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Late Wisconsinan deglaciation and glacial lake development in the Appalachians of southeastern Québec
Déglaciation et évolution des lacs glaciaires dans les Appalaches du Québec méridional au Wisconsinien supérieur
Enteisung im Spät-Wisconsin und glaziale Seen-Entwicklung in den Appalachen von Südost-Québec

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Article abstract

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LATE WISCONSINAN DEGLACIATION AND GLACIAL LAKE DEVELOPMENT IN THE APPALACHIANS OF SOUTHEASTERN QUÉBEC[*]

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ABSTRACT   Late Wisconsinan deglaciation in southeastern Québec was preceded by a northward ice-flow reversal that was recorded in the northeastern part of the region. The reversal event was generated by flow convergence toward the St. Lawrence Ice Stream, a northeastward-flowing ice stream which formed in the St. Lawrence estuary prior to 13 000 years BP and lasted until at least 12 400 years BP. In the Bois-Francs uplands, the flow reversal event led to the formation of a semi-detached ice mass that underwent widespread stagnation and downwasting. In the southwestern region, northward retreat of the margin of the Laurentide Ice Sheet was marked by the formation of a series of discontinuous recessional moraines and by the development of ice-dammed lakes in the main valleys. The level of these lakes fell as progressively lower outlets became ice-free. The main episodes are (1) the Sherbrooke Phase of Glacial Lake Memphremagog, (2) an unnamed transitional lake and (3) Glacial Lake Candona, a large lake which had expanded northeastward from the deglaciated regions of the Upper St. Lawrence (Lake Iroquois) and Ottawa valleys to the Lake Champlain (Glacial Lake Vermont) basin. As recorded by the Danville Varves, Lake Candona lasted about 100 years following deposition of the Ulverton-Tingwick Moraine. Subsequent ice retreat along the Appalachian piedmont led to final drainage of Lake Candona and allowed Champlain Sea waters to invade much of these glaciolacustrine terrains about 12 000 years BP. On the basis of the Danville Varves record, a regional rate of ice retreat of about 200 m·a^{-1} is inferred. The age of the earliest moraine, the Frontier Moraine, is thus about 12 550 years BP, while the ages of the subsequent Dixville, Cherry River–East-Angus, Mont Ham and Ulverton-Tingwick moraines are estimated at 12 500, 12 325, 12 200 et 12 100 years BP, respectively.

RÉSUMÉ   *Déglaciation et évolution des lacs glaciaires dans les Appalaches du Québec méridional au Wisconsinien supérieur.* La déglaciation du Wisconsinien supérieur a été précédée d'une inversion d'écoulement glaciaire vers le nord. Cette inversion, enregistrée dans le nord-est de la région, résulte de la convergence des glaces vers le Courant glaciaire du Saint-Laurent, lequel s'était formé dans l'estuaire du Saint-Laurent avant 13 000 ans BP et a duré au moins jusque vers 12 400 ans BP. Dans les hautes-terres des Bois-Francs, cette inversion d'écoulement a mené à l'isolement d'une masse glaciaire partiellement détachée et qui se dissipa tardivement. Dans le secteur sud-ouest, le retrait de l'Inlandsis est ponctué par la mise en place d'ensembles morainiques discontinus et par le développement de lacs proglaciaires barrés dans les vallées principales. Le niveau de ces lacs chutait à mesure que le retrait glaciaire libérait des cols de plus en plus bas. Les principaux épisodes sont: la Phase Sherbrooke du Lac glaciaire Memphrémagog ; un lac de transition ; le Lac glaciaire Candona, formé de la coalescence des lacs glaciaires endigués dans les vallées du haut Saint-Laurent, des Outaouais, du lac Champlain (Lac Vermont) et du Saint-François. Le Lac Candona a subsisté quelque 100 ans après le dépôt de la Moraine d'Ulverton-Tingwick, comme le démontrent les Varves de Danville. Le retrait glaciaire subséquent sur le piémont appalachien causa la vidange finale du Lac Candona, permettant l'incursion de la Mer de Champlain vers 12 000 ans BP dans plusieurs des terrains glaciolacustres. Les Varves de Danville permettent d'estimer le taux de retrait glaciaire régional à environ 200 m·a^{-1}. Ainsi, l'âge de la moraine la plus ancienne de la région, la Moraine de la Frontière, est estimé à environ 12 550 ans BP, et celui des moraines subséquentes, Dixville, Cherry River–East-Angus, Mont Ham et Ulverton-Tingwick, à environ 12 500, 12 325, 12 200 et 12 100 ans BP, respectivement.

ZUSAMMENFASSUNG   *Enteisung im Spät-Wisconsin und glaziale Seen-Entwicklung in den Appalachen von Südost-Québec.* Der Enteisung im Spät-Wisconsin ging eine Umkehrung des Eisflusses nach Norden voraus, die im nordöstlichen Teil der Region aufgezeichnet ist. Das Umkehrungs-Ereignis wurde durch konvergierendes Fließen in Richtung des St. Lorenz-Eisstroms bewirkt, eines nordostwärts fließenden Eisstroms, der sich in der St. Lorenz-Mündung vor 13000 Jahren v.u.Z. bildete und mindestens bis 12 400 Jahre v.u.Z. dauerte. Im Hochland der Bois-Francs führte das Umkehrungs-Ereignis des Fließens zur Bildung einer halb losgelösten Eismasse, die Stagnation und Abzehrung durchlief. Im südwestlichen Gebiet war der Rückzug des Rands der laurentischen Eisdecke in Richtung Norden markiert durch die Bildung einer Serie von diskontinuierlichen Rückzugsmoränen sowie die Entwicklung von Eisabdämmungsseen in den Haupttälern. Die Hauptepisoden sind (1) die Sherbrooke-Phase des glazialen Sees Memphremagog, (2) ein namenloser Übergangssee und (3) der glaziale See Candona, ein breiter See, der sich norostwärts ausgedehnt hat, von den ent-eisten Gebieten des oberen St. Lorenz (Lake Iroquois) und den Ottawa-Tälern zum Champlain-See-Becken (glazialer Vermont-See). Wie in den Danville-Warwen aufgezeichnet dauerte der Candona-See ungefähr 100 Jahre nach der Ablagerung der Ulverton-Tingwick-Moräne. Der darauffolgende Eisrückzug entlang des Vorlands der Appalachen führte zu einer endgültigen Drainage des Candona-Sees und erlaubte den Wassern des Champlain-Meeres in einen großen Teil dieser Gebiete glazialer Seen um etwa 12 000 Jahre v.u.Z. einzudringen. Das Alter der frühesten Moräne, der Frontier-Moräne ist etwa 12 560 Jahre v.u.Z., während die Alter der darauffolgenden Dixville-, Cherry River–East-Angus-, Mont Ham- und Ulverton-Tingwick-Moränen jeweils auf 12 500, 12 325 und 12 100 Jahre v.u.Z. geschätzt werden.

