **Portlandia arctica** association.

DISCUSSION

The Champlain Sea did not occupy lower Ottawa and upper St. Lawrence valleys immediately following deglaciation. A conformable sequence of rhythmically laminated sediments (freshwater?) and marine clay is recognized in boreholes and in sections in lower Ottawa Valley and near Cornwall (Terasmae, 1965; MacClintock and Stewart, 1965; Gadd, 1986b). Early workers (see Prest, 1970) have used the presence of these sediments and later workers (see Anderson et al., 1985) have used the occurrence of freshwater ostracodes in these deposits to suggest that parts of the western basin of the Champlain Sea were occupied by a lake following deglaciation but prior to invasion by the Champlain Sea.

A succession from till to rhythmically laminated sediments to marine clay to fossiliferous sand is exposed near Sparrowhawk Point, New York, on the south shore of the St. Lawrence River (Fig. 1). At one spot along the exposure the rhythmically laminated sediments are truncated and underlain by marine clay which in turn is underlain by rhythmically laminated sediments. The presence of marine clay below the laminated sediments is apparently related to postdepositional slumping. Radiocarbon dates on shells of **Portlandia arctica,** 11 900 ± 100 BP (GSC-3767; Rodrigues and Richard, 1985) for the base of marine clay

Table 6. Faunal data and radiocarbon dates for late Pleistocene sediments exposed on the south shore of St. Lawrence River near Sparrowhawk Point, New York

Lithology (Position in Section)	[1]Macrofaunal Association (Number)	^{14}C Age (BP) (Lab No.)	[1]Foraminiferal Assemblage (Number)	[1]Ostracode Assemblage (Number)
Fossiliferous sand (top of section)	**Macoma balthica** (3)	11 300 ± 100 (GSC-3788)	**Elphidium clavatum Haynesina orbicularis** (11)	
Marine clay (base of marine clay overlying rhythmically laminated sediments)	**Portlandia arctica** (2)	11 900 ± 100 (GSC-3767)	**Pateoris hauerinoides Elphidium clavatum** (10)	**Cytheropteron pseudomontrosiense** (2)
Marine clay (below rhythmically laminated sediments)	**Portlandia arctica** (2)	11 900 ± 140 (GSC-4044)	**Elphidium clavatum Haynesina orbicularis** (11)	**Cytheromorpha macchesneyi** (9)
Rhythmically laminated sediments (underlain by till)				**Candona** sp.

[1]See Table 2 and Figure 12

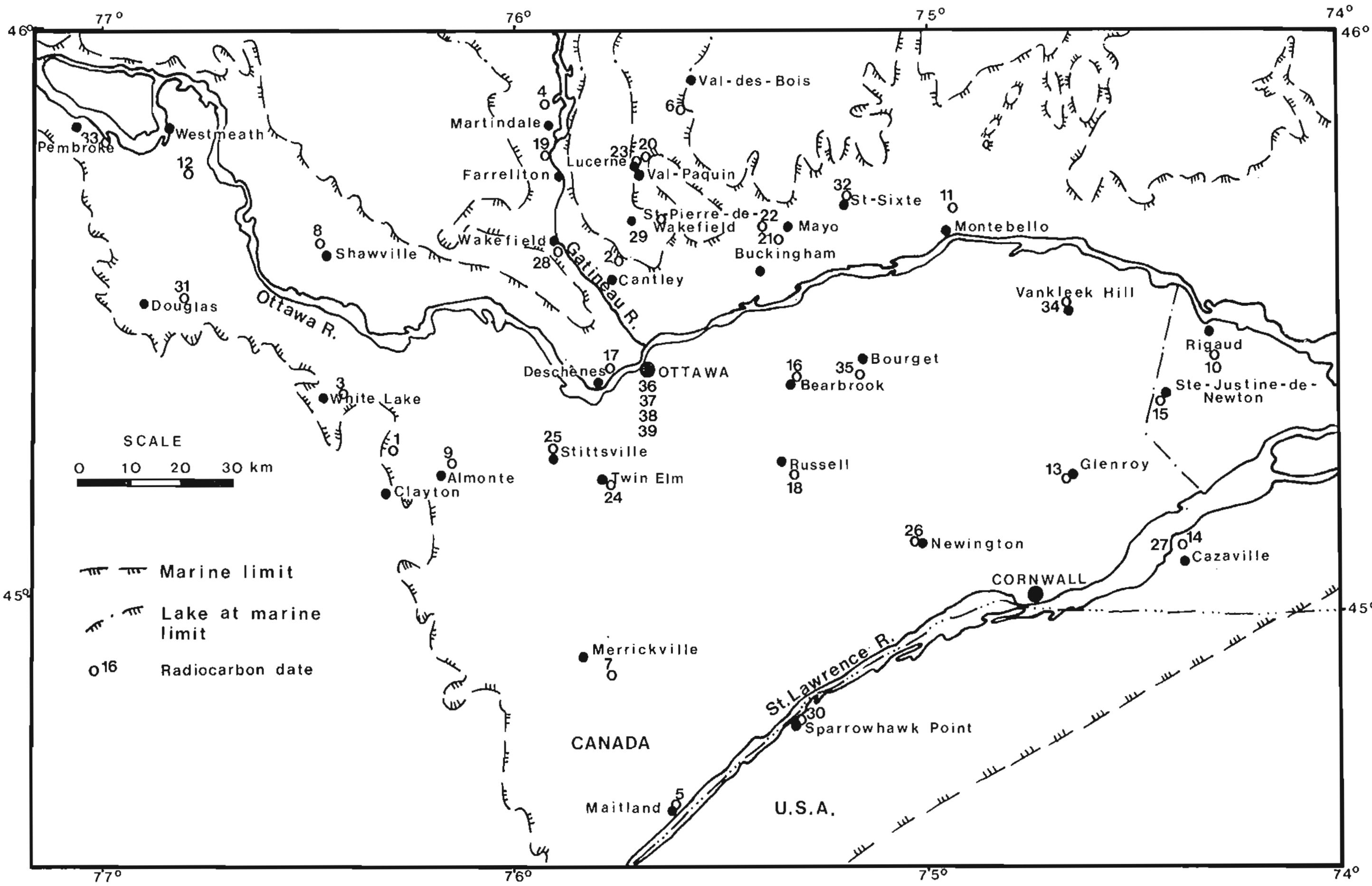

Figure 13. Location map for radiocarbon-dated sites (Table 7).

overlying the laminated sediments and 11 900 ± 140 BP (GSC-4044) for the marine clay below the laminated sediments, are consistent with the slumping hypothesis. The faunal assemblages for the units overlying till at the Sparrowhawk Point site are shown in Table 6. The freshwater ostracode **Candona** is present in the rhythmically laminated sediments. The foraminiferal assemblages show that the salinity of the bottom water was less than 30‰ and probably between 15 and 20‰ during deposition of the marine clay and fossiliferous sand (Table 2 and Fig. 12).

The timing of invasion of the Champlain Sea is a much debated point (Gadd, 1980a; Hillaire-Marcel, 1981; Karrow, 1981; Clark and Karrow, 1984; Anderson et al., 1985). Radiocarbon dates for shells of **Macoma balthica** from deposits near the maximum extent of the Champlain Sea in Ottawa Valley are 12.7 (Clayton site, Fig. 13) and 12.2 ka (Cantley and White Lake, Fig. 13). Therefore it appears that the change from fresh to marine conditions occurred before 12.2 ka in Ottawa Valley. Radiocarbon dates of 11.9 ka on shells of **Portlandia arctica** from the lower part of the marine clay overlying rhythmically laminated sediments at the Sparrowhawk Point site provide an approximate age for the change from freshwater to marine conditions in upper St. Lawrence Valley. The Sparrowhawk Point dates are comparable to dates of 11 800 ± 210 BP (GSC-1013; Lowdon and Blake, 1970) and 11 800 ± 100 BP (GSC-3523; Blake, 1984) for shells of **Macoma balthica** from beach deposits near Maitland and Merrickville in upper St. Lawrence Valley. The question which remains unanswered is whether the 300 year or more apparent difference between the arrival of marine waters in lower Ottawa Valley and in upper St. Lawrence Valley is real or whether it is due to incorporation of old carbon by shells in the Ottawa area.

Pioneer macrofaunal associations in the western basin of the Champlain Sea were the **Portlandia arctica** and **Macoma balthica** associations. Six foraminiferal and two ostracode assemblages accompanied the pioneer macrofaunal associations. The **Cassidulina reniforme-Islandiella helenae** (No. 2; Table 2 and Fig. 12), and **Cassidulina reniforme** (No. 4) foraminiferal assemblages, **Cytheropteron pseudomontrosiense** (No. 2) ostracode assemblage, and the **Portlandia arctica** association colonized cold, high salinity bottom water (30-34‰) in the deep parts of the sea. The **Pateoris hauerinoides-Elphidium clavatum** (No. 10), **Elphidium clavatum-Haynesina orbicularis** (No. 11), **Elphidium clavatum** (No. 12), and **Elphidium sp.-Haynesina orbicularis-Elphidium clavatum** (No. 14) foraminiferal assemblages, **Cytheropteron pseudomontrosiense** (No. 2) and **Cytheromorpha macchesneyi** (No. 9) ostracode assemblages, and the pioneer macrofaunal associations colonized cold low salinity (<20‰) bottom water in the shallow parts of the sea. Most pioneer assemblages also occur in younger deposits but foraminiferal assemblage No. 10 and the ostracode assemblages No. 2 and 9 appear to be restricted entirely to the early part of the marine episode.

During the middle part of the marine episode, boreal water characterized by the **Mya arenaria** association migrated into the basin at shallow depths along the southern margin of the sea and did not penetrate west of 74°47'W (Rodrigues and Richard, 1983). The arrival of the

Mya arenaria association, in the shallow parts of the sea ca. 11.1 ka, is almost synchronous with that of the **Balanus hameri** association, in the cold high salinity bottom water in the deep parts of the western basin of the Champlain Sea (Fig. 10). The high salinity bottom water retreated from the western basin of the Champlain Sea during the late part of the marine episode ca. 10.5 ka and was replaced by lower salinity water which eventually became fresh ca. 10.1 ka. The **Lampsilis** association colonized freshwater environments in the Cornwall and Huntingdon map areas and in abandoned channels of Ottawa River in the Russell and Hawkesbury map areas. Freshwater pelecypod shells were also found in the Cobden map area near the western margin of the basin. The date of 11 400 ± 400 BP (GSC-3852; Rodrigues and Richard, 1985) for these indicates that a freshwater surface layer was present along the western margin of the basin during the middle part of the marine episode. The freshwater layer may be related to the outflow from Lake Algonquin into the Champlain Sea basin.

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CHRONOLOGY OF LATE QUATERNARY EVENTS IN THE OTTAWA REGION

R.J. Fulton[1] and S.H. Richard[1]

Quaternary materials of various kinds from the vicinity of Ottawa have provided numerous radiocarbon dates. Although some of the dated samples were such materials as freshwater shells, gyttja, peat, and whale bone, the vast majority have been marine shells. For the sake of uniformity in this report and to avoid the necessity of assessing the reliability of dates on different types of materials, we have used dates on marine shells, almost exclusively. There remains, however, the possibility that shells from some areas may have lived in marine water in which the $^{12}C/^{14}C$ ratio was in equilibrium with the atmosphere, whereas in other areas marine shells may have grown in waters where concentrations of old carbon (from bedrock, meltwater, old marine water etc.) may have upset the carbon isotope equilibrium. Some shells may thereby give radiocarbon dates older than the calendar age of the enclosing sediments.

With one exception, all dates used in this report were run by the Geological Survey of Canada. In this way we have avoided the difficulties of comparing numbers from laboratories that use different analytical and reporting techniques. From the large number of dates available from that source, only dates used in this report are listed in Table 7. Additional dates are presented in Rodrigues and Richard (1985). Figure 13 shows the sites of radiocarbon-dated material.

In this report, we have divided the late Quaternary into arbitrary but apparently meaningful time slices and have described the chronology of events within them. Despite the large number of dates available, it is not always possible to date specific events accurately; as a result, there may appear to be conflicts in the chronologies of different areas. Such correlation problems may be due to variation in seawater carbon isotope contents in different parts of the basin, misinterpretation of the history of collection sites, or to our attempting to define the chronology at a level of accuracy beyond the limits of reliability of shell radiocarbon dates. Despite these possible limitations, we are presenting the first relatively comprehensive chronology of late Quaternary events for the western basin of the Champlain Sea.

EARLIEST DEGLACIATION

The oldest date on marine shells in the Ottawa region, 12 700 ± 100 BP[2] (GSC-2151; 1 of Table 7), is on **Macoma balthica** collected near Clayton, west of Ottawa (Richard, 1978; Fig. 13). This date from shells at an elevation of 168 m (near marine limit), is 500 ^{14}C years older than other samples from the area and is in fact the oldest date on marine shells anywhere in the Champlain Sea basin. In addition, it is 600 ^{14}C years older than GSC-3110 (3 of Table 7) which was collected from littoral sediments at about 170 m elevation at White Lake approximately 15 km to northwest. Because the Clayton date is so far out of line with dates on the same pelecypod species in the same stratigraphic position, it is regarded as an unreliable estimate of the age of the enclosing sediments and therefore is not considered further in this paper. A subsequent accelerator date on part of this same collection (12 180 ± 90 BP, TO-245) is comparable to the White Lake date and casts additional doubt on the 12.7 ka age of the Clayton shells.

Two other sites in the Ottawa region have yielded dates of >12 ka at White Lake and Cantley (3 and 2a of Table 7, respectively) where the dated shells (**Macoma balthica**) are from sand and probably lie near marine limit.

Taken at face value these dates suggest that the Ottawa area was undergoing deglaciation and was being invaded by the Champlain Sea as early as 12.2 ka. This does not fit with the interpretation of Karrow (in press) and Karrow et al. (1961) that Lake Iroquois was present in the Lake Ontario basin until after 12 ka and that the Champlain Sea did not invade the area until after Lake Iroquois had drained. Karrow (1981) explained this discrepancy by suggesting that old carbon in the Champlain Sea water has resulted in unreliable shell ages. Gadd (1981, 1987) argues that development of a calving bay permitted Champlain Sea to reach the Ottawa region while ice remained in upper St. Lawrence Valley, ponding Lake Iroquois in the Lake Ontario basin. It has, however, been argued that even though a calving bay developed in the lower St. Lawrence, this did not advance upstream from Quebec City (Chauvin et al., 1985). Unfortunately these differences of interpretation remain unresolved because there is no means of identifying "anomalous" dates and if these could be singled out, we would not know how much error to assign to them. In addition, because sedimentological and geomorphological evidence have not been used to verify the proposed methods of deglaciation, the mechanism and sequence of deglaciation are not known.

Rhythmically bedded sediments, containing the freshwater ostracode **Candona** conformably underlie Champlain Sea sediments at several places in upper St. Lawrence Valley and in the Ottawa area (Gadd, 1977, 1986a; Anderson et al., 1985; Rodrigues, 1987; Anderson, 1987). Hence if the ^{14}C dates on marine shells mentioned above are accepted as accurate maximum ages for Champlain Sea sediments, then the freshwater phase of deposition occurred before 12.2 ka.

PERIOD OF ESTABLISHMENT OF THE CHAMPLAIN SEA

Between 12.1 and 11.1 ka the Ottawa region was undergoing deglaciation and the Champlain Sea was characterized by "pioneer" fossil associations (Rodrigues and Richard, 1983). By the end of this time the western basin of the Champlain Sea was free of ice except in valleys where the sea extended far north into the Laurentian Highlands.

It is known that upper St. Lawrence Valley was free of ice and that Lake Iroquois had drained near the beginning of this time because of dates of 11.8 and 11.9 ka obtained for shells in the area from Maitland, Merrickville and Sparrowhawk Point (5, 7, and 30 of Table 7). An estimated age of 11.7 ka for basal gyttja from Lambs Pond (Anderson et al., 1985) corroborates this conclusion. Dates near this same age from Martindale and Val-des-Bois (4 and 6 of Table 7) suggest that ice had retreated as much as 50 km up Gatineau Valley at this time. An age of ca. 11.7 ka for stony clay thought to be of glaciomarine origin (ice proximal) at Saint-Pierre-de-Wakefield, 15 km east of the Gatineau valley and 20 km north of Ottawa (29 of Table 7) and an age

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[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

[2] A date of 12 800 ± 220 BP (GSC-1859) was first obtained for shells from this site (Richard, 1974). The inner fraction of a second collection made from this site gave the date of 12 700 ± 100 BP. A subsequent accelerator date, run on material from the same collection as GSC-2151, gave an age of 12 180 ± 90 BP (TO-245).

of ca. 11.8 ka for glaciomarine diamicton near Wakefield (28 of Table 7), 20 km northwest of Ottawa, suggest that the dates from farther north may be anomalously old. If a rate of recession of the ice margin of 0.25-0.3 km/year is assumed, as has been calculated for a marine-based ice sheet for the coastal areas of central Maine (Stuiver and Borns, 1975), it would have taken 500 to 600 years for ice receding south to north to retreat from Maitland to Val-des-Bois (150 km). Hence southern deposits would be expected to be 500-600 years older than those at Val-des-Bois. But since shells from upper St. Lawrence Valley in the south and the Val-des-Bois site near the northern margin of the basin give equivalent dates, either contamination or presence of old carbon has caused the shells to give unreliable ages or the Champlain Sea did not occupy the Ottawa region progressively from south to north as ice disappeared from the area.

