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Late Wisconsinan Deglaciation and Champlain Sea Invasion in the St. Lawrence Valley, Québec
Le retrait glaciaire et l'invasion de la Mer de Champlain à la fin du Wisconsinien dans la vallée du Saint-Laurent, Québec
Enteisung im späten Wisconsin und der Einbruch des Meeres von Champlain in das Tal des Sankt-Lorenz-Stroms, Québec

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Article abstract

Champlain Sea history is directly linked to Late Wisconsinan deglacial episodes. Champlain Sea Phase I (Charlesbourg Phase) began in the Québec area at about 12.4 ka. It represented a western extension of the Goldthwait Sea between remnant Appalachian ice masses and the Laurentide Ice Sheet. Further south, at about the same time, in the Appalachian uplands and piedmont, high-level glacial lakes were impounded by the ice-front during glacial retreat toward NNW: lakes Vermont, Memphrémagog and Mégantic. Lowlands of the Upper St. Lawrence and Lake Champlain valleys were progressively deglaciated and inundated by Lake Iroquois and Lake Vermont. At about 12.1 ka, these two lakes coalesced and formed a single water-body, here referred to as Lake Candona. After the Ulverton-Tingwick Moraine was constructed, this lake extended northeastward onto the Appalachian piedmont where varved sediments containing Candona subtriangulata underlie marine clays. Current data and interpretations bring into question the former concept of the Highland Front Moraine System. The invasion of the main basin, or Champlain Sea Phase II, began around 12 ka. Replacement of Lake Candona by the sea resulted in a fall of about 60 m in water levels. Champlain Sea Phase III began at the end of the Saint-Narcisse episode, at about 10.8 ka. At this time marine waters were able to enter valleys of the Laurentian Highlands where brackish or fresh paramarine basins developed.

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LATE WISCONSINAN DEGLACIATION AND CHAMPLAIN SEA INVASION IN THE ST. LAWRENCE VALLEY, QUÉBEC

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ABSTRACT Champlain Sea history is directly linked to Late Wisconsinan deglacial episodes. Champlain Sea Phase I (Charlesbourg Phase) began in the Québec area at about 12.4 ka. It represented a western extension of the Goldthwait Sea between remnant Appalachian ice masses and the Laurentide Ice Sheet. Further south, at about the same time, in the Appalachian uplands and piedmont, high-level glacial lakes were impounded by the ice-front during glacial retreat toward NNW: lakes Vermont, Memphrémagog and Mégantic. Lowlands of the Upper St. Lawrence and Lake Champlain valleys were progressively deglaciated and inundated by Lake Iroquois and Lake Vermont. At about 12.1 ka, these two lakes coalesced and formed a single water-body, here referred to as Lake Candona. After the Ulverton-Tingwick Moraine was constructed, this lake extended northeastward onto the Appalachian piedmont where varved sediments containing *Candona subtriangulata* underlie marine clays. Current data and interpretations bring into question the former concept of the Highland Front Moraine System. The invasion of the main basin, or Champlain Sea Phase II, began around 12 ka. Replacement of Lake Candona by the sea resulted in a fall of about 60 m in water levels. Champlain Sea Phase III began at the end of the Saint-Narcisse episode, at about 10.8 ka. At this time marine waters were able to enter valleys of the Laurentian Highlands where brackish or fresh paramarine basins developed.

RÉSUMÉ *Le retrait glaciaire et l'invasion de la Mer de Champlain à la fin du Wisconsinien dans la vallée du Saint-Laurent, Québec.* L'histoire de la Mer de Champlain est directement liée à la déglaciation du Wisconsinien supérieur. La phase I de la Mer de Champlain (Phase de Charlesbourg) débute dans la région de Québec vers 12,4 ka. Elle représente le prolongement de la Mer de Goldthwait entre l'Inlandsis laurentidien et les glaces résiduelles appalachiennes. Plus au sud et approximativement en même temps, le retrait glaciaire vers le NNW sur les plateaux et le piémont appalachiens est marqué par des moraines et les lacs proglaciaires Vermont, Memphrémagog et Mégantic; les terres basses du haut Saint-Laurent et du lac Champlain étaient progressivement déglacées et inondées par les lacs Iroquois et Vermont. Vers 12,1 ka, ces deux lacs forment par coalescence le Lac Candona. Après l'épisode de la Moraine d'Ulverton-Tingwick, ce lac inondait le piémont appalachien vers le NE, où des varves à *Candona subtriangulata* reposent sous les argiles marines. Ces données remettent en question le concept de *Highland Front Moraine System*. L'invasion du bassin principal (Phase II de la Mer de Champlain) débute vers 12 ka. Le remplacement du Lac Candona par la mer provoque une chute d'environ 60 m du niveau du plan d'eau. La Phase III de la Mer de Champlain commence à la fin de l'épisode de Saint-Narcisse, vers 10,8 ka; les eaux marines pénètrent dans les vallées des Laurentides et sont coalescentes à des bassins paramarins saumâtres ou non salés.

ZUSAMMENFASSUNG *Enteisung im späten Wisconsin und der Einbruch des Meeres von Champlain in das Tal des Sankt-Lorenz-Stroms, Québec.* Die Geschichte des Meeres von Champlain ist direkt mit Enteisungs-Episoden im späten Wisconsin verknüpft. Die Phase I des Champlains-Meeres (Charlesbourg Phase) began im Gebiet von Québec um etwa 12.4 ka. Sie stellt eine westliche Ausdehnung des Goldwaith-Meeres dar, zwischen restlichen Eismassen der Appalachen und der laurentidischen Eisdecke. Während des glazialen Rückzugs nach NNW wurden weiter südlich etwa zur selben Zeit im Hochland und am Fuss der Appalachen glaziale Seen mit hohem Wasserspiegel durch die Eisfront geformt: der Vermont-, Memphrémagog-, und der Mégantic-See. Das Tiefland und die oberen Sankt-Lorenz- und Champlainsee-Täler wurden allmählich enteist und durch den Iroquois-See und den Vermont-See überschwemmt. Um etwa 12.1 ka vereinigten sich diese beiden Seen und bildeten einen einzigen See, der hier Candona-See genannt wird. Nachdem die Ulverton-Tingwick-Moräne geformt war, dehnte sich dieser See nach Nordosten aus bis zum Fuss der Appalachen, wo sich Warwen-Sedimente, die *Candona subtriangulata* enthalten, unter marinem Ton befinden. Diese Daten und Interpretationen stellen das frühere Konzept des Highland-Front-Moräne-Systems in Frage. Der Einbruch des Hauptbeckens oder die Phase II des Meeres von Champlain begann etwa um 12 ka. Als der Candona-See durch das Meer ersetzt wurde, senkte sich das Wasserniveau um etwa 60 m. Die Phase III des Champlain-Meeres begann am Ende der Episode von Saint-Narcisse, um etwa 10.8 ka. Zu diesem Zeitpunkt konnten marine Wasser in die Täler des laurentidischen Hochlands eindringen, wo sich mit Salzwasser gemischte oder frische paramarine Becken bildeten.

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INTRODUCTION

The Champlain Sea (Fig. 1) was an epicontinental, ephemeral sea of glacial isostatic origin that extended west of Québec at the end of the last glaciation. Sea waters flooded a basin of about 55 000 km^2 (Elson, 1969a) that occupied parts of the St. Lawrence, Ottawa and modern Lake Champlain lowlands. Deep valleys along the edge of the Laurentian Highlands, such as the Gatineau and the Saint-Maurice (Fig. 2), were occupied by waters which were relatively fresh due to the influx of abundant glacial meltwaters (Lamothe, 1977; Pagé, 1977; Occhietti, 1977). The areal extent of the marine basin underwent temporal changes that were controlled by three coeval and interdependent factors: 1) retreat of the ice sheet from the valley, 2) glacial isostatic recovery and 3) globally rising sea level. The elevation of the marine limit varies between 250 m north of Montréal and 55 m south of Lake Champlain (Fig. 1). The marine limit is diachronic, with its age ranging from about 12 ka to about 10.5 ka.

Champlain Sea sediments provide a record of a short-lived (about 2000 years) sedimentary cycle. Vertical and lateral facies changes are numerous and generally abrupt. High-energy, rapidly changing depositional environments were favored by the irregular topography of much of the submerged landscape, by the presence of extensive ice margins within

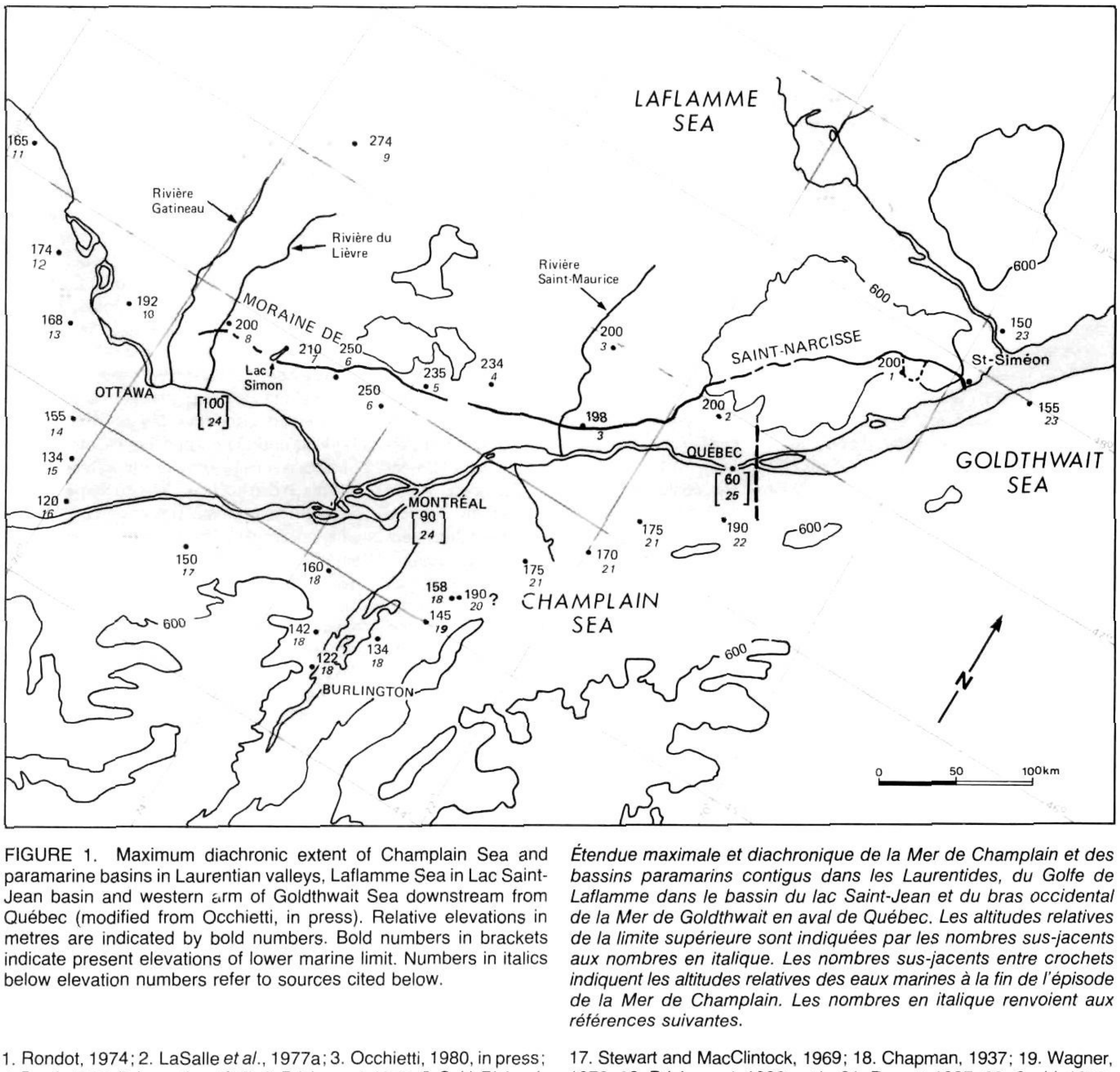


FIGURE 1. Maximum diachronic extent of Champlain Sea and paramarine basins in Laurentian valleys, Laflamme Sea in Lac Saint-Jean basin and western arm of Goldthwait Sea downstream from Québec (modified from Occhietti, in press). Relative elevations in metres are indicated by bold numbers. Bold numbers in brackets indicate present elevations of lower marine limit. Numbers in italics below elevation numbers refer to sources cited below.

Étendue maximale et diachronique de la Mer de Champlain et des bassins paramarins contigus dans les Laurentides, du Golfe de Laflamme dans le bassin du lac Saint-Jean et du bras occidental de la Mer de Goldthwait en aval de Québec. Les altitudes relatives de la limite supérieure sont indiquées par les nombres sus-jacents aux nombres en italique. Les nombres sus-jacents entre crochets indiquent les altitudes relatives des eaux marines à la fin de l'épisode de la Mer de Champlain. Les nombres en italique renvoient aux références suivantes.

1. Rondot, 1974; 2. LaSalle *et al.*, 1977a; 3. Occhietti, 1980, in press; 4. Pagé, 1977; 5. Lamothe, 1977; 6. Prichonnet, 1977; 7. S. H. Richard, 1980; 8. Allard, 1977; 9. Dadswell, 1974; 10. Fulton, pers. comm., 1983; 11. Gadd, 1963; 12. Barnett, 1980; 13. Barnett and Clarke, 1980; 14. Henderson, 1973; 15. Sharpe, 1979; 16. Henderson, 1970; 17. Stewart and MacClintock, 1969; 18. Chapman, 1937; 19. Wagner, 1972; 20. Prichonnet, 1982a et b; 21. Parent, 1987; 22. Gadd, 1978; 23. Dionne, 1977; 24. Brown Macpherson, 1967; 25. Occhietti, unpublished.

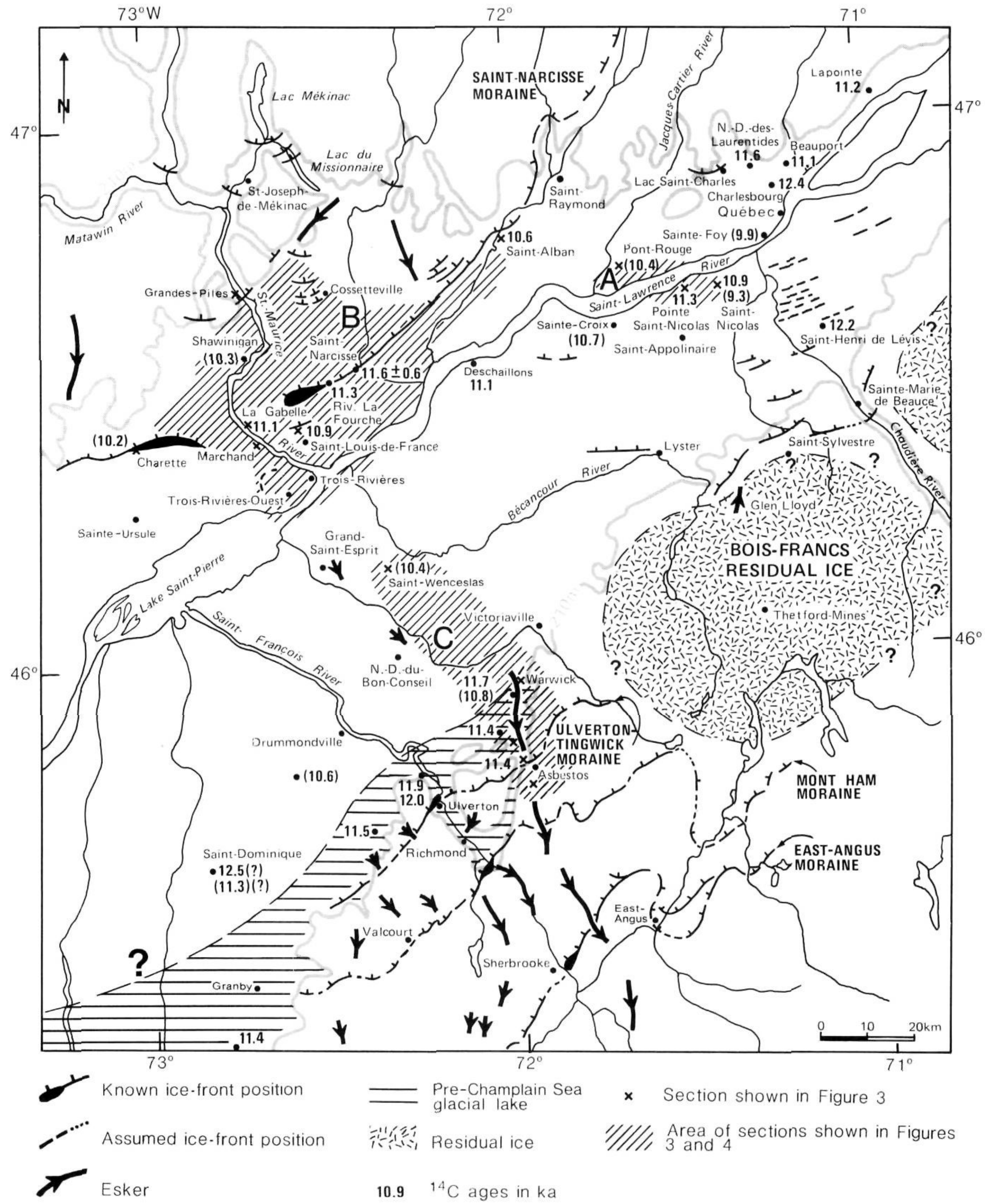


FIGURE 2. Recessional ice-front positions and chronological data for the eastern part of Champlain Sea. All ^{14}C dates are from marine shells; dates in brackets refer to regressive facies. The 210 m contour gives the approximate limit of the St. Lawrence Valley.

Étapes du retrait glaciaire et données chronologiques de la marge orientale de la Mer de Champlain. Toutes les datations proviennent de coquilles marines. Les âges (ka) entre parenthèses datent des faciès régressifs. La courbe de niveau de 210 m représente la limite approximative de la vallée du Saint-Laurent.

the basin and by high rates of glacial isostatic rebound. Rates of emergence locally reached 12 cm per year (Elson, 1969a; Hillaire-Marcel and Occhietti, 1980).

Although glacial and marine sediments in southern Québec and adjacent regions have been studied for over 100 years, the question of the timing and context of the onset of Champlain Sea invasion remains controversial. The first comprehensive assessment of deglacial events and chronology in southern Québec was that of Gadd *et al.* (1972b). The main points of their deglaciation model are the following ones:

- Between 14 ka and 12 ka, the margin of the Laurentide Ice Sheet retreated northwestward across the Appalachian uplands and deposited a series of recessional moraines.
- Proglacial lakes were impounded in Appalachian valleys until the ice margin had retreated from the position of the Highland Front Morainic System (older than about 12.2 ka).
- As the ice margin retreated north of the narrows at Québec, about 12 ka, marine waters invaded the isostatically depressed St. Lawrence Valley, thus forming the Champlain Sea.

However, reports of northward-trending striations in vast areas of the Appalachians of southern Québec (Lamarche, 1971, 1974; Gauthier, 1975; Lortie, 1976; LaSalle *et al.*, 1977a) together with reports of ^{14}C ages older than 12 ka on marine shells in the Ottawa Valley (Richard, S. H., 1974, 1978) soon suggested that substantial modifications should be made to this first deglaciation model. To fit this new data into a deglaciation model, most investigators then inferred that an ice-flow reversal had taken place in Late Wisconsinan time and that the northward flow event was caused by development (prior to 12.5 ka) of a calving bay in the St. Lawrence Valley (Gauthier, 1975; Lortie, 1976; Gadd, 1976; Shilts, 1976; Thomas, 1977), that perhaps extended into the Ottawa Valley as well (Gadd, 1980). In this revised deglacial framework, the Appalachian uplands of southern Québec remained covered by a substantial ice mass at the time of Champlain Sea invasion (Shilts, 1981; Gadd, 1983).