INTRODUCTION

The question of deglaciation style and patterns in northern New England and adjacent southeastern Québec has been debated for nearly a century. A useful introduction on the development of early concepts, which we do not intend to review here, may be found in the latest book on the Late Pleistocene history of the northern Appalachians (Borns *et al.*, 1985). The latest chapters came about when reports of "old shell dates" (12800 ± 220 years BP, GSC-1859; Richard, 1974; 1978), together with reports of late-glacial northward and westward ice-flow in the Thetford-Mines and Asbestos areas (Lamarche, 1971, 1974), brought into question the concept of an active ice-front retreating northwestward from northern New England and across adjacent southern Québec (McDonald, 1967, 1968; Shilts, 1970; Gadd *et al.*, 1972). At that time, as pointed out by Borns (1985), the debate on Appalachian ice masses and deglacial patterns was rekindled as it appeared that the ice margin was actively depositing end moraines in glaciomarine environments of coastal Maine (Stuiver and Borns, 1975) at about the same time (12700-12800 years BP) the late glacial Champlain Sea had apparently invaded the central St. Lawrence Lowland to the vicinity of Ottawa (Richard, 1974, 1978). Although the validity and meaning of the "old shell dates" from the Ottawa region have been questioned in several papers (Hillaire-Marcel, 1981; Harington and Occhietti, 1988; Parent and Occhietti, 1988; Fulton and Richard, 1987; LaSalle and Chapdelaine, 1990), the fact remains the Champlain Sea incursion is only a few hundred years younger than the end of deglaciation of coastal Maine. The distance between the two regions being about 300 km (Fig. 1), there is little doubt that large Appalachian ice masses were isolated southeast of the St. Lawrence valley during Late Wisconsinan deglaciation. The best approach to resolve these long-standing questions is to reconstruct robust local and regional deglacial histories and to establish interregional correlations based on benchmark events such as late-glacial marine incursions and glaciolacustrine water bodies.

Subsequently to the early reports by Lamarche (1971, 1974), new fieldwork in the southern Québec Appalachians (*e.g.* Gauthier, 1975; Lortie, 1976; LaSalle *et al.*, 1977a) revealed the considerable areal extent of the northward-trending striae. These findings led to the well known concept of a late-glacial ice-flow reversal in southern Québec (Gadd, 1976; Shilts, 1976, 1981), presumably driven by calving bay dynamics in the Gulf of St. Lawrence (Thomas, 1977), as well as to rejuvenated concepts of local or Appalachian ice masses in southeastern Québec and adjacent Maine (Chauvin *et al.*, 1985; Lowell, 1985; Lortie and Martineau, 1987). However, because the northward-trending striae extend well into an area where the deglacial record is documented by several recessional morainic belts and by a series of ice-dammed lakes (McDonald, 1968; Clément and Parent, 1977; Parent, 1987), showing that the ice-front last retreated northwestward across the region (Fig. 2), it has proven difficult to fully reconcile the ice-flow reversal event with the local deglacial history.

FIGURE 1. Location map.
Carte de localisation.

In an earlier paper (Parent and Occhietti, 1988), we discussed at some length the key relationships between the deglacial record and the incursion of the Champlain Sea. In this paper we intend (1) to discuss briefly the nature of the deglacial record from the Dixville Moraine, near the Québec-Vermont boundary, to the Ulverton-Tingwick Moraine along the northern margin of the Appalachian uplands, (2) to present the related development of ice-dammed lakes in the upper and middle St-François valley and in adjacent regions, (3) to propose the Danville Varves, exposed at the Rivière Landry section, as the stratotype for pre-Champlain Sea varves, (4) to assess regional rates of ice retreat and chronology, and (5) to reassess the origin of the northward ice-flow reversal which is directly linked to the St. Lawrence Ice Stream.

ICE-MARGINAL ENVIRONMENTS

MORAINIC BELTS

A series of four main recessional morainic belts extend across this area of the Applachian uplands (Fig. 2). From oldest to youngest, these are (1) the Dixville Moraine, (2) the Cherry River–East-Angus Moraine, (3) the Mont Ham Moraine and (4) the Ulverton-Tingwick Moraine. As shown in Figure 2, it seems reasonable to correlate the oldest of these morainic belts, the Dixville Moraine of the Coaticook region, with the Ditchfield Moraine (Shilts, 1970) of the adjacent Lac-Mégantic region. The Frontier Moraine (Shilts, 1970, 1981), which extends slightly into the region shown in Figure 2 and which has yet to be traced into northern New Hampshire and Vermont, would then be somewhat older than the Dixville Moraine. These are discontinuous end moraines which
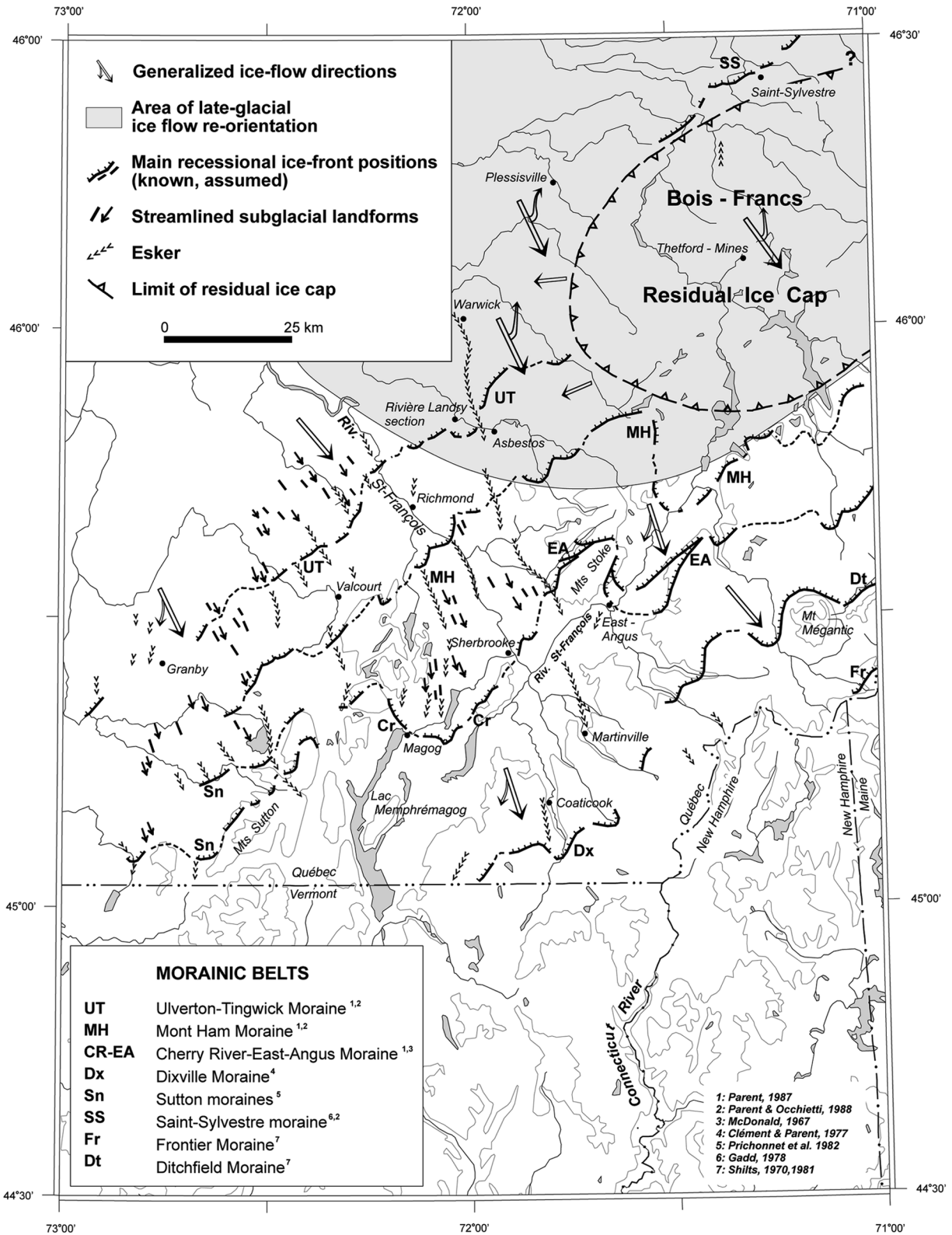
FIGURE 2. The main recessional morainic belts and glacial landforms show that the ice margin retreated northward across the region. The map also shows the area affected by a late-glacial ice-flow reversal (northward and northeastward).