Shells from the top of the marine clay sequence at several sites in valleys within the Laurentian Highlands (2b, 19, 20, 21, 23, and 32 of Table 7) provide approximate but maximum dates for retreat of Champlain Sea from these valleys. These indicate that the top of the marine succession in Gatineau Valley at Farrellton is ca. 11.7 ka and that marine clay was still being deposited at an elevation of 195 m at Cantley at ca. 11.8 ka. At Val-Paquin and Mayo (30 and 20 km north of Ottawa, respectively) marine clay was still being deposited at elevations of 195 and 182 m at 11.5 ka and at Lucerne, 2 km northwest of Val-Paquin, marine clay was being deposited at 175 m, 11.2 ka. By 11.4 ka (8 of Table 7) ice had retreated west of Shawville (60 km northwest of Ottawa), and sea level had dropped from a marine limit of between 190 and 185 m to 170 m. Also to the west, at Douglas (100 km west of Ottawa) ice had disappeared and Champlain Sea sediments were being deposited by 11.7 ka (31 of Table 7).

In several places in the central part of the western basin of the Champlain Sea, sediments interpreted as glaciofluvial sediments in esker-like deposits contain marine shells. This is puzzling because several of these sites (24a, 25 and 40 of Table 7 among them) lie just south of the area through which Gadd (1980a) proposed passage of a calving bay. Entrance of the Champlain Sea could not predate development of the calving bay and yet development of the calving bay would have truncated the ice lobe plumbing system cutting off the flow of meltwater so that glaciofluvial deposition could not have occurred at these sites after development of the calving bay. If the ice front had retreated in a "window blind" fashion as proposed by Prest (1970) and Clark and Karrow (1984), the ice margin in the Ottawa area would have been fronted by freshwater so that ice contact deposits would have been deposited in a glaciolacustrine and not a glaciomarine environment. Alternatively, the glaciofluvial sediments might be interpreted as having been deposited during a readvance such as that proposed by Richard (1975b). However, as discussed below, several lines of evidence make such a readvance unlikely. The explanation favored by one author (R.J.F.) is that the marine shells in these sediments are not contemporaneous with the glaciofluvial deposits but are somewhat younger; possibly the sediments in which they are enclosed were slumped into their present position during melting of residual buried ice or during slumping of steep sided ice contact faces. Consequently, the dates obtained for these shells (ca. 11.3 ka) are only a minimum age for the deposition of the associated glaciofluvial deposits.

According to Rodrigues (1987) sediments deposited in the Champlain Sea during this time are mainly characterized by **Macoma balthica**, and **Portlandia arctica** associations with **Hiatella arctica** and **Balanus hameri** associations occurring in deposits laid down near the end of this period. **Mya arenaria** association also appeared in the western basin of the Champlain Sea at the end of this time.

If the deglaciation ideas of Prest (1970) and Clark and Karrow (1984) are followed, the lowlands south and east of Ottawa were occupied by a glacial lake confluent with those in the Lake Ontario basin and Lake Champlain valley (Belleville-Fort Ann Phase of Muller and Prest, 1985) at the time of entrance of the Champlain Sea shortly after 12 ka. Water levels in the Ontario basin quickly dropped to below present levels as St. Lawrence Valley was opened to the Great Lakes (Anderson and Lewis, 1985). Next, lakes in the entire Great Lakes basin fell to low levels as ice retreat from the Madawaska Highland opened ice marginal channels so that drainage from Lake Huron and western Great Lakes basins entered Ottawa Valley via Barron and Petawawa valleys west of Pembroke (Fig. 5) – and possibly at an earlier time by more southerly routes. The sequence of the opening of channels has been discussed by Harrison (1972) but unfortunately the exact chronology is not known. In addition to this large flow of water from glacial lakes to the west, significant quantities of meltwater entered the basin directly from the ice margin which was retreating northward into the Laurentian Highlands.

REGRESSION OF THE CHAMPLAIN SEA

Regression of the Champlain Sea would have begun as soon as marine limit was achieved but because valley walls were generally steep in the vicinity of marine limit, early withdrawal of the sea did not greatly affect the size or shape of the basin. After about 11.2 ka (23 of Table 7), when the sea began to retreat from the valleys of the Laurentian highlands, the area available for sediment deposition became more restricted and emerging sediments were subjected to subaerial and littoral marine erosion. A littoral bench was cut in glaciofluvial sediments at an elevation of 170 m near Shawville by 11.4 ka (8 of Table 7) and to the west of Ottawa sea level had dropped to near the top of glaciofluvial deposits near Westmeath by 11 ka (elevation 158 m; 12 of Table 7). At Almonte, west of Ottawa, sea level was near 154 m by 11.2 ka (9 of Table 7) and at Twin Elm to the south of Ottawa it was still above 100 m at 10.6 ka (24b of Table 7); however, it apparently did not drop below 94 m at Ottawa until after 10.1 ka (17 of Table 7). At the northern edge of the western Champlain Sea basin, sea level was about 180 m at Buckingham at 11.4 ka (22 of Table 7), was still above 175 m at Lucerne at 11.2 ka (23 of Table 7), and was near 167 m at Montebello at 11.1 ka (11 of Table 7). Few dates are available for higher levels of the sea from the lowland east of Ottawa; however at Rigaud the water level was ca. 160 m at 11.2 ka (10 of Table 7). There are many dates related to lower levels of the Champlain Sea in the lowlands. Some of these are: Glenroy, 79 m at 10.7 ka (13 of Table 7); Cazaville, 55 m at 10.6 ka (14 of Table 7); Sainte-Justine-de-Newton, 75 m at 10.3 ka (15a of Table 7); Bearbrook, 69 m at 10.2 ka (16 of Table 7); and Russell, 70 m at 10 ka (18 of Table 7). All of these are from shells which appear to be related to littoral sediments and, therefore, sea level was at or slightly higher than these levels at the time indicated by these dates.

The date at Russell (18 of Table 7) is the youngest date on marine shells in this area and indicates that locally marine waters were present until after 10 ka. The error term on this date is however large (± 320) so it is not significantly younger than several other dates.

According to Gadd (this publication), a delta developed near Petawawa in Ottawa Valley (large area of unit A upstream from Pembroke on Fig. 4). This delta emerged and the front migrated eastward as the Champlain Sea regressed.

Table 7. Selected radiocarbon dates on materials collected in the Ottawa Region

No.	Site	Elev. (m a.s.l)	Lab. No.	Age (years BP)	Material Dated	$\delta^{13}C$ ‰	Collector	Reference	Comment
A. Dates from beach and nearshore sediments									
1.	Clayton	168	GSC-2151	[1] 12 800 ± 100 [2] 12 700 ± 100	**Macoma balthica** (Linné)	-0.2 -0.2	S.H. Richard and W. Blake, Jr.	Richard, 1978	Marine limit
			TO-245	12 180 ± 90				Unpublished	
2a.	Cantley	194	GSC-1646	12 200 ± 160	**Macoma balthica**		R. Romanelli	Romanelli, 1975	Marine limit (198 m?)
3.	White Lake	170-71	GSC-3110	[1] 12 100 ± 100 [2] 12 200 ± 100 [3] 12 100 ± 100	**Macoma balthica**	-0.6 -0.5 -0.6	S.H. Richard	Rodrigues and Richard, 1983	Marine limit
4.	Martindale	176	GSC-1772	11 900 ± 160	**Macoma balthica**		R. Romanelli	Lowdon and Blake, 1973	Marine limit
5.	Maitland	103	GSC-1013	11 800 ± 210	**Macoma balthica**		E.P. Henderson	Lowdon and Blake, 1970	Near marine limit
6.	Val-des-Bois	180	GSC-2769	11 800 ± 100	**Macoma balthica**	-1.8	S.H. Richard	Richard, 1980	High marine level
7.	Merrickville	118	GSC-3523	11 800 ± 100	**Macoma balthica**	-0.7	C.G. Rodrigues and S.H. Richard	Blake, 1984	Near marine limit
8.	Shawville	170	GSC-3670	11 400 ± 190	**Macoma balthica**	1.7	R.J. Fulton	Blake, 1983	20 m below marine limit (?)
9.	Almonte	154	GSC-1672	11 200 ± 160	**Macoma balthica**		S.H. Richard	Lowdon and Blake, 1973	Intermediate level
10.	Rigaud	160	GSC-2296	11 200 ± 90	**Hiatella arctica** (Linné)	1.7	S.H. Richard	Richard, 1978	Near marine limit
11.	Montebello	167	GSC-2590	11 100 ± 120	**Hiatella arctica**	2.3	S.H. Richard	Richard, 1980	Intermediate level
12.	Westmeath	158	GSC-1664	11 000 ± 160	**Macoma balthica**	-1.6	P.J. Howarth	Lowdon and Blake, 1979	Intermediate level
13.	Glenroy	79	GSC-3845	10 700 ± 100	**Mya arenaria** (Linné)	-2.7	C.G. Rodrigues and S.H. Richard	Rodrigues and Richard, 1985	Low level
14.	Cazaville	55	GSC-2423	10 600 ± 140	**Macoma balthica**		S.H. Richard	Richard, 1978	Low level
15a.	Ste-Justine-de-Newton	75	GSC-2261	10 300 ± 100	**Mya truncata** (Linné)	1.5	S.H. Richard and W. Blake, Jr.	Richard, 1978	Low level
16.	Bearbrook	69	GSC-3907	[1] 10 200 ± 110 [2] 10 200 ± 90	**Hiatella arctica**	-0.8 0.3	C.G. Rodrigues	Rodrigues and Richard, 1985	Low level
17.	Deschênes	94	GSC-2189	10 100 ± 130	**Hiatella arctica**	1.3	S.H. Richard	Richard, 1978	Low level
18.	Russell	70	GSC-1553	10 000 ± 320	**Macoma balthica**		S.H. Richard	Lowdon and Blake, 1973	Youngest marine date in region
B. Dates from high level fine grained marine sediments									
2b.	Cantley	195	GSC-3844	11 800 ± 170	**Macoma balthica**		S.H. Richard	Rodrigues and Richard, 1985	Near marine limit (198 m?)
19.	Farrellton	180	GSC-3862	11 700 ± 100	**Macoma balthica**	-1.4	S.H. Richard	Rodrigues and Richard, 1985	Maximum for emergence
20.	Val-Paquin	195	GSC-3865	11 500 ± 130	**Macoma balthica**	-1.7	S.H. Richard	Rodrigues and Richard, 1985	Maximum for emergence
21.	Mayo	182	GSC-2878	11 500 ± 210	**Macoma balthica**	-0.6	S.H. Richard	Richard, 1980	Maximum for emergence
22.	Buckingham	180	GSC-2763	11 400 ± 140	**Hiatella arctica**	1.9	S.H. Richard	Richard, 1980	Maximum for emergence
23.	Lucerne	175	GSC-3997	11 200 ± 130	**Macoma balthica**	0.7	S.H. Richard	Unpublished	Maximum for emergence

As falling sea level lowered base level, Ottawa River cut channels into the emerged part of the delta and into what had been the floor of the Champlain Sea. There is little chronological information on progradation of this delta or the subsequent channel cutting phase. If the available dates are taken at face value, the delta quickly passed through some segments of the flat floored valley; marine shells at an elevation of 139 m at Pembroke give a date of 10.9 ka (33 of Table 7) and at Vankleek Hill, 200 km to the east, a date on **Lampsilis radiata** s.l. suggests freshwater conditions in the river channel at 61 m by 10.3 ka (34 of Table 7). Hence the river delta apparently prograded this 200 km in less than 600 years. The faith that can be placed in the accuracy of ^{14}C dates and the validity of comparing freshwater shell dates with marine shell dates should, however, be questioned. **Lampsilis** sp. at 53 m at Bourget (40 km west of Vankleek Hill, 35 of Table 7) was dated at 10.2 ka, and **Hiatella arctica** shells from littoral deposits at 94 m at Deschênes are dated at 10.1 ka (17 of Table 7). This is 90 km upstream from the Vankleek Hill site where the receding Champlain Sea shoreline had supposedly passed some 200 years earlier. Probably the only conclusion that should be drawn from these dates is that the delta moved rapidly through the flat floored basin.

At the beginning of this time (11.1 ka) drainage from the Great Lakes basin was probably still entering upper Ottawa Valley via Petawawa River (see above). Northward retreat of the ice permitted the Champlain Sea to extend possibly as far up valley as Deep River (Fig. 1; N.R. Gadd, personal communication, 1985) and certainly as far as Chalk River (Fig. 1; Catto et al., 1981), and retreat opened the depressed North Bay outlet between the Great Lakes basin and Ottawa Valley. Unfortunately the timing of opening of the North Bay outlet is not closely known but is estimated by Terasmae and Hughes (1960) to have occurred between 11 and 10 ka, by Harrison (1972) between 10.9 and 10.1 ka, and by

Table 7. (cont'd)

No.	Site	Elev. (m a.s.l)	Lab. No.	Age (years BP)	Material Dated	$\delta^{13}C$ ‰	Collector	Reference	Comment
C. Dates from glaciomarine(?), glaciofluvial(?), and glacial(?) sediments									
24a.	Twin Elm	105	GSC-3641	11 200 ± 200	**Portlandia arctica** (Gray)	-2.8	S.H. Richard	Blake, 1983	Marine clay interbedded with glaciofluvial (?) gravel
24b.	Twin Elm	104	GSC-587	10 620 ± 200	**Macoma balthica**		R.J. Mott	Mott, 1968	Glaciofluvial (?) sand
25.	Stittsville	130	GSC-2448	11 300 ± 120	**Hiatella arctica**		N.R. Gadd	Gadd, 1978	Sands (glaciofluvial?) with ball and pillow structure
26.	Newington	106	GSC-2108	11 200 ± 100	**Hiatella arctica**		S.H. Richard	Richard, 1975a	From compact diamicton
15b.	Ste-Justine-de-Newton	74	GSC-2391	10 500 ± 110	**Hiatella arctica**	1.5	S.H. Richard	Richard, 1978	From diamicton
27.	Cazaville	71	GSC-3882	[1] 10 300 ± 90 [2] 10 500 ± 90	**Hiatella arctica**	-0.1 1.2	S.H. Richard	Rodrigues and Richard, 1985	From diamicton
28.	Wakefield	140	TO-112R[4]	11 760 ± 120	**Portlandia arctica**		R.J. Fulton	Unpublished	From glaciomarine diamicton
29.	Saint-Pierre-de-Wakefield	160	GSC-3834	11 700 ± 150	**Portlandia arctica**	-1.1	S.H. Richard	Rodrigues and Richard, 1985	From glaciomarine diamicton
40.	Ottawa	107	GSC-4166	11 100 ± 130	**Hiatella arctica**	1.5	B.R. Rust	Unpublished	From glaciofluvial (?) gravel
D. Other dates									
30.	Sparrowhawk Point	76	GSC-3767	11 900 ± 100	**Portlandia arctica**	0.2	C.G. Rodrigues and S.H. Richard	Rodrigues and Richard, 1985	Marine clay overlying rhythmically laminated sediments
31.	Douglas	120	GSC-3872	11 700 ± 120	**Macoma balthica**	-3.6	C.G. Rodrigues, S.H. Richard and R.J. Fulton	Rodrigues and Richard, 1985	From clay clasts in sand
32.	Saint-Sixte	145	GSC-2863	11 500 ± 200	**Macoma balthica**	-4.8	S.H. Richard	Richard, 1980	Maximum age of emergence
33.	Pembroke	139	GSC-90	10 870 ± 130	Marine shells		J. Terasmae	Dyck and Fyles, 1963	Maximum age of emergence
34.	Vankleek Hill	61	GSC-3235	10 300 ± 90	**Lampsilis radiata** (Gmelin) s.l.		S.H. Richard	Lowdon and Blake, 1981	Early Ottawa R. terrace
35.	Bourget	53	GSC-1968	10 200 ± 90	**Lampsilis** sp.		N.R. Gadd	Gadd, 1976	Alluvium in abandoned channel
36.	Ottawa	60	GSC-546	8 830 ± 190	Marly gyttja		R.J. Mott	Lowdon et al., 1967	Basal organic in channel
37.	Ottawa	70	GSC-547	8 220 ± 150	Woody peat		N.R. Gadd and J. Terasmae	Lowdon et al., 1967	Dates channel filling
38.	Ottawa	46	GSC-4059	8 140 ± 100	Gyttja		R. McNeely, S.R. Brown, and J.P. Smol	Unpublished	Dates low terrace
39.	Ottawa	67	GSC-628	7 870 ± 160	Marly gyttja		J. Terasmae	Lowdon et al., 1967	Dates channel filling

[1] Outer fraction
[2] Inner fraction
[3] Middle fraction
[4] Originally IsoTrace normalized this date to $\delta^{13}C$ = -25‰ and reported an age of 12 160 ± 120. The revised date is normalized to 0‰ and hence is directly comparable to GSC shell dates.