A third framework of deglacial events is now emerging as a result of fieldwork recently carried out along the southern shores of the Champlain Sea basin (Parent, 1984c, 1987; Prichonnet *et al.*, 1982a, 1982b; Parent *et al.*, 1985; Clark and Karrow, 1984) and in the central part of the basin (Occhietti, 1980, in press). This framework, in which the main Champlain Sea basin is not invaded until about 12 ka, is based on several lines of evidence which are presented in this paper.

This paper discusses (1) the opening of the lowlands to marine waters and (2) deglacial events and ice-retreat patterns prior to and during the Champlain Sea episode. The main area of interest includes the eastern part of the Champlain Sea basin and extends from Québec to Lac Saint-Pierre (Fig. 2).

Much earlier information on the eastern part of the Champlain Sea may be found in regional geological reports (Gadd, 1971; Gadd *et al.*, 1972a; LaSalle, 1978; LaSalle *et al.*, 1976, 1977a, 1977b; Dubé, 1971). These data will be discussed and summarized in this paper, but more emphasis will be placed on original data acquired by the present authors. The reconstruction of Champlain Sea phases is preceded by a summary of marine sedimentary facies, a general discussion of the stratigraphic record and a description of deglacial patterns.

SEDIMENTARY FACIES: INDICATORS OF GLACIAL RETREAT AND MARINE SUBMERGENCE

Champlain Sea facies analysis has had its own history of development. At first, Champlain Sea sediments were subdivided into *Leda* clays (*Portlandia arctica*), and *Saxicava* sands (*Hiatella arctica*) (Logan, 1863; Dawson, 1893). Gadd (1971) summarized much of the evidence presented during the previous 100 years and his work, together with that of Wagner (1970) on invertebrate marine faunas, laid a solid foundation for subsequent studies. The diversity of sediment facies and the importance of detailed local sedimentological and biostratigraphic studies were demonstrated by Hillaire-Marcel (1972, 1974) and co-workers (Hillaire-Marcel *et al.*, 1974) and by Rust and Romanelli (1975). During the last decade, fundamental contributions toward systematic facies analysis were made by several authors. Paleoecological studies of the invertebrate marine fauna allowed Hillaire-Marcel (1977, 1980) to first identify a series of typical faunal communities that commonly occur in a variety of Champlain Sea sedimentary facies. Faunal assemblages were later studied in the Ottawa basin of Champlain Sea (Rodrigues and Richard, 1983; Rodrigues, 1987). Hillaire-Marcel (1977, 1979, 1981) also developed the geochemical study of stable isotopes (^{18}O and ^{13}C) in typical faunal communities, in relation to the paleohydrological evolution of the marine basin. The biostratigraphy of ostracods and foraminifers (Cronin, 1977, 1979a, 1979b; Guilbault, 1980; Rodrigues, 1987) is now known and can be used jointly with the sedimentological and geochemical data. Several regional studies, such as those of Lamothe (1977), Pagé (1977), Donovan (1977), and Occhietti (1977, 1980), have placed much emphasis on marine and glaciomarine sedimentation, particularly in the vicinity of the Saint-Narcisse Moraine. In the Ottawa valley, drilling programs have also provided valuable local records of subsurface marine lithofacies sequences (Fransham *et al.*, 1976; Gadd, 1977, 1986). Therefore, a well defined set of sedimentological, paleontological and geochemical criteria are available as useful tools for Champlain Sea facies identification and paleoenvironmental reconstruction (Table I).

Through analysis of a large number of sections, three groups of primary sediments have been distinguished: (1) glacial facies, (2) glaciomarine facies, and (3) marine facies. One group of secondary deposits, consisting of sediments reworked by marine waters was also recognized. While description or discussion of glacial facies is beyond the scope of this paper, the other facies groups are briefly described, mainly in terms of their paleoenvironmental significance.

GLACIOMARINE FACIES

This facies group consists of sediments which were deposited at the glacier-sea interface and which were emplaced directly from the ice or by meltwater.

TABLE I

Main characteristics of sedimentary facies associated with Champlain Sea invasion

Facies group	Facies	Grain size	Sedimentary structures	Dominant macrofossils*	$\partial^{18}O$* (General trend)	Genesis
GLACIOMARINE	- marine allo-tills	- diamicton, mostly silt and clay	- crude stratification	Portlandia arctica	> 0	floating ice margin
	- pushed marine allo-tills	- diamicton and clay	- deformation structures	broken Portlandia arctica	> 0	glacial overriding
	- proximal glaciomarine sediments	- clayey and silty diamicton	- coarsely stratified and disrupted structures	Portlandia arctica Balanus hameri	> 0 ≥ 3	fluctuating ice margin
	- marine subaquatic outwash sediments	- cobble gravel, sand	- well stratified	unfossiliferous		meltwater undercurrents
	- pebbly clays and distal glaciomarine deposits	- clay, silt with ice-rafted stones and diamictons	- massive to well stratified	Portlandia arctica Balanus hameri Macoma calcarea	± 1 > 0 > 0	distal iceberg sedimentation, ice-rafted clasts and pods
MARINE	- transgressive facies	- gravel to clay	- stratified, turbidite bedding, rhythmites	Portlandia arctica Hiatella arctica	~ 0 ~ 0	strong bottom currents
	- inundation facies	- clay, silt, fine sand	- laminated, massive	Macoma balthica	~ -2	rain out, prodelta
	- regressive facies	- sand and gravel	- stratified	Hiatella arctica Mytilus edulis Macoma balthica Mya arenaria	~ -5 - 8 - 7 -10	prodelta, delta, littoral
REWORKED SEDIMENTS	- deep-water reworked sediments	- heterogeneous, diamictons	- coarsely stratified structures over an erosional surface	Balanus hameri Mya truncata Macoma calcarea	> 0 ~ 0 ± 2	bottom currents, tidal currents
	- shallow-water, sub-littoral and littoral reworked sediments	- coarse sediments	- stratified, winnowed	Mytilus edulis Balanus crenatus Hiatella arctica Mya arenaria	- 8 - 8 - 5 to -10 - 10	tidal and littoral currents, waves

* Mainly after Hillaire-Marcel (1980, 1981)

Proximal glaciomarine sediments include crudely stratified or massive, compact, matrix-dominated diamictons. They contain *Portlandia arctica* with articulated, broken or deformed valves and, less commonly, *Balanus hameri*, a species that inhabited cold, deep water. *Portlandia* shells usually have positive $\delta^{18}O$ values (Table I). In the field, these facies vary from crudely stratified, undisturbed allo-tills to massive diamictons or to diamictons deformed by ice-marginal or grounding line oscillations. Such facies are present in the Yamachiche Diamicton at the base of the Saint-Narcisse Moraine in the Saint-Maurice region (Occhietti, 1980). Proximal glaciomarine facies can be used to determine former ice-frontal positions.

Marine subaquatic outwash deposits consist of sediments directly deposited into the sea at points of high meltwater discharge. They form mounds or ridges that rise slightly above the level of surrounding clay plains or which may be buried beneath Champlain Sea clays. They apparently contain no fossils, except at their summit or in distal positions. These sediment bodies which are elongated more or less parallel to glacier flow have been well described by Rust and Romanelli (1975) and by Rust (1977).

Distal glaciomarine deposits and pebbly clays occur at the base of many sections, notably at the edge of the Saint-Narcisse Moraine, such as at Marchand and Saint-Alban (Occhietti, 1980; Fig. 3). They consist of uncompacted, faintly stratified silt and clay containing ice-rafted clasts. They contain species which inhabited cold, polyhaline-euhaline waters: *Portlandia arctica, Balanus hameri, Macoma calcarea* and *Macoma balthica*. Distal glaciomarine sediments mark the transition from ice-marginal to true marine sediments. The distance between the ice front and the ice-rafted clasts is variable. For instance, in the Saint-Nicolas section (Fig. 3), the lower clays with dropstones and diamicts probably accumulated several tens of kilometres from the ice front while at Saint-Alban the ice front may have been only about 2 km away (Occhietti, 1980). Consequently, the occurence of pebbly clays and distal glaciomarine diamictons cannot be retained as a criterion to identify ice marginal positions.

MARINE SEDIMENTS

A threefold sequence has been proposed for strictly marine sediments: transgressive facies, inundation facies and regressive facies (Occhietti, 1980). Transgressive facies occur at the base of marine sequences. They include stratified deposits that show strong granulometric variability. Inundation facies comprise marine clays and silts that are usually faintly stratified or, in places, massive. These deep-water sediments contain few fossils; however, faunal and microfaunal abundance

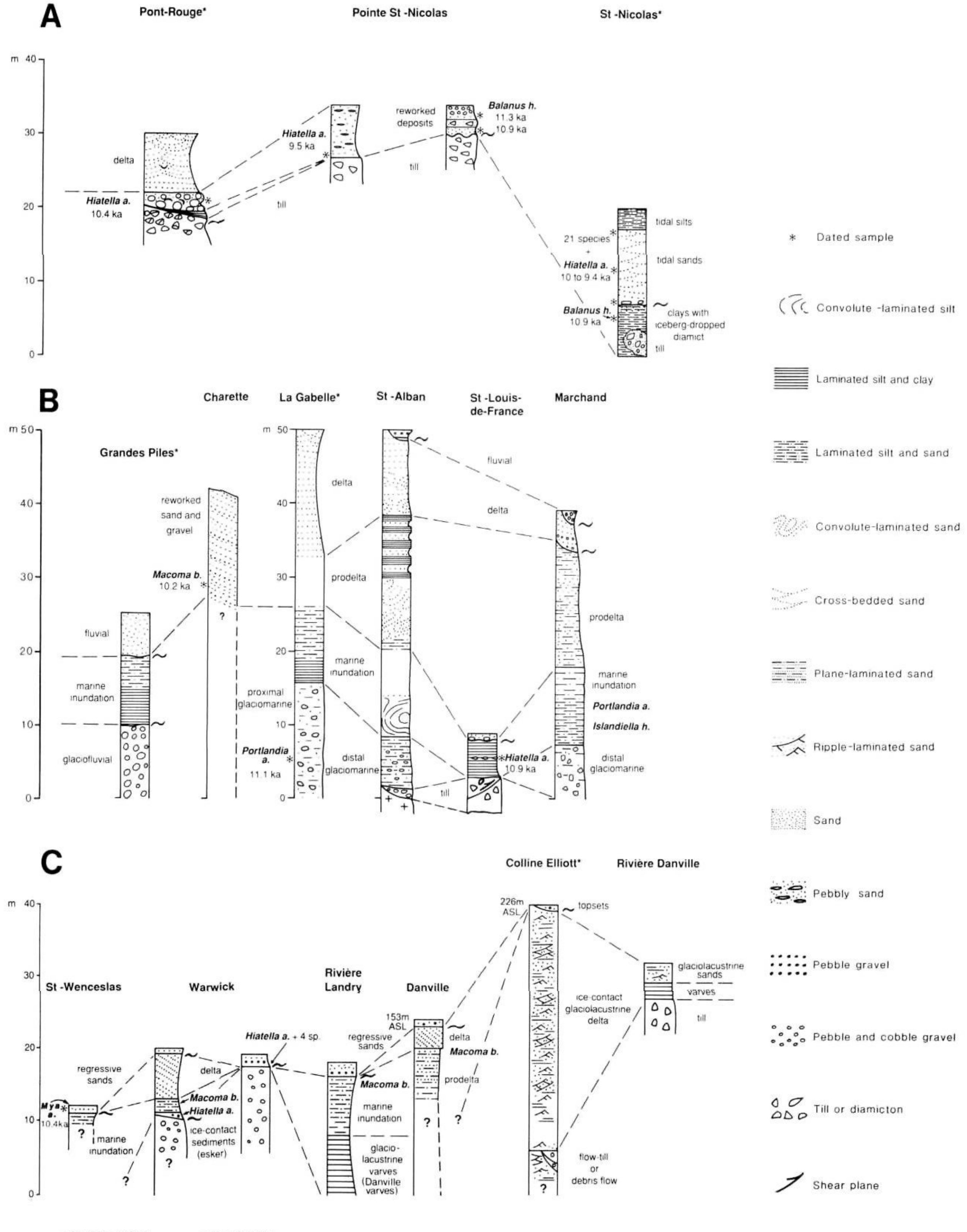


FIGURE 3. Representative sections in the area of interest, and proposed correlations. Marine inundation facies refers to laminated or massive silt and clay. Sections in the Québec region (group A) are from Occhietti and Hillaire-Marcel (1982, unpublished). Sections in the Trois-Rivières region (group B) are from Occhietti (1980). Sections in the Asbestos region (group C) are from Parent (1987). Locations of sections are shown in Figure 2; names of four sections in the Asbestos region are given in Figure 5.

Lithostratigraphie et corrélations de coupes représentatives de la région étudiée. Les faciès d'inondation marine sont composés d'argiles et de silts massifs ou stratifiés. Les coupes de la région de Québec (groupe A) sont de Occhietti et Hillaire-Marcel (1982, non publié); les coupes de la région de Trois-Rivières (groupe B) sont de Occhietti (1980); les coupes de la région d'Asbestos (groupe C) sont de Parent (1987). Les coupes sont localisées sur la figure 2; les noms de quatre coupes de la région d'Asbestos sont indiqués sur la figure 5.

and diversity tend to increase upward in these sequences (Parent, 1987, Fig. 5-5; Rodrigues, 1987). Regressive facies commonly include fossiliferous littoral sands, sandy alluvial fans and bars (called high-terrace sands in Gadd, 1971), stratified silts deposited in intertidal zones, and deltaic sediments. The malacological fauna and microfauna have strongly depleted $\delta^{18}O$ values (Table I); assemblages are indicative of significantly reduced salinities.

Non-fossiliferous deposits of the Laurentian paramarine basins are a variety of marine sediments. The clays in these basins are derived from glacial flour as are those in the main Champlain Sea basin. These sediments are stratified or massive but contain no marine fauna or microfauna. Deposits in the paramarine basin near Saint-Joseph-de-Mékinac (Fig. 2) are a good example of this type of material.

GLACIAL SEDIMENTS REWORKED BY THE SEA

In the field, these secondary deposits may be confused with glaciomarine deposits. *Deep-water reworked sediments* may resemble either coarse glaciofluvial deposits or glaciomarine diamictons. The absence of *Portlandia arctica*, a species typical of glaciomarine facies, is quite characteristic. In species assemblages, the most common species seems to be *Balanus hameri* accompanied by *Hiatella arctica, Macoma calcarea* and *Mya truncata*. Contacts between these stratified sediments and underlying or adjacent glacial deposits are erosional. The reworking may be attributed to gravity- or tide-driven bottom currents. These deposits do not record ice-marginal positions but simply indicate reworking of previous glacial deposits.

Shallow-water reworked deposits are easier to identify. In general, they consist of stratified sandy or gravelly sediments and may also include boulder lags. Their grain size is a direct reflection of their source. At Charette (Fig. 2), stratified sands on the side of the Saint-Narcisse Moraine containing *Hiatella arctica, Macoma balthica* and *Mya arenaria* originated from reworking of glaciofluvial sands. At Pont-Rouge (Fig. 2), reworked sediments containing numerous boulders and shells of *Balanus* cf *crenatus, Mytilus edulis* and *Hiatella arctica* come from the erosion of a till mass in shallow water. At Warwick (Fig. 2), the summit of the esker has been reworked by waves (Figs. 3c and 4). In general, the strongly depleted $\delta^{18}O$ values of the fauna confirm an intertidal paleoenvironment and exclude the possibility of sedimentation near an ice front (Hillaire-Marcel, 1977, 1980).

THE STRATIGRAPHIC RECORD

Stratigraphic records described in this paper are from the eastern part of the Champlain Sea basin, in 1) the Appalachian piedmont and adjacent uplands between Valcourt and Victoriaville, 2) the Saint-Maurice region and 3) the Québec region (Fig. 2). These regions are separated by a sizeable area in which there is a paucity of radiocarbon-dated faunal assemblages (Fig. 2) and where glacial and deglacial sediments are commonly concealed under a blanket of marine sediments. Representative sections from these three regions are shown in Figure 3 while schematic stratigraphic relationships are depicted in Figure 4. Features and deposits described in the text are mainly those that are used to link the lithostratigraphic and morphostratigraphic record with deglaciation and Champlain Sea history (Table II).

THE ULVERTON-TINGWICK MORAINE

Glacial deposits of the Ulverton-Tingwick Moraine (Parent, 1984b, Fig. 2) rest directly on Appalachian hills. They grade laterally into ice-contact deltas and subaquatic outwash bodies that were deposited into pre-Champlain Sea glaciolacustrine water bodies (McDonald, 1968; Parent, 1987). At Ulverton, the moraine lies locally below the marine limit; the ice-contact deposits were thus slightly reworked during the marine episode. After the moraine was formed, small stagnant ice blocks persisted until the marine invasion and were partially buried by deltaic and prodeltaic sediments, for instance, near Danville and Warwick (Parent, 1984c, Fig. 3c). The Tingwick segment, east of Asbestos, is mainly composed of till ridges with intervening meltwater channels, both of which lie consistently on north- or northwest-facing slopes. Ice-contact deltas, downstream of the main channels, indicate that the level of the ice-dammed lake was at an elevation of about 230 m. Construction of these deltas is unequivocally contemporaneous with moraine-building (Parent, 1987).

The Ulverton segment, southwest of Asbestos (Fig. 2), was formed in a large proglacial lake whose shorelines also reached an elevation of about 230 m near the moraine. This segment includes a few till ridges, but is mainly composed of bodies of subaquatic outwash sediments supplied by subglacial channels, as indicated by the presence of several eskers (Fig. 2). Lateral correlation between the two discontinuous moraine segments can be established in two ways: (1) this was the last series of morainic deposits that formed in a pre-Champlain Sea ice-dammed lake; and (2) glaciolacustrine deltas as high as 220 or 230 m in elevation are commonly found along both segments of the morainic belt. This does not exclude the possibility that the two moraine segments may be slightly diachronic. Since varve series near Richmond and at Rivière Landry record only about 150 years of glaciolacustrine sedimentation before and after construction of the moraine (Parent *et al.*, 1986; Parent, 1987), this diachronism should be minor. Field evidence indicates that the Ulverton-Tingwick Moraine is a recessional feature, marking merely a pause during retreat of the Laurentide Ice Sheet.