Les principales moraines de retrait et formes glaciaires montrent que la marge glaciaire s'est retirée du sud vers le nord dans la région. La carte montre aussi l'aire touchée par une inversion de l'écoulement tardiglaciaire (vers le nord et le nord-est).

mainly consist of ice-contact sediment bodies and of assem-
blages of morainic ridges and intervening ice-marginal melt-
water channels. These recessional features were described
in earlier papers (McDonald, 1968; Gadd *et al.*, 1972; Clé-
ment and Parent, 1977; Boissonneault and Gwyn, 1983;
Larocque, A. *et al.*, 1983; Larocque, G. *et al.*, 1983; Parent
and Occhietti, 1988) and there is no need to redescribe them
here. Correlations between discontinuous moraine seg-
ments were also discussed in our earlier paper (Parent and
Occhietti, 1988). However it may be useful to point out that,
although connecting moraine segments a few kilometers
apart is a relatively straightforward exercise, the only sure
way of connecting more distant segments of these reces-
sional features, from one valley to the next for instance, is
through their relationships with coeval glaciolacustrine water
planes. The reconstruction of the Mont Ham Moraine and of
the maximum extent of Sherbrooke Phase of Glacial Lake
Memphrémagog (Table I, Fig. 3) provides a good example of
such correlations. This method has proved quite useful in the
region, particularly in view of the lack of radiocarbon control
on deglacial events. Unlike the Saint-Narcisse Moraine
which is a prominent feature that can be easily traced both at

a regional scale (LaSalle *et al.*, 1972; LaSalle and Elson,
1975; Occhietti, 1977, 1980; Prichonnet, 1977) and at a local
scale (Gadd, 1971; Occhietti, 1980; Govare, 1995; Cloutier
et al., 1997), the morainic belts of the Appalachian uplands
do not form distinct drift boundaries, and are thus thought to
be essentially recessional features. This does not preclude
however the possibility of local oscillations during the course
of generalized ice-marginal retreat.

On the basis of the extent and depth of proglacial lakes
impounded by the retreating ice-front, three main types of
ice-marginal environments can be recognized in the region:
glaciolacustrine, transitional and terrestrial.

GLACIOLACUSTRINE ICE-MARGINAL ENVIRONMENTS

In terrains where large, deep proglacial lakes were
impounded at the ice-margin, such as in the Rivière Saint-
François valley downstream from Sherbrooke and on low
adjacent uplands, morainal sediments consist mainly of
subaquatic outwash deposits, several of which are evidently
esker-fed sediment bodies. The Colline Elliot delta-kame
and the associated Warwick-Asbestos esker (Parent and

TABLE I

Elevation of strandline features related to the Sherbrooke Phase of Glacial Lake Memphrémagog

| Feature #	Elevation(m asl) and error	Departure (m) from Sherbrooke Phase isobases *	Method of Measurement	Feature	Locality	Reference
1	249 ± 2	−1	Altimeter	Floor of spillway	Lac Nick, Qc	McDonald, 1967
2	233 ± −4	−1	Map contours	Delta (fluvial)	Ayer's cliff, Qc	McDonald, 1967
3	248 ± −2	0	Altimeter	Delta (fluvial)	Castle Brook, Qc	McDonald, 1967
4	249 ± 2	0	Altimeter	Delta (ice-contact)	Cherry River, Qc	McDonald, 1967
5	249 ± 2	−1	Altimeter	Delta (ice-contact)	Cherry River, Qc	McDonald, 1967
(6)	246 ± 3	N.A.	Altimeter	Delta (ice-contact)	Cherry River, Qc	McDonald, 1967
7	255 ± 2	−1	Altimeter	Terrace	St-Denis-de-Brompton, Qc	McDonald, 1967
8	259 ± 4	0	Map contours	Wave-cut terrace	Côte St-Joseph, Qc	Parent, 1987
(9)	254 ± 2	N.A.	Altimeter	Delta (fluvial)	Lac Brompton, Qc	McDonald, 1967
10	273 ± 2	1	Spot elevation	Ice-contact lake terrace	St-François-Xavier-de Brompton,Qc	Parent, 1987
11	240 ± 4	0	Map contours	Delta (ice-contact)	North Hatley, Qc	Parent, 1987
12	233 ± 2	−1	Altimeter	Delta (fluvial)	Waterville, Qc	McDonald, 1967
13	248 ± 2	3	Altimeter	Delta (ice-contact)	Ascot Corner, Qc	McDonald, 1967
14	244 ± 2	3	Altimeter	Delta (ice-contact)	Brookbury,Qc	Parent, 1987
15	242 ± −2	0	Altimeter	Delta (ice-contact)	Bishopton, Qc	Parent, 1987
16	254 ± 2	0	Altimeter	Beach gravel	Mont Carrier, Qc	Parent, 1987
17	255 ± 2	−1	Altimeter	Beach gravel	Windsor, Qc	Parent, 1987
18	251 ± 4	−1	Map contours	Delta (fluvial)	Ruisseau Niquette, Qc	Parent, 1987
(19)	262 ± 2	N.A.	Altimeter	Delta (ice-contact)	St-Claude, Qc	Parent, 1987
20	267 ± −4	0	Map contours	Delta (ice-contact)	Wotton, Qc	Parent, 1987
21	240 ± −4	1	Map contours	Delta (ice-contact)	Bury, Qc	Parent, 1987

* Elevation of feature minus elevation of isobase; mean departure = 0.1 m (s.d. = 1.3).
( ) These features were not used for constructing isobases.
N.A. Not applicable.

FIGURE 3. Glacial Lake Memphremagog (Sherbrooke Phase) during formation of the Mont Ham Moraine. Isobases (m ASL) show that differential isostatic rebound has tilted the former shoreline features up to the northwest at a rate of 1.2 m·km^{-1}. Elevations of shoreline features are shown in Figure 4 and listed in Table I.