GSC – Geological Survey of Canada
TO – IsoTrace (University of Toronto)

Karrow et al. (1975) between 10.5 and 10.4 ka. According to Teller and Thorleifson (1983) and Clayton (1983) Lake Agassiz drained into Lake Superior between ca. 10.8 and 9.9 ka. As Great Lakes water was draining into Ottawa Valley at this time, this means that all glacial meltwaters and nonglacial drainage waters from the Canadian Prairies were passing for a period of time through the North Bay outlet and entering the regressing Champlain Sea. At the same time, most glacial and nonglacial drainage from Ontario (except for extreme southern Ontario which drained into upper St. Lawrence Valley via lakes Erie and Ontario) would also have flowed into lakes Superior and Huron and through this eastern outlet. In addition to western waters, northern meltwater from glacial Lake Barlow entered from the Lake Temiskaming basin (Vincent and Hardy, 1979) and meltwater from the north entered through valleys draining the Laurentian Highlands. This means that meltwater from 3000 km of receding ice front poured into the western basin of the Champlain Sea at this time.

During this period the receding ice constructed the St. Narcisse Moraine. LaSalle and Elson (1975) have traced this feature from Lac Simon (50 km east of Ottawa and 30 km north of Ottawa River; Fig. 1) to Saint-Siméon on St. Lawrence River, about 450 km to the northeast. They suggested that this feature represents a climatically controlled glacial event. Hillaire-Marcel et al. (1981) argued that it represents an equilibrium phase which occurred when the ice was adjusting its profile from retreat in the Champlain Sea basin to retreat over the land surface of the Laurentian Highlands. The age of the St. Narcisse Moraine has generally been referred to as slightly younger than 11 ka (Occhietti and Hillaire-Marcel, 1977). In most areas the St. Narcisse ice margin lay outside the western basin of the Champlain Sea but it might have been responsible for some of the outwash plains built in valleys of the Laurentian Highlands (such as the prominent terrace at Kazabazua; Fig. 1), and according to Catto et al. (1981), the advancing ice possibly deposited till in Ottawa River valley north of Chalk River.

Richard (1975b) reported finding marine fossils in the matrix of till and suggested that this was verification of a post-Fort Covington readvance (the "Newington ice readvance") which had been postulated by Terasmae (1965). In subsequent years of regional mapping in the Ottawa area, many other sites where marine shells are enclosed by diamicton were discovered. In almost all cases the presence of the shells in the diamicton can be explained without bringing ice back into the area to rework Champlain Sea sediments; in some places the shells were probably incorporated in diamicton during the slumping of ice contact and marine deposits; in others, the upper few centimeters of till may have been converted into a shell-rich diamicton by bioturbation and other processes at the seabed. One factor which suggests that these do not relate to a single ice readvance phase is the wide range in ages of shells at different sites (from 11.2 ka to 10.3 ka; 26 and 27 of Table 7). Other factors which appear to rule out a major ice advance after the Champlain Sea invaded this area are that till has not been found overlying marine clay and cores from the Ottawa Valley display a conformable sequence of fine grained Champlain Sea sediments without any evidence of a sudden change in depositional environment which might be related to advance or overriding by ice. Another line of evidence which appears to rule out a late ice readvance is that marine sediments dated from 12.2 to 11.2 ka, lying north of Ottawa River, are not overlain by till nor do they show any evidence of ice overriding. If a readvance had occurred it would have originated in the highlands north of Ottawa River. One further point against a late readvance is that according to Rodrigues and Richard (1985), boreal water, characterized by the **Mya arenaria** association, migrated into the western basin of the Champlain Sea and was present at least as far west as Vankleek Hill and Cornwall (Fig. 13) between 11.1 and 10.5. This warmer water episode of the Champlain Sea is not compatible with ice shelves and tidewater glaciers. Hence, the occurrence of a "Newington ice readvance" appears to be ruled out.

Despite these points, one author (S.H.R.) continues to believe that a late ice advance covered much of the western basin of the Champlain Sea. This is based on the interpretation of the depositional environment of fossiliferous sediments which occur mainly in ridges that contain cores of glacial or ice contact glaciofluvial deposits. Of the sites that have been dated, in 13 places the dated shells came from pebbly mud or clayey sands which intertongue with sediments interpreted as being deltaic or subaqueous outwash; at 10 sites these shells were interpreted as being autochthonous in deltaic or subaqueous outwash deposits and in 4 cases the enclosing diamicton was interpreted to be till. These interpretations of the fossiliferous sediments as being closely associated with glacier ice leads to the hypothesis that a late glacial advance(s?) caused coalescing ice lobes or tidewater glaciers emanating from the north to push out into the marine embayment between 11.3 and 10.3 ka. It should be noted that these dates correspond well with the age of the Younger Dryas, a time of cooler climate and glacier readvance in northern Europe. Evidence that eastern Canada also felt the effects of this cooling has been reported from Atlantic Canada (Brown-Macpherson and Anderson, 1985; Mott, 1985). These ridges clearly result from and have been subjected to complex and poorly understood processes. A clear understanding of all the factors that have been involved in their formation will be necessary before the controversy surrounding their origin and interpretation will be settled to everyone's satisfaction.

DEVELOPMENT OF THE MODERN ENVIRONMENT

As mentioned above, 10 ka is the youngest date on marine shells in the Champlain Sea basin (18 of Table 7). During earlier regression and following this, Ottawa River developed a complex channel system as it cut into the former floor of the Champlain Sea. High level channels and terraces were cut and abandoned as the flow shifted away from channels hung up on till or bedrock ridges and as discharge into the system fluctuated. Landsliding occurred as channels were cut into the saturated marine sediments and local lakes and extensive peat bogs developed in abandoned channels.

As noted above, at the beginning of this period Ottawa River carried meltwater from approximately 3000 km of ice margin. Shortly after this, however, the Marquette advance in the Lake Superior basin (Teller and Thorleifson, 1983) diverted Lake Agassiz waters southward into Mississippi drainage reducing the length of ice front drained by Ottawa River to ca. 2000 km. This only lasted until ca. 9.5 ka when ice retreat in the Lake Superior basin again permitted Lake Agassiz waters to flood into the Great Lakes basins and to drain into Ottawa Valley via the North Bay outlet. The initial breakthrough of Agassiz waters may have produced high rates of flow in the Ottawa River because Teller and Thorleifson (1983) estimated that flow rates into the Superior basin were in excess of 100 000 m^3/s and that early floods involved 4000 km^3 of water; possibly high terrace levels in lower Ottawa Valley are in some way related to this flood (Catto et al., 1982).

Flow from Lake Agassiz and also from glacial lakes to the north (mainly Barlow-Ojibway) would have continued to enter Ottawa Valley until ca. 8 ka when opening of Hudson Bay permitted drainage to the north (Dredge and Cowan, in press). The upper Great Lakes drainage continued to pass exclusively through the North Bay outlet until ca. 5.5 ka. At this time isostatic uplift had raised this outlet to the level of the Chicago and Port Huron outlets so that the Great Lakes entered the three outlet Lake Nipissing Phase (Lewis, 1969; Prest, 1970). Great Lakes waters no longer entered Ottawa Valley at 4.7 ka, the age obtained for basal gyttja in the bottom of the North Bay outlet channel (Lewis, 1969). Hence, by 4.7 ka the basin drained by Ottawa River was reduced to approximately its present size.

Little data related to the discharge history of the river have been obtained so far from within the western basin of the Champlain Sea. It is known, however, that a number of early river channels were abandoned prior to 8 ka (36, 37, and 39 of Table 7). It is not known whether these channel abandonments were related to discharge events or to the transfer of flow from a channel hung up on till or bedrock to one cutting into clay. A date of 8.1 ka (38 of Table 7), from basal organic sediments collected from a plunge pool basin at the level of a low terrace in Ottawa, suggests that Ottawa River may have been near its present level and position at that time. This date may be anomalously old, however, due to hardwater conditions in this small lake basin (R.N. McNeely, personal communication, 1985).

SHORE LEVEL DISPLACEMENT – GLACIAL ISOSTASY

Analysis of data in Table 7 gives a south to north upward tilt of 0.5 m/km from upper St. Lawrence Valley towards the Laurentian Highlands for the 11.8 ka water plane. This assumes that sediments enclosing shells dated 11.8 ka in upper St. Lawrence Valley are the same age as sediments enclosing shells radiocarbon dated 11.8 ka in Gatineau Valley. The slope compares with tilts of 0.53 to 0.75 m/km in the Lake Simcoe area (Deane, 1950) and of 0.5 to 1.2 m/km in the basin of lakes Barlow and Objiway (Vincent and Hardy, 1979) and consequently appears low for a water plane immediately postdating deglaciation. If instead of comparing elevations on the apparent 11.8 ka water plane in Gatineau and upper St. Lawrence Valley areas, the marine limit date within each area is used, the apparent tilt towards the north is 1.6 m/km. This illustrates the difficulty associated with using ^{14}C dates when the validity of dates and the relative timing of deglaciation in two areas are in question. If the ^{14}C dates are accepted, then deglaciation must have occurred 300 years earlier in the Gatineau area than in the upper St. Lawrence, marine limit in the two areas is not synchronous, and hence tilting of this surface is an unreliable measure of isostatic tilting. If the marine limits in the two areas are assumed to fall on a synchronous water plane, then the sediment ages as indicated by the shell dates are being ignored and it is assumed that the two areas were inundated by the Champlain Sea at the same time. This may not be too far from being correct because, if deglaciation is assumed to have occurred by regular northward retreat of the ice front and a proglacial lake occupied much of the Ottawa Lowland prior to marine inundation, occupation of the former lake basin by the Champlain Sea would have been instantaneous so that marine limit in at least part of the area would be synchronous.

The data listed in Table 7 are areally scattered and consequently do not lend themselves well to shoreline displacement analysis. Data from a single site near Wakefield, however, indicates uplift may have been as great as 10 m/century (see stop C-2 of Fulton, 1987). An emergence curve based on selected Table 7 data is presented as Figure 14. Construction of this curve is seriously handicapped by a lack of younger and lower level fixes on sea level. An important question is the shape of the curve connecting the lowest segment which is based on data points, to the 0 origin of the diagram. A sharp inflection is necessary to keep the curve above present sea level (line A of Fig. 14). If the uplift rate did not make this sudden and uncharacteristic change then the curve would pass below present sea level (B of Fig. 14). If the uplift curve for this area is shaped like curve B then it suggests that following initial uplift the area underwent subsidence. Such a history of isostatic movement might explain the development of lakes in channels excavated by Ottawa River and the apparent presence of submerged river channels in the floors of these lakes (Shilts, 1984).

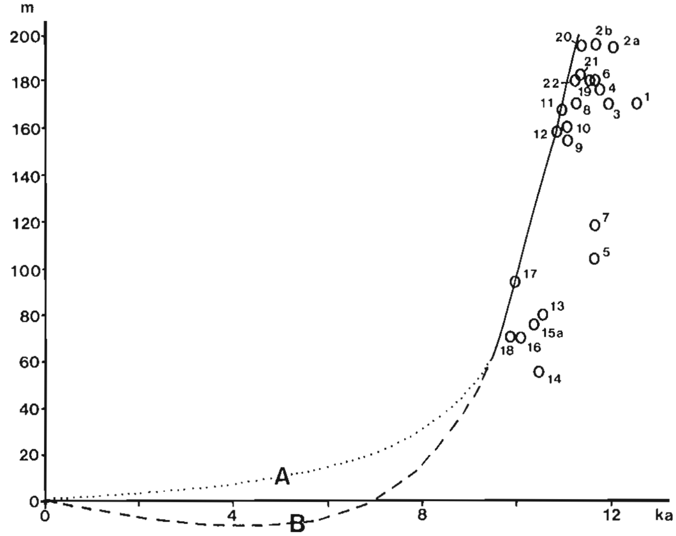


Figure 14. Emergence curve for the Ottawa region. The diameter of the circles represents 200 years or the approximate error term associated with most dates. Numbers refer to dates listed in Table 7. Hypothetical extension "A" shows the shape of the curve if the area has only been subject to isostatic rebound; extension "B" is drawn such that the curve is more similar in shape to emergence curves from other areas, but suggests that the area has undergone a period of subsidence.

REGIONAL CORRELATIONS

Figure 15 shows events that occurred in southern glacial Lake Agassiz, and in the Lake Huron, Lake Ontario, and Ottawa regions. There has been no physical correlation of stratigraphic units between these areas and the columns are merely aligned using the suggested ages of events. This provides a deceptively well integrated picture. In fact, most events in the various areas are not as closely dated as the columns imply; error terms of ^{14}C dates allow for a 100-200 year shifting of boundaries and the chronologies for the different columns are based on ^{14}C dates on different types of organic materials; the Lake Agassiz chronology is based on wood dates, the Lake Huron chronology on peat and gyttja dates, and the Champlain Sea chronology on marine shell dates. Wood has generally been considered to supply the most reliable ages, peat can contain materials of mixed ages, gyttja can give anomalous dates where mixed with marl, and marine shell ages tend to be old compared with those for other organic materials (Karrow, 1981). In lieu of physical tracing of units or events from basin to basin, however, this is the best that can be done.

Probably the most interesting aspect of Figure 15 is the potential for correlation of drainage events. As mentioned above, glacier retreat and advance and differential isostatic uplift caused wide fluctuations in the size of the basin draining into the Champlain Sea. These changes in flow should have left an imprint on the deposits and the terraces and channels developed in the Ottawa area. Catto et al. (1982) did some preliminary work along this line in the Chalk River areas but unfortunately these aspects of Quaternary history of most of the Ottawa region have not yet been studied.

Finally, Figure 15 contains a column showing the late Quaternary chronostratigraphic units used in Norden. No correlation is implied but this is added as an aid to those familiar with the European terminology.

SUMMARY

Late Quaternary deglaciation of the Ottawa area commenced at some time prior to 12 ka and at about or slightly later than this time ice retreat permitted the Champlain Sea to enter the area. The conventional interpretation is that much of Ottawa Valley lowland was occupied by a glacial lake as the ice retreated northward and that the Champlain Sea broke through an ice barrier to the east and replaced the freshwater. An alternative hypothesis, however, is that the Champlain Sea entered the area by means of a calving bay that extended westward up the deep Ottawa Valley leaving residual ice in the shallow upper St. Lawrence Valley.

The entire western basin of the Champlain Sea was deglaciated by 11.1 ka and relatively stable marine conditions prevailed with new and slightly warmer water faunas invading the basin. Regression began at the time of deglaciation and a delta migrated eastward through the basin. Construction of this delta began at the western extremity of the Champlain Sea ca. 11 ka and it reached a point about 200 km to the east in ca. 1 ka. Apparently the Champlain Sea had receded from its western basin by 10 ka.