THE WARWICK-ASBESTOS ESKER

The Warwick-Asbestos Esker consists of a series of segments, each grading into subaquatic outwash, which are characterized mainly by plane- and cross-laminated sandy lithofacies (Fig. 5: sites 2 and 3). The youngest segment, near Warwick, also shows distinct grain-size gradation, ranging from cobble gravels upstream to sandy gravels downstream. At all sites, primary structures indicate southward meltwater flow (Fig. 5). Vertical and lateral sequences are similar to those observed in eskers that emerged from the base of glaciers in deep water (Banerjee and McDonald, 1975). At its southern extremity, the esker grades into a large flat-topped kame-delta, Colline Elliott (Figs. 3c, 4, 5), which forms

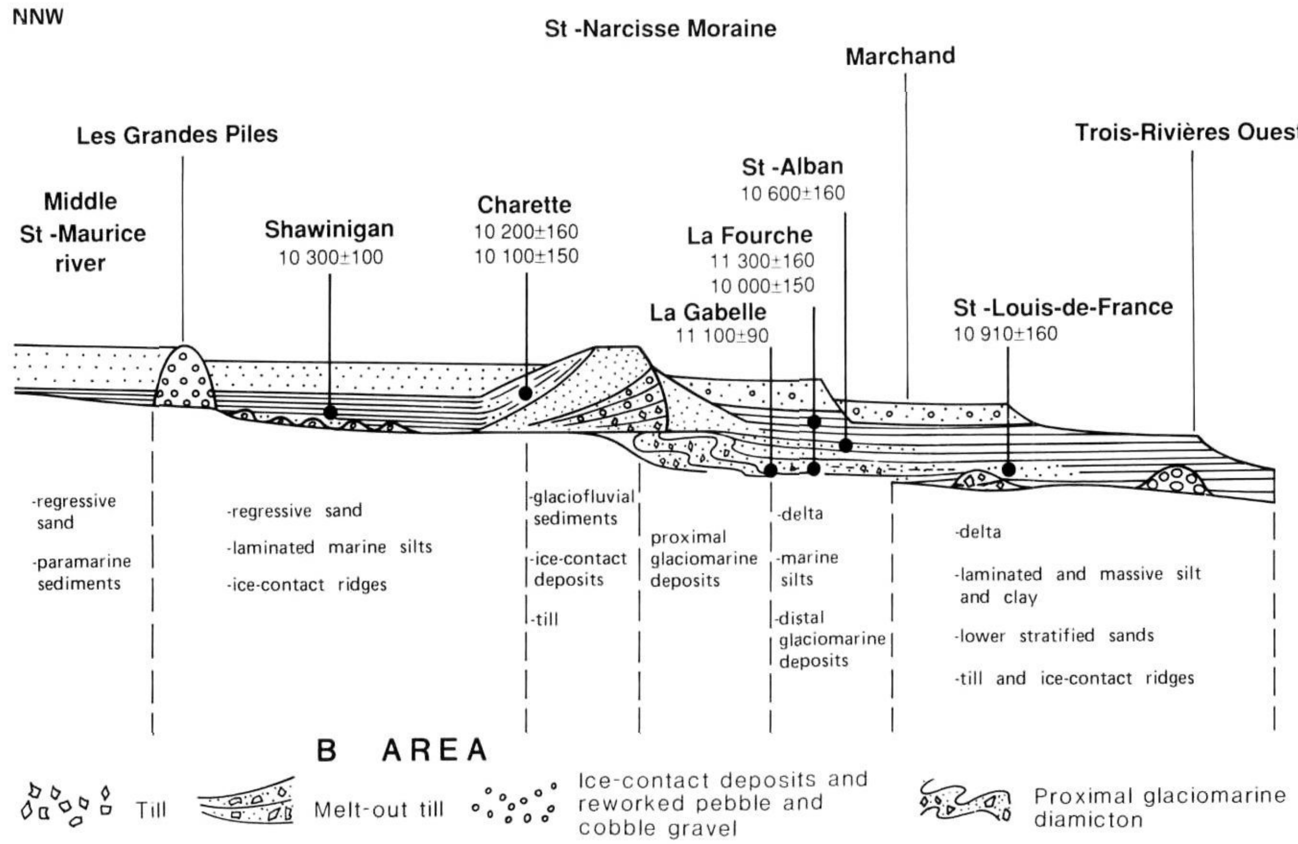


FIGURE 4. Schematic cross-section showing lateral relations of facies sequences and chronology of marine units across the St. Lawrence Valley, between the Appalachian piedmont and the middle and lower Saint-Maurice region.

Variations latérales des séquences de faciès et chronologie au ¹⁴C des milieux glaciomarins et marins, transversalement à la vallée du Saint-Laurent, entre le piémont des Appalaches et la Mauricie (région du bas et du moyen Saint-Maurice).

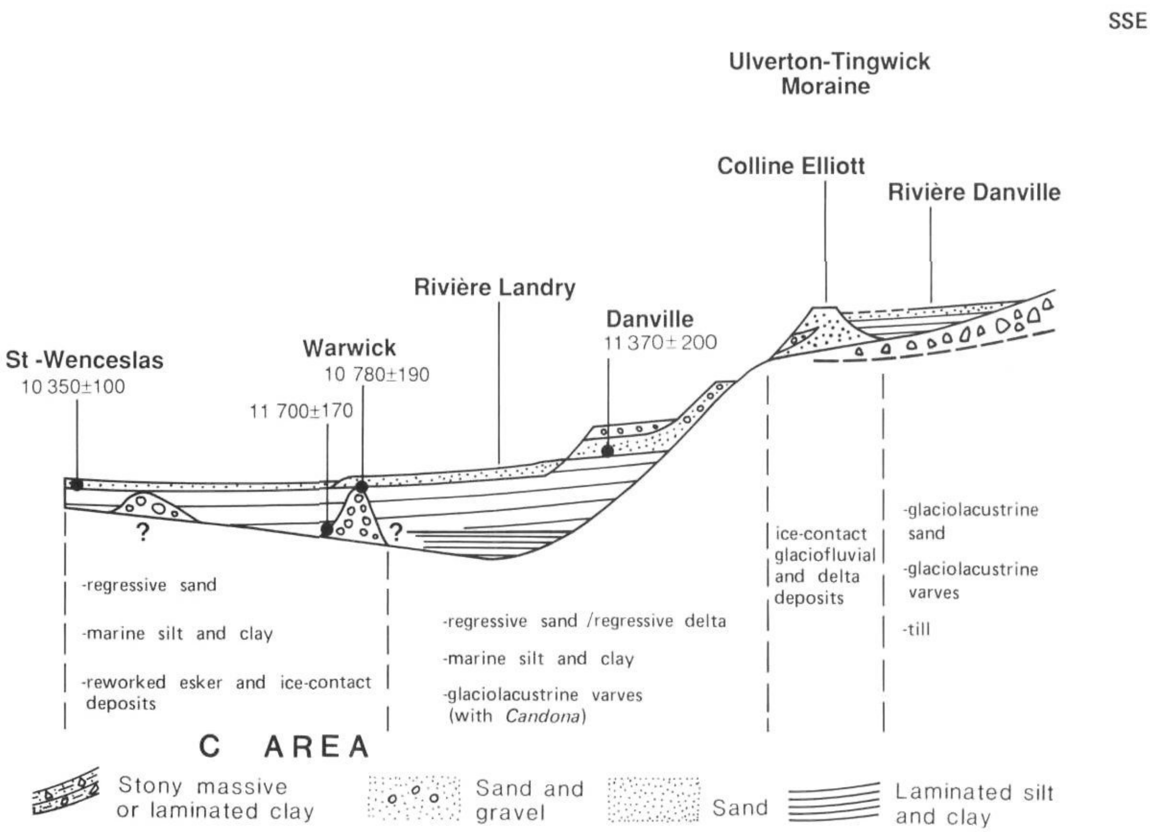
part of the Ulverton-Tingwick Moraine. At this locality, over 40 m of sandy subaquatic outwash sediments, which also include a few flow till or debris flow beds, are capped by discontinuous, up to 1 m-thick, topset beds consisting of trough-cross-bedded sandy gravel (Fig. 3). It is inferred that the high rate of sediment supply during construction of the Ulverton-Tingwick Moraine allowed the subaquatic outwash fan to build up to lake level, at which time the cross-bedded gravels were deposited in shallow channels that formed at the surface of the flat-topped sediment body, at an altitude of about 226 m.

Since the Warwick-Asbestos Esker lies below marine limit, its surface was locally rounded by waves. Elsewhere, it was buried under inundation facies clays and regressive facies sands. At Warwick (Figs. 4, 5), coarse reworked sediments underlying marine inundation facies clays with *Macoma balthica* contain an almost undisturbed community of *Hiatella arctica* and *Balanus crenatus*, dated 11 700 ± 170 BP (I-13342). This date provides a minimum age for the marine invasion. The reworked gravel and sand deposited during marine regression contain a more diverse fauna (*Hiatella arctica, Macoma balthica, Balanus crenatus, Mytilus edulis* and *Mya pseudo-arenaria*), dated 10 780 ± 190 BP (UQ-289).

GLACIOLACUSTRINE VARVES WITH *CANDONA*

Varves have been found in several sections on both the proximal and distal sides of the Ulverton-Tingwick Moraine. Some sites lie close to the upper marine limit: either above, as at Rivière Danville (Figs. 3, 4, 5) or definitely below, as at Rivière Landry (Figs. 3, 4, 5). These varves provide further key evidence indicating that glaciolacustrine water bodies were impounded in the Appalachian uplands and piedmont prior to Champlain Sea invasion (Parent, 1984b, 1986, 1987). These lakes extended northeastward at least as far as Warwick (Fig. 2).

The most significant section is located on the banks of Rivière Landry near Danville. An 8 m-thick varve series (the Danville varves of Parent, 1987), is conformably overlain by fossiliferous marine inundation facies clays, which grade upward into prodeltaic silt and sand rhythmites. The varves contain 103 couplets. Silty summer layers may be characterized as part of a fining-and-thinning-upward turbidite sequence, while winter layers consist of typical rain-out sediments (80 to 90% clay). The varves contain only one species of ostracod, *Candona subtriangulata* (Parent *et al.*, 1986). Unlike the overlying marine clays, the varves contain no foraminifers and can be attributed to an Arctic lacustrine paleoenvironment according to the paleoecological studies of Cronin (1977). Because the shorelines of the last recognized pre-Champlain Sea lacustrine water-body in the Asbestos-Warwick area reach an elevation of about 230 m, it must be concluded that the varves were deposited in this ice-dammed lake. An age of 11 700 ± 170 BP (I-13342) on *Hiatella arctica* from the



Warwick site provides a maximum age for the end of glaciolacustrine sedimentation at nearby Rivière Landry (Parent, 1987).

THE SAINT-LOUIS-DE-FRANCE MORAINIC RIDGE

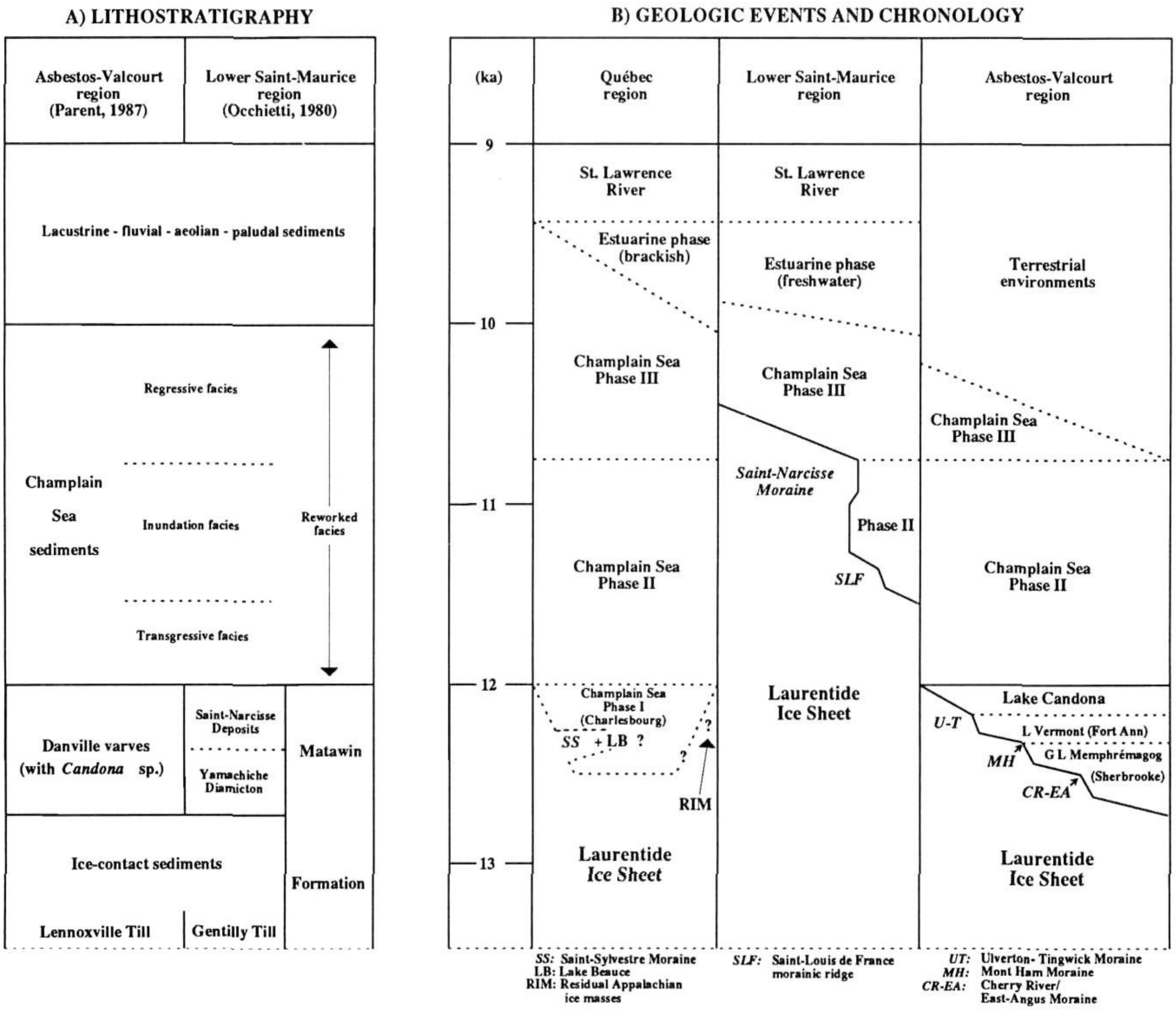
Except for information on units pertaining to Champlain Sea regression, and for a few wave-reworked ice-contact deposits (Gadd, 1971), there are few pertinent stratigraphic observations from the lowlands lying just south of Lac Saint-Pierre (Fig. 2). Areas north of the St. Lawrence River, however, are well known (Gadd, 1971; Occhietti, 1977, 1980). The 24 m-wide and 5 m-high Saint-Louis-de-France morainic ridge lying 12 km north of Trois-Rivières (Figs. 2, 4; Occhietti, 1977, 1980) consists of two imbricated, northward-dipping thrust slices. The proximal flank of the ridge has a steep slope gradient while the distal flank has smoother slopes and is underlain by a more bouldery till. The ridge is definitely older than 10 910 ± 160 BP (I-9484; *Hiatella arctica*), since it is stratigraphically older than the shell-bearing cobbly layer intercalated in marine clays that overlie the distal flank of the ridge. The shear structures indicate that the ridge was formed at the margin of a grounded glacier undergoing compressive ice-flow. This morainic ridge, which is distinctly older than the age of marine shells included in the Saint-Narcisse Moraine (see next section), records a recessional ice-front position that is probably slightly older than 11.3 ka. The dated fossiliferous cobbly layer records an episode of extensive ice-rafting, presumably shortly prior to or during construction of the nearby Saint-Narcisse Moraine (Figs. 2, 4).

THE SAINT-NARCISSE MORAINE COMPLEX

Deposits of the Saint-Narcisse Moraine complex, which extend over a distance of roughly 550 km between Lac Simon (northeast of Ottawa) and Saint-Siméon, about 150 km northeast of Québec (Fig. 1), have been described by several authors since the early 1950s (see Occhietti, 1980, for references). In the Saint-Maurice region, the morainic complex lies along the southern edge of the shield and is composed of two partly superposed sediment bodies: the Yamachiche Diamicton and the Saint-Narcisse Deposits (Occhietti, 1980). The Yamachiche Diamicton consists of fossiliferous proximal glaciomarine deposits and of probable subaquatic allo-tills. The unit outcrops at the base of sections along the Rivière Saint-Maurice (La Gabelle section in Fig. 3) and other nearby rivers. It contains *Portlandia arctica* shells dated 11 300 ± 160 BP (GSC-1729) and 11 100 ± 90 BP (GSC-2045) (Fig. 4). These ages confirm the age of 11 600 ± 600 BP (GSC-1526) obtained by Gadd on foraminifers (Gadd *et al.*, 1972a). The Saint-Narcisse Deposits include melt-out tills or tills with scale structures (Lavrushin, 1971), and a suite of ice-contact and proglacial deposits which form the core of the Saint-Narcisse Moraine ridges in the Saint-Maurice region (Fig. 4). *Macoma balthica* and *Hiatella arctica* shells sampled at Charette in sediments reworked from the moraine were dated at 10 100 ±

TABLE II

Late Wisconsinan and early Holocene stratigraphy and events in the eastern region of the Champlain Sea basin. Notice that the timing of events prior to and during the Charlesbourg Phase in the Québec region is conjectural (see text).



150 BP (GSC-1444; Gadd *et al.*, 1972a) and 10 200 ± 160 BP (GSC-1700; Occhietti, 1976).

QUÉBEC REGION

In the Québec narrows, where oceanic waters entered and glacial meltwaters exited, the observed sections contain essentially reworked facies and tidally influenced regressive facies (Fig. 3; Occhietti and Hillaire-Marcel, 1982). Other deposits are exposed at the base of some sections but lateral correlations remain to be established for these underlying glacial and non-glacial units. Based on the Pointe Saint-Nicolas sections, deglaciation is known to have occurred before 11 340 ± 180 BP (UQ-40) (Fig. 3; Occhietti and Hillaire-Marcel, 1982). At Charlesbourg, on the north shore of the Saint-Lawrence River, fossiliferous clayey sands, dated 12 400 ± 160 BP (GSC-1533), are overlain by a 3 m-thick diamicton (LaSalle *et al.*, 1977a, photo 19; Fig. 5-2 in Gadd *et al.*, 1972a). The dated shells, intact specimens of *Balanus hameri*, at an elevation of 109 m, suggest that initial deglaciation had occurred by 12.4 ka. Depending on the origin assigned to the overlying diamicton, marine sedimentation may have been subglacial or may have preceded a glacial readvance. The region was certainly deglaciated before 11 600 ± 160 BP (GSC-1235), which is the age of a *Mya truncata* biocenotic assemblage found at the top of ice-marginal deposits at Notre-Dame-des-Laurentides (Fig. 2; LaSalle *et al.*, 1977a).