Le Lac glaciaire Memphrémagog (Phase Sherbrooke) durant la mise en place de la Moraine du Mont Ham. Les isobases montrent que les paléorivages de ce lac ont subi un relèvement isostatique différentiel de 1,2 m·km^{-1}. Les altitudes des paléorivages sont illustrées à la figure 4 et données au tableau I.
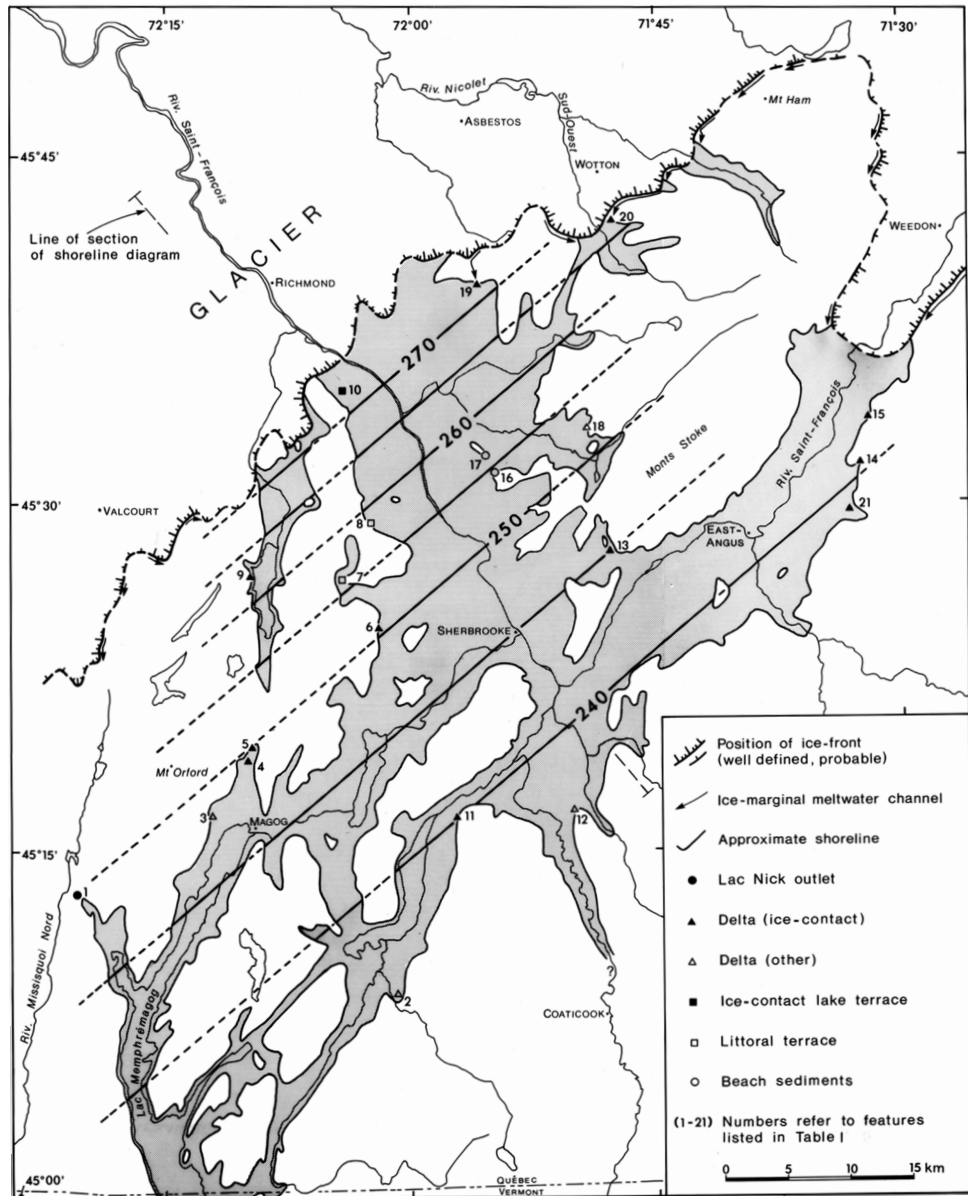
Occhietti, 1988) are typical assemblages formed in such deglacial environments. End-moraine segments are particularly discontinuous in such terrains; however, numerous esker ridges and small streamlined subglacial landforms (drumlins, drumlinoid ridges, crag-and-tail ridges, flutings) are present in low-relief areas between morainic segments (Fig. 2). The area of occurrence of these glaciolacustrine ice-marginal environments is approximately given by the area of occurrence of eskers and streamlined subglacial landforms, roughly west of the series of eskers running from Warwick to Martinville. The surficial mapping work of McDonald (1966, 1967) was mostly carried out in such areas, which may explain why he initially thought that morainic belts in the region are "... largely composed of ice-contact stratified drift..." (Gadd *et al.*, 1972). As originally discussed by Clément and Parent (1977), adjacent terrains are characterized by morainic belts with different landform-sediment assemblages.

TRANSITIONAL ICE-MARGINAL ENVIRONMENTS

Transitional terrains are those in which only small or shallow proglacial lakes were impounded at the ice-margin. In such terrains, morainic belts consist mainly of assemblages of till ridges and intervening meltwater channels. These transverse morainic landforms, which most commonly lie on north- or northwest-facing slopes, locally extend over distances up to 8 km; their trend reflects quite closely that of the retreating ice-margin (Clément et Parent, 1977; Parent, 1978). Small bodies of ice-contact stratified drift, which locally include deltas, commonly occur at the downstream extremity of ice-marginal meltwater channels; when present,

these deltas allow specific ice-front positions to be tied with the level of coeval ice-dammed lakes. In terrains between the main morainic belts, similar assemblages of transverse landforms also occur, but they are much more scattered. This type of deglacial terrain, in which eskers are notably scarce, occurs mainly in the eastern part of the region (Fig. 2).

TERRESTRIAL ICE-MARGINAL ENVIRONMENTS

The third type of deglacial terrain occurs within the area occupied by the Bois-Francs residual ice cap (Parent and Occhietti, 1988). In this terrestrial deglacial environment, very few ice-marginal landforms and deposits have been reported, and there are almost no eskers. In fact, only one esker has been reported from within the limits of the residual ice cap (Fig. 2): The Glen Lloyd esker, a small esker deposited by northward-flowing meltwaters (Lortie, 1976; Gadd, 1978) near the northern margin of the residual ice cap. Available evidence suggests that deglaciation took place mainly by regional stagnation and downwasting; meltwaters were simply evacuated from ice margins through drainage routes similar to present. In the Thetford-Mines area, unlike adjacent regions to the southwest, a thick melt-out till unit is commonly present at the top of stratigraphic sections (Chauvin, 1979); this provides further evidence for downwasting and stagnation of the residual ice cap. Late deglaciation of the central area covered by the Bois-Francs residual ice cap is also suggested by a date of only 10930 ± 140 years BP (δ ^{13}C = −24.3‰, Beta-50190; P.J.H. Richard, written comm., nov. 1998).