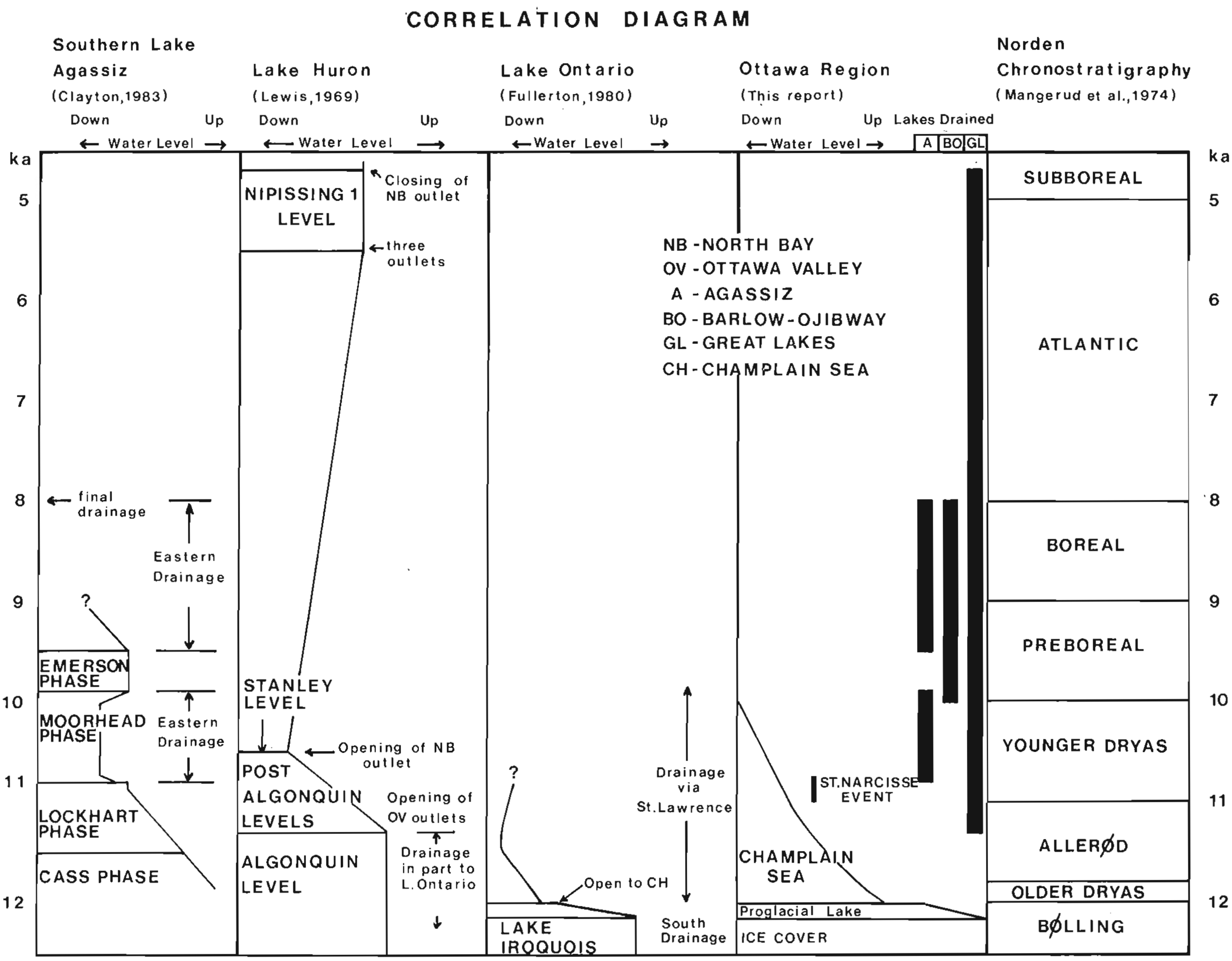


Figure 15. Correlation diagram showing time relationship between events of the western basin of the Champlain Sea and those in southern Lake Agassiz, and the Lake Huron and Lake Ontario basins.

High and variable discharges passed through Ottawa Valley and at times the valley's drainage basin included as much as 3000 km of retreating ice front. A noteworthy drainage event that should have left an imprint on the fluvial sediments and the terrace and channel configuration is the rapid drainage of Lake Agassiz. This apparently released 4000 km^3 of water which should have flowed through the area ca. 9.5 ka.

Marine limit is tilted up towards the north and emergence rates as high as 10 m/century appear to have occurred. The isostatic adjustment record may have been such that subsidence followed uplift. This might be an explanation for drowning of sections of Ottawa River.

The chronology presented in this paper makes it possible to compare the timing of deglaciation and later events in this area with those in the Lake Agassiz basin and the adjacent Great Lakes.

ACKNOWLEDGMENTS

We must first acknowledge the careful work of W. Blake, Jr. in operating the Geological Survey of Canada Radiocarbon Laboratory; without his co-operation and discussion of the geological interpretation placed on many dates, this report would not have been possible. C.G. Rodrigues helped with the collection of many of the dated samples, interpreted the paleoecology of many of the sites, and provided thought-provoking critiques of our geological interpretations. D.R. Grant and A.S. Dyke provided useful advice and discussion which aided in preparation of the shore level displacement part of this report. Even though this report may differ in detail from some of the ideas of N.R. Gadd, much of the general Quaternary information on which it is based stems from his earlier work and from numerous discussions both in the office and field.

TERRESTRIAL ENVIRONMENTS AND AGE OF THE CHAMPLAIN SEA BASED ON POLLEN STRATIGRAPHY OF THE OTTAWA VALLEY-LAKE ONTARIO REGION

T.W. Anderson[1]

The earliest palynological analyses of the Ottawa Valley-Lake Ontario region (Auer, 1930; Potzger, 1953; Potzger and Courtemanche, 1956, 1957) were restricted mainly to southern Quebec and lower St. Lawrence Valley area and lacked chronological control. Later studies by Terasmae (1965, 1980), Mott and Camfield (1969), and Camfield (1969) in eastern Ontario and Richard (1977), Savoie and Richard (1979), Mott and Farley-Gill (1981), Mott et al. (1981), and Delage et al. (1985) in southernmost Quebec provide regional coverage and each contains at least one basal radiocarbon date and, at some sites, up to five dates on major pollen boundaries. On-going palynological studies of sites outside the Champlain Sea basin and in the highlands of Ottawa Valley, east-central Ontario, and adjacent New York State are aimed towards filling gaps in the regional pollen stratigraphic coverage.

The majority of the pollen-analyzed sites are Late Wisconsinan and Holocene (post-23 ka, Fulton, 1984) with the exception of the sediments at the Pointe-Fortune site which are believed to correlate with the Late Sangamonian or Early Wisconsinan St. Pierre Interval of the St. Lawrence Lowlands (Veillette and Nixon, 1984). The Late Wisconsinan-Holocene records are from Champlain Sea sediments, underlying glaciolacustrine clay, overlying lake and bog deposits, and sediments from lake basins located outside the limit of the Champlain Sea.

Several lake and bog sites were selected as being representative of the region (Fig. 16). Pollen records at these sites provide data on regional patterns of deglaciation, vegetation and climatic history, and environments of deposition. Pollen counts from Champlain Sea sediments and the underlying glaciolacustrine unit in a core from adjacent to Mer Bleue Bog on the east side of Ottawa are compared with the upland pollen stratigraphy in an attempt to estimate the time of arrival of marine water in the western basin of the Champlain Sea.

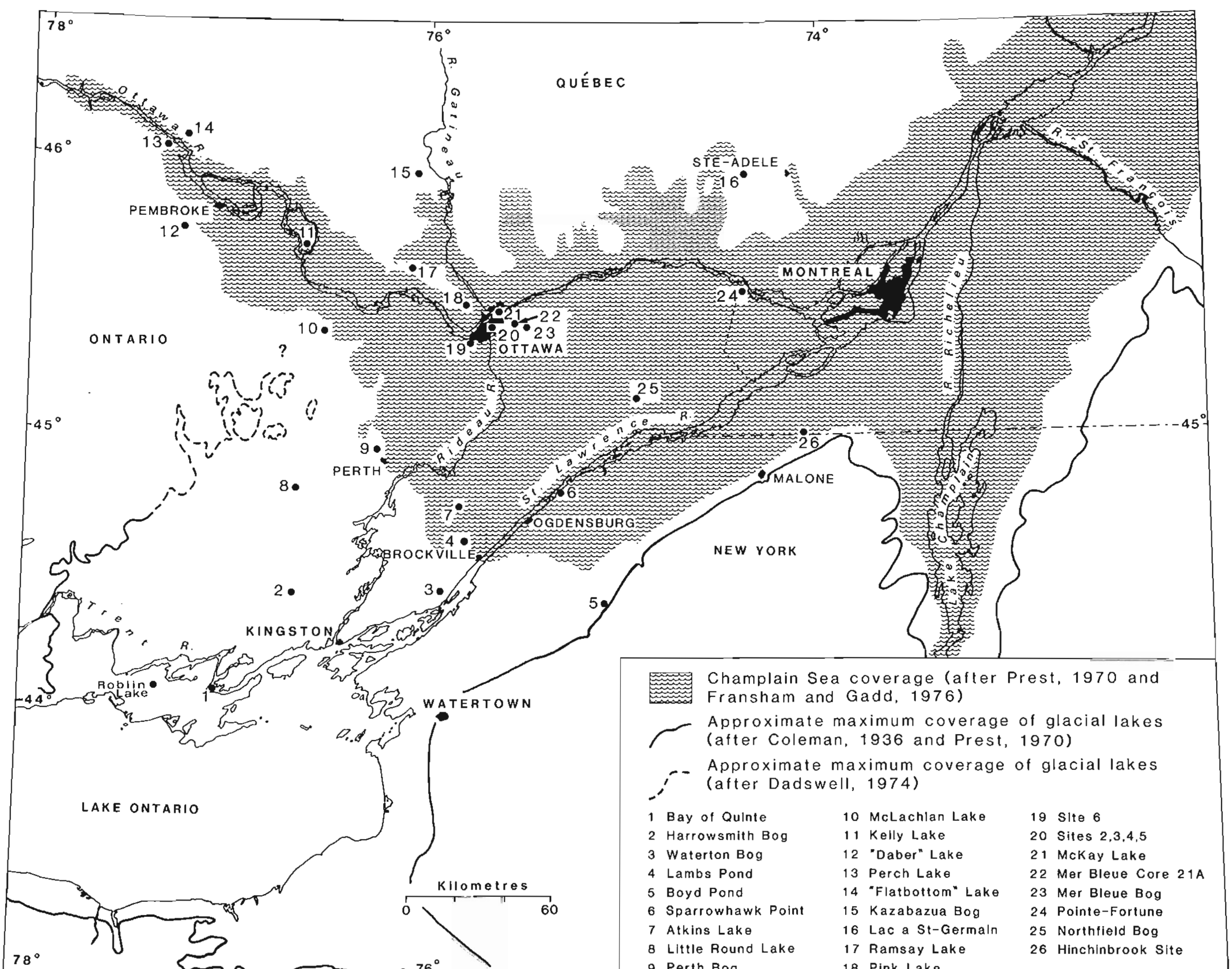


Figure 16. Lake Ontario-upper St. Lawrence-Ottawa Valley region showing approximate extent of glacial lake coverage, Champlain Sea shoreline, and locations of sites discussed in text.

[1] Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8

SETTING AND PRESENT-DAY VEGETATION

The geological setting and Quaternary deposits and features of the Ottawa Valley-Lake Ontario region are described by Gadd (this publication). Fulton and Richard (this publication) outline the general chronology of late Quaternary events based largely on ^{14}C dates on marine shells.

The Ottawa-St. Lawrence Lowlands stand in sharp contrast with the highlands to the north, west, and southeast providing a variety of lowland and upland habitats for plant colonization. Because this region is transitional between the Deciduous Forest Region to the south and southeast and the Boreal Forest Region to the north, it comprises a mixture of southern elements and those which are more typical of northern areas. Some plant taxa such as **Celtis occidentalis** and **Cephalanthus occidentalis** reach their northern limits in this region (I. Bayly, personal communication, 1985).

The present-day forest of the Ottawa Valley-upper St. Lawrence Valley region falls in the Great Lakes-St. Lawrence Forest Region of Rowe (1977). A mixed forest dominated by sugar maple (**Acer saccharum**), beech (**Fagus grandifolia**), yellow birch (**Betula lutea**), red maple (**Acer rubrum**), and eastern hemlock (**Tsuga canadensis**), almost always accompanied by white and red pine (**Pinus strobus** and **P. resinosa**), characterizes the upland areas to the north and west of the Champlain Sea basin. Within the basin the forest is predominantly deciduous, consisting of sugar maple and beech, with red maple, yellow birch, basswood (**Tilia americana**), white ash (**Fraxinus americana**), largetooth aspen (**Populus grandidentata**), and red and bur oak (**Quercus rubra** and **Q. macrocarpa**). Local occurrences in the Champlain Sea basin include white oak (**Quercus alba**), red ash (**Fraxinus pennsylvanica**), grey birch (**Betula populifolia**), rock elm (**Ulmus thomasii**), blue-beech (**Carpinus caroliniana**), and bitternut hickory (**Carya cordiformis**). Varying amounts of white spruce (**Picea glauca**), balsam fir (**Abies balsamea**), trembling aspen (**Populus tremuloides**), and white birch (**Betula papyrifera**) are common at higher elevations outside the basin. Black spruce (**Picea mariana**), black ash (**Fraxinus nigra**), white elm (**Ulmus**), and eastern white cedar (**Thuja occidentalis**) dominate hardwood and mixed-wood swamps; white butternut (**Juglans cinerea**), eastern cottonwood (**Populus**), and slippery elm (**Ulmus rubra**) are sporadically distributed along river valleys.

PRE-LATE WISCONSINAN RECORD

The only pre-Late Wisconsinan deposits in the area were described by Veillette and Nixon (1984) from a site near Pointe-Fortune. At this site, till is overlain by crossbedded sand and silt which is overlain by organic-rich clay. The organic-rich clay is overlain by stratified silt and fine sand which in turn is overlain by till. A date of >42 ka was obtained on wood from the upper sand-silt sequence (GSC-2932; Gadd et al., 1981).

Preliminary pollen and plant macrofossil analyses were undertaken on the organic clay interval and underlying and overlying silt and fine sand (Fig. 17). The pollen assemblage is dominated by **Picea**, **Pinus**, **Salix**, **Alnus**, Gramineae and Cyperaceae and minor occurrences of **Abies** and thermophilous tree pollen. Plant macrofossils consist mainly of seeds of **Carex**, **Eleocharis**, **Potentilla**, and **Hippuris vulgaris**.

The overall pollen assemblage indicates that cooler than present climatic conditions prevailed in the area during deposition of the organic bed. This supports the assignment of the interval to the St. Pierre Interstade of the St. Lawrence Lowlands by Gadd et al. (1981). The overlying till is believed to correlate with the Gentilly Till (LaSalle, 1984).

LATE WISCONSINAN-HOLOCENE RECORD

Pollen diagrams from Lambs Pond (Fig. 18), McKay Lake (Fig. 19), Ramsay Lake (Fig. 20), and Lac à St-Germain (Fig. 21) were chosen as being representative of the late glacial and Holocene pollen stratigraphy of the Ottawa Valley-St. Lawrence River valley-Lake Ontario region. These are supplemented by partial (late glacial) diagrams from Waterton Bog (Fig. 22), Boyd Pond (Fig. 23), Northfield Bog (Fig. 24), Kelly Lake (Fig. 25), and "Daber" Lake (Fig. 26). Figure 16 shows the locations of these sites.

Pollen stratigraphy

The pollen stratigraphy and radiocarbon dates for sites in the Ottawa Valley-Lake Ontario region are summarized in Figure 27. The lowermost zones (zones 9 to 6) are clearly time transgressive from south to north. Zones become synchronous for the first time everywhere in the region in the **Pinus** zone (5).

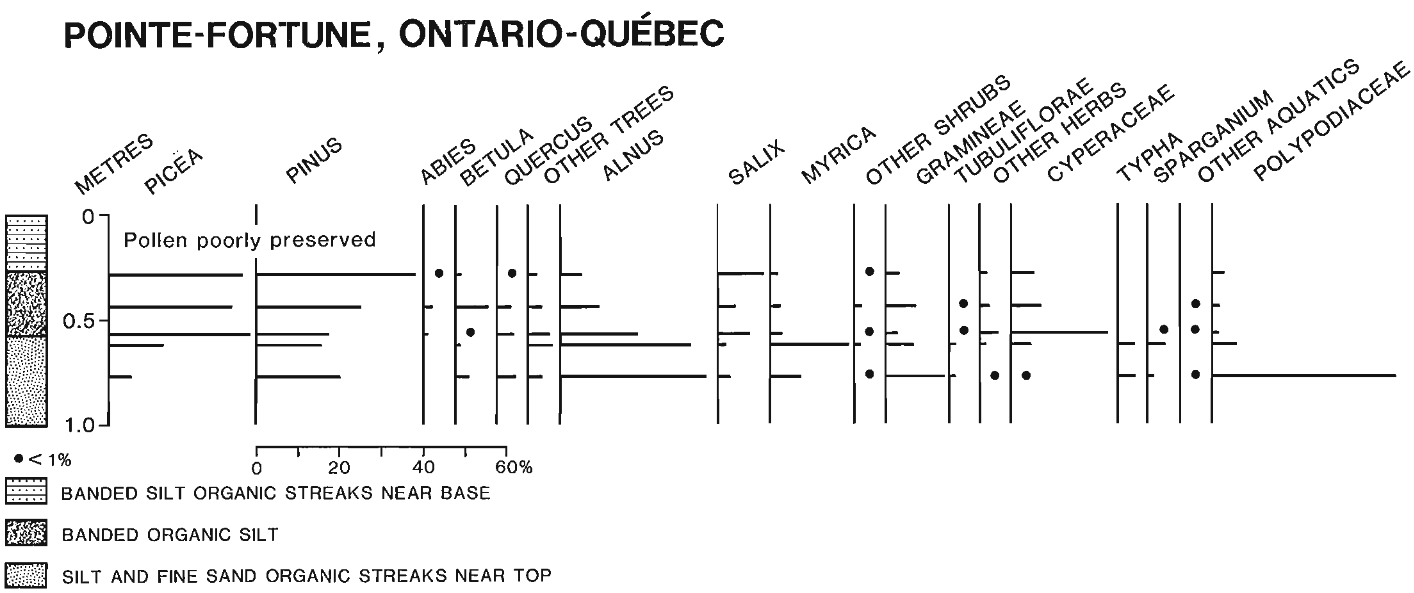


Figure 17. Pollen diagram of the Pointe-Fortune section, Ontario-Quebec border.