In conclusion, the deglacial and stratigraphic records from the eastern sector of the Champlain Sea show a sequence of time-transgressive events: glacial retreat, glacial-glaciolacustrine transition, glaciolacustrine-marine transition and glacial-marine transition, marine cycle, lacustrofluvial or estuarine phase, and fluvial erosion (Table II and Fig. 4). The

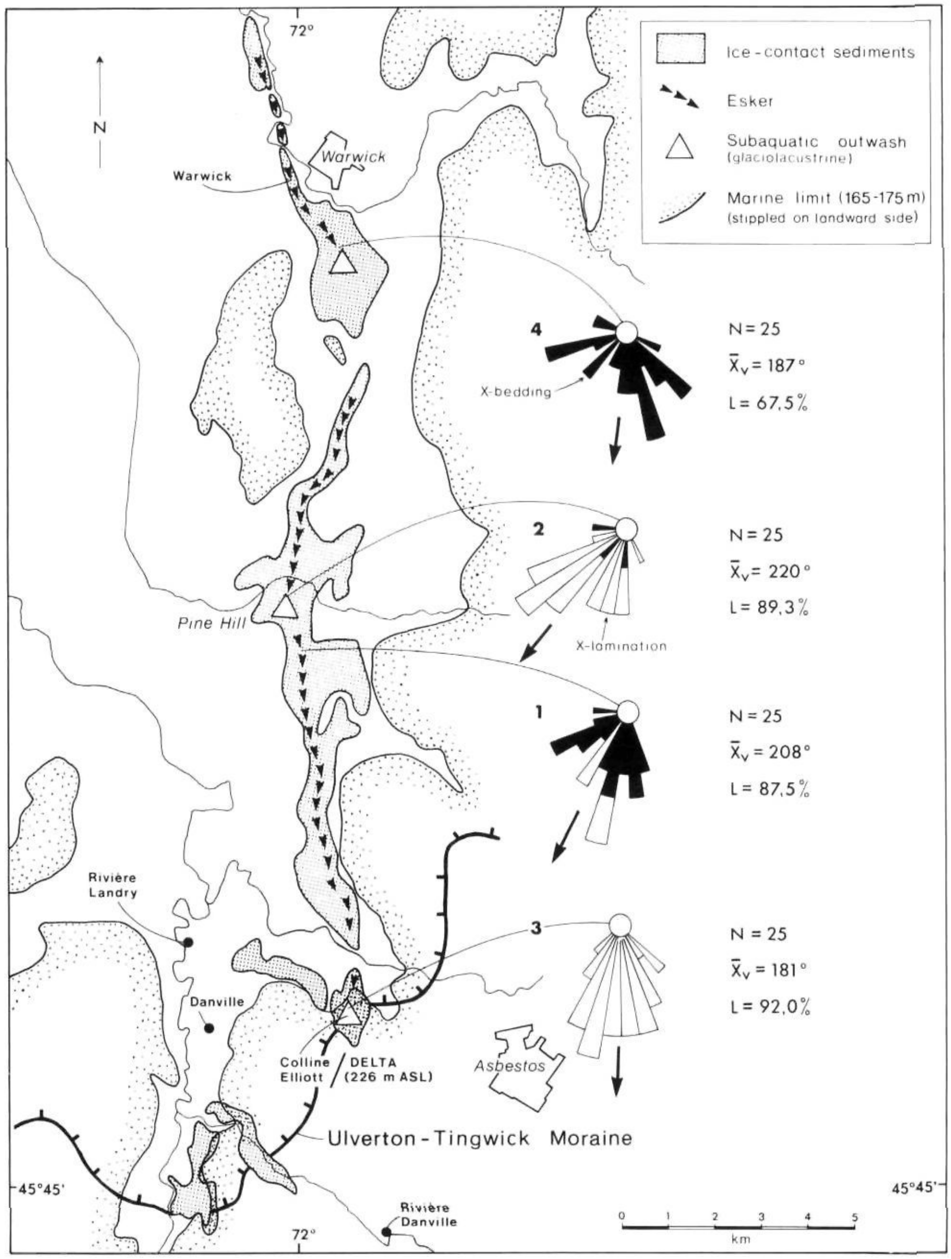


FIGURE 5. The Warwick-Asbestos Esker and its relationship with the Ulverton-Tingwick Moraine and with the upper marine limit. Paleocurrent data indicate that the esker was deposited by southward-flowing meltwaters (from Parent, 1987).

L'esker de Warwick-Asbestos, en relation avec la Moraine d'Ulverton-Tingwick et la limite marine supérieure de la Mer de Champlain. Les indicateurs de paléocourants représentés par les vecteurs moyens indiquent que l'esker a été construit par des eaux de fonte s'écoulant vers le sud (de Parent, 1987).

marine cycle consisted of a threefold time-transgressive sequence, transgression/inundation/regression.

DEGLACIAL PATTERNS IN THE APPALACHIAN UPLANDS

Mapping of deglacial landforms and deposits in the Appalachian uplands of southern Québec (McDonald, 1967; Gadd *et al.*, 1972b; Dubé, 1971; Shilts, 1981, Clément and Parent, 1977, Parent, 1978, 1987; Gadd, 1978; Chauvin, 1979; Prichonnet *et al.*, 1982a, 1982b; Chauvin *et al.*, 1985) indicates that backwasting of an active ice-front, as evidenced by the construction of recessional morainic belts, was a characteristic deglaciation process in several parts of the region, while downwasting of residual ice masses was the dominant process in other parts, such as the Bois-Francs region (Fig. 6).

RECESSIONAL MORAINES AND GENERAL PATTERN OF GLACIAL RETREAT

The recessional morainic belts shown in Figure 6 commonly consist of discontinuous segments that are laterally correlated on the basis of their altitude and trend and of their inferred relationships with proglacial lakes (Parent, 1987). In terrains where large, deep proglacial lakes were impounded at the ice margin, such as in the middle Saint-François and upper Chaudière valleys, end-moraine segments consist mainly of ice-contact bodies of stratified sediments that were deposited subaquatically; these ice-contact bodies are locally distinctly

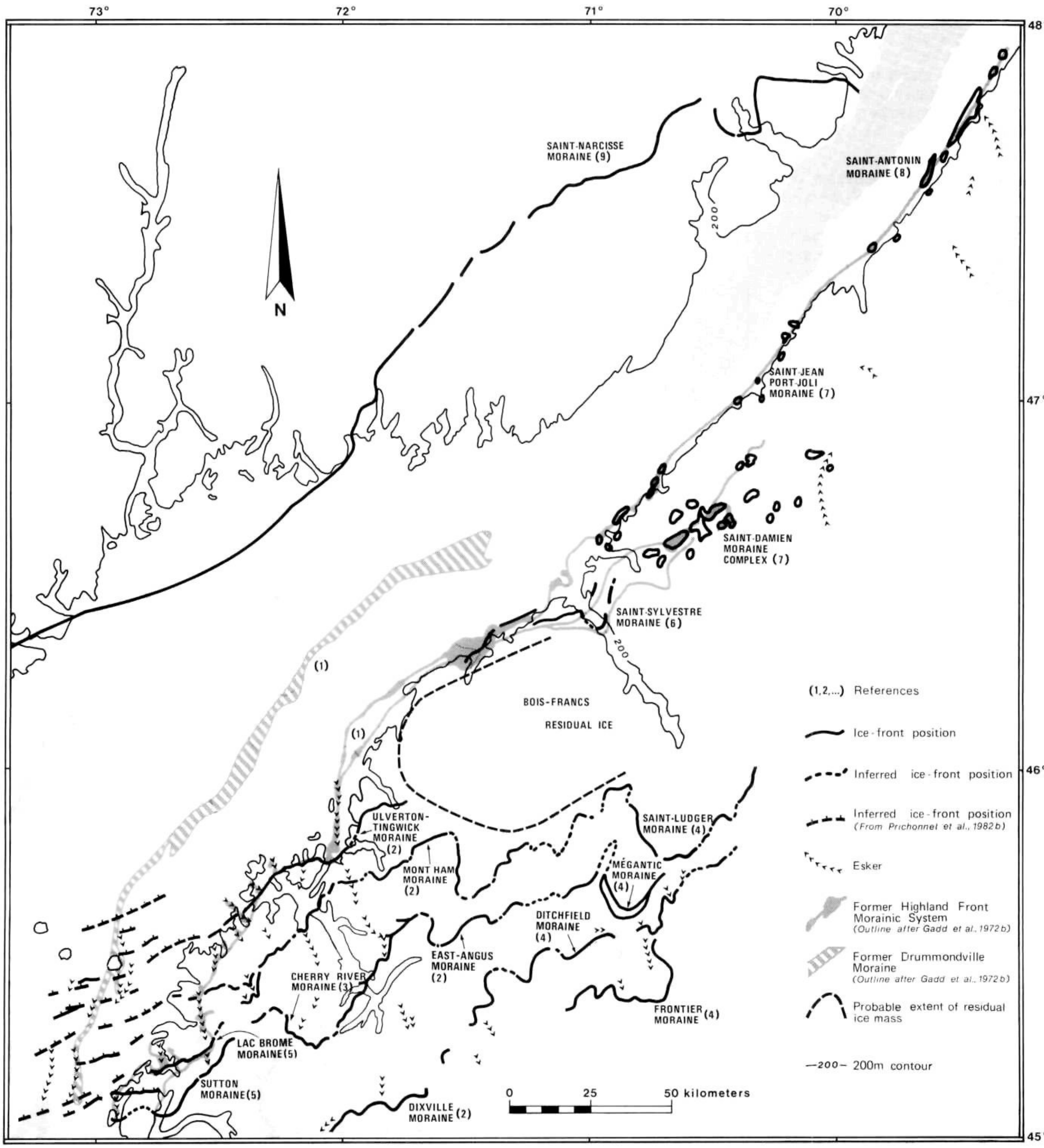


FIGURE 6. Deglacial patterns in southern Québec. Recessional ice-front positions in the Appalachian uplands and piedmont were compiled from several sources (see below). Lac Brome, Sutton and Saint-Ludger moraines are informal terms introduced in this paper for location purposes. Notice that known or inferred ice-front positions in the southwest part of the area are almost perpendicular to the trend of features which were referred to as the Drummondville Moraine and as the Highland Front Moraine System by Gadd *et al.* (1972b).

1. Gadd *et al.*, 1972b; 2. Parent, 1987; 3. McDonald, 1967; 4. Shilts, 1981; 5. Prichonnet *et al.*, 1982a, 1982b; 6. Gadd, 1978; 7. Chauvin *et al.*, 1985; 8. Lee, 1962; 9. Occhietti, 1980.

Le mode de retrait glaciaire au Québec méridional. Les positions du front glaciaire en retrait sur les reliefs et le piémont des Appalaches sont extraites des références citées plus bas. Les moraines du lac Brome, de Sutton et de Saint-Ludger sont des termes non formels introduits dans cet article pour des raisons pratiques. À noter que dans la partie sud-ouest de la carte, les positions déduites ou observées du front glaciaire en retrait sont transversales à la Moraine de Drummondville et au Highland Front Morainic System *proposés par Gadd* et al. (1972b).

ridged. In terrains where only small or shallow proglacial lakes were impounded at the ice-front, morainic segments consist mainly of assemblages of till ridges with intervening ice-marginal meltwater channels (Clément and Parent, 1977; Parent, 1978, 1987).

The oldest morainic belts are the Frontier Moraine and the Ditchfield Moraine of the Lac Mégantic region (Shilts, 1981). These moraines are thought to have formed some time prior to 13 ka (Shilts, 1981). The Dixville Moraine (Parent, 1987), a morainic belt whose features were first reported and discussed by Clément and Parent (1977), is probably a correlative of the Ditchfield Moraine or perhaps, of the slightly older Frontier Moraine (Fig. 6).

The next younger group of morainic belts are the Mégantic Moraine (Shilts, 1981; Fig. 6), the East Angus Moraine (Clément and Parent, 1977; Parent, 1978; Fig. 2), the Cherry River Moraine (McDonald, 1967, 1968; Fig. 6) and belts of morainic deposits on the northwest flank of the Sutton Mountains (Prichonnet *et al.*, 1982b), herein called Sutton and Lac Brome moraines for convenience (Fig. 6). Although interconnecting these morainic belts has generally proven to be a difficult task in the field, Parent (1987) believes that they are at least broadly synchronous. Because of significant gaps between deposits of the Cherry River Moraine in the vicinity of Sherbrooke, the outline of the moraine has been the subject of somewhat different interpretations as new evidence was gathered (McDonald, 1967, 1968; Clément and Parent, 1977; Boissonnault and Gwyn, 1983; Parent, 1987). The outline given in Figure 6 closely resembles that originally suggested by McDonald (1967, 1968). The Cherry River and East Angus moraines are thought to be almost synchronous features since related ice-contact deltas indicate that both morainic belts formed shortly after water levels of glacial Lake Memphrémagog had fallen to the level of the Sherbrooke phase (Parent, 1987; see section on lacustrine phases prior to the marine invasion). Ice-contact deltas and terraces associated with the Sutton and Lac Brome moraines (Prichonnet *et al.*, 1982a) suggest that waters of glacial Lake Vermont (Chapman, 1937) fell from the level of the Coveville phase to that of the Fort Ann phase during glacial retreat from the Sutton moraine to the Lac Brome moraine. The level of the Sherbrooke phase of glacial Lake Memphrémagog was controlled by the Lac Nick outlet (McDonald, 1968), a spillway which lies some 10 km south of the Sutton moraine. Because this spillway was being utilized during construction of the Cherry River and East Angus moraines, it may be inferred that these moraines are at least partly, and perhaps wholly, correlatives of the Sutton moraine. Although the correlation of the Mégantic Moraine with the Cherry River and East Angus moraines cannot be made through relationships with late-glacial water bodies, they are tentatively correlated on the basis of discontinuous morainic deposits located in the intervening uplands (Fig. 6).

The Mont Ham Moraine (Parent, 1978, 1984d, 1987) and the Ulverton-Tingwick Moraine (Parent, 1987) are morainic belts formed during ice-retreat north of Sherbrooke. Although these morainic belts are locally discontinuous, lateral relationships with regional and local proglacial lakes indicate that the various segments of each belt were formed nearly synchronously. The Mont Ham Moraine formed just prior to final drainage of glacial Lake Memphrémagog while the younger Ulverton-Tingwick Moraine was built during the Fort Ann phase of glacial Lake Vermont (Parent, 1987). The trend of the Mont Ham Moraine strongly suggests that it is younger than the Lac Brome Moraine (Fig. 6). Since morainic deposits north of this moraine were formed during the Fort Ann phase (Cloutier, 1982), it may be inferred that the Mont Ham Moraine and the Sherbrooke phase of glacial Lake Memphrémagog are at least partly correlative with the Fort Ann phase. Recessional morainic deposits near Saint-Ludger in the upper Chaudière Valley (Shilts, 1981), shown as the Saint-Ludger moraine in Figure 6, are tentatively correlated with the Mont Ham Moraine on the basis of discontinuous morainic segments located in the intervening uplands (Parent, 1987).

These morainic belts show a general ENE-WSW trend, with lobes and re-entrants related to local topography (Fig. 6). This general trend is oblique to Appalachian structures and landscape features. Except for the Bois-Francs area, where morainic belts are conspicuously absent (Chauvin, 1979; Parent, 1987), the trend of morainic belts described above indicates that the Laurentide Ice Sheet retreated toward NNW across the Appalachian uplands and piedmont of southern Québec.

East of the Chaudière Valley, deglacial patterns seem more complex. This may be because the Notre-Dame Mountains and adjacent uplands were covered by a remnant Appalachian ice mass (Chauvin *et al.*, 1985), which was an extension of residual ice in the northern Maine uplands (Borns, 1985; Lowell, 1985). This probability is confirmed by Martineau and Corbeil (1983) and by Chauvin *et al.*, (1985), who described the Saint-Antonin and Saint-Jean-Port-Joli moraines as being built at the northern margin of a remnant Appalachian ice mass. The Saint-Damien Moraine complex was formed during subsequent melting of this remnant glacier (Chauvin *et al.*, 1985).

RE-INTERPRETATION OF DEPOSITS ASSIGNED TO THE HIGHLAND FRONT MORAINE SYSTEM

The Highland Front Moraine System was described by Gadd *et al.* (1972b) as a "...complex of moraine features..." and of "...large, although separate, accumulations of ice-contact gravel and sand" (p. 12); the moraine system was considered as extending along the Appalachian front from the vicinity of Saint-Raphaël, about 40 km east of Québec, to that of Dunham, about 12 km north of the Vermont border (Fig. 6). Gadd *et al.* (1972b) thought that the morainic system formed at the margin of the main (Laurentide) ice sheet and that "general southeasterly flow maintained the ice in contact with the hills while the glacier surface was lowered by melting several hundred feet of thickness" (p. 12). The moraine was thought to have formed about 12.5 ka and to be a correlative of the Saint-Antonin Moraine of Lee (1962). Gadd *et al.* (1972b, p. 12) inferred that: "During this time, ice-marginal drainage grew from minor local drainages of isolated basins into an integrated drainage system connecting the Lake Champlain basin to the Gulf of St-Lawrence and releasing northward waters that were ponded in major valleys, such as the St-Francis and Chaudière, and that had previously drained eastward and southward". It was further inferred that "The wasting of ice in the St-Lawrence Lowlands *after* (our italics) construction of the moraine was the key control factor in the northward expansion of glacial lakes in the Lake Champlain basin and also during the inland migration of salt water of the Champlain Sea" (p. 12).

More detailed work by several authors has shown that the morainic system as defined and mapped by Gadd *et al.* (1972b) includes many diverse elements. Among these are:

— bedrock-controlled landscape features (map in Prichonnet, 1984a);
— diachronic morainic belts (Prichonnet, 1984a; Parent, 1987; Chauvin *et al.*, 1985; Fig. 6);
— eskers (Prichonnet, 1984a; Parent, 1987; Figs. 2, 6);
— early Champlain Sea deltas formed at the mouth of Appalachian rivers, such as the Danville delta (Parent, 1984c, Figs. 3, 4).

It is the opinion of the authors that grouping a variety of deposits and features that lie along the front of the Appalachian uplands as a single "morainic system" gives a misleading picture of deglaciation in southeastern Québec. Consequently we suggest that the terms Highland Front Moraine or Highland Front Moraine System be abandoned. As shown in Figures 6 and 8, ice-marginal features are oblique to the edge of the Appalachian uplands; the deposits formerly assigned to the Highland Front Moraine System are markedly time-transgressive. Moreover, these deposits were formed by meltwaters flowing mainly toward *south* (Figs. 5, 6), not toward northeast as an "integrated drainage system", as suggested by Gadd *et al.* (1972b). Glacial Lake Vermont, which drained southward via the Hudson Valley (Chapman, 1937), expanded northward from the Lake Champlain Valley into southeastern Québec *throughout*, not after, glacial retreat along the edge of the Appalachian uplands (see references in Appendix I). Morainic belts, such as the Ulverton-Tingwick and the Mont Ham (Parent, 1987; Fig. 2) and other belts (Lac Brome and Sutton) in the vicinity of Sutton and Granby (Prichonnet *et al.*, 1982a, 1982b), provide a much more accurate picture of deglacial events leading to the Champlain Sea incursion (Figs. 6, 8, 9) and hence should be used in place of the all encompassing Highland Front Moraine System.

The only major segment of the old Highland Front Moraine System that does not now have a different name is a morainic belt constructed by the Laurentide Ice Sheet near Saint-Sylvestre (Gadd, 1978). These moraines which are characterized by sediment containing much erratic material derived from the Canadian Shield lie in the area referred to as the type area for the "Highland Front Moraine System" (Gadd, 1964). A new name, Saint-Sylvestre Moraine, is proposed for these features (Fig. 6).

NORTHWARD ICE-FLOW AND REMNANT APPALACHIAN ICE

In the Notre-Dame Mountains and adjacent uplands east of the Chaudière Valley, northward-trending striations occur together with deglaciation features, including eskers and other ice-contact deposits, that also suggest northward meltwater flow (Chauvin *et al.*, 1985; LaSalle *et al.*, 1977a; Martineau and Corbeil, 1983; Fig. 6). In the Appalachian uplands of northern Maine, ice-flow reversal also occurred and was followed by isolation of remnant ice masses (Lowell, 1985).

In uplands west of the middle Chaudière Valley, northward- and westward-trending striations have led several authors to propose a late-glacial ice-flow reversal in this region (Lamarche, 1974; Lortie, 1976; Shilts, 1981). However, other deglacial features that are incompatible with an ice-flow reversal have been recorded throughout much of the area where the northward-trending striations have been mapped (Parent, 1987; Shilts, 1981). The main ones are:

1. Glacial lake sediments in the Chaudière Valley (Shilts, pers. comm., 1986), and in the Saint-Sylvestre (Gadd, 1978) and Asbestos regions (Parent, 1987) indicate that high-level lakes were impounded at the ice front in areas that should have been occupied by late ice.
2. The Warwick-Asbestos Esker which was obviously deposited by southward-flowing meltwaters extends well into the area of northward-trending striations (Parent, 1987; Fig. 5).
3. The Ulverton-Tingwick and Mont Ham moraines also extend into an area of north-pointing striations but were evidently built during northward retreat of the Laurentide Ice Sheet.
4. Stratigraphic data from the Asbestos-Valcourt region (Parent, 1984d, 1987) show that till deposited by westward- and northward-flowing ice underlies till deposited by southeastward-flowing ice (Lennoxville Till). This suggests that at least some of the northward- and westward-trending striations predate the last glacial maximum.
5. The common occurrence of an upper melt-out till in the Thetford Mines area (Chauvin, 1979) and the absence of recessional moraines (Parent, 1987) suggest that a remnant ice-mass which dissipated mainly through downwasting became isolated on the Bois-Francs uplands during deglaciation (Figs. 2, 6). Some northward-trending striations may have been produced by this Bois-Francs residual ice-mass, particularly along its northern edges. The small Glen Loyd esker which was deposited by northward-flowing meltwater (Lortie, 1976; Gadd, 1978) may have formed near the northern margin of the Bois-Francs residual ice-mass (Fig. 2).