GLACIAL LAKES

GLACIAL LAKE MEMPHREMAGOG: SHERBROOKE PHASE

The development of proglacial lakes remains the cornerstone of the deglacial record of the Appalachians of southeastern Québec. McDonald (1968) first recognized that because the Laurentide ice front retreated northwestward across the region, normal stream drainage was blocked and proglacial lakes were impounded in valleys and adjacent uplands southeast of the ice front. As lower outlets became ice-free during northwestward glacial retreat, the level of ice-dammed lakes fell; individual water planes were thus generally short-lived. One of the largest ice-dammed lakes that formed in the region is Glacial Lake Memphremagog. This high-level lake was originally recognized by Hitchcock (1907) on the basis of discontinuous shoreline features and sediments at elevations ranging from 391 to 296 m in the upper Lake Memphremagog watershed in Vermont. As the ice margin retreated northwestward from the Vermont-Québec boundary, lake level fell in three main steps to 365, 320 and 305 m levels (Boissonnault and Gwyn, 1983). Further retreat of the ice margin uncovered a lower outlet (249 m) at Lac Nick (Fig. 3); this marked the beginning of the Sherbrooke Phase of Glacial Lake Memphremagog (McDonald, 1967, 1968). During this phase, the glacial lake drained via the Lac Nick spillway and the Missisquoi River system into the Lake

Champlain basin, which was occupied by Glacial Lake Vermont prior to the Champlain Sea incursion (Chapman, 1937; Connally and Sirkin, 1973; Parrott and Stone, 1972).

The Sherbrooke Phase was in existence at the time of emplacement of the Cherry River Moraine and its level was maintained until the ice margin had retreated from high bedrock ridges near Richmond (McDonald, 1968). Subsequent fieldwork by Parent (1987) revealed that the East-Angus Moraine is a correlative of the Cherry River Moraine and that an additional recessional morainic belt, the Mont Ham Moraine (Fig. 2), formed during the Glacial Lake Memphremagog episode. In the Saint-François River valley, this morainic belt consists of a series of cross-valley ridges mainly composed of stratified drift and deposited in deep waters of the Sherbrooke Phase. Water levels fell rapidly in the region (Fig. 4) after the ice margin retreated from the position of the Mont Ham Moraine, which thus marks the maximum extent of the Sherbrooke phase. The presence of ice-contact deltas in the upper Saint-François valley (Figs. 3 and 4, sites 14 and 15) that were deposited at elevations corresponding to the level of the Sherbroke Phase confirms the correlation of morainic segments between the upper and middle Saint-François valley (Figs. 2, 3).

The shoreline diagram (Fig. 4) shows that glacioisostatic rebound has tilted upward to the northwest the shoreline features of Glacial Lake Memphremagog at a rate of about 1.2 m·km^{-1}. A transitional water plane, controlled by an outlet at 251 m near Valcourt (Figs. 4, 5: site C), formed about 40 m below the strandlines of the Sherbrooke Phase. This transitional glacial lake was fairly short-lived and rather local in extent. Further ice retreat caused lake level in the Saint-François valley to fall by another 20 meters to the level of Glacial Lake Candona (Fig. 4).

GLACIAL LAKE CANDONA

Regional correlations, which were fully discussed previously (Parent and Occhietti, 1988), have shown that the level of the next series of glaciolacustrine features in the Saint-François valley coincides with the Fort Ann Phase of Glacial Lake Vermont (Chapman, 1937). Our reconstruction of Glacial Lake Candona (Fig. 5), a simplified version of our earlier one (Parent, 1987; Parent and Occhietti, 1988), shows that this regional water plane extended not only to the Lake Champlain valley but also to lowlands southwest of Montréal (Clark and Karrow, 1984; Anderson et al., 1985). Subsequent work on invertebrate microfossils in early postglacial sediments (Rodrigues, 1992) confirmed the extension of this freshwater lake to the Ottawa and upper St. Lawrence valleys. The shoreline features assigned to the Fort Ann Phase in Figure 4 are from the Rivière Saint-François valley; they were listed with the same numbering in our previous paper (Parent and Occhietti,1988) and will thus not be listed here, nor will features assigned to a local transitional water-plane (Fig. 4). Details of the latter may be found in Parent (1987). Our reconstruction shows that the Lake Candona (Fort Ann) shoreline features are tilted up to the northwest at a rate of about 1.2 km·km^{-1}, much the same as that observed for the shorelines of Glacial Lake Memphremagog. This suggests

FIGURE 4. Shoreline diagram for the Sherbrooke Phase of Glacial Lake Memphremagog (a) and for Glacial Lake Candona (b). Line of section and location of shoreline features of Glacial Lake Memphrémagog are given in Figure 3 with the same symbols. Lake Candona features follow the same numbering as in Parent and Occhietti (1988). Dashed vertical lines show the probable measurement error of elevation data.

Diagramme des paléorivages du Lac glaciaire Memphrémagog (Phase Sherbrooke) et du Lac glaciaire Candona. Le tracé de la coupe et la localisation des paléorivages sont donnés à la figure 3 avec les mêmes symboles. Les paléorivages du Lac Candona suivent la même numérotation que dans Parent et Occhietti (1988). Les lignes verticales tiretées montrent l'erreur de mesure probable des données d'altitude.
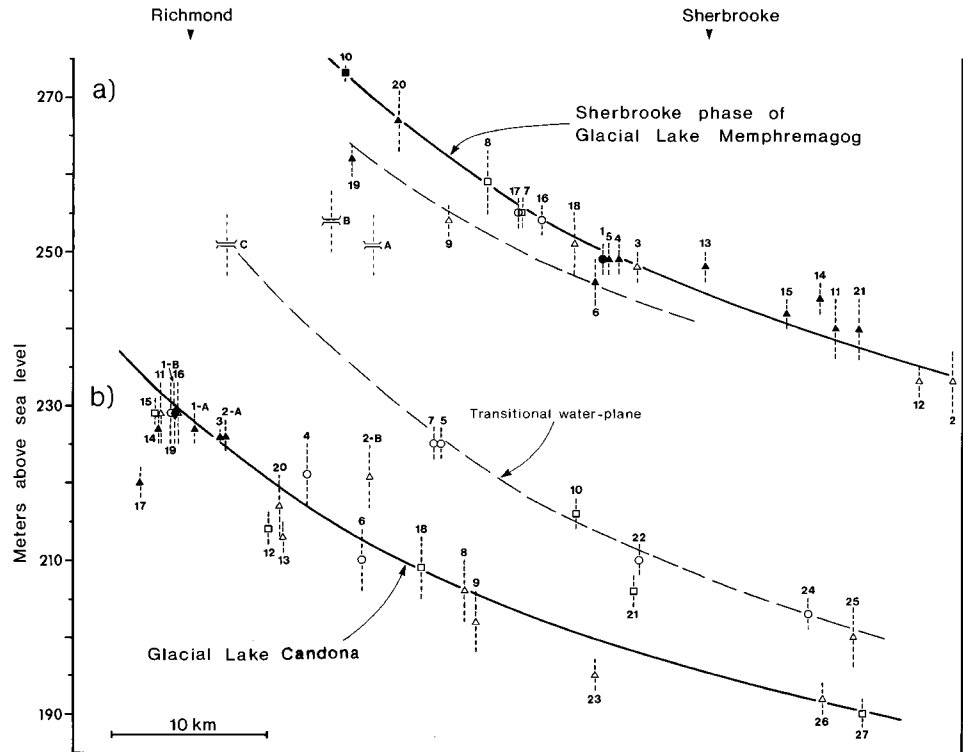
that little change in glacial isostatic unloading conditions had taken place during the time interval from Glacial Lake Memphremagog to Lake Candona and thus that these two sets of shoreline features formed in rapid succession.