LAMBS POND, ONTARIO

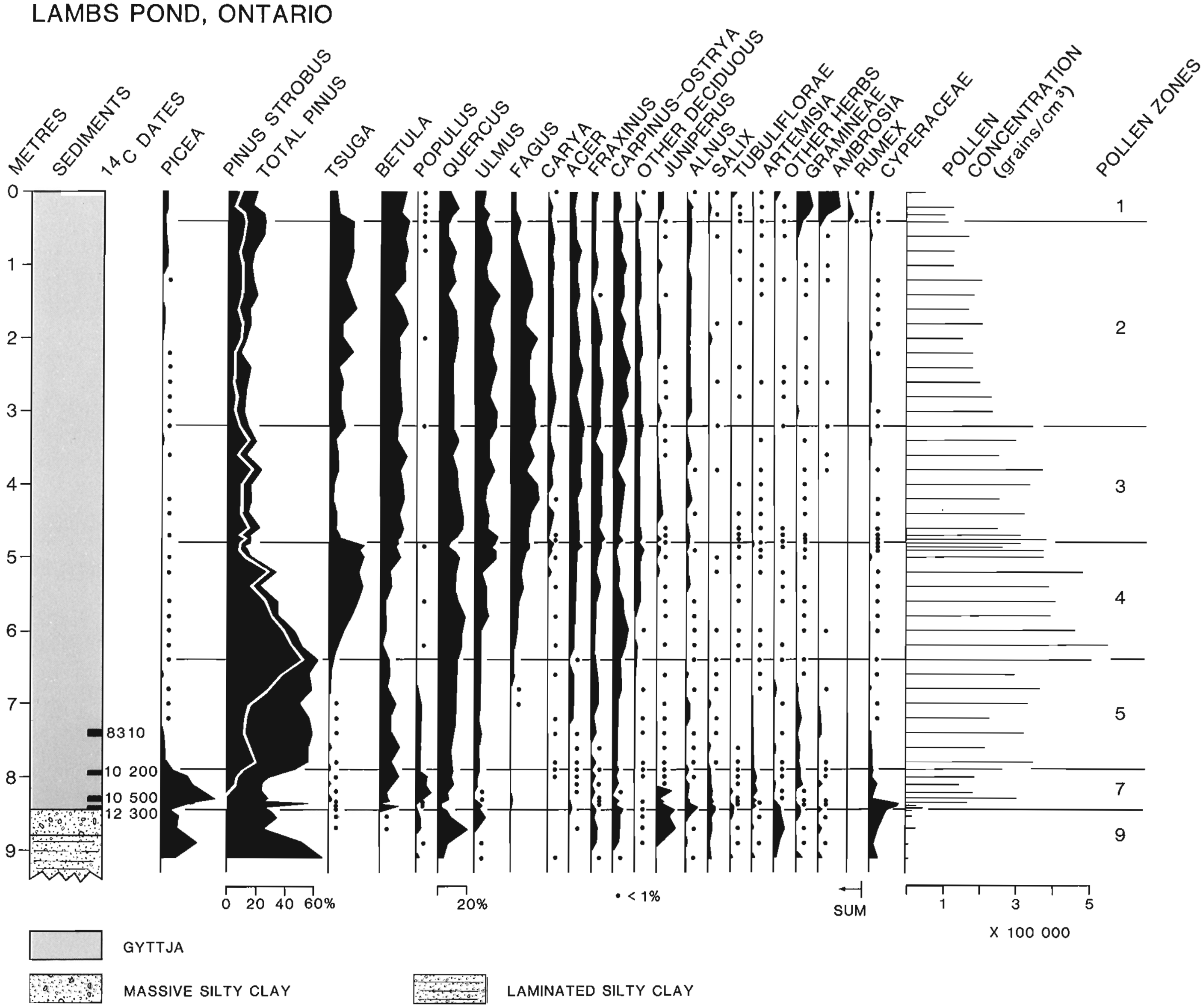


Figure 18. Abbreviated pollen diagram for Lambs Pond, Ontario.

The earliest pollen assemblage zone (9) is characterized by a shrub-herb component of **Salix, Artemisia**, Gramineae and Cyperaceae. In the southern part of the region the herb-shrub assemblages are succeeded first by a **Picea** or **Picea-Populus** zone (7) whereas in Ottawa Valley and areas to the north they give way exclusively to **Populus** (8). The progressive succession from **Picea** or **Picea-Populus** to **Pinus** in the south is replaced by one from **Populus** to **Picea** to **Betula** to **Pinus** in Ottawa Valley and sites northeast of St. Lawrence River. **Picea** becomes less prominent towards the northeast, especially at sites north of St. Lawrence River, for example at Lac à St-Germain (16, Fig. 16; Fig. 21) the **Populus** zone is succeeded by a zone dominated by **Betula** (6) which, in turn, gives way to **Pinus** (5). The **Pinus** zone is characterized initially (lower part) by **Pinus banksiana/resinosa** type pollen and higher in the zone by **Pinus strobus** type (see Lambs Pond diagram).

Zones 4, 3, and 2 are defined on the basis of two peaks in the **Tsuga** curve. The **Tsuga canadensis** zone (4) represents the first peak in **Tsuga**, the **Betula-Pinus strobus** zone (3) marks the **Tsuga** minimum, and the **Tsuga canadensis**-Mixed hardwoods zone (2) denotes the second peak in **Tsuga**. The uppermost zone, the **Ambrosia** zone (1), is represented by sharp increases in weed pollen such as **Ambrosia**, Gramineae and **Rumex**, and by decreases in tree pollen of **Pinus, strobus, Tsuga** and **Fagus**.

Vegetation History

Pollen evidence shows that the initial upland vegetation bordering the western Champlain Sea basin was herb-shrub tundra intermixed with woodlands of spruce, poplar, juniper, and shrub birch and alder. The high values of pine, oak, and other thermophilous tree pollen in zone 9 are attributed to long distance transport from sources to the south and deposition in a forest-tundra landscape. Tundra woodland vegetation prevailed in the southern uplands around the Champlain Sea from shortly after deglaciation to as late as 11.2 ka at Boyd Pond and grew to the north of the Champlain Sea and on certain islands in the sea to as late as 10 ka (Webb et al., 1983).

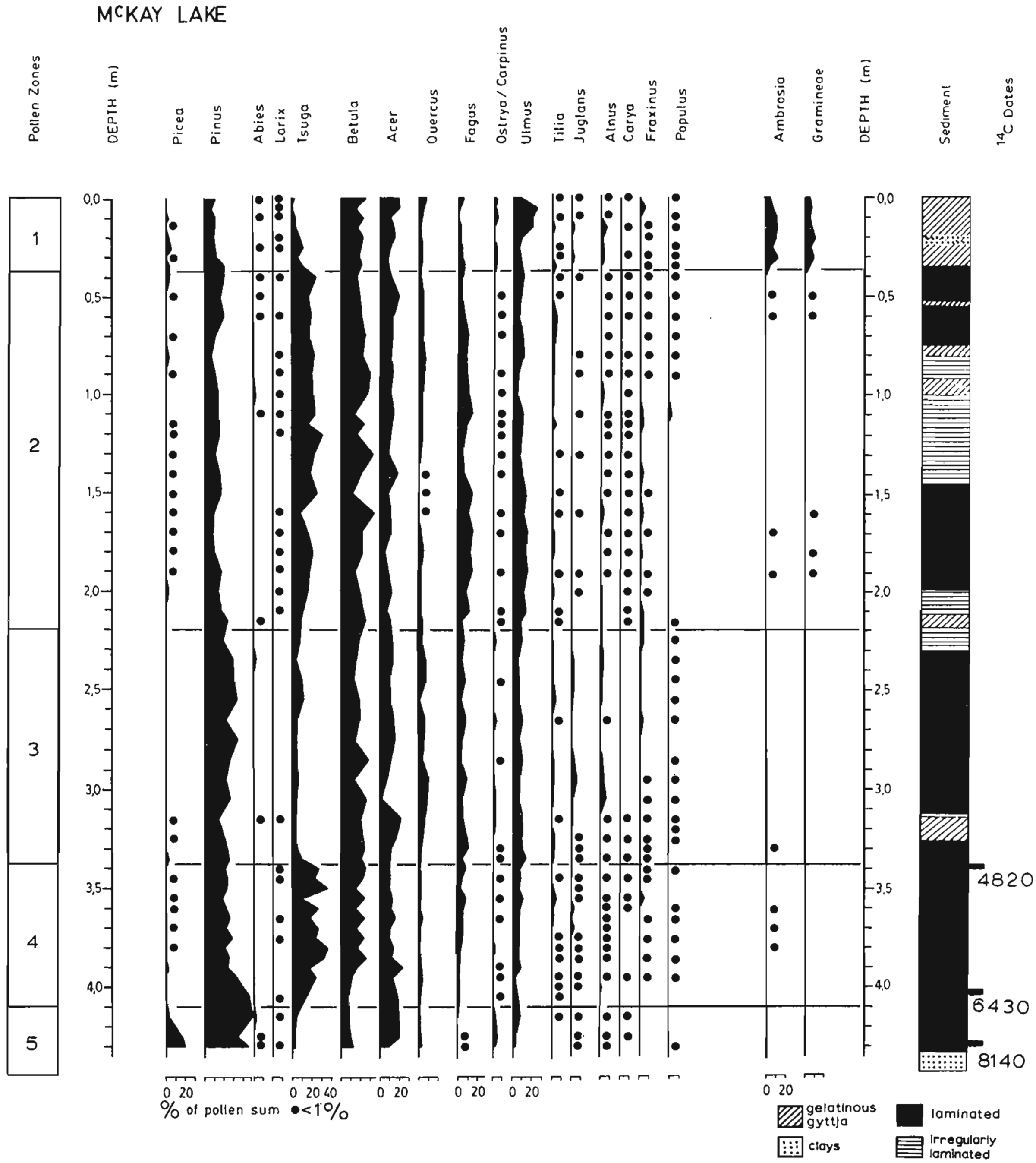


Figure 19. Abbreviated pollen diagram for McKay Lake, Ottawa, Ontario courtesy of R. McNeely, Geological Survey of Canada, and T. Oliver, Department of Biology, Queen's University.

Tundra-woodland gave way to spruce-poplar woodland in the southern uplands of the Champlain Sea and to poplar woodland on the recently deglaciated terrain bordering the Champlain Sea in Ottawa Valley and the southern Laurentians north of Montreal (Webb et al., 1983). At Ramsay Lake pollen influx increased two-fold across the shrub and herb-poplar boundary indicating a sudden increase in overall vegetation productivity associated with the movement of trees into the region.

By ca. 11 ka spruce had become abundant at localities to the south and southeast of the Champlain Sea. The spruce forest eventually moved northward and replaced poplar woodland in Ottawa Valley by ca. 10.2 ka and slightly later (ca. 10-9.5 ka) at sites in the southern Laurentians (Webb et al., 1983). Thus, by ca. 10 ka a spruce forest occupied the entire western basin of the Champlain Sea and surrounding uplands from Lake Ontario to the southern Laurentians.

Spruce forests dominated for about 1 ka until birch and pine migrated into the region from the south. Birch (mainly white birch) occupied upland sites in Ottawa Valley prior to pine. Jack Pine was widespread throughout the Ottawa Valley-Lake Ontario region apparently as early as ca. 9.4 ka but it was later replaced (after ca. 9 ka) by white pine. By ca. 8-7.5 ka the pine populations had shifted northward into the southern Laurentians (Webb et al., 1983).

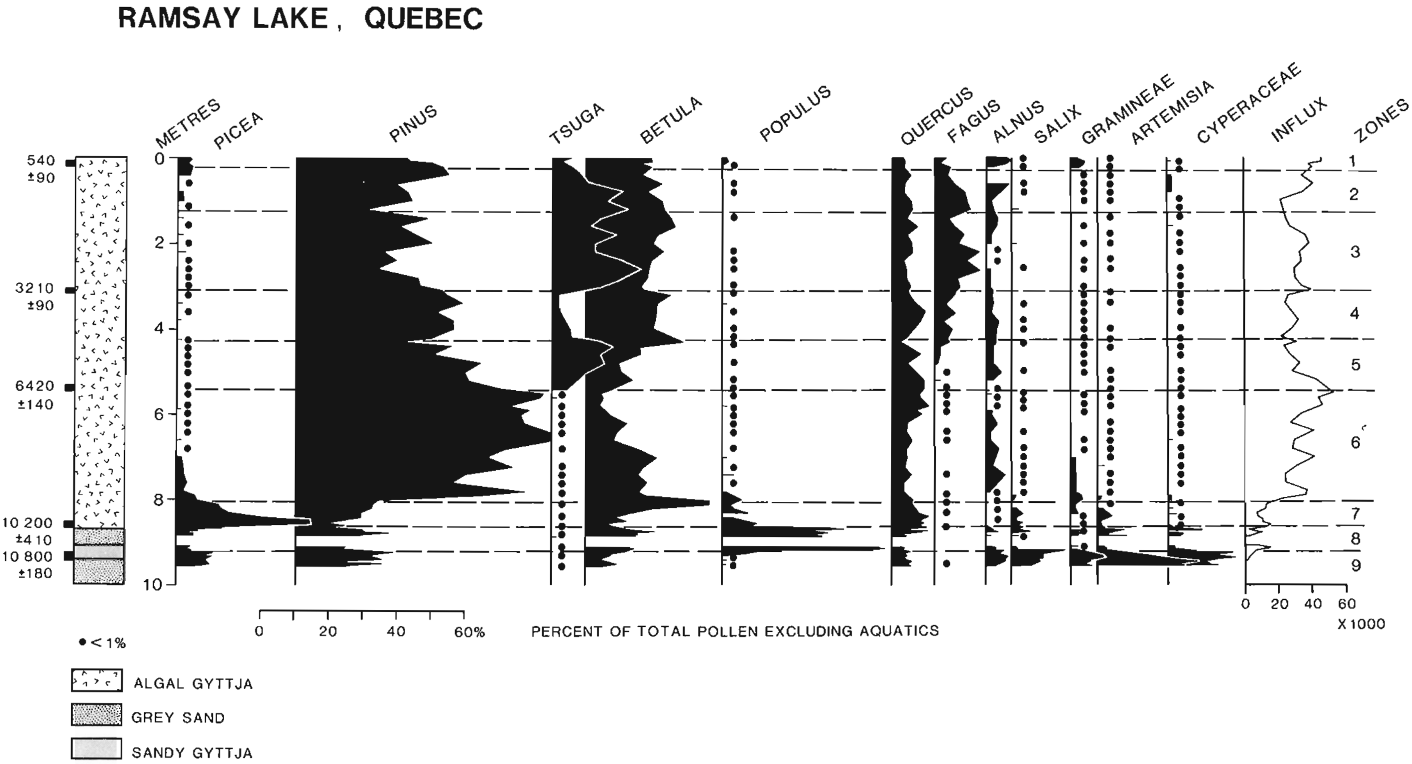


Figure 20. Abbreviated pollen diagram for Ramsay Lake, Quebec (modified from Mott and Farley-Gill, 1981).