Deglacial patterns that are now recognized in the Appalachians of southern Québec differ significantly from those recognized at the time when regional evidence for a northward ice-flow reversal was first reported (Lamarche, 1971, 1974). As shown above, ice margins retreated toward NNW in the southern part of the Appalachian uplands and piedmont while large residual ice masses dissipated, mainly through downwasting, in the northern part of the Appalachian uplands (Fig. 6). The above re-assessment of deglacial patterns places further constraints on attempts to reconstruct deglacial events leading to Champlain Sea incursion. For instance, glacial retreat north of the Saint-Sylvestre Moraine rather than northwest of the entire Highland Front Moraine System is now seen as one of the key events that allowed the incursion of marine waters in the main Champlain Sea basin. Similarly, a glacial readvance along the entire Highland Front Moraine System, as suggested by Shilts (1981), is no longer required in order to provide a coherent picture of deglacial events in southern Québec. This however does not exclude the possibility of a local readvance to the position of the Saint-Sylvestre Moraine; such a readvance would necessarily predate marine incursion in the main Champlain Sea basin.

GOLDTHWAIT SEA AND CHARLESBOURG PHASE OF CHAMPLAIN SEA

A key question is: when and how did the Champlain Sea basin open to oceanic waters? Analysis of glacial and marine data seems to point to progressive expansion of the west arm of the Goldthwait Sea up the St. Lawrence to the vicinity of Québec, between about 12.8 and 12.4 ka, with the first or Charlesbourg Phase of the Champlain Sea, forming in the Québec region, about 12.4 ka. At this same time, lakes were developing and extending northward at the southern margin of ice that occupied the main Champlain Sea basin, but a barrier created by the Saint-Maurice lobe (Occhietti, 1980, p. 146) on the north shore and by an ice remnant in the Bois-Francs region on the south separated the marine from the lacustrine part of the basin until about 12 ka.

WESTERN ARM OF THE GOLDTHWAIT SEA (12.8 to 12.4 ka)

Ice-flow events in the middle estuary of the St. Lawrence River between Québec and the mouth of the Saguenay are not fully documented. Northeast-trending glacial striations in lowlands of the south shore indicate that the latest ice-flow event was parallel to the axis of the estuary (Dionne, 1972). On the north shore, east-west striations suggest a late, local shift of ice-flow direction (Rondot, 1974; É. Govare, Université de Montréal, pers. comm., 1986). Progressive marine invasion of the estuary by a calving bay was proposed by Gauthier (1975), Gadd (1976), Shilts (1976), LaSalle *et al.* (1977b) and Thomas (1977) and the concept is now commonly accepted. There is also general agreement as to the existence of late-glacial northward ice flow, from the Appalachians toward the middle estuary (Shilts, 1976, 1981; Martineau and Corbeil, 1983; Chauvin *et al.*, 1985). It appears that the calving bay progressed some 200 km between 12.7 and 12.4 ka, according to an age of 12 720 ± 170 BP (GSC-102), obtained on *Portlandia arctica* shells collected near Trois-Pistoles, at the edge of the Saint-Antonin Moraine (Lee, 1962), and the 12.4 ka age at Charlesbourg (Table III). It is quite possible, however, that the marine invasion in the estuary began earlier as an age of 13 390 ± 690 BP (QU-271) on *Hiatella arctica* obtained from a marine-limit delta at Saint-Flavien, 35 km downstream

TABLE III

^{14}C ages of shells in the eastern and central basin of Champlain Sea. Locations of group 1 samples are shown in Figure 10. Locations of most of the samples of groups 2 and 3 are shown in Figure 2

Laboratory number	Age (years BP)	Altitude (m a.s.l.)	Locality	Dated material	Sediment	Collector	Reference
1. Early Champlain Sea fauna, southeastern part of central basin							
I-13342	11 700 ± 170	127	Warwick, Qc	*Hiatella arctica* **Balanus crenatus*	Reworked esker gravel and silty sand	Parent	Parent, 1984d
GSC-187	11 410 ± 150	122	Kingsey Falls, Qc	*Macoma calcarea* *Mya truncata* *Hiatella arctica*	Stratified silty sand and clay	Gadd	Dyck *et al.* 1965
**UQ-290	11 370 ± 200	149	Danville, Qc	*Macoma balthica*	Deltaic sand	Parent	Parent, 1984c
GSC-505 GSC-936	11 880 ± 180 12 000 ± 230	122	L'Avenir, Qc	*Macoma balthica* (90%) *Hiatella arctica* (8%) *Mya* sp. (1%) *Yoldia* sp. (1%)	Pebbly gravel and sand	McDonald	Lowdon and Blake, 1970
GSC-475-2 GSC-475	11 500 ± 160 11 530 ± 160	145	Ste-Christine, Qc	*Hiatella arctica* (mostly) *Macoma* sp. (fragments) *Yoldia* sp. (fragments) *Mytilus edulis* (fragments)	Silt	McDonald	Lowdon and Blake, 1970
UQ-29	11 360 ± 110	105	Adamsville, Qc	*Macoma balthica*	Sand	Prichonnet	Prichonnet, 1982a, 1984
I-4489	11 740 ± 200	145	Frelighsburg, Qc	*Macoma balthica*	Lens of sand and clay in deltaic material	?	Parrott and Stone, 1972 Wagner, 1972
QC-200	11 665 ± 175	79 (95?)	Plattsburg, NY	*Macoma balthica*	Unspecified	Cronin	Cronin, 1977
GSC-2338	11 900 ± 120	101	Peru, NY	*Macoma balthica*	Pebbly sand	Cronin	Lowdon and Blake, 1979
GSC-2366	11 800 ± 150	96	Plattsburg, NY	*Macoma balthica*	Silty sand	Cronin	Lowdon and Blake, 1979
QC-475	*12 480 ± 240*	*61*	*St-Dominique, Qc*	*Mya sp.*	*Unspecified*	*Prichonnet*	*Prichonnet, 1982a*
UQ-1429	11 250 ± 100	63	St-Dominique, Qc	*Mya arenaria*	Sand and silt	Occhietti	Unpublished
GrN-1697	11 490 ± 110	171	Mont Royal, Qc	Unidentified shells	Unspecified	Elson	Elson, 1962, 1969a
***W-2311	11 420 ± 350	58	Burlington, Vt	Unidentified pelecypods	Silt and clay	Hunt and Wagner	Wagner, 1972 Spiker *et al.*, 1978

Laboratory number	Age (years BP)	Altitude (m a.s.l.)	Locality	Dated material	Sediment	Collector	Reference
2. Champlain Sea fauna, Québec City region							
GSC-1533	12 400 ± 160	110	Charlesbourg, Qc	*Portlandia arctica*	Clayey sand	LaSalle	Lowdon and Blake, 1973
QU-93	12 230 ± 250	104	St-Henri-de-Lévis, Qc	*Hiatella arctica*	Interbedded gravel and clay	LaSalle	Samson *et al.*, 1977
GSC-1235	11 600 ± 160	176	N.-D.-des-Laurentides, Qc	*Mya truncata*	Glaciomarine sediment	LaSalle	Lowdon and Blake, 1976
UQ-39	10 890 ± 125	~44	St-Nicolas, Qc	*Balanus hameri*	Stony clay and silt	Occhietti	Unpublished
GSC-1451	10 000 ± 150	~48	St-Nicolas, Qc	*Hemithyris psittacea*	Regressive sands	LaSalle	Lowdon and Blake, 1979
UQ-64	9 355 ± 115	~48	St-Nicolas, Qc	*Hiatella arctica*	Regressive sands	Occhietti	Unpublished
UQ-40	11 340 ± 180	~57	Pointe St-Nicolas, Qc	*Balanus hameri*	Deep-water reworked sediment	Occhietti	Unpublished
GSC-1476	11 200 ± 170	~55	Pointe St-Nicolas, Qc	*Balanus hameri*	Deep-water reworked sediment	Occhietti	Unpublished
UQ-206	9 460 ± 190	~50	Pointe St-Nicolas, Qc	*Hiatella arctica*	Shallow-water reworked sediment	Occhietti	Unpublished
QU-447	10 390 ± 130	~90	Pont-Rouge, Qc	*Hiatella arctica* **Mytilus edulis* **Balanus* sp.	Shallow-water reworked sediment	LaSalle	LaSalle *et al.*, 1977a
UQ-1287	9 900 ± 100	38	Ste-Foy, Qc	*Hiatella arctica*	Regressive sands	Occhietti	Unpublished
UQ-651	11 130 ± 180	35	Deschaillons, Qc	*Balanus hameri*	Massive pebbly clay	Lamothe	Lamothe, 1985
QU-94	10 700 ± 200	84	Ste-Croix-de-Lotbinière, Qc	*Hiatella arctica* *Macoma balthica*	Shallow-water reworked sediment	LaSalle	Samson *et al.*, 1977
GSC-1295	11 200 ± 160	67	Lapointe, Qc	*Balanus* cf *hameri*	Diamicton (glaciomarine?)	LaSalle	Lowdon and Blake, 1976
GSC-1232	11 100 ± 160	106	Beauport, Qc	*Balanus* sp.	Diamicton (glaciomarine?)	LaSalle	Lowdon and Blake, 1976
QU-97	10 650 ± 240	58	Beauport, Qc	*Hiatella arctica*	Littoral sediment	LaSalle	Samson *et al.*, 1977
QU-20	11 260 ± 290	70	Pointe St-Nicolas, Qc	*Mya truncata*	Shallow-water reworked sediment	LaSalle	Samson *et al.*, 1977
QU-98	11 120 ± 220	60	Pointe St-Nicolas, Qc	*Balanus hameri*	Gravel (ice-contact?)	LaSalle	Samson *et al.*, 1977
3. Dates pertaining to marine regression, Asbestos-Valcourt region							
****UQ-289	10 780 ± 190	132	Warwick, Qc	*Hiatella arctica* *Macoma balthica* **Balanus crenatus* **Mytilus edulis* **Mya pseudo-arenaria*	Reworked esker gravel and sand	Parent	Parent, 1987
UQ-947	10 350 ± 100	81	St-Wenceslas, Qc	*Mya arenaria* **Macoma balthica*	Littoral sand	Parent	Parent, 1987
GrN-2034	10 590 ± 100	81	Duncan, Qc	*Mya arenaria* **Mytilus edulis* **Macoma balthica*	Littoral sand	Elson	Elson, 1969b
4. Champlain Sea fauna, Trois-Rivières-Shawinigan region							
GSC-1526	11 600 ± 630	91	Riv. Batiscan, Qc	Foraminifera	Glaciomarine sediment	Gadd	Lowdon and Blake, 1975
GSC-1729	11 300 ± 160	81	Riv. La Fourche, Qc	*Portlandia arctica*	Glaciomarine sediment	Occhietti	Occhietti, 1976
GSC-2045	11 100 ± 90	11	La Gabelle, Qc	*Portlandia arctica*	Glaciomarine sediment	Occhietti	Occhietti, 1976
I-9484	10 910 ± 160	60	St-Louis-de-France, Qc	*Hiatella arctica*	Cobbly layer in laminated clay	Occhietti	Occhietti, 1980
GSC-2090	10 600 ± 160	71	St-Alban, Qc	*Balanus hameri*	Pebbly silt	Occhietti	Occhietti, 1980
GSC-2101	10 300 ± 100	129	Shawinigan, Qc	*Mya arenaria*	Littoral silt	Occhietti	Occhietti, 1980
GSC-2150	10 200 ± 90	78	St-Alban, Qc	*Macoma calcarea*	Inundation silt	Occhietti	Occhietti, 1980
GSC-1700	10 200 ± 160	128	Charette, Qc	*Macoma balthica*	Reworked sand on St-Narcisse Moraine	Occhietti	Occhietti, 1980
GSC-1444	10 100 ± 150	137	Charette, Qc	*Hiatella arctica* *Macoma balthica*	Reworked sand on St-Narcisse Moraine	Gadd	Lowdon and Blake, 1975
GSC-1739	10 000 ± 150	98	Riv. La Fourche, Qc	*Hiatella arctica*	Inundation silt	Occhietti	Occhietti, 1980

* Other species present, but not submitted for dating.
** This date relates to a regressive sea level stand at 153 m a.s.l.
*** Wood from the same silt and clay unit was dated 10 950 ± 300 years BP (W-2309).
**** This date relates to a regressive sea level stand at 136 m a.s.l.

GrN: Groningen
GSC: Geological Survey of Canada
I: Teledyne Isotope
QC: Queens College
QU: Min. de l'Énergie et des Ressources du Québec
UQ: Université du Québec à Montréal (GEOTOP)
W: United States Geological Survey

from Trois-Pistoles, may suggest (Locat, 1977). The rate of ice retreat or of calving bay extension would be either 650 m per year or 235 m per year, depending on which age is used.

CHARLESBOURG PHASE: EASTERN SECTOR OF THE CHAMPLAIN SEA

A new name, Charlesbourg Phase or Champlain Sea phase I, is proposed for an early phase of Champlain Sea invasion that was restricted to the Québec region. The Charlesbourg Phase began around 12.4 ka and terminated when marine waters invaded the main Champlain Sea basin, around 12 ka.

The existence of this phase is based mainly on two ^{14}C-dated sites. Near Saint-Henri-de-Lévis, on the south shore of the St. Lawrence River, minimum age for deglaciation is given by an age of 12 230 ± 250 BP (QU-93) on *Hiatella arctica* in ice-marginal sediments at 104 m a.s.l. (LaSalle, 1974). On the north shore, the age for the Charlesbourg site (12 400 ± 160 BP, GSC-1533) was obtained on *Balanus hameri*, a species which commonly lives at depths of 90 m. This date may, much as other dates on deep-water species from the Champlain or Goldthwait seas, be affected by apparent aging caused by slow exchanges between oceanic and atmospheric CO_2 reservoirs (Mangerud and Gulliksen, 1975; Hillaire-Marcel, 1977, 1981).

The areal extent of the Champlain Sea during the Charlesbourg Phase cannot be defined at this time. For instance, marine limit on the Appalachian piedmont near Saint-Sylvestre, west of the Rivière Chaudière, may be as high as 195 m, if the ice-contact deposits were reworked by marine action as suggested by Gadd (1978), but this cannot be proven because the deposits lack fossils. At the foot of the Laurentian Highlands, on the north shore, we know that marine limit lies at about 200 m from the work of LaSalle (1974) but because this limit has not yet been dated, we do not know whether or not this area was ice-free during the Charlesbourg Phase.

Available data seem to fit a plausible model of deglaciation. Just prior to opening of the area to seawater, a glacial lake apparently developed in the middle and lower Chaudière Valley as Laurentide ice retreated northward (LaSalle *et al.*, 1977*a*; Shilts, pers. comm. 1986). We suggest calling this water body "Lake Beauce" (*Lac de la Beauce*). Its southern limit as well as its possible relationships with a glacial lake in the upper Chaudière Valley (Shilts, 1981; Parent *et al.*, 1985), called "glacial Lake Mégantic" for convenience (Fig. 8), are not known. The existence of Lake Beauce is supported by the presence of varves underlying marine clays (LaSalle *et al.*, 1977a; LaSalle, 1984), and the WSW-ENE trend of subsequent ice-retreat positions of the Laurentide Ice Sheet. The age of Lake Beauce is not known nor are its temporal relationships with Appalachian ice masses.

The opening of the Charlesbourg basin to seawater and the drainage of Lake Beauce possibly occurred around 12.4 ka but no later than about 12.2 ka. Drainage of Lake Beauce marks the beginning of the Charlesbourg Phase, which was restricted to the easternmost margin of the marine basin. The ice front subsequently retreated northward to the vicinity of Québec (Fig. 9). The extent of the Charlesbourg Phase around 12 ka is not known. Based on the dates of 11.6 ka at Notre-Dame-des-Laurentides, 11.1 ka at Beauport and 11.2 ka at Lapointe (Table III, Fig. 2), it is possible that the edge of the Laurentian Highlands was not deglaciated during the Charlesbourg Phase.

LACUSTRINE PHASES PRIOR TO THE MARINE INVASION OF THE MAIN BASIN

When and how marine invasion occurred in the main basin of the Champlain Sea have been the subject of various interpretations (Antevs, 1925; McDonald, 1968, Prest, 1970; Gadd, 1983; Parent *et al.*, 1985). In light of the work of McDonald (1967) and Parent (1987) and other recently published data, in particular by Prichonnet (1982a, 1982b, 1984a, 1984b), Clark and Karrow (1984) and Anderson *et al.* (1985), the hypothesis of a late marine invasion, around 12 ka, now seems most plausible in the paleoenvironmental context of the St. Lawrence Valley. The invasion was preceded by lacustrine phases, the last two of which (Figs. 8, 9) occurred over a very short time span, probably less than 400 or 500 years.

STRANDLINE FEATURES OF THE FORT ANN LEVEL PHASE OF GLACIAL LAKE VERMONT

Glaciolacustrine shoreline features at elevations well above the marine limit were first recorded by McDonald (1967, 1968) in the vicinity of Richmond. Further work by Parent (1987), mainly in areas between Valcourt and Warwick, has not only revealed that this proglacial lake was coeval with formation of the Ulverton-Tingwick Moraine but it also indicates that this water body represents a northeastern extension of glacial Lake Vermont (Fort Ann phase).

Near the edge of the Appalachian uplands, distinctive strandline features of the former water body reach altitudes on the order of 230 m (Fig. 7 and Appendix I) but their altitude decreases southeastward at a rate of about 1 m per kilometre (Parent, 1987). In the vicinity of Sherbrooke, features on the warped strandline now lie at about 190 m a.s.l. (Parent, 1987). The reconstructed water plane in the Saint-François Valley differs somewhat from that suggested by McDonald (1967, 1968) for his "lower glacial-lake system". Because the number of sites available to him was quite limited, he was unable to separate strandline features of a transitional glacial lake of local extent from a series of features which lie about 20 m lower. Features numbered 1 to 30 in Figure 7 belong to this lower water body. At the edge of the uplands, glacial lake shoreline features lie 50 to 60 m above shoreline features used to define marine limit. Because of their separation from Champlain Sea features and the fact that the lake was present during ice retreat, these strandline features provide an excellent regional morphostratigraphic marker which constrains regional deglaciation models. Large remnant ice masses could not have been present when and where shoreline features were forming nor where varves were being deposited.