Parent and Occhietti (1988) proposed a new name, Glacial Lake Candona, to designate the large glacial lake (Fig. 4) that preceded the late-glacial marine incursion (Champlain Sea) in the central St. Lawrence Lowlands. This name was chosen on the basis of a key ostracod species, *Candona subtriangulata*, found in its sediments at Rivière Landry (Parent and Occhietti, 1988). In a subsequent paper, Rodrigues (1992) designated the same water body (his Fig. 17) by a different name, Lake St. Lawrence, a name which he reintroduced from a paper by Upham (1895). In this paper, Upham referred to an ice barrier in the vicinity of Québec City[1] that impounded the glacial lake and prevented marine waters from entering the lowlands, a concept that has been repeated in subsequent Quaternary literature. However, it is precisely the ill-defined, poorly documented nature of this pre-Champlain Sea glacial lake that led to its near-dismissal (*e.g.* Gadd, 1983, 1988a, 1988b) and that led us to reinvestigate glaciolacustrine shoreline features and sediments south of the Champlain Sea basin (Parent, 1984a, 1987). In fact, one of the key points of our earlier paper (Parent and Occhietti, 1988) was that the ice barrier was not at all in the vicinity of Québec, as had been so often shown or proposed in earlier reconstructions (*e.g.*, Prest, 1970; LaSalle *et al.*, 1977b; Dyke and Prest, 1987). Rather,

our reconstruction showed that the ice-front which impounded Lake Candona extended from the Montréal region to the vicinity of Warwick (Parent and Occhietti, 1988; Fig. 5 of this paper), roughly 130 km south of Québec City. Our reconstruction also showed that the Québec City region had already been invaded by an early arm of the Champlain Sea, the Charlesbourg phase (12.4-12.0 ka), at the time of the Lake Candona episode. The paleontologic work of Rodrigues (1992) confirms that our previous reconstruction of the limits of Lake Candona was essentially correct, both in terms of the inferred position of the impounding ice margin and in terms of shoreline correlation. Therefore we propose maintaining the name Lake Candona to designate the glacial lake which resulted from the coalescence of glacial lakes Vermont, Iroquois and Memphremagog and which extended northward as the ice retreated from Covey Hill and the Appalachian uplands (Parent and Occhietti, 1988). This name has the distinct advantage of providing a firm link between the shoreline record of the glacial lake and its distinctive faunal record, to which Rodrigues (1992) has made significant additional contributions.

DANVILLE VARVES

RIVIÈRE LANDRY SECTION

The occurrence of pre-Champlain Sea glaciolacustrine sediments in southeastern Québec has been recognized for some time (McDonald, 1967; Gadd *et al.*, 1972; Warren and Bouchard, 1976); however, few, if any, good varve sections have been described in published reports. The most significant varve section reported by Parent (1987) is at

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1. The lake proposed by Upham (his Plate I) actually extended from the Abitibi region to the highlands north of Québec City.
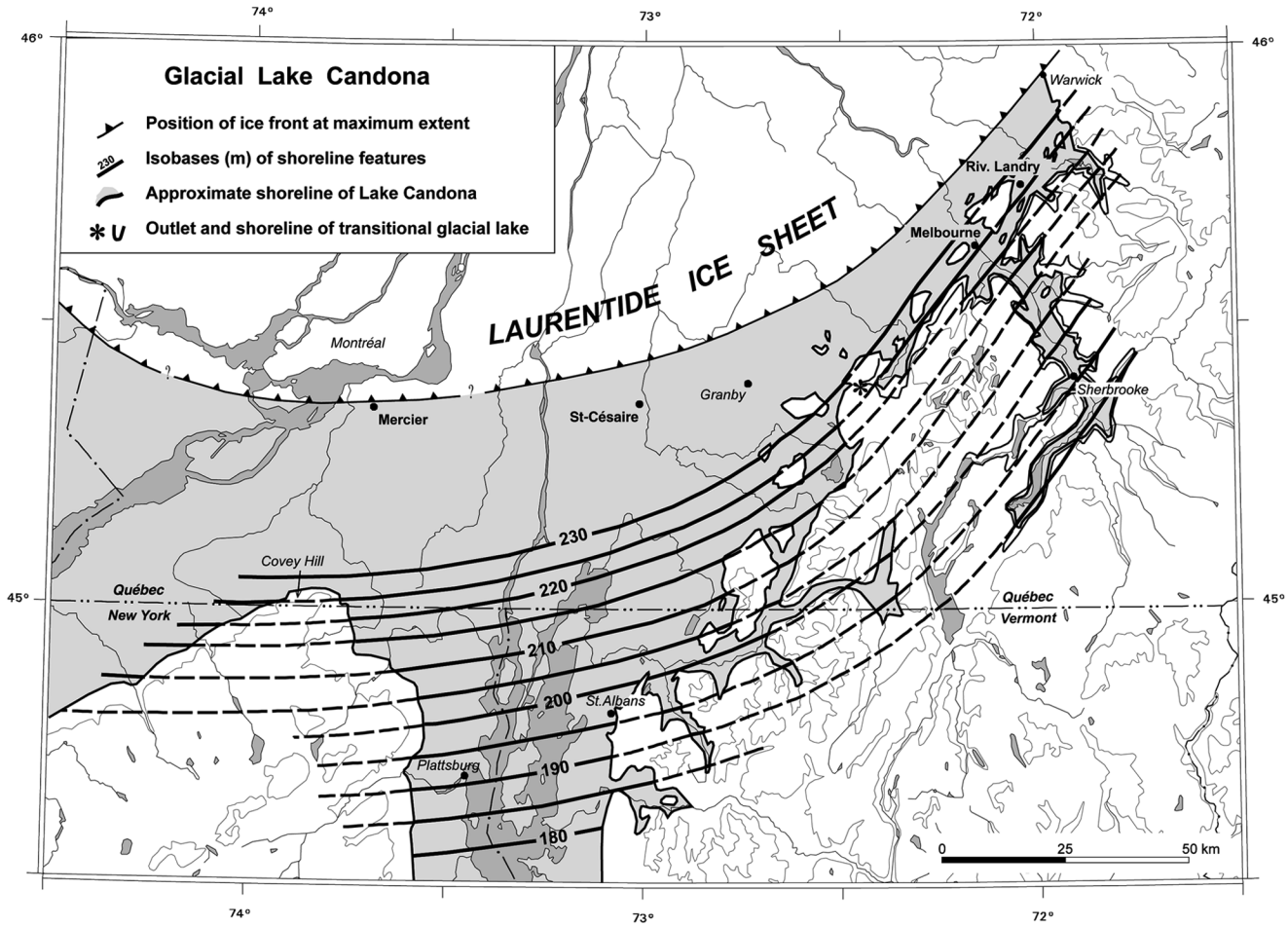
FIGURE 5. Glacial Lake Candona, showing isobases of the warped shorelines features and the location of sections discussed in text (after Parent and Occhietti, 1988). The lake outlet was located at Fort Ann (Chapman, 1937), near Glens Falls, about 75 km south of map area.