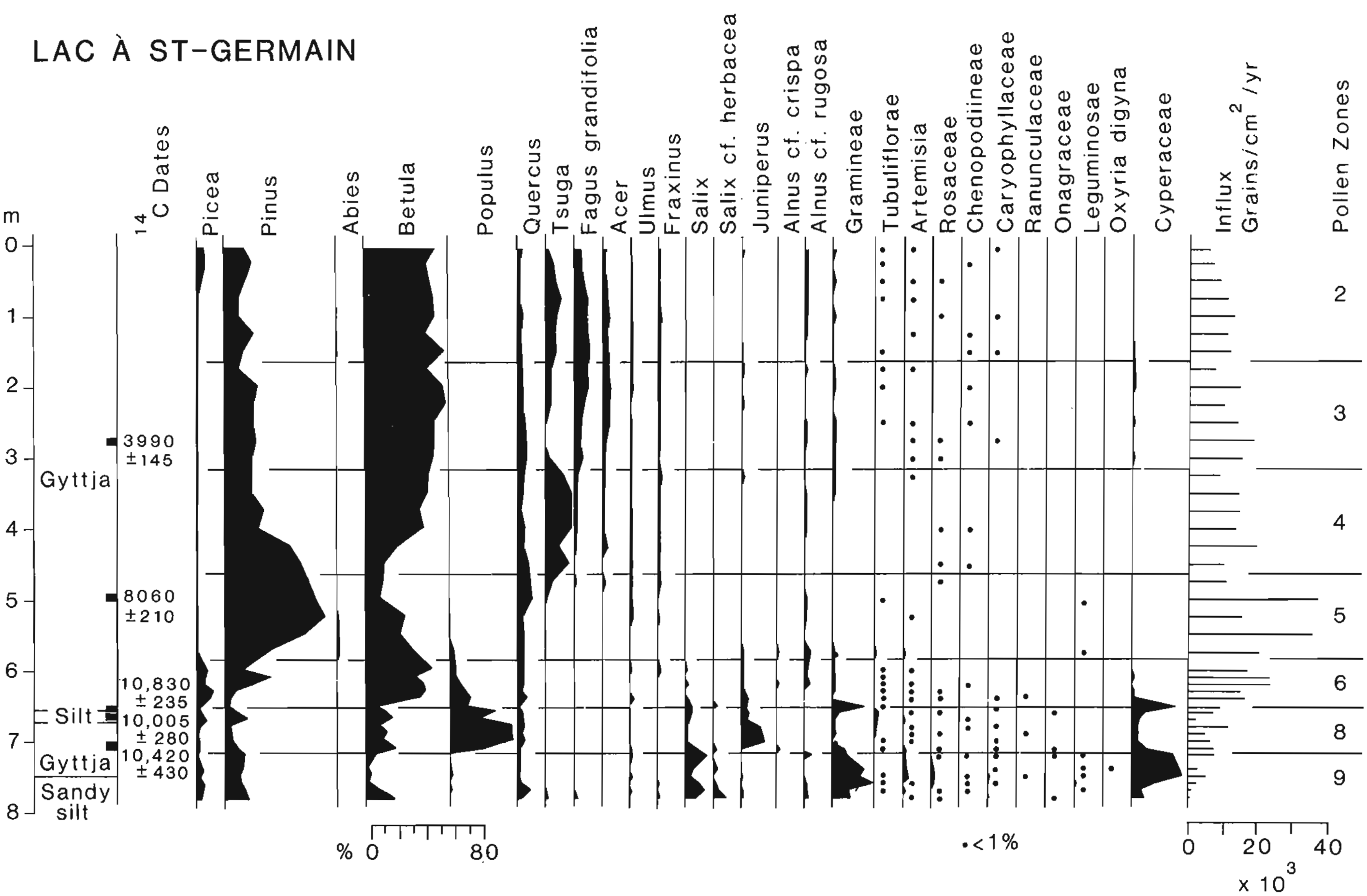


Figure 21. Abbreviated pollen diagram for Lac à St-Germain, Quebec (modified from Savoie and Richard, 1979).

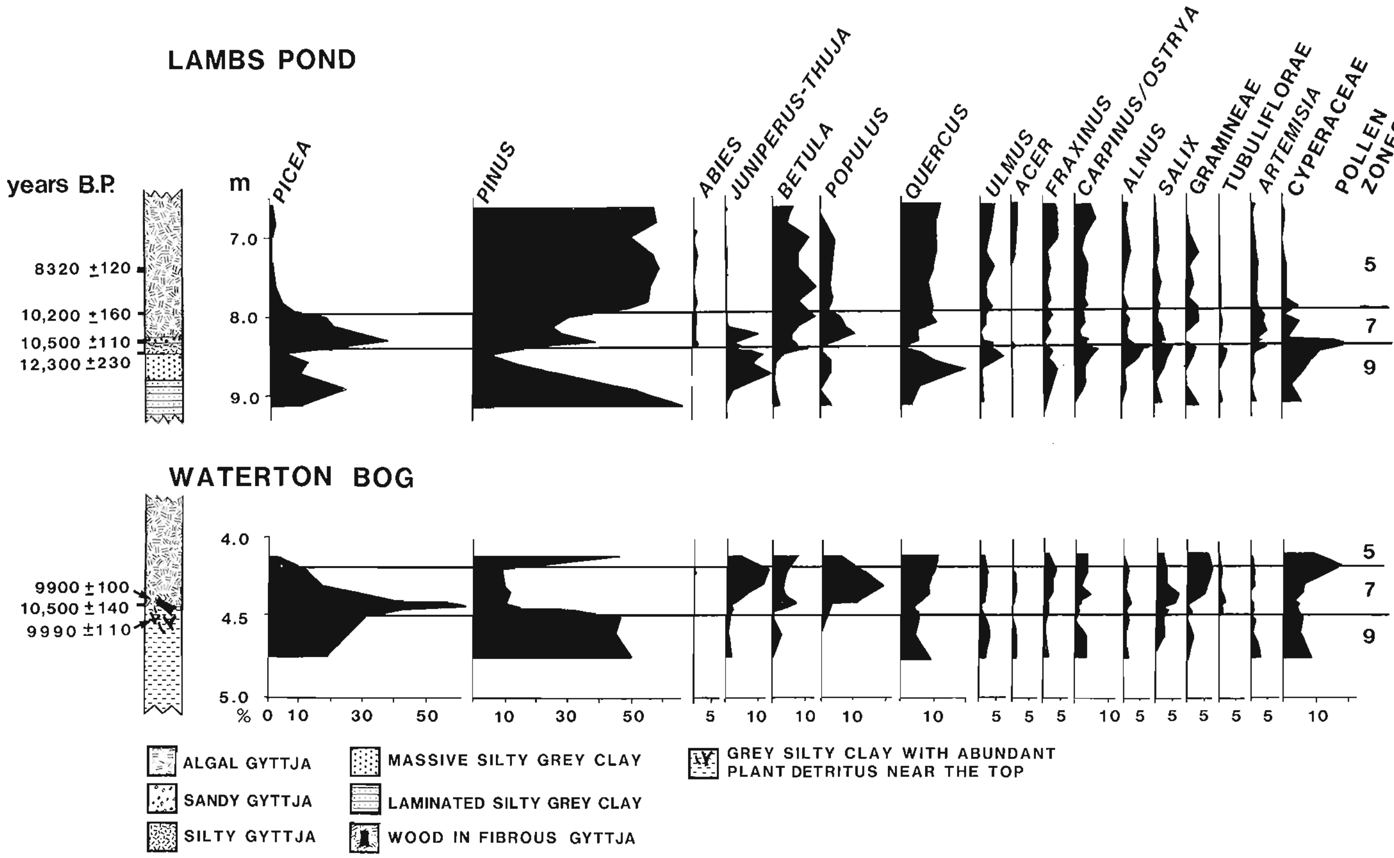


Figure 22. Abbreviated pollen diagram of basal sediments in Lambs Pond and Waterton Bog, Ontario.

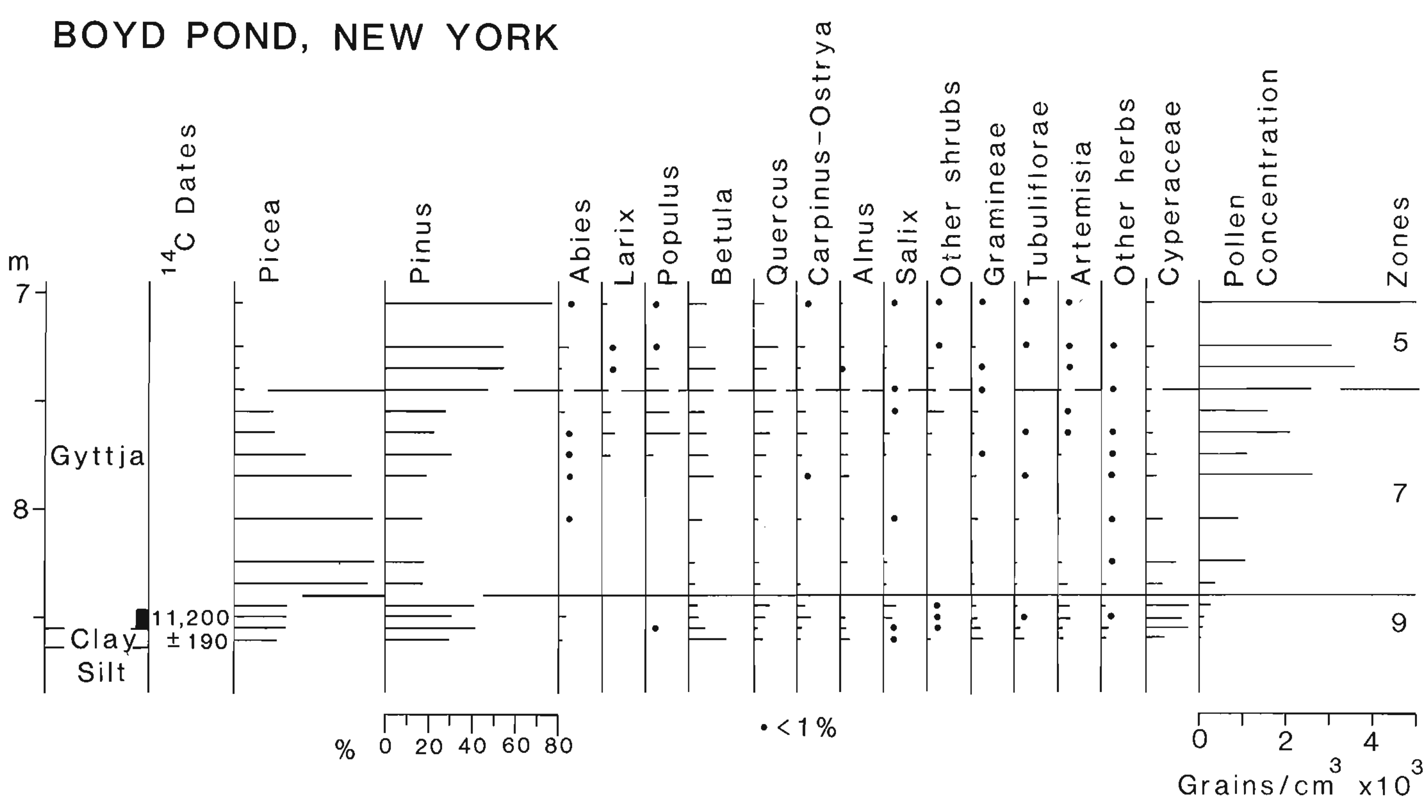


Figure 23. Abbreviated pollen diagram of basal sediments in Boyd Pond, New York State.

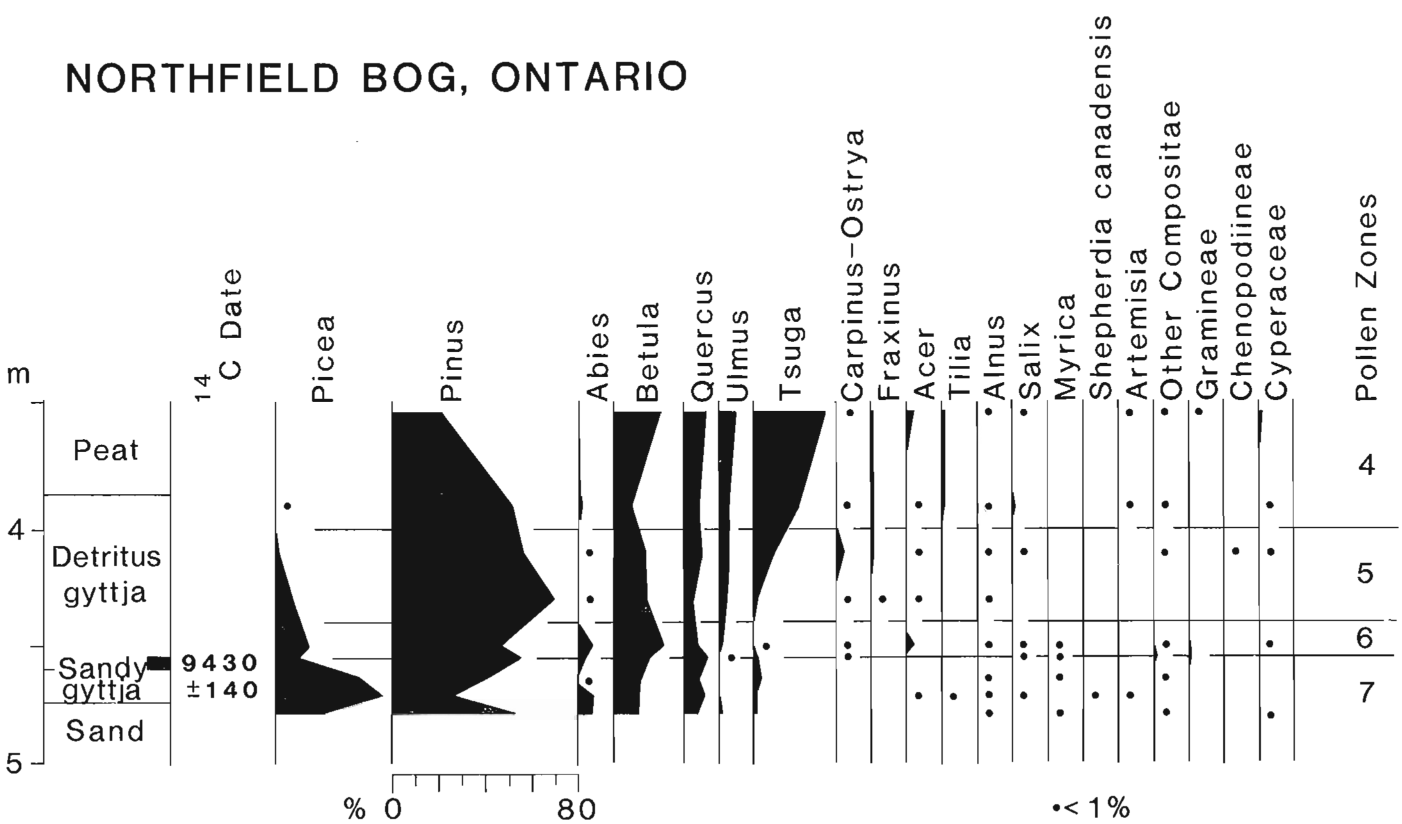


Figure 24. Abbreviated pollen diagram of basal sediments in Northfield Bog, Ontario (modified from Terasmae, 1965).

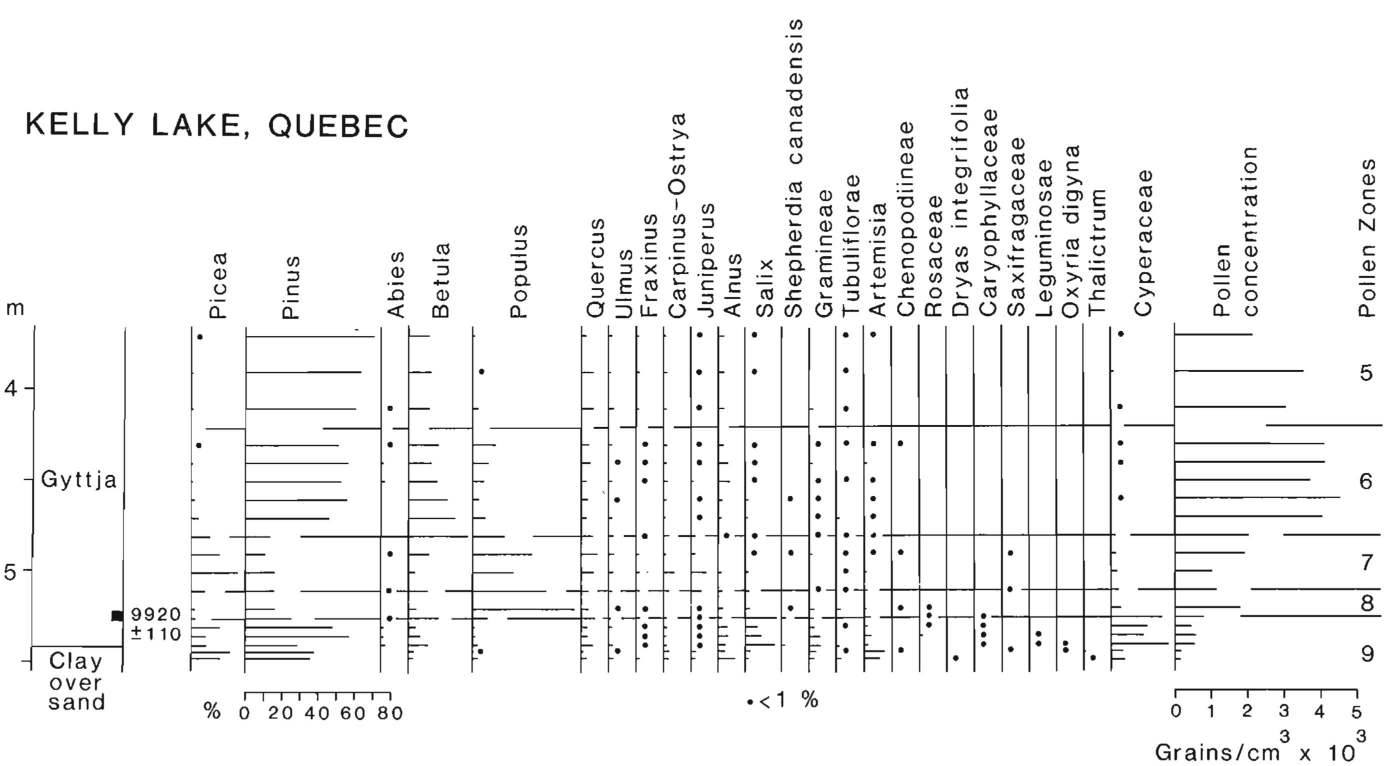


Figure 25. Abbreviated pollen diagram of basal sediments in Kelly Lake, Quebec.