Correlation of Late Wisconsinan water bodies in the middle Saint-François Valley with those in Lake Champlain Valley is seen as a very distinct possibility, because the altitude

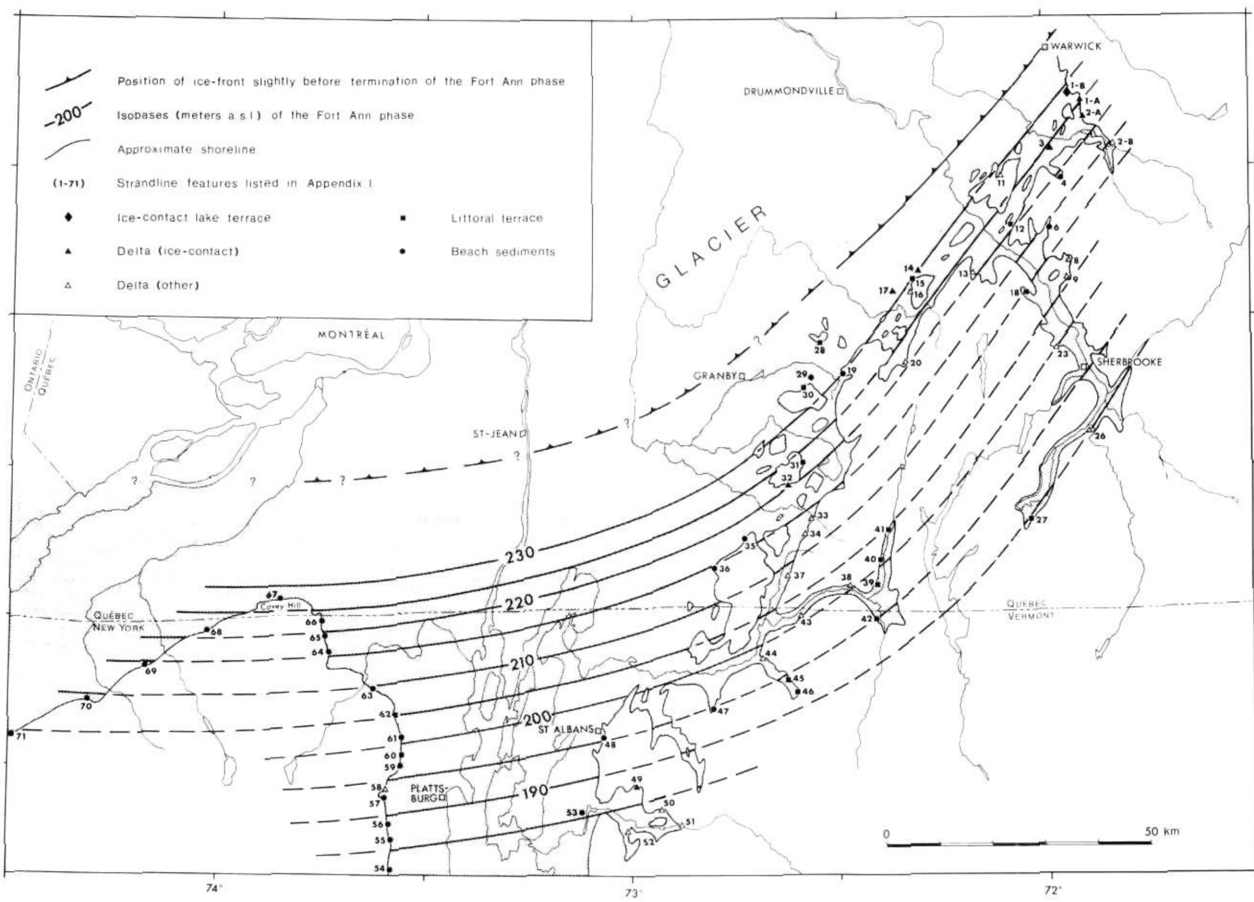


FIGURE 7. Fort Ann level phase of glacial Lake Vermont. Isobases on the former water plane indicate an average tilting of 1 m/km. Lake level fell when low terrains northeast of Warwick became ice-free (after Parent, 1987).

Phase du niveau Fort Ann du Lac glaciaire Vermont. Les isobases de l'ancien plan d'eau indiquent une inflexion géoïdale moyenne d'environ 1 m/km. Le niveau du lac s'est abaissé lorsque les terrains au nord-est de Warwick ont été déglacés (de Parent, 1987).

difference between the latest glacial lacustrine shoreline features and upper marine shoreline features is on the order of 60 m. In order to determine whether glacial lake strandline features of the middle Saint-François Valley correspond to the Upper or to the Lower Fort Ann water plane of Clark and Karrow (1984), strandline features recorded in several published sources were compiled for the area extending between Granby (Québec) and St. Albans (Vermont), east and northeast of Lake Champlain (Appendices I and II). Features numbered #54 to 71 in Figure 7 belong to the well established Fort Ann strandline in New York (Chapman, 1937; Denny, 1974; Clark and Karrow, 1984) and those numbered #43 to 53 have been assigned to this phase in Vermont (Parrot and Stone, 1972; Wagner, 1972). Features that are located upstream of narrow valleys, however, were deleted, since they may have formed in small, local lake basins. In Québec, because most of the recorded shoreline features have not been assigned to a given phase of glacial Lake Vermont by earlier authors, only features which could also be identified on air photographs have been included in the reconstruction (#29 to #41).

Since there is general agreement (references in Appendix I) that glacial Lake Vermont expanded northward and northeastward into southern Québec as the Laurentide ice margin retreated, the Fort Ann strandline is expectedly somewhat diachronous over the area shown in Figure 7. As a consequence, a perfect fit of shoreline features with reconstructed isobases was neither expected nor found. The best fit was obtained by correlating shoreline features of southeastern Québec with Fort Ann features recorded by Chapman (1937) in New York and these have been correlated by Clark and Karrow (1984) with their Upper Fort Ann level. A number of Fort Ann Phase features, particularly in Vermont, lie 10 to 15 m below the reconstructed water-plane (Appendix I). These features possibly formed during the Lower Fort Ann phase of Clark and Karrow (1984).

DEGLACIATION OF THE SOUTHERN APPALACHIANS OF QUÉBEC AND FORT ANN PHASE OF LAKE VERMONT

The Fort Ann phase of Lake Vermont provides a means of correlating deglacial events in the southern Appalachians of Québec with events in the St. Lawrence Lowland. Around 12.5 ka, Lake Ontario basin was inundated by Lake Iroquois (MacClintock and Stewart, 1965; Denny, 1974; Fullerton, 1980; Clark and Karrow, 1984). High level of Lake Iroquois extended into the St. Lawrence Valley until the margin of the Laurentide

Ice Sheet had retreated from highlands south of Covey Hill (Clark and Karrow, 1984). Retreat of the glacial lobe that occupied the Lake Champlain Valley allowed waters of glacial Lake Vermont to expand northward (Chapman, 1937) and when ice retreated from the Covey Hill heights, lakes Iroquois and Vermont became confluent.

Several end-moraines were formed during ice retreat in the Appalachian uplands. General trend of these moraines is WSW-ENE, with distinct lobes in the main valleys. Ice-flow toward SSE was maintained during construction of the Cherry River-East Angus Moraine (McDonald, 1967; Parent, 1987) in the Sherbrooke area and glacial lakes formed in the valleys south of the retreating ice-front. The main lakes were glacial Lake Memphrémagog (McDonald, 1968; Clément and Parent, 1977; Boissonnault and Gwyn, 1983) and glacial Lake Mégantic (Fig. 8; see also Shilts, 1981). During formation of the Mont Ham Moraine, glacial Lake Memphrémagog reached its maximum extent, known as the Sherbrooke Phase (McDonald, 1968), and its waters overflowed into Lake Vermont through the Lac Nick spillway (Parent *et al.*, 1985; Fig. 8). Strandline features of the Sherbrooke Phase are tilted upward to the northwest, from 240 m a.s.l. near Sherbrooke to 270 m a.s.l.

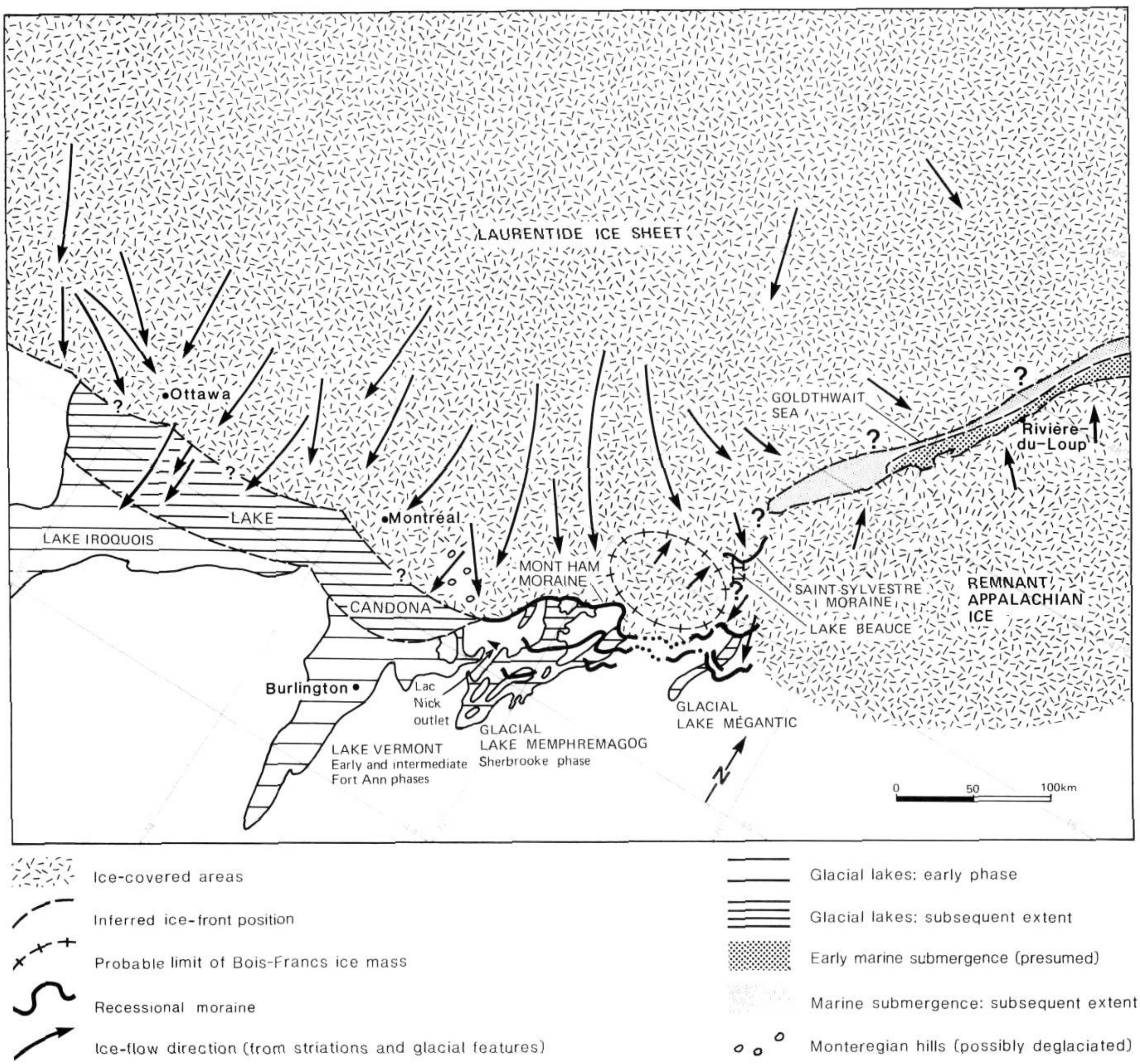


FIGURE 8. Proposed paleogeography of St. Lawrence Valley during the early and intermediate stages of the Fort Ann level phase of glacial Lake Vermont and early Lake Candona. Undifferenciated episodes of the Goldthwait Sea into the St. Lawrence middle estuary seem more or less coeval with the development of these glacial lakes.

Essai de reconstitution paléogéographique de la vallée du Saint-Laurent pendant les phases précoces et intermédiaires du niveau Fort Ann du Lac glaciaire Vermont et le début du Lac Candona. Il est probable que la Mer de Goldthwait a progressé dans le moyen estuaire du Saint-Laurent au cours d'épisodes non différentiables et plus ou moins synchrones à ces épisodes lacustres.

at points along the Mont Ham Moraine (Parent, 1984b). Given the position and trend of the Mont Ham Moraine (Fig. 2), it is probable that the ice front north of Lake Champlain Valley had already retreated to near Covey Hill at this time. It may thus be concluded that the Mont Ham Moraine and consequently, the end of the Sherbrooke phase of Lake Memphrémagog, are nearly contemporaneous with the coalescence of lakes Iroquois and Vermont (Fort Ann phase), as defined by Clark and Karrow (1984).

Glacial lake levels in the Saint-François Valley fell during ice retreat that followed construction of the Mont Ham Moraine and stabilized at the level of the Fort Ann phase of Lake Vermont during deposition of the Ulverton-Tingwick Moraine. This is shown by the presence of several ice-contact deltas at points along the moraine, and by beaches and bluffs south of the moraine that are considered to correlate with Fort Ann phase water levels (Parent, 1987). Glaciolacustrine shorelines at the mouth of the middle Saint-François Valley lie some 60 m above the upper marine limit. This correlation is made by extending isobases (average tilting 1 m/km), from the Asbestos and Granby regions towards the southwest, where they appear to connect with those constructed by Chapman (1937) for the Lake Champlain Valley. The paleogeographic reconstruction shown in Figure 7 thus corresponds to the upper Fort Ann level of Clark and Karrow (1984).

In the Saint-François Valley, there are a few ill-defined littoral deposits, which lie about 10 m below the main Fort Ann phase level. These may correspond to the lower Fort Ann level described by Clark and Karrow (1984). Correlative glacial lake shorelines are possibly present in the Granby area (Prichonnet, 1982a, 1982b). The narrowness of bays and arms on the eastern shore of Lake Vermont and the short duration of the lake in southern Québec, provide an explanation for the less pronounced development of littoral forms in these areas in comparison with those observed by Clark and Karrow (1984) on the northwest slope of the Adirondacks.

LAKE CANDONA

A glacial lake, formed by coalescence of Lake Vermont and Lake Iroquois, extended northward as the ice front retreated from Covey Hill. We propose naming this water body Lake Candona. The name comes from the common occurrence of *Candona* ostracodes, specifically *Candona subtriangulata*, in sediments deposited in this lake. Lake Candona encompasses the "lower" (*i.e.* later) Fort Ann phase of the Iroquois-Vermont Lake (Clark and Karrow, 1984), the Belleville phase of Lake Iroquois in the upper St. Lawrence valley (Muller and Prest, 1985), the glacial lacustrine phase of the Ottawa Valley lowlands (Anderson *et al.*, 1985) and a lacustrine arm that extended north along the front of the Appalachians of Québec (Fig. 9). Lacustrine sediments of the Chambly Formation (LaSalle, 1981), which were deposited at altitudes (<40 m a.s.l.) well below those of previously recognized Fort Ann shoreline features in this region (Denny, 1974; Parrott and Stone, 1972), were probably deposited in this lake. The presence of clays containing *Candona subtriangulata* below marine clays in the lowlands between the St. Lawrence and Ottawa rivers suggests that the ice front may have retreated almost as far north as Ottawa during the life of this freshwater body (Anderson *et al.*, 1985). In the Appalachian piedmont, ice retreat that followed deposition of the Ulverton-Tingwick Moraine allowed glaciolacustrine waters to expand northeastward so the Danville varves, containing *Candona subtriangulata*, could be deposited at Rivière Landry (Parent, 1987; Fig. 3).

Fossiliferous Lake Candona sediments are an important marker unit below Champlain Sea sediments. Laminated (or varved) sediments are exposed below Champlain Sea sediments at many sites. However, only where these materials contain freshwater fossils can it unequivocally be stated that a glacial lake occupied the Champlain Sea basin prior to marine invasion.

CHAMPLAIN SEA, PHASE II

The phase of development of the Champlain Sea that followed the initial Charlesbourg Phase is here referred to as Phase II. It encompasses the history of the Champlain Sea from the time it broke through the ice barrier just upstream from Québec (12 ka) until retreat of ice from the Saint-Narcisse Moraine (10.8 ka). During this period, marine waters replaced Lake Candona in the main basin of the Champlain Sea and the Laurentide Ice Sheet withdrew from the St. Lawrence Lowland.

In the area of interest (Fig. 2), the onset of the marine invasion in the main Champlain Sea basin is defined stratigraphically at Rivière Landry (Parent, 1987). Along the Appalachian piedmont, between Valcourt and Warwick, water levels fell from 230 m a.s.l. (Fort Ann level) to about 170 m a.s.l. as a result of opening to marine waters. Similar drops of water-levels were also recognized further southwest along the Appalachian piedmont (Prichonnet, 1982a, 1982b, 1984a; McDonald, 1968) and at the foot of the Adirondacks (MacClintock and Terasmae, 1960; Chapman, 1937; Denny, 1974; Clark and Karrow, 1984). The microfaunal record at Rivière Landry (Parent, 1987) and the sudden fall of water levels suggest that glaciolacustrine waters were replaced by marine waters almost instantaneously, around 12 ka, in the Appalachian piedmont, southwest of Victoriaville. This invasion occurred shortly before fossiliferous sediments at L'Avenir, found at an altitude of approximately 120 m and dated 12 000 ± 230 BP (GSC-936) and 11 880 ± 180 BP (GSC-505) (McDonald, 1968; Gadd *et al.*, 1972b), were deposited. At Warwick, at an altitude of 117 m, a slightly reworked community of *Hiatella arctica* and *Balanus crenatus*, dated 11 700 ± 170 BP (I-13 342; Parent, 1984a), provides an age for early deep-water sedimentation. Deltaic sediments deposited at the marine limit (145 m a.s.l.) near Frelighsburg contain shells of *Hiatella arctica*, dated 11 740 ± 200 BP (I-4489; Wagner, 1972; Parrott and Stone, 1972; see Fig. 10 and Table III). To the authors' knowledge, these four ages are the oldest ones for faunal assemblages recording early occurrence of *Hiatella arctica* in the main Champlain Sea basin.

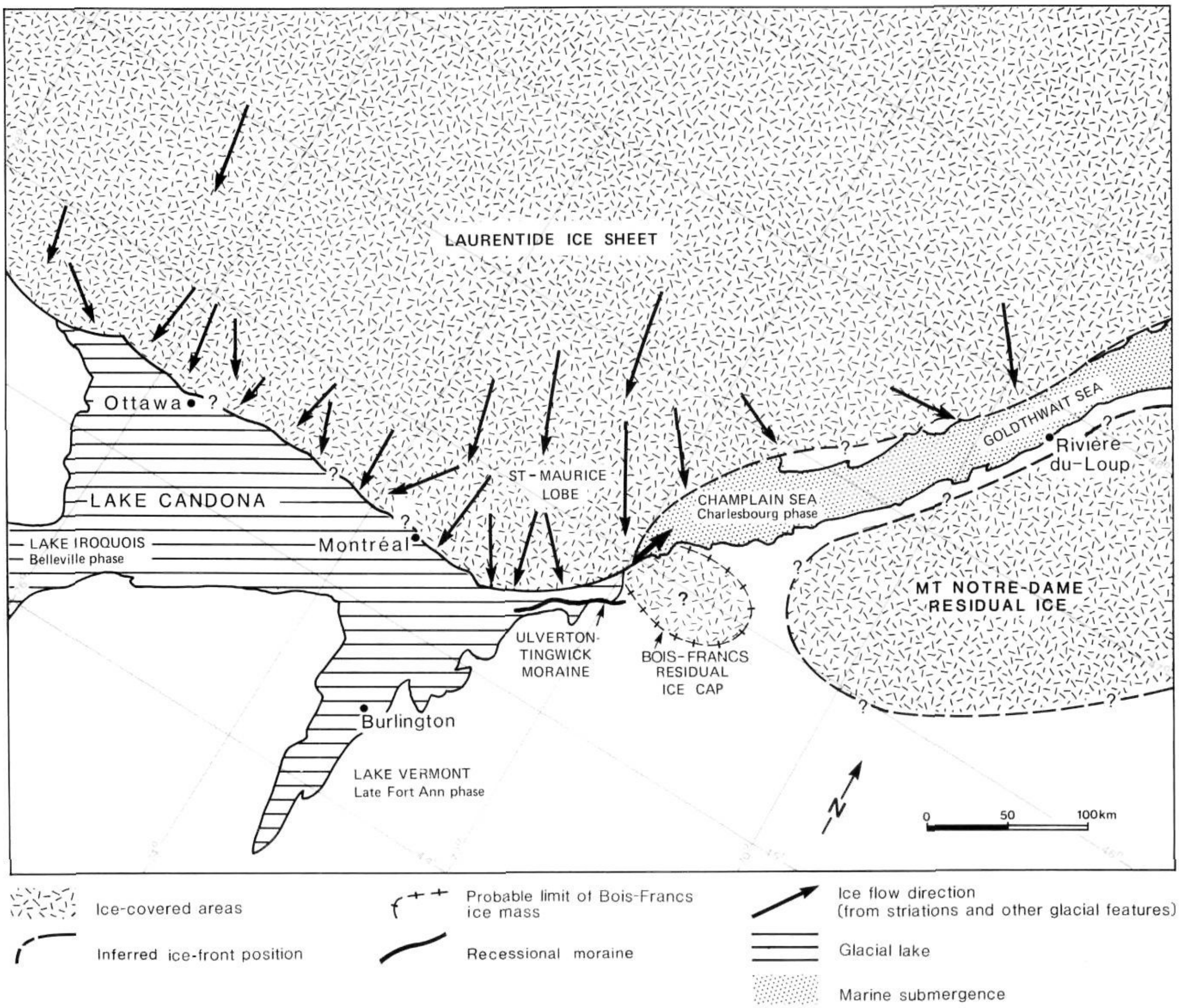


FIGURE 9. Proposed paleogeography of St. Lawrence valley immediately before marine invasion of the main Champlain Sea basin, and approximate maximum extent of Lake Candona.