Le Lac glaciaire Candona. La carte montre les isobases du paléorivage gauchi par suite du relèvement isostatique différentiel ainsi que la localisation des coupes dont parle le texte (d'après Parent et Occhietti, 1988). L'exutoire du lac était situé à Fort Ann (Chapman, 1937), près de Glens Falls, environ 75 km au sud de la région montrée.

Rivière Landry (Fig. 6), a section which was subsequently resampled by Rodrigues (1992). This varve series was deposited in the northeastern part of Lake Candona (Fig. 5), about 5 km northwest (upglacier) of the Ulverton-Tingwick Moraine. The section is located on the right bank of Rivière Landry, in a sharp meander bend, about 1.2 km northwest (333°) from the 4-way intersection in Danville (Québec); coordinates are 45°47'44" N and 72°01'49" W. At Rivière Landry, the glaciolacustrine sequence has an exposed thickness of 9 m and consists of silt and clay varves, the upper part of which contains sparse valves of *Candona subtriangulata* (Fig. 6). Hand-augering has revealed that the coarse basal varves extend to a depth of 2 m below river level, at which point they grade into a massive fine sand unit. This unit is similar to sandy turbidites which directly overlie the regional surface till in a small riverbank section located about 200 m upstream from the main section. The varves are overlain by 8 m of faintly laminated to laminated, fossiliferous (*Macoma balthica* and

Hiatella arctica, plus microfossils) marine silt which are in turn overlain by a 2 m-thick regressive sand unit (Fig. 6).

The exposed varve sequence consists of 103 couplets whose average thickness (5-year running mean) decreases from about 25 cm near the base to about 2.5 cm near the top (Fig. 7); however, this overall thinning-upwards trend, which is dominated by the summer layer record, shows several secondary cycles. Summer layers consist mainly of plane-laminated, light gray, calcareous, slightly bioturbated silt. Partings of very fine sand are present within almost all summer layers; in two varves, these partings show distinct, NW-SE trending, parallel lineations that were presumably produced by turbidity currents. On the basis of examples given by Banerjee (1973) and Ashley (1975), these varves are interpreted as turbidites with (C)DE$_{(t,h)}$-divisions (nomenclature of Walker, 1984). The attenuated covariation of winter and summer layer thicknesses in the entire varve series (Fig. 7) suggest that the winter layers consist of a combination of distal turbiditic mud (E_t) and hemipelagic mud (E_h). In the lower 3.5 m of the varve series (varves #4,
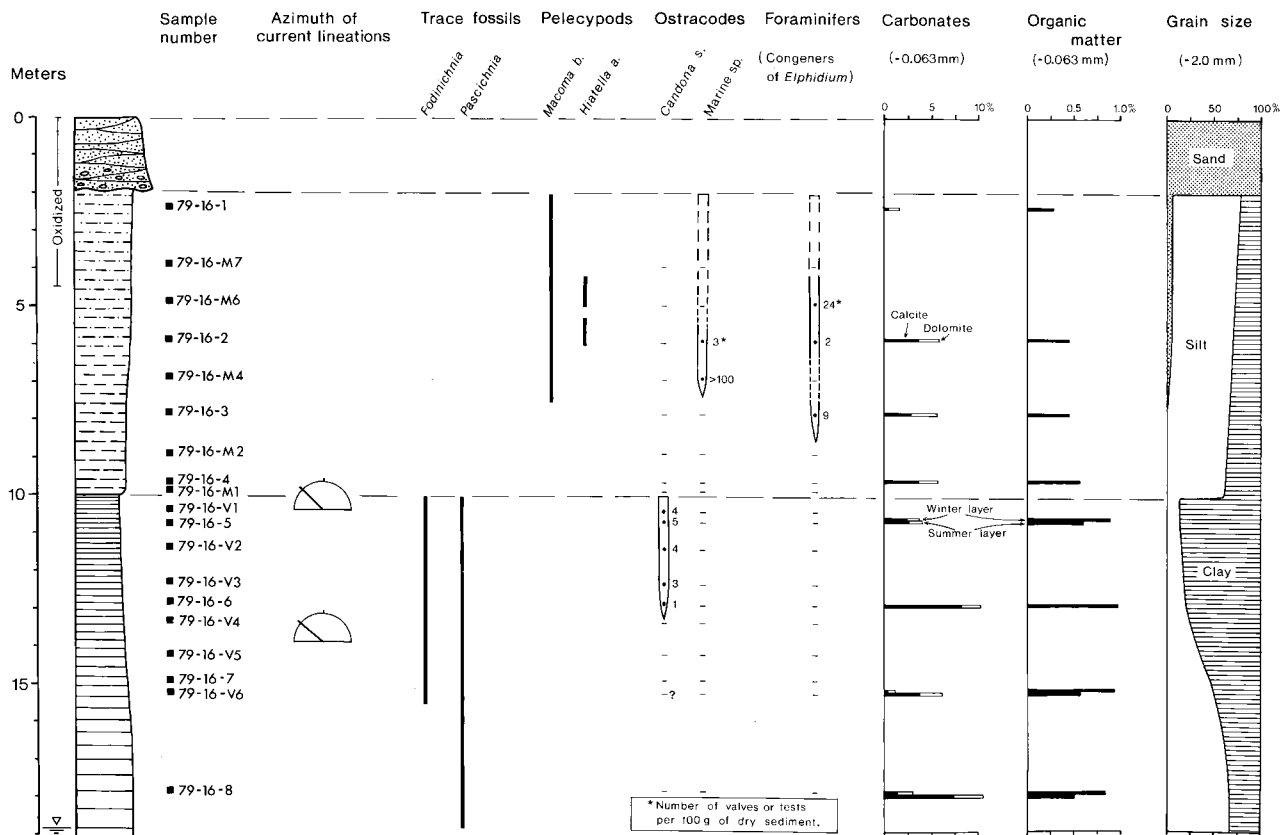
FIGURE 6. Faunal and sedimentologic record of the Rivière Landry type section. The Danville Varves, which extend from a depth of 10 m down to river level, are overlain by Champlain Sea silt (2-10 m) and regressive sand (0-2 m).

Enregistrement faunique et sédimentologique de la coupe type de la rivière Landry. Les Varves de Danville, qui s'étendent depuis une profondeur de 10 m jusqu'au niveau de la rivière, sont surmontés par les silts de la Mer de Champlain (2-10 m) et des sables régressifs (0-2 m).

7, 12 and 17), thick convoluted summer layers indicate extensive slumping in the earlier history of the basin. A plot of mean grain size (M_z) versus depth shows that the thinning-upwards trend is paralleled by a fining-upwards trend which is best recorded within summer layers (Fig. 8). The winter layers consist mainly of thinly laminated or faintly laminated, stiff, non-calcareous, dark gray clay.