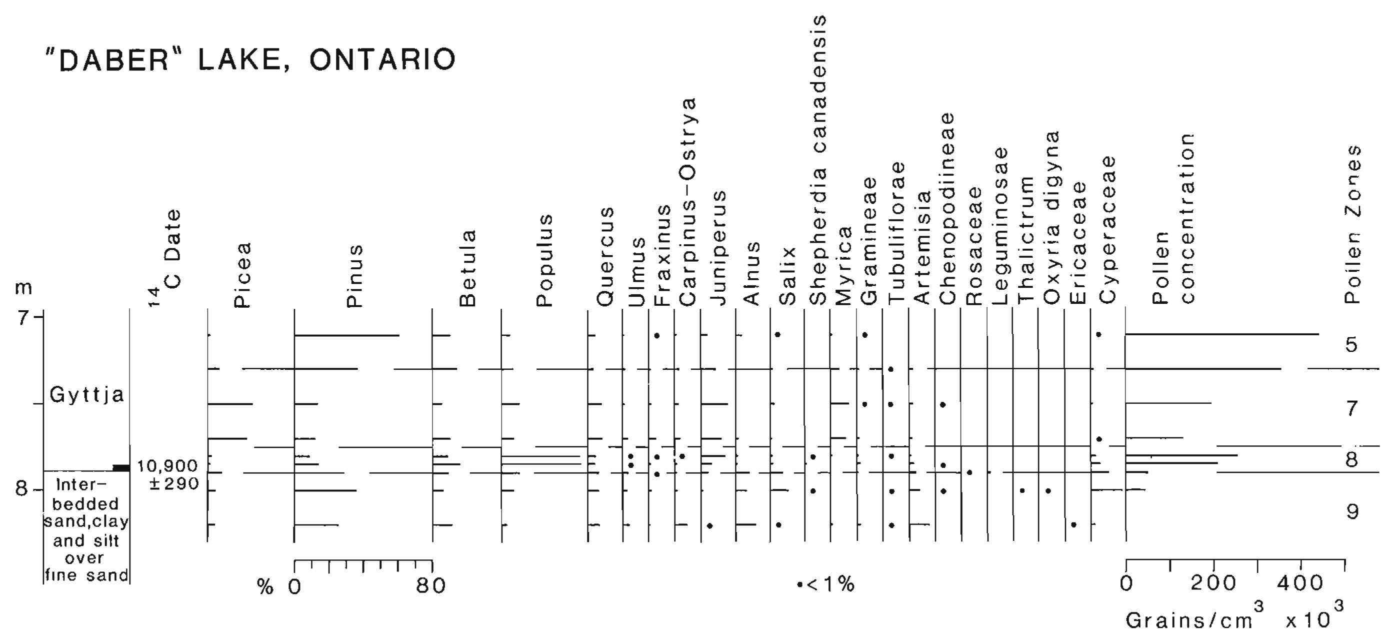
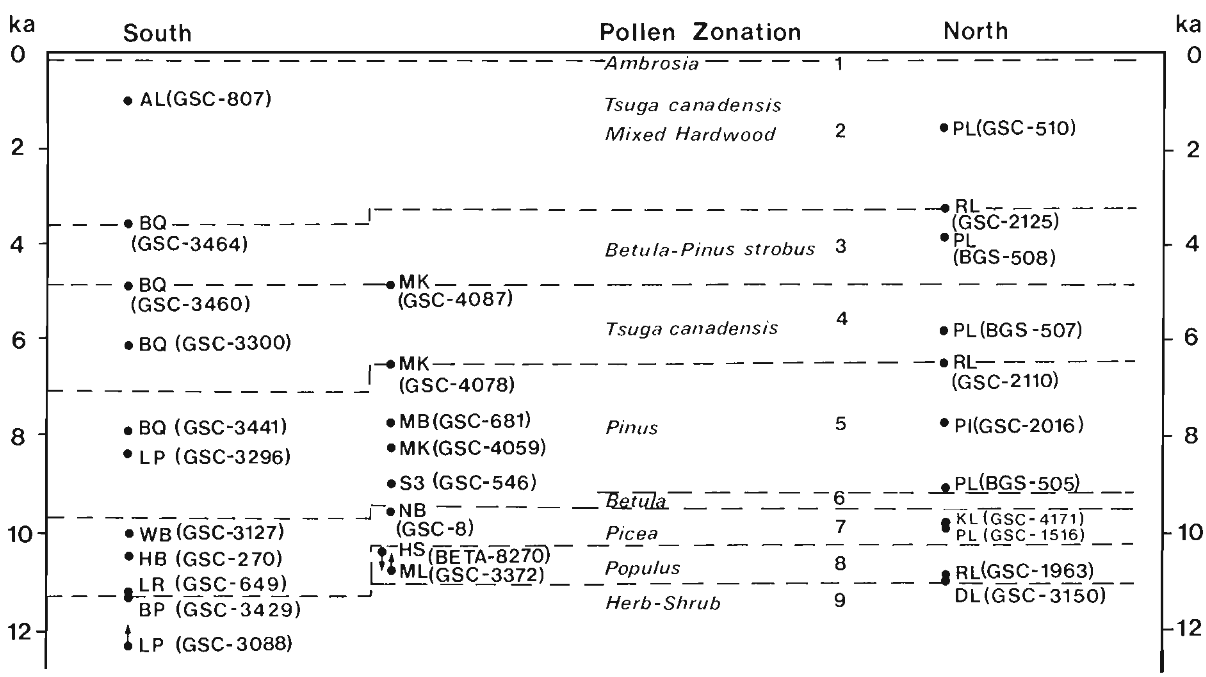


Figure 26. Abbreviated pollen diagram of basal sediments in "Daber" Lake, Ontario.



• Control date
Date considered too old
Date considered too young

Sources of control dates

AL Atkins Lake (Terasmae,1980)
BP Boyd Pond (unpublished)
BQ Bay of Quinte (Anderson and Lewis,1985)
DL "Daber" Lake (unpublished)
HB Harrowsmith Bog (Terasmae,1968)
HS Hinchinbrook Site (Delage et al.,1985)
KL Kelly Lake (unpublished)
LP Lambs Pond (unpublished)
LR Little Round Lake (Terasmae,1980)
MB Mer Bleue Bog (Camfield,1969)
MK McKay Lake (R.McNeely,pers.comm.,1985)
ML McLachlan Lake (unpublished)
PI Pink Lake (Mott and Farley-Gill,1981)
PL Perch Lake (Terasmae and McAtee,1979; Terasmae,1980)
RL Ramsay Lake (Mott and Farley-Gill,1981)
S3 Ottawa Site 3 (Mott and Camfield,1969)
WB Waterton Bog (unpublished)

Figure 27. Summary diagram of pollen stratigraphy and radiocarbon dates for lake and bog sites – Ottawa Valley-Lake Ontario region. Dates indicated are listed in Table 8.

The vegetation of the region changed dramatically between about ca. 8 ka and 6 ka when other tree taxa invaded and displaced pine. Hemlock, in particular, and maple and birch (most likely yellow birch) increased perceptibly as early as ca. 7.6 ka at Roblin Lake (Terasmae, 1980) southwest of the region and not until about ca. 6.4 ka in Ottawa Valley. Even though hemlock, white pine, birch, and oak may be unequally represented in the pollen profiles between ca. 7.5 and 4.8 ka, hemlock is considered to have been the more dominant tree species at this time. Based on its modern-day silvical characteristics (Hough, 1960), the shade tolerant hemlock became established under dense stands of white pine and slowly advanced to a dominant position in the forest stand, thus behaving as a climax species. At ca. 4.8 ka, however, the hemlock population was suddenly and drastically reduced, possibly as a result of the spread of a forest pathogen (Davis, 1981). Beech and maple populations migrated northwards into the region and probably occupied the openings left by hemlock. Shade intolerant hardwoods such as elm, ash, hickory, and basswood were also more prominent at this time. The inferred vegetation during the time of the hemlock pollen minimum (ca. 4.8 to 3.5 ka) probably resembled a mixed conifer-hardwood forest with white pine, white and yellow birch, beech, and maple the dominant taxa. Hemlock populations increased again as early as ca. 3.5 ka in the southern part of the region. The lower hemlock percentages of pollen zone 2, compared with those of pollen zone 4, suggest that hemlock did not regain its earlier Holocene dominance in the mixed forest. The inferred vegetation from ca. 3.5 ka to present consisted of a hemlock-white pine-mixed hardwoods association with hemlock, white pine, beech, and maple the dominant taxa.

AGE OF THE CHAMPLAIN SEA BASED ON POLLEN STRATIGRAPHY

Pollen analyses are being undertaken on Champlain Sea sediments and underlying glaciolacustrine clay at two sites in the Ottawa Valley-St. Lawrence River area for the purpose

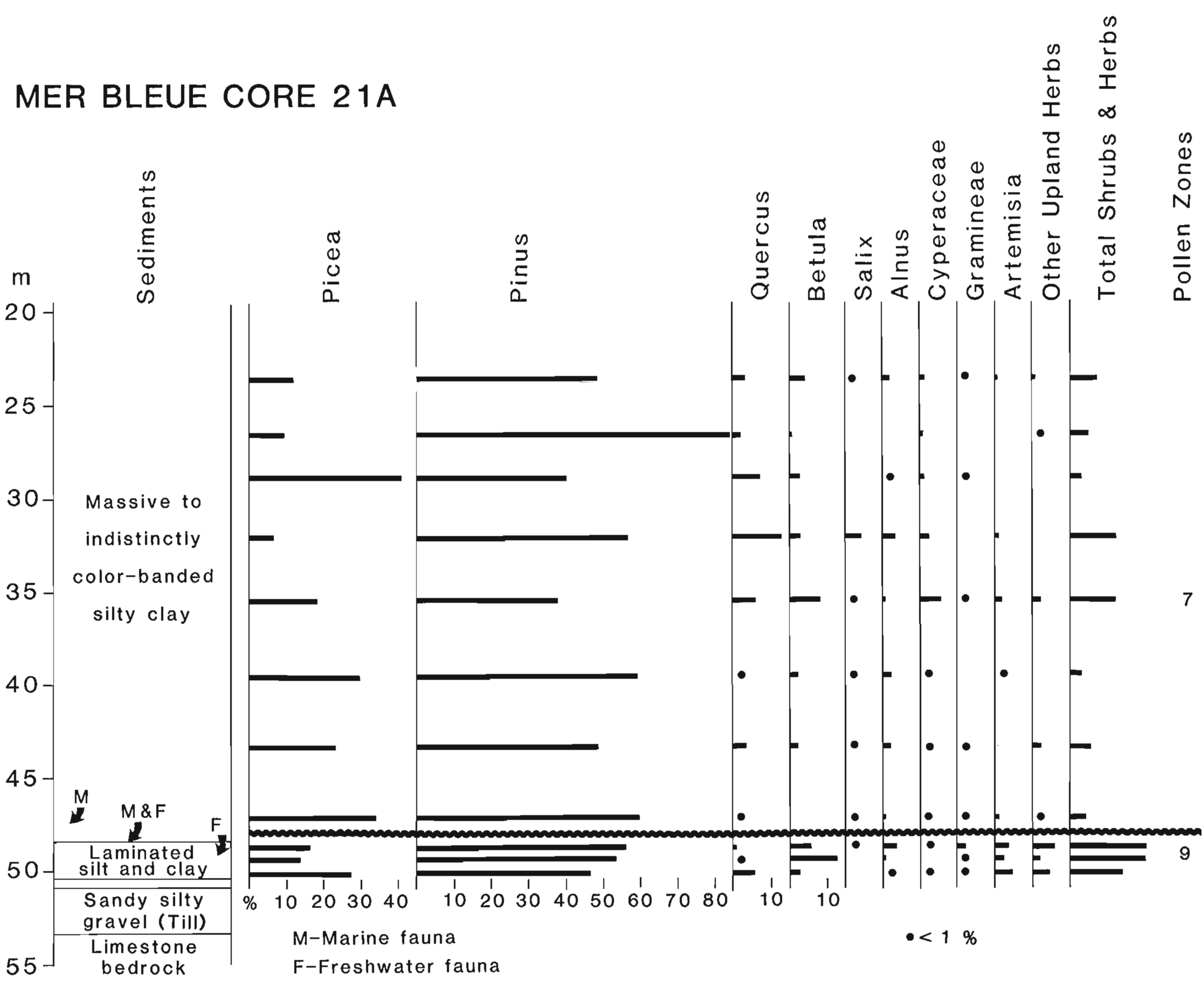


Figure 28. Abbreviated pollen diagram for Mer Bleue core (21A). Core depth is in metres below ground surface. F-freshwater ostracodes; M-marine foraminifers.

of using pollen stratigraphy as an independent method to estimate the age of the Champlain Sea as opposed to dating marine shells. The two sites are the Central Research Forest located adjacent to the west side of Mer Bleue, east of Ottawa (22, Fig. 16) and a section on St. Lawrence River near Sparrowhawk Point, located 17 km northeast of Ogdensburg, New York (6, Fig. 16).

Preliminary results from the Mer Bleue Central Research Forest core (21A) are reported here. The box containing this core had been erroneously labelled "LaRose Forest" and the core was discussed as such in Anderson et al. (1985). However, archived documents show that the core was obtained from the Central Research Forest, Mer Bleue Bog area by the Testing Laboratories, Department of Public Works in 1969 as requested by M.J.J. Bik formerly of the Geological Survey of Canada. The core was extruded and described by Gadd (1986b). Briefly, 48.26 m of massive to indistinctly colour-banded light and dark grey, blue-grey, and reddish brown to brownish grey clay of the Champlain Sea overlies 2.22 m of laminated (varved?) grey clay and silt, which overlies 2.9 m of silty, sandy gravel (till). Limestone bedrock was reached at 53.38 m below surface. The contact between the marine sediments of Champlain Sea and underlying freshwater clay is based on upward changes from freshwater ostracodes to marine ostracodes and foraminifera at 48.35 m depth in the core (Anderson et al., 1985).

Pollen spectra from the 20 to 50 m interval of Mer Bleue core (21A) fall into two assemblage zones (Fig. 28). The lowermost zone occurs between about 47 and 50 m and is differentiated on the basis of low **Picea** (10 to 20%), high **Pinus** (50 to 60%) and **Betula** (5 to 10%), and up to 20% shrubs and herbs notably **Betula**, **Artemisia**, Compositae and Gramineae. An upper zone is recognized between about 28 and 47 m on the basis of maximum values in **Picea** (30-40%), lower but fluctuating values in **Pinus** and lower percentages (up to 12%) of shrub and herb pollen than in the previous zone.