Essai de reconstitution paléogéographique de la vallée du Saint-Laurent immédiatement avant l'invasion du bassin principal par la Mer de Champlain, c'est-à-dire avant le début de la phase II, et extension approximative du Lac Candona.

Dating the onset of the marine invasion around 12 ka is apparently contradicted by three age measurements:

1. An age of 12 480 ± 240 BP (QC-475) on *Mya* sp. has been obtained from a site near Saint-Dominique (Figs. 2, 10; Prichonnet, 1982a). The dated beds are now submerged by groundwater and the dated specimens were not identified to species level (*Mya arenaria? Mya pseudoarenaria*?). Little information is available on the nature (biocenosis/thanatocenosis) and sedimentary facies of the deposit. Such an early age for a *Mya* assemblage appears somewhat surprising (Parent *et al.*, 1985) and a visit to the site revealed that the species is *Mya arenaria* and that it occurred in a 5 m-thick sequence of reworked sand and silt, at an elevation of about 60 m. A new ^{14}C age of 11 250 ± 100 BP (UQ-1429) contradicts the earlier date.
2. Basal gyttja in a core from Mont Saint-Bruno was dated 13 000 ± 290 BP, (GSC-1344; Terasmae and LaSalle, 1968). The coring site, at an altitude of 125 m, definitely lies below marine limit. Lack of reworking of the core material indicates that lacustrine deposition took place only after marine water levels had receded below 125 m a.s.l., thus suggesting that this age is too old by several hundred years. An age of 11 400 ± 340 BP (I-8841) on organic debris at the base of a core from Mont Shefford (P. Richard, 1978), at an altitude of 282 m, well above marine limit, confirms that GSC-1344 probably is not an accurate age for the sediments.

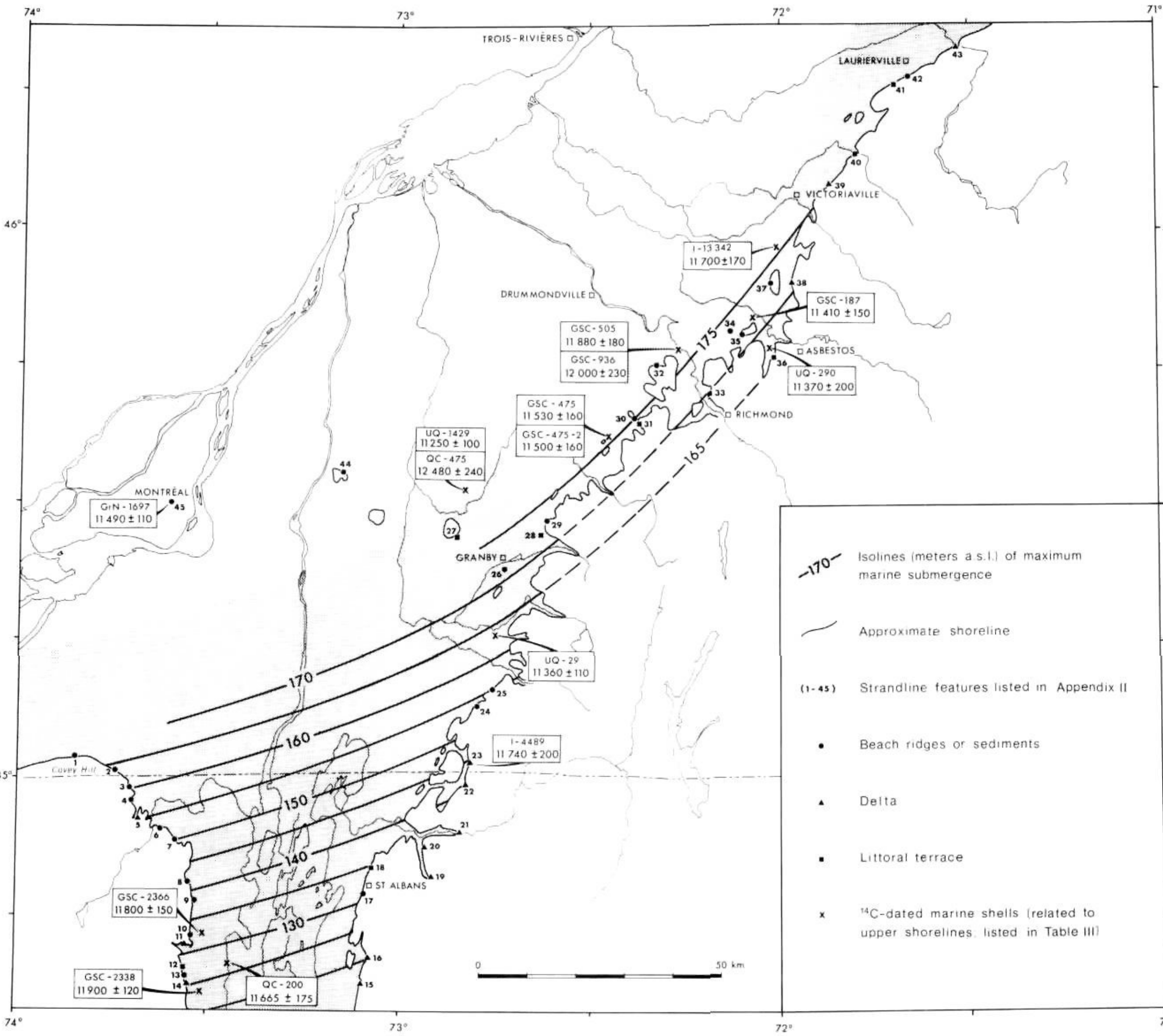


FIGURE 10. Early Champlain Sea submergence in the main basin (Phase II). Isolines on upper marine features suggest a tilting of about 0.9 m/km. Apparent ^{14}C age of early marine faunas (*Hiatella arctica* and *Macoma balthica*), several of which were obtained from regressive sediments, averages 11.7 ka; QC-475 and UQ-1429 are from *Mya* sp. assemblages (after Parent, 1987).

*Début de l'invasion de la Mer de Champlain dans le bassin principal (Phase II). Les isolignes de la limite marine supérieure indiquent une inflexion géoïdale de l'ordre de 0,9 m/km. L'âge moyen apparent des premières faunes marines (*Hiatella arctica *et* Macoma balthica*) datées au ^{14}C est de 11,7 ka. Certaines des coquilles datées proviennent de sédiments régressifs. Les datations QC-475 et UQ-1429 ont été obtenues sur des assemblages de* Mya sp. *(de Parent, 1987).*

3. Gadd (1980) used a radiocarbon age of 12.8 ka (GSC-2151: 12 800 ± 100 BP, outer fraction, and 12 700 ± 100 BP on the inner fraction) as part of his argument for a calving bay carrying the Champlain Sea into the Ottawa area very early during deglaciation. This age exceeds by 500 to 600 years other early dates from the western Champlain Sea basin (Fulton and Richard, 1987). GSC-2151 on *Macoma balthica* is contradicted by an age of 11 400 ± 90 BP (GSC-2269; Harington and Occhietti, 1988) on bone collagen from a finback whale (*Balaena mysticetus*) found at the same altitude in a nearby site. Since dates obtained on collagen from mammal bones have been found to be internally consistent (Harington and Occhietti, 1988), this shell date seems strongly suspect. Moreover, an accelerator date, subsequently run on the same material as that used for GSC-2151, gave an age of 12 180 ± 90 BP (TO-245; Fulton and Richard, 1987).

In summary, ^{14}C dates that may have suggested marine invasion into the main basin of the Champlain Sea prior to 12 ka now appear unreliable. Moreover, several studies (Prest and Hode-Keyser, 1977; Gadd, 1971; Prichonnet, 1984a; Occhietti, 1980; Parent, 1987) have reported no stratigraphic data to support the early invasion hypothesis. As a consequence, the paleogeographic sequence seems clear: Lake Vermont and its successor, Lake Candona, directly precede the marine invasion which occurred about 12 ka. This interpretation, which resembles an earlier suggestion by McDonald (1967, 1968), is proven by an excellent stratigraphic record, the Rivière Landry section (Parent *et al.*, 1986; Parent 1987).

Additional confirmation that invasion of the main basin and commencement of Phase II of the Champlain Sea occurred about 12 ka is supplied by clustering of ages from early migrating mollusks (Fig. 10; Table III). With one exception theses ages were obtained on *Hiatella arctica* and *Macoma balthica* specimens. These species are believed to have migrated fairly early into the Champlain Sea basin (Hillaire-Marcel, 1979; 1980). The dated faunas were found at altitudes close to the marine limit and although many of the dated faunas are from regressive facies (*e.g.* UQ-290 and UQ-29, Table III), their average age is 11.7 ka, which obviously is only the minimum age for the marine invasion.

The location and extent of the glacial barrier that stood between Lake Candona and waters of the Charlesbourg Phase just prior to Phase II of Champlain Sea (see Fig. 9) are known only from indirect evidence. The last barrier was conceivably located in the Appalachian piedmont where it probably extended from the southwest end of the Saint-Sylvestre Moraine, near Lyster, to the vicinity of Victoriaville. While glacial retreat toward NW and NNW, and coeval development of glacial lakes are distinctly recorded in terrains located to the northeast (Saint-Sylvestre area and lower Chaudière Valley) and to the southwest (Asbestos-Valcourt region) of the inferred barrier, no such record is present in the intervening area. Field investigations in the area of the inferred barrier (Dubé, 1971; Chauvin, 1979; Parent, 1987) consistently indicate that glacial sedimentation there was immediately followed by marine sedimentation, without an intervening glaciolacustrine event. In this 50 km-wide area, Bois-Francs residual ice presumably remained in contact with the Saint-Maurice glacial lobe until the barrier was breached by the incoming marine waters. The absence of northeastward-trending spillway channels or outwash bodies in this area (Parent, 1987) further suggests that Lake Candona finally drained down to sea level across the glacial barrier rather than alongside the retreating margin of the main ice sheet.

In contradiction to the ideas presented here, Gadd (1983) postulated that a large remnant ice mass prevented early Champlain Sea waters from invading the Appalachian piedmont in the vicinity of Granby. Three main lines of reasoning were given to support his thesis: (1) the absence of marine fauna predating the *Mya* biotic phase of the Champlain Sea, (2) the presence of unfossiliferous ice-contact sediment bodies below the local marine limit and (3) the existence of ice-marginal and proglacial drainage systems which apparently formed subaerially at altitudes well below the marine limit recognized by earlier authors in this area (Figs. 1 and 10; see references in Appendix II). The first line of reasoning must be rejected as there is excellent evidence indicating that early marine fauna had indeed migrated into the area of the postulated remnant ice mass: (1) deltaic sediments containing shells of *Hiatella arctica*, dated 11.7 ka (Wagner, 1972; Parrott and Stone, 1972; see Fig. 10 and Table III) were deposited at marine limit (145 m a.s.l.) near Frelighsburg; (2) regressive deltaic sediments near Granby, at an altitude of 105 m, contain shells of *Macoma balthica*, dated 11 360 ± 110 BP (UQ-29; Prichonnet, 1982a, 1984a; see Fig. 10 and Table III). The validity of the second line of reasoning was previously discussed by Parent (1984c). Unfossiliferous ice-contact sediments of the Warwick-Asbestos Esker provide yet further evidence to indicate that the presence of ice-contact sediments below marine limit *does not* "preclude the existence of seawater at those places and elevations" (Gadd, 1983: p. 411). His third line of reasoning requires thorough demonstration based on actual field data, since previous investigations in this and adjacent areas (Prichonnet, 1982a, 1982b, 1984a, 1984b; Prichonnet *et al.*, 1982a, 1982b; McDonald, 1967, 1968; Parent, 1987) do not report evidence to support the postulated episode of subaerial meltwater flow. On the contrary, these and other investigations provide ample evidence for regional glaciolacustrine water bodies prior to Champlain Sea invasion (Fig. 7 and Appendix I).

During the latter part of Phase II of the Champlain Sea, the retreating Laurentide Ice Sheet formed the Saint-Narcisse Moraine. This complex of morainic ridges was deposited near the southern margin of the Laurentian Highlands over a distance of 500 km between the Gatineau River, north of Ottawa, and Saint-Siméon, in the Charlevoix region (LaSalle and Elson, 1975; Occhietti, 1980, Hillaire-Marcel *et al.*, 1981). Synchroneity of various segments of the morainic complex however has yet to be firmly established. In the study area, the morainic complex occurs as two distinct segments. In the lower Saint-Maurice Valley, the ice-front in the Champlain Sea deposited ice-contact and glaciomarine sediments in a 150 km-long arcuate salient. In the Parc des Laurentides highlands (Dionne *et al.*, 1968; Fig. 6), a discontinuous series of morainic ridges were deposited well outside of the marine basin.

The stratigraphic record (Figs. 3, 4) shows that the base of the Saint-Narcisse morainic complex in the lower Saint-Maurice region is composed of proximal and distal glaciomarine deposits containing *Portlandia arctica* and *Balanus hameri*, dated between 11.3 and 11.1 ka. This morainic complex was apparently deposited at a grounded ice margin that stabilized at the contact zone between the shield and the sedimentary platform. The possibility that the ice may have readvanced from a more northerly position is not excluded. The margin of the Laurentide Ice sheet may have extended into the Champlain Sea at this place because of glacial streaming within the Saint-Maurice lobe. Consistent SSE trend of glacial striae throughout the upper Saint-Maurice Valley (Occhietti, 1980) suggests that ice was channelled between the Mont Tremblant highlands to the west and the Parc des Laurentides highlands to the east (Occhietti, 1980). As a consequence, around 11 ka, the Saint-Maurice lobe of the Laurentide Ice Sheet was dynamic enough in the Saint-Maurice Valley to maintain the 150 km-long marine front. The occurrence of compression structures in melt-out till within the morainic complex suggests that the ice front remained active and grounded. The huge Mont Carmel raised delta and the kettled outwash train abutting the Laurentian Highlands north of Saint-Raymond-de-Portneuf (Occhietti, 1980) record glaciofluvial sedimentation toward the end of the episode around 10.8 ka. Evidence for a stabilization of the ice-front, equivalent to that recorded in the lower Saint-Maurice Valley, has not been observed north of Montréal; instead, a series of small subparallel ridges were formed during ice retreat (Prichonnet, 1977). The end of the Saint-Narcisse episode marked the beginning of Phase III of the Champlain Sea.

CHAMPLAIN SEA, PHASE III

Phase III of the Champlain Sea is here defined as the interval that followed retreat of Laurentide ice from the Saint-Narcisse Moraine. Events included are retreat of ice from major valleys in the Laurentian Highlands, invasion of these by the Champlain Sea and regression of the Champlain Sea from its entire basin.

In the lower Saint-Maurice Valley, post-Saint-Narcisse retreat of an active ice-front is recorded by the Cossetteville parallel ridges (Fig. 2). Transverse ridges of the middle Saint-Maurice Valley (Occhietti, 1980) and Lac du Missionnaire (Gagnon and Morelli, 1986) indicate the presence of an active ice-front in these valleys. Elsewhere, ice-frontal constructions and raised outwash trains were formed at the mouth of valleys, e.g. the proglacial delta perched at an altitude of 200 m, at the mouth of Lac Mékinac (Fig. 1, site 3).

The Champlain Sea extended into the Laurentian Highlands in the form of brackish or freshwater bodies considered as paramarine basins (Occhietti, 1977). Examples are the middle Saint-Maurice basin and the tributary basins of Saint-Joseph-de-Mékinac and Rivière Matawin, other shallow basins north and northeast of Montréal (Lamothe, 1977; Pagé, 1977), in the Gatineau valley north of Ottawa (Dadswell, 1974), and the depression north of Lac Simon (Occhietti, 1980, Fig. 1).

North of the Saint-Narcisse Moraine, sedimentation rates were high, ranging from 9 to 18 cm per year. At the mouth of the Saint-Joseph-de-Mékinac paramarine basin, single layers in basal rhythmites are up to 20 cm thick. In the fjord-like valley of the middle Rivière Saint-Maurice, 30 km north of Shawinigan, 57 m of silt and sand were deposited during an interval lasting from 300 to 500 years. In the Shawinigan embayment (Fig. 2), marine clays and silts commonly reach thicknesses on the order of 30 to 60 m and were deposited in a few centuries, between approximately 10.8 and 10 ka, at an average sedimentation rate of 7.5 cm per year. Sedimentation rates were up to 1 cm per year, in lower pebbly clays at Saint-Alban, between reference horizons with *Balanus hameri* dated 10.6 ka and with *Macoma calcarea* dated 10.2 ka (Occhietti, 1980).

In the Saint-Maurice region, sea level stood at an altitude of 200 m during construction of the proglacial delta south of Lac Mékinac shortly after the beginning of Phase III of the Champlain Sea. At about the same time, water levels had already receded to about 135 m a.s.l. on the south side of the St. Lawrence Valley, as indicated by the age of a shallow-water faunal assemblage near Warwick (UQ-289: 10 780 ± 190 BP; Parent, 1987). By 10.5 ± 0.1 ka, water levels had further receded to about 81 m a.s.l., which is the altitude of faunal assemblages dominated by *Mya arenaria* at Duncan (10 590 ± 100 BP, GrN-2034; Elson, 1969b) and at Saint-Wenceslas (10 350 ± 100 BP, UQ-947; Parent, 1987) on the south shore of Champlain Sea. North of the St. Lawrence River, the high terraces of the middle Saint-Maurice Valley and the first inset deltas of the lower Saint-Maurice were built during the regression of marine waters. By 10.2-10.1 ka, relative sea level had fallen to about 120 m a.s.l., based on a biocenotic assemblage present in reworked deposits of the Saint-Narcisse Moraine at Charette (Occhietti, 1980).

At the close of Phase III, considerable shoaling had already taken place and waters in the Champlain Sea basin turned fresh upstream of Québec. Elson (1969a, 1969b) called this freshwater body Lake Lampsilis, "after a freshwater clam commonly found as a fossil near its shores" (1969a, p. 253). Development of Lake Lampsilis in the basin was coeval with time-transgressive retreat of the saltwater wedge. Surface waters had turned fresh in the lower reaches of the Ottawa Valley by about 10.3 ka, as suggested by the ages of *Lampsilis* faunas (10 300 ± 90 BP, GSC-3235, Lowdon and Blake, 1981; 10 200 ± 90 BP, GSC-1968; Gadd, 1976). However, ^{14}C-dated marine fauna suggest that waters at depth may have remained saline until about 10 ka in that area (10 000 ± 320 BP, GSC-1553; 10 100 ± 130 BP, GSC-2189; Rodrigues and Richard, 1983). Ages of freshwater shells near Montréal (9750 ± 150 BP, GSC-2414; S. H. Richard, 1978) and near Québec (9730 ± 140 BP, GSC-1976; LaSalle *et al.*, 1977b) indicate that Lake Lampsilis lasted until at least 9.7 ka in the broad lowlands upstream of Québec. Marine waters however continued to penetrate into the fresh waters of Lake Lampsilis until very late in the Québec area, presumably because of strong tides (Occhietti and Hillaire-Marcel, 1982). The youngest tidal silts, containing shells of *Hiatella arctica* and *Macoma balthica* in growth position were found in the Saint-Nicolas sand pit (Fig. 2), up to an altitude of 60 m. They lie at the summit of a thanatocenotic assemblage containing 22 species of marine shells dated between 10 ka and 9.3 ka (Table III). Retreat of the saltwater wedge, which marked the end of Champlain Sea Phase III, was therefore diachronic at the scale of the basin; its age ranges from about 10 ka in the Ottawa Valley to about 9.3 ka in the vicinity of Québec.

CONCLUSION

The Champlain Sea invasion may be subdivided into three phases which correspond to distinct episodes of glacial retreat: Phase I (Charlesbourg), between 12.4 and 12 ka, limited to the Québec region; Phase II in the main basin, between 12 ka and approximately 10.8 ka, and Phase III, posterior to the Saint-Narcisse episode between 10.8 and 9.3 ka. Marine waters receded diachronously between 10 ka in the Ottawa Valley, in the westernmost part of the basin, and 9.3 ka near Québec, the easternmost part. The transition from Champlain Sea to Lake Lampsilis may have occurred as the saltwater wedge retreated progressively toward the Québec narrows.

The glacial context provides an explanation for the delay of some 200 to 400 years between the beginning of the Charlesbourg phase and the invasion of the main Champlain Sea basin. In the St. Lawrence Valley and adjacent piedmont, patterns of ice-marginal retreat are oblique to the margin of the Appalachian uplands, as shown in particular by the Mont Ham and Ulverton-Tingwick moraines. Because of this configuration, invasion of the main basin only occurred once the ice front, controlled by the Saint-Maurice lobe, had retreated northward to the vicinity of the early marine embayment of the Charlesbourg phase. The Highland Front Morainic System consists of a suite of strongly time-transgressive deposits. Because referring to these features by a single name does not aid in decephering deglacial patterns and events, this term should be abandoned. A remnant ice mass in the Bois-Francs uplands may have reinforced the glacial barrier between incoming marine waters of the Charlesbourg phase to the north and waters of Lake Candona to the south.

The marine invasion of Phase II began around 12 ka. Much of the main basin was already ice-free and inundated by a vast proglacial lake, Lake Candona, which was at least partly correlative with glacial Lake Iroquois of the Lake Ontario basin and with the Fort Ann level phase of Lake Vermont. Marine invasion caused Late Wisconsinan water levels to fall by 50 to 60 m in southern Québec and adjacent Lake Champlain Valley. The proposed age, 12 ka, for the beginning of the invasion is in agreement with falling Lake Iroquois levels and is consistent with reliable shell dates from the main basin or close to the controversial 12.2 $\pm$ 0.1 ka ages on *Macoma balthica* from the Ottawa area. Due to lack of field evidence, deglaciation patterns and events during Phase II are little known. Rates of ice retreat between the Appalachian piedmont and the Saint-Narcisse Moraine, may have averaged only about 90 m per year. If so, the ice sheet remained thick and active in a fairly large area of the middle St. Lawrence Valley.

Between approximately 11.3 and 11 ka, a 150 km-long glaciomarine front stabilized in the Saint-Maurice region and deposited the fossiliferous Yamachiche Diamicton. Deposits of the Saint-Narcisse Moraine, dated between approximately 11 and 10,8 ka, then covered the glaciomarine deposits. Phase III of the Champlain Sea followed glacial retreat from the Saint-Narcisse Moraine. Marine waters invaded the Shawinigan embayment and then extended northward to form paramarine basins in valleys of the Laurentian Highlands.

This paper has also revealed some significant uncertainties. Our proposed model for deglaciation events in the Québec region and lower Chaudière Valley needs further testing in the field, particularly because the relationships between the retreating Laurentide Ice Sheet and the Appalachian residual ice masses are not yet satisfactorily resolved, nor are their relationships with Lake Beauce. Deglaciation patterns during Champlain Sea Phase II remains poorly documented in the middle St. Lawrence Valley.

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APPENDIX I

Elevation of strandline features related to the Fort Ann level phase of Glacial Lake Vermont; numbers refer to Figure 7 (from Parent, 1987)

Feature #	Elevation (m a.s.l.) and error	Departure (m) from Fort Ann isobases*	Method of measurement	Feature	Locality	Reference
1-A	227 +/− 2	−1	Altimeter	Delta (ice-contact)	Tingwick, Qc	Parent, 1987
1-B	229 +/− 4	0	Map contours	Ice-contact lake terrace	Tingwick, Qc	Parent, 1987
2-A	226 +/− 2	+3	Altimeter	Delta (ice-contact)	Tingwick, Qc	Parent, 1987
2-B	221 +/− 4	+6	Map contours	Delta (fluvial)	Moulin Samson, Qc	Parent, 1987
3	226 +/− 1	+1	Spot elevation	Delta (ice-contact)	Colline Elliott, Qc	Parent, 1987
4	221 +/− 4	+3.5	Map contours	Beach sediments	Pinnacle Shipton, Qc	Parent, 1987
6	210 +/− 4	−3	Map contours	Beach sediments	St-Claude, Qc	Parent, 1987
8	206 +/− 4	−1	Map contours	Delta (fluvial)	Riv. Watopéka, Qc	Parent, 1987
9	202 +/− 4	−4	Map contours	Delta (fluvial)	Riv. Stoke, Qc	Parent, 1987
11	229 +/− 4	−3	Map contours	Delta (fluvial)	Denison Mills, Qc	Parent, 1987
12	214 +/− 2	−6	Altimeter	Wave-cut terrace	Richmond, Qc	McDonald, 1967 (Fig. 14: #18)
13	213 +/− 2	−6	Altimeter	Delta (fluvial)	Kingsbury, Qc	McDonald, 1967 (Fig. 14: #19)
14	227 +/− 2	−5	Altimeter	Delta (ice-contact)	Dalling, Qc	McDonald, 1967 (Fig. 14: #16)
15	229 +/− 2	−3	Altimeter	Wave-cut terrace	Dalling, Qc	Parent, 1987
16	229 +/− 4	0	Map contours	Delta (fluvial)	Valcourt, Qc	Parent, 1987
17	220 +/− 2	−15	Altimeter	Delta (ice-contact)	Béthanie, Qc	McDonald, 1967 (Fig. 14: #17)
18	209 +/− 4	0	Map contours	Wave-cut terrace	Petit Lac St-François, Qc	Parent, 1987
19	229 +/− 4	−1	Map contours	Littoral boulders	Martin Corner, Qc	Doiron, 1981
20	217 +/− 4	−2	Map contours	Delta (flluvial)	Lawrenceville, Qc	Parent, 1987
23	195 +/− 2	−5	Altimeter	Delta (fluvial)	Key Brook, Qc	McDonald, 1967 (Fig. 14: #22)
26	192 +/− 2	+1	Altimeter	Delta (fluvial)	Capelton, Qc	McDonald, 1967 (Fig. 14: #24)
27	190 +/− 2	0	Altimeter	Wave-cut terrace	Ayer's Cliff, Qc	Boissonneault *et al.*, 1981 (p. 29)
28	229 +/− 4	−7	Map contours	Wave-cut terrace	St-Joachim-de-Shefford, Qc	Parent, 1987

Feature #	Elevation (m a.s.l.) and error	Departure (m) from Fort Ann isobases*	Method of measurement	Feature	Locality	Reference
29	223 +/− 4	−12	Map contours	Beach sediments	Savage Mills, Qc	Prichonnet, 1982a (Fig. 29-6)
30	229 +/− 4	−4	Map contours	Wave-cut terrace	Mont Shefford, Qc	Parent, 1987 (also Doiron, 1981)
31	224 +/− 4	0	Map contours	Wave-cut terrace	Pine Mountain, Qc	Parent, 1987
32	216 +/− 4	−4	Map contours	Delta (ice-contact)	Brome Centre, Qc	Prichonnet *et al.*, 1982a
33	213 +/− 4	−1	Map contours	Delta (fluvial)	Brome, Qc	Prichonnet *et al.*, 1982a
34	213 +/− 4	0	Map contours	Delta (fluvial)	Sutton Jonction, Qc	Prichonnet *et al.*, 1982a
35	219 +/− 4	+3	Map contours	Beach ridges	Farnham Corners, Qc	Prichonnet, 1982b (Fig. 3)
36	215 +/− 4	0	Map contours	Beach ridges	Lac Selby, Qc	Prichonnet, 1982b (Fig. 3)
37	210 +/− 4	+1	Map contours	Delta (fluvial)	Alva, Qc	Prichonnet *et al.*, 1982a
38	200 +/− 4	0	Map contours	Delta (fluvial)	Dunkin, Qc	Dubé, 1983, p. 106
39	187 +/− 4	−11	Map contours	Terrace (fluvial?)	Mansonville, Qc	Parrott and Stone, 1972 (Fig. 1: #31)
40	191 +/− 4	−11	Map contours	Terrace (fluvial?)	Travor Road, Qc	Parrott and Stone, 1972 (Fig. 1: #34)
41	203 +/− 4	−2	Map contours	Terrace (fluvial?)	Potton Springs, Qc	Parrott and Stone, 1972 (Fig. 1: #39)
42	186 +/− 4	−9	Map contours	Littoral terrace	Troy, Vt	Parrott and Stone, 1972 (Fig. 1: #28)
43	198 +/− 3	−3	Map contours	Delta-terrace	East Richford, Vt	Parrott and Stone, 1972 (Fig. 1: #29)
44	186 +/− 3	−12	Map contours	Delta, terrace	East Berkshire, Vt	Parrott and Stone, 1972 (Fig. 1: #27)
45	183 +/− 3	−11	Map contours	Terrace (fluvial?)	Montgomery, Vt	Parrott and Stone, 1972 (Fig. 1: #26)
46	186 +/− 3	−6	Map contours	Terrace (fluvial?)	Montgomery Center, Vt	Parrott and Stone, 1972 (Fig. 1: #25)
47	183 +/− 3	−12	Map contours	Beach sediments	West Enosburg, Vt	Parrott and Stone, 1972 (Fig. 1: #23)
48	183 +/− 3	−12	Map contours	Beach	Bellevue Hill, Vt	Wagner, 1972 (Fig. 1: F-32)
49	186 +/− 3	−1	Map contours	Delta (ice-contact)	Buck Hollow, Vt	Wagner, 1972 (Fig. 1: F-31)
50	174 +/− 3	−10	Map contours	Delta	Binghamville, Vt	Wagner, 1972 (Fig. 1: F-30)
51	168 +/− 3	−12	Map contours	Delta	Cambridge, Vt	Wagner, 1972 (Fig. 1: F-28)
52	162 +/− 3	−18	Map contours	Delta	Fairfax, Vt	Wagner, 1972 (Fig. 1: F-26)
53	171 +/− 3	−14	Map contours	Beach	Arrowhead Mountain, Vt	Wagner, 1972 (Fig. 1: F-29)
54	180	−2	Leveling	Beaches	Peru, NY	Chapman, 1937 (Fig. 15: #23)
55	182	−4	Leveling	Beach	Schuyler Falls, NY	Chapman, 1937 (Fig. 15: #24)
56	183	−6	Leveling	Beach	Beckwith School, NY	Chapman, 1937 (Fig. 15: #25)
57	190	−3	Leveling	Beaches	Morrisonville, NY	Chapman, 1937 (Fig. 15: #26)
58	192	−2	Leveling	Delta	Morrisonville, NY	Chapman, 1937 (Fig. 15: #27)
59	197	0	Leveling	Beaches	West Beekmantown, NY	Chapman, 1937 (Fig. 15: #28)
60	198	0	Leveling	Littoral terrace	West Beekmantown, NY	Chapman, 1937 (Fig. 15: #29)
61	203	+2	Leveling	Beaches	Shelter's Corners, NY	Chapman, 1937 (Fig. 15: #30)
62	206	+1	Leveling	Beaches	Cobblestone Hill, NY	Chapman, 1937 (Fig. 15: #31)
63	210	0	Leveling	Beach	Pine Ridge, NY	Chapman, 1937 (Fig. 15: #32)
64	215	+1	Leveling	Beaches	Dear Pond, NY	Chapman, 1937 (Fig. 15: #34)
65	222	+3	Leveling	Beaches	Cannon Corners, NY	Chapman, 1937 (Fig. 15: #35)

Feature #	Elevation (m a.s.1.) and error	Departure (m) from fort ann isobases*	Method of measurement	Feature	Locality	Reference
66	223	0	Map contours	Beaches	Cannon Corners Road, NY	Denny, 1974 (Fig. 6)
67	226.5 +/− 1.5	−1	Map contours	Beach	Maritana, Qc	Clark and Karrow, 1984 (Fig. 2: #18)
68	220.5 +/− 1.5	−1	Map contours	Beach	Earlville, NY	Clark and Karrow, 1984 (Fig. 2: #17)
69	214.5 +/− 1.5	0	Map contours	Beach	Burke Center, NY	Clark and Karrow, 1984, (Fig. 2: #16)
70	208.5 +/− 1.5	−1	Map contours	Beach	Malone, NY	Clark and Karrow, 1984, (Fig. 2: #15)
71	205.5 +/− 1.5	+1	Map contours	Beach	East Dickinson, NY	Clark and Karrow, 1984, (Fig. 2: #14)

* Elevation of feature minus elevation of isobase. Mean departure = −3.4 m (s.d. = 5.1); when features from the Vermont-Missisquoi region (#39 through #53) are excluded, mean departure falls to −1.6 m (s.d. = 3.5). Isobases are shown in Figure 7.

APPENDIX II

Elevation of strandline features defining the marine limit on the south shore of the main Champlain Sea basin; numbers refer to Figure 10 (from Parent 1987)

Feature #	Elevation (m a.s.l.) and error	Departure (m) from marine limit isolines*	Method of measurement	Feature	Locality	Reference
1	161.5 +/− 1.5	−6	Map contours	Beach	Maritana, Qc	Clark and Karrow, 1984 (Fig. 2: #41)
2	160	−4	Leveling	Beaches	Covey Hill, Qc	Chapman, 1937 (Fig. 15: #57)
3	164	+3	Leveling	Beaches	English River, NY	Chapman, 1937 (Fig. 15: #56)
4	162	+4	Leveling	Beaches	Cannon Corners, NY	Chapman, 1937 (Fig. 15: #55)
5	157	+1	Leveling	Delta	North Branch, NY	Chapman, 1937 (Fig. 15: #54)
6	152	−1	Leveling	Beach	Woods Falls, NY	Chapman, 1937 (Fig. 15: #53)
7	148	−2	Leveling	Beaches	Sciota, NY	Chapman, 1937 (Fig. 15: #52)
8	142	0	Leveling	Beaches	West Chazy, NY	Chapman, 1937 (Fig. 15: #51)
9	138	0	Leveling	Beaches	West Beekmantown, NY	Chapman, 1937 (Fig. 15: #50)
10	131	−1.5	Leveling	Beaches	West Beekmantown, NY	Chapman, 1937 (Fig. 15: #49)
11	130	−1.5	Leveling	Delta	Morrisonville, NY	Chapman, 1937 (Fig. 15: #48)
12	123	−5	Leveling	Terrace	Schuyler Falls, NY	Chapman, 1937 (Fig. 15: #47)
13	121	−5	Leveling	Beaches	Schuyler Falls, NY	Chapman, 1937 (Fig. 15: #46)
14	122	−3	Leveling	Delta	Schuyler Falls, NY	Chapman, 1937 (Fig. 15: #45)
15	110	−7	Leveling	Delta	Milton, Vt	Chapman, 1937 (Fig. 16: #22)
16	120	0	Leveling	Delta	East Georgia, Vt	Chapman, 1937 (Fig. 16: #24)
17	119 +/− 3	−12	Map contours	Beach	St. Albans, Vt	Wagner, 1972 (Fig. 1: S-55)
18	134	−1	Leveling	Terrace	St. Albans, Vt	Chapman, 1937 (Fig. 16: #25)
19	119 +/− 3	−11	Map contours	Delta	Fairfield Station, Vt	Wagner, 1972 (Fig. 1: S-60)
20	125 +/− 3	−10	Map contours	Delta	Sheldon, Vt	Wagner, 1972 (Fig. 1: S-66)

Feature #	Elevation (m a.s.l.) and error	Departure (m) from marine limit isolines*	Method of measurement	Feature	Locality	Reference
21	119 +/− 3	−15	Map contours	Delta	South Franklin, Vt	Wagner, 1972 (Fig. 1: S-69)
22	137 +/− 3	−4	Map contours	Delta	East Franklin, Vt	Wagner, 1972 (Fig. 1: S-87)
23	145 +/− 4	0	Map contours	Delta	Frelighsburg, Qc	Wagner, 1972 (Fig. 1: S-88)
24	155	+1	Leveling	Beach	Dunham, Qc	Chapman, 1937 (Fig. 16: #26)
25	158	+3	Leveling	Beach	Oak Hill, Qc	Chapman, 1937 (Fig. 16: #27)
26	171	0	Leveling	Beach	Granby, Qc	Chapman, 1937 (Fig. 16: #28)
27	175	−3	Leveling (?)	Terrace	Mont Yamaska, Qc	Goldthwait, 1911 (in Brown Macpherson, 1967)
28	168	−4	Leveling	Terrace	Ste-Prudentienne, Qc	Chapman, 1937 (Fig. 16: #29)
29	171 +/− 4	−2	Map contours	Terrace	Ste-Prudentienne, Qc	Parent, 1987
30	175 +/− 2	0	Altimeter	Beach	South Durham, Qc	Parent, 1987
31	176 +/− 2	+2	Altimeter	Wave-cut terrace	South Durham, Qc	Parent, 1987
32	175 +/− 4	+2	Map contours	Beach sediments	L'Avenir, Qc	Parent, 1987
33	170 +/− 2	0	Altimeter	Wave-cut terrace	Trenholm, Qc	Parent, 1987
34	171 +/− 2	−2	Altimeter	Beach gravel	Mareuil, Qc	Parent, 1987
35	164 +/− 2	−8	Altimeter	Beach	Mont Proulx, Qc	Parent, 1987
36	165 +/− 2	0	Altimeter	Wave-cut terrace	Danville, Qc	Parent, 1987
37	**164 +/− 2	−9	Altimeter	Beach gravel	Ste-Elizabeth-de-Warwick, Qc	Parent, 1987
38	174 +/− 2	+2	Altimeter	Delta	Warwick, Qc	Parent, 1987
39	172.5 +/− 2	−2.5	Altimeter	Delta	Mont St-Michel, Qc	Parent, 1987
40	175 +/− 2	0	Altimeter	Terrace	Norbertville, Qc	Parent, 1987
(41)	191 +/− 4	N.A.	Map contours	Wave-cut terrace	Laurierville, Qc	Parent, 1987
(42)	191 +/− 4	N.A.	Map contours	Beach sediments	Laurierville, Qc	Parent, 1987
(43)	183 +/− 4	N.A.	Map contours	Delta	Rectory Hill, Qc	Parent, 1987 (also Dubé, 1971)
(44)	181	N.A.	Unknown	Beach	Mont St-Hilaire, Qc	Brown Macpherson, 1967: Table 1
(45)	188	N.A.	Unknown	Highest marine shells	Mont Royal, Qc	Stanfield, 1915 (in Brown Macpherson, 1967: Table 1)

* Elevation of feature minus elevation of isoline; mean departure = −2.5 m (s.d. = 4.4). Isolines are shown in Figure 8.
** A sand blanket of either eolian or littoral origin extends up to 179 m a.s.l.
( ) These features were not used for constructing isolines shown in Figure 10.
N.A. Not appllicable.