The pollen trends compare closely with those in pollen zones 9 and 7 at Lambs Pond and Boyd Pond (4 and 5, Fig. 16) which bordered the Champlain Sea in the south. The **Picea** counts from near the top of the herb-shrub zones at Lambs Pond (Fig. 18) and Boyd Pond (Fig. 23) decrease to minimum trends like those in the Mer Bleue core. **Pinus** is high at both sites but it fluctuates more in Boyd Pond in a manner similar to that at Mer Bleue. The decline in herb and shrub pollen in the basal marine sediments of the Mer Bleue core

Table 8. Radiocarbon dates of pollen stratigraphic sites in the Ottawa Valley-Lake Ontario Region

	Site and No. (cf. Fig. 16)	Latitude	Longitude	^{14}C Age BP Uncorrected	^{14}C Age BP Corrected (based on $\delta^{13}C$)	Laboratory Dating No.	Elevation ^{14}C Sample (m a.s.l.)	Sample Interval (cm)	Reference
		(to the nearest minute)							
1	Bay of Quinte	44°02'	77°05'	3610 ± 80	3550 ± 80	GSC-3464	60.9	215-220	Anderson and Lewis, 1985
				4850 ± 80	4790 ± 80	GSC-3460	59.9	320-325	Anderson and Lewis, 1985
				6140 ± 100	6080 ± 100	GSC-3300	59.3	378-382	Anderson and Lewis, 1985
				7960 ± 120	7920 ± 120	GSC-3441	59.1	393-397	Anderson and Lewis, 1985
2	Harrowsmith Bog	44°25'	76°42'	10 390 ± 160		GSC-270	152.8	560-570	Terasmae, 1968
3	Waterton Bog	44°25'	75°58'	9930 ± 100	9900 ± 100	GSC-3127	83.8	440-443	This study
				10 600 ± 140	10 500 ± 140	GSC-3146	83.8	441-443	This study
				10 000 ± 110	9990 ± 110	GSC-3163	83.7	448-460	This study
4	Lambs Pond	44°39'	75°48'	8410 ± 120	8320 ± 120	GSC-3296	106.2	737-743	This study
				10 300 ± 160	10 200 ± 160	GSC-3259	105.6	793-796	This study
				10 600 ± 110	10 500 ± 110	GSC-3273	105.3	828-831	This study
				12 300 ± 230	12 300 ± 230	GSC-3088	105.2	841-843	This study
5	Boyd Pond	44°23'	75°05'	11 200 ± 190	11 200 ± 190	GSC-3429	250.0	847-856	This study
7	Atkins Lake	44°45'	75°51'	1020 ± 160		GSC-807	114.7	0.69-0.82	Terasmae, 1980
				11 100 ± 270		GSC-762	111.6	376-388	Terasmae, 1980
8	Little Round Lake	44°48'	76°42'	11 180 ± 180		GSC-649	186.5	322-327	Terasmae, 1980
9	Perth Bog	44°57'	76°16'	8790 ± 100	8770 ± 100	GSC-3298	138.6	467-472	This study
10	McLachlan Lake	45°22'	76°33'	10 700 ± 150	10 700 ± 150	GSC-3372	167.7	1100-1104	This study
11	Kelly Lake	45°40'	76°39'	9920 ± 110		GSC-4171	185.3	523-528	This study
12	"Daber" Lake	45°45'	77°17'	10 900 ± 290	10 900 ± 290	GSC-3150	148.2	785-788	This study
13	Perch Lake	46°02'	77°32'	1340 ± 100		BGS-510	151.5	100-110	Terasmae and McAtee, 1979
				2510 ± 130		BGS-509	150.5	200-210	Terasmae and McAtee, 1979
				3790 ± 120		BGS-508	149.5	300-310	Terasmae and McAtee, 1979
				5780 ± 160		BGS-507	148.5	400-410	Terasmae and McAtee, 1979
				7800 ± 300		BGS-506	147.5	500-510	Terasmae and McAtee, 1979
				9030 ± 220		BGS-505	147.0	550-560	Terasmae and McAtee, 1979
				9830 ± 250		GSC-1516	146.5	595-605	Terasmae, 1980
14	"Flatbottom" Lake	46°03'	77°16'	9490 ± 160	9460 ± 160	GSC-3659	315.4	501-509	This study
15	Kazabazua Bog	45°57'	76°04'	9910 ± 200		GSC-680	176.0	895-905	Terasmae, 1980
16	Lac à St. Germain	45°56'	74°22'	3990 ± 145		GX-5229	464.0	270-280	Savoie and Richard, 1979
				8060 ± 210		GX-5230	461.8	490-500	Savoie and Richard, 1979
				10 830 ± 235		GX-5231	460.3	647-657	Savoie and Richard, 1979
				10 005 ± 280		GX-5232	460.2	658-668	Savoie and Richard, 1979
				10 420 ± 430		GX-5233	459.2	700-712	Savoie and Richard, 1979
17	Ramsay Lake	45°36'	76°06'	3320 ± 90	3210 ± 90	GSC-2125	185.9	311-316	Mott and Farley-Gill, 1981
				6510 ± 140	6420 ± 140	GSC-2110	183.6	542-547	Mott and Farley-Gill, 1981
				10 300 ± 410	10 200 ± 410	GSC-2122	180.4	860-864	Mott and Farley-Gill, 1981
				10 900 ± 180	10 800 ± 180	GSC-1963	179.6	930-942	Mott and Farley-Gill, 1981
18	Pink Lake	45°28'	75°49'	3410 ± 260	3270 ± 260	GSC-2014	141.3	132-136	Mott and Farley-Gill, 1981
				7920 ± 170	7750 ± 170	GSC-2016	140.0	222-227	Mott and Farley-Gill, 1981
				10 600 ± 150	10 600 ± 150	GSC-1956	139.0	295-300	Mott and Farley-Gill, 1981
19	Site 6 - Ottawa	45°21'	75°48'	8220 ± 150		GSC-547	63.3	500-550	Mott and Camfield, 1969
20	Site 3 - Ottawa	45°24'	75°42'	8830 ± 190		GSC-546	57.2	380-390	Mott and Camfield, 1969
20	Site 4 - Ottawa	45°25'	75°42'	7870 ± 160		GSC-628	61.7	680-695	Mott and Camfield, 1969
21	McKay Lake	45°27'	75°18'	4910 ± 80	4820 ± 80	GSC-4087	32.0	343-347	R.N. McNeely, pers. comm.
				6510 ± 80	6430 ± 80	GSC-4078	31.4	401-403	R.N. McNeely, pers. comm.
				8260 ± 100	8140 ± 100	GSC-4059	31.1	429-430	R.N. McNeely, pers. comm.
23	Mer Bleue Bog	45°24'	75°30'	7650 ± 210		GSC-681	63.4	515-525	Mott and Camfield, 1969
25	Northfield Bog	45°08'	74°56'	9430 ± 140		GSC-8	94.2	475-485	Terasmae and Mott, 1959
26	Hinchinbrook Site	45°01'	74°04'	10 480 ± 140		Beta-8270	110.0	78-80	Delage et al., 1985

GSC - Geological Survey of Canada
BGS - Brock University
GX - Geochron Laboratories
Beta - Beta Analytic Inc.

is comparable to that across the shrub and herb-**Picea** boundary in Boyd and Lambs ponds. Accompanying the decrease in herb and shrub pollen is the sudden increase in **Picea** which remains relatively high in the marine sediments.

The top of the herb-shrub zones at Boyd and Lambs ponds coincides with the end of glaciolacustrine deposition dated 11 200 ± 190 BP (GSC-3429, Table 8) at Boyd Pond. Thus, on the basis of palynological zonation, the change from proglacial lake sedimentation to marine deposition in the Champlain Sea at the Mer Bleue site is placed between 11 and 11.5 ka.

The retreat of marine waters from the western basin of the Champlain Sea was followed closely by the migration of spruce-dominated vegetation into Ottawa Valley. The spruce pollen peak at the base of Northfield Bog (Fig. 24) begins in eolian sand which postdates the Champlain Sea (Terasmae and Mott, 1959). The base of the spruce peak occurs at a level dated 10 200 ± 410 BP (GSC-2122, Table 8 and Fig. 20) at Ramsay Lake. The spruce pollen rise thus provides a convenient marker horizon to date the recession phase of the Champlain Sea and subsequent dune development in eastern Ottawa Valley.

DISCUSSION

The late glacial pollen stratigraphy and hence implied vegetation is time transgressive in a north-south transect across the western Champlain Sea basin. After ca. 9.5 ka, pollen zonations are everywhere synchronous for the first time.

The late glacial variations in pollen stratigraphy may be explained partly on the basis of the regional patterns of ice retreat, regional changes in late glacial climate, and the effect of the Champlain Sea on the composition of the late glacial vegetation, and on vegetation migration. These considerations bear on the use of pollen stratigraphy as an independent means of dating the Champlain Sea and are discussed below.

Deglaciation of the Ottawa Valley-Lake Ontario region

Late Wisconsinan ice recession in upper St. Lawrence Valley allowed glacial Lake Iroquois and the post-Iroquois glacial lakes of the Ontario basin to extend northeastward into the Central St. Lawrence Lowland and to drain to lower lake levels (Prest, 1970). The Ontario basin extension existed for a time in contact with the north and west retreating ice margin. Dates on basal gyttja in small lake and bog basins, located outside the Champlain Sea and at or beyond the limits of the glacial lakes, provide minimum ages for deglaciation of the Ottawa Valley-Lake Ontario uplands. The basal date of 11 200 ± 190 BP (GSC-3429, Table 8) at Boyd Pond provides a minimum estimate for ice retreat from the Adirondack Highlands south of the Champlain Sea. Basal gyttja dates from the Madawaska Highland west of the Champlain Sea are 10 700 ± 150 BP (GSC-3372) at McLachlan Lake (10, Fig. 16) and 10 900 ± 290 BP (GSC-3150) at "Daber" Lake (12, Fig. 16). Dates from the uplands north of the Champlain Sea are 10 420 ± 430 BP (Gx-5233) at Lac à St-Germain (16, Fig. 16), 10.2 ka at Ste. Agathe, 10 ka at Tania (Webb et al., 1983), 10 800 ± 180 BP (GSC-1936) at Ramsay Lake (17, Fig. 16), 9910 ± 200 BP (GSC-680) at Kazabazua (15, Fig. 16) and 9460 ± 160 BP (GSC-3659) at "Flatbottom" Lake (14, Fig. 16). The range of dates from ca. 11.2 ka in the southeast to 9.5 ka in the north and northwest confirms a northwesterly retreat of ice from the western Champlain Sea basin and surrounding Madawaska and Laurentian highlands.

According to Terasmae (1980) the ice retreated north of Ottawa before ca. 11 ka and west of Pembroke ca. 10 ka. The basal dates of 9.9 ka at Kazabazua and 9.5 ka at "Flatbottom" Lake indicate therefore that a lapse time of about 1 ka may have occurred between deglaciation and initiation of accumulation of organic sediment deposition in these lake basins. However, 1 ka seems more than adequate for deposition of the stratified silt and clay underlying gyttja at "Flatbottom" Lake.

Influence of the Champlain Sea and climate on the late glacial vegetation

Tundra woodland existed in contact with the Champlain Sea in the southern part of the basin until ca. 11.2 ka and formed the initial vegetation of the newly deglaciated areas to the north, west, and northwest of the Champlain Sea. It was replaced by spruce-poplar woodlands in the southern part of the basin and by poplar woodlands to the west and north of the sea. Poplar grew at the northwest margin of the Champlain Sea at ca. 10.9 ka, to the north of the sea at ca. 10.7 ka, and slightly later (ca. 10.0 ka) at sites in Laurentian Highland areas farther east. Poplar was able to reach areas to the north and west of the Champlain Sea and islands within the Champlain Sea prior to other trees because its seeds were more easily transported by wind currents and in the water of the Champlain Sea than those of other tree species including spruce (Richard, 1977; Mott, 1978; Webb et al., 1983). In addition, poplar seeds, especially those of **Populus balsamifera, P. grandidentata, P. deltoides** and **P. tremuloides**, prefer wet, bare mineral soil seedbeds that have abundant sunlight (Fowells, 1965). Such conditions prevailed in the recently deglaciated sites around the sea and poplar was quick to colonize these habitats.

By ca. 11 ka spruce dominated the uplands bordering the south shore of the Champlain Sea. The earliest arrival time for spruce north of the Champlain Sea was ca. 10.2 ka at Ramsay Lake. In contrast to poplar, spruce experienced a migration lag of about 1.5 ka in reaching the north shore of the Champlain Sea. The migrational lag may be attributed to the presence of the Champlain Sea, which represented a physiographic barrier preventing spruce seeds from dispersing to the north and northwest. Once the sea barrier had decreased in areal extent, spruce advanced northward and eventually occupied the entire western basin of the Champlain Sea. The peak in **Picea** pollen above Champlain Sea sediments at Northfield Bog (Terasmae, 1965; Terasmae in Dyck and Fyles, 1963) shows that spruce formed the pioneer forest of the Champlain Sea basin.

Alternatively, the lag in spruce migration may have been climatically induced. Convincing evidence now exists at several pollen-analyzed sites in Nova Scotia and New Brunswick (Mott, 1985) and at two areas in Newfoundland (Brown Macpherson and Anderson, 1985) for a climatic oscillation which interrupted deglacial warming between about ca. 11 and 10 ka. It may be more than coincidental that the timing of the spruce lag corresponds closely with that for the climatic oscillation in eastern Canada and possibly with the climatically induced St. Narcisse event in Quebec (LaSalle and Elson, 1975) bracketed between ca. 11 and 10.3 ka (LaSalle, 1984). The spruce lag resulted from the presence of the Champlain Sea perhaps in combination with regional climatic cooling even though evidence for climatic cooling at this time is not clear in pollen diagrams from the Ottawa Valley-Lake Ontario region.

Correlation of Champlain Sea chronologies and implications

The palynological findings reported here are in agreement with a previous correlation between the Champlain Sea and the post-Iroquois-Early Lake Ontario transition in the Lake Ontario basin (Anderson and Lewis, 1985). The Champlain Sea may have been confluent with water levels in the Ontario basin after the fall of the post-Iroquois lake phases to the low water phase of Early Lake Ontario (Anderson and Lewis, 1985). The low water phase is palynologically dated at ca. 11.4 ka in shallow water deposits from western Lake Ontario (Anderson and Lewis, 1985). Since Lake Ontario may have been confluent with sea level just prior to or at its lowest level, i.e. prior to or during early isostatic uplift of the controlling Duck-Galloo sill in eastern Lake Ontario (Anderson and Lewis, 1985), the 11.4 ka estimate is therefore a minimum age for the Champlain Sea at its maximum. Ontario basin and Champlain Sea waters separated as a result of isostatic uplift of the upper St. Lawrence outlet area.

The oldest dates on the Champlain Sea range from ca. 12.1 to 12.7 ka for marine shells collected at or near marine limit in the western basin of the Champlain Sea (Fulton and Richard, this publication). The shell dates differ by as much as 1.7 ka from estimates of ca. 11.0 to 11.5 ka for the beginning of the Champlain Sea as determined from pollen stratigraphy. Younger shell dates, however, overlap with age estimates derived from pollen stratigraphy. For example, one of the youngest marine shell dates on the recession phase of the Champlain Sea (10 000 ± 320 BP, GSC-1553; Table 7; Fulton and Richard, this publication) corresponds closely with the estimated 10.2 ka age for the post-Champlain Sea spruce arrival in Ottawa Valley.

The discrepancy between the older shell dates and the pollen-dated chronologies might be attributed to the possibility that the oldest marine shells incorporated carbon from carbonate-charged glacial meltwater that flowed into the western basin of the Champlain Sea (Hillaire-Marcel, 1981). With time, the carbonate-rich glacial meltwater became diluted by the marine water of the Champlain Sea and eventually reached equilibrium with atmospheric CO_2. Younger dated shells were therefore less affected.

Basal gyttja from McLachlan Lake, located in the same general area as the oldest marine shells, also appears to have been affected by hardwater dating error. The gyttja yielded a date of 10 700 ± 150 BP (GSC-3372) which is about 1.3 ka older than the date anticipated from pollen correlation with other sites in the area. In a similar manner basal gyttja from Lambs Pond is too old by about 1000 years.

Thus a chronological framework for the western basin of the Champlain Sea based on marine shell dates greater than about 10 ka and dated pollen stratigraphy in carbonate-rich areas should be regarded with suspicion. Pollen studies on Champlain Sea sediments correlated with ^{14}C dated pollen zones from the Adirondack Highlands where Proterozoic igneous and metamorphic rocks predominate (Isachsen and Fisher, 1970) perhaps offer the most reliable means to date the Champlain Sea incursion. Additional lake sediment coring and basal sediment dating in the Adirondack Highlands are recommended in order to establish better chronological control for the late glacial interval of the Lake Ontario-Ottawa Valley region.

SUMMARY AND CONCLUSIONS

Following deglaciation, herb-shrub tundra-woodland represented the earliest colonizing vegetation of the Ottawa Valley-Lake Ontario region. In the south, spruce began arriving about ca. 11.2 ka and dominated a spruce-poplar woodland ca. 11 ka. Poplar woodlands replaced tundra woodland in areas bordering the Champlain Sea to the west and north between ca. 10.9 to 10.7 ka. Spruce did not migrate northward to replace poplar until ca. 10.2 ka.

Invasion of the Champlain Sea appears to correspond approximately with movement of spruce into the area (ca. 11.2 ka). The end of the Champlain Sea correlates closely with the replacement of poplar by spruce on the north shore of the basin (ca. 10.2 ka).

The oldest shell dates on the Champlain Sea (12.7 to 12.1 ka) are too old by up to 1.7 ka on the basis of the pollen stratigraphy presented here and correlation between Champlain Sea and the post-Iroquois water-level history of Lake Ontario basin. Younger marine shell dates overlap with those derived from pollen correlations. Pollen studies on Champlain Sea sediments and underlying glaciolacustrine deposits correlated with dated pollen profiles from the Adirondack Highlands probably represent the most reliable potential method for dating the Champlain Sea.

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