Annual rhythmicity of the varves is demonstrated by the fact that the occurrence of trace fossils is restricted to the coarser, light-coloured, summer layers. Trace fossils are most common in sandy partings that are particularly abundant at the base and in the lower part of summer layers throughout the varve series; they consist mainly of grazing traces (*Pascichnia*), but crawling traces (*Repichnia*) were also observed together with grazing traces in a few varves. Feeding structures (*Fodinichnia*) which are present within laminated silt layers above a depth of 15.5 m disappear abruptly at the contact with the overlying marine silt, much like *Pascichnia* and *Repichnia* (Fig. 6).

Above a depth of 13 m, the varves contain sparse valves of *Candona subtriangulata*, a benthic ostracode species which typically inhabits arctic freshwater lakes but may also be tolerant to low salinity environments (Cronin, 1977;

Rodrigues, 1992). However, the characteristic sedimentology of the varves and the absence of other ostracode or foraminifer species indicate the varves were deposited in a typical glaciolacustrine environment, a conclusion that is supported by the microfaunal record from other Lake Candona sections (Rodrigues, 1992). The Rivière Landry record shows that *Candona subtriangulata* first appears at a depth of 12.8 m (varve #41), while the sedimentation rate was still very high at 10 cm·a^{-1} (Fig. 7). A fragment of *Candona* sp. which was found in a sample in the lower varve series (#79-16-V6 at a depth of 15.2 m) may result from sample-to-sample contamination during laboratory work. Sedimentation rates fell to about 3 cm·a^{-1} during the second half of the Lake Candona episode. Varve samples were found to contain abundant insoluble micro-concretions and a few globular, agglutinated forms of probable biogenic origin.

Together with *Candona subtriangulata*, trace fossils and agglutinated forms disappear at the contact with the overlying soft marine silt; no ostracodes nor foraminifers were found in samples collected in the lower 2 m of the marine unit. This indicates that the onset of marine conditions virtually eliminated the *Candona subtriangulata* fauna from deepwater benthic environments and that a significant time inter-
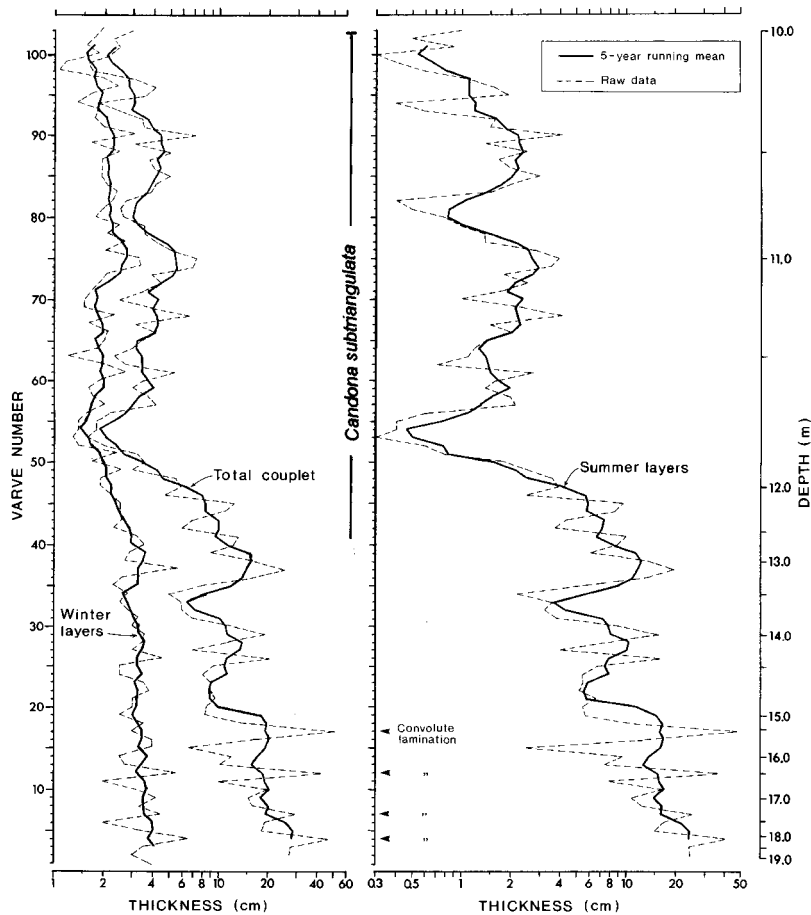
FIGURE 7. Varve diagram at the Rivière Landry section. The Danville Varves record a significant decrease of the sedimentation rate during the last 50 years of Lake Candona. (Note that the thickness data are plotted on a logarithmic scale.)

Diagramme varvaire à la coupe de la rivière Landry. Les Varves de Danville ont enregistré une chute importante du taux de sédimentation durant les 50 dernières années du Lac Candona. (N.B. L'échelle des épaisseurs est logarithmique.)

val was required before a benthic marine fauna could migrate into this part of the marine basin. The onset of marine conditions also caused an abrupt change of sedimentary conditions. Stiff varved clay layers with a thickness of about 2 cm and mean grain size of 1 to 2 μm were replaced by soft, faintly laminated silty clays with a mean grain size of 5 to 6 μm (Fig. 8). The transition to the typical marine silt is marked by a 20 cm-thick bed of thin sand and silt rhythmites.

The earliest marine microfossils recorded in the Rivière Landry section are a few tests of *Elphidium* sp. (sample 79-16-3); except for a single test of *Virgulina loeblichi* in sample 79-16-M6, other foraminiferal tests found in the marine silt samples are all congeners of *Elphidium* and probably belong to the same species (*Elphidium clavatum*). Ostracodes were found in two samples from the marine unit; *Cytheropteron montrosiense* (form 1) and *Cytheropteron inflatum* are the most abundant species. Several valves of *Eucytheridea punctillata* and two valves of *Cytheromorpha macchesneyi* were also found in sample 79-16-M4; however, the next sample above (79-16-2) contained, in addition to *Elphidium* sp., only a few valves of *Cytheropteron latissimum*. According to Cronin (1981), these ostracode species belong to faunal assemblages that characterized frigid to subfrigid, polyhaline to euhaline environments during the

early Champlain Sea episode (prior to about 11 000 years BP). Subsequent work by Rodrigues (1988, 1992) led to a significant refinement of Champlain Sea microfaunal assemblages. The assemblage described above corresponds to his zone 3 assemblage (Rodrigues, 1992), which characterized early low-salinity marine environments. Resampling of the Rivière Landry section by Rodrigues (1992) led essentially to the same record as that of our earlier report (Parent, 1987).

Above a depth of 4.5 m, the marine silt unit contains sandy interbeds whose thickness varies from 2 to 3 cm. The upper part of the marine unit is oxidized and seems to be barren of microfossils, presumably as a result of carbonate dissolution by meteoric waters. However several molds of *Macoma balthica* (in growth position) were observed in silt layers up to the base of the regressive sand unit (Fig. 6).

MELBOURNE SECTION

Pre-Champlain Sea varves were also exposed in a borrow pit located near Melbourne (Fig. 5), in the Rivière Saint-François valley, about 18 km southwest of the Rivière Landry site. The Melbourne section (45°40'02" N and 72°09'46" W), which was subsequently destroyed as a result of pit operations, is located close to the valley floor, at an elevation of 122 m, well below the upper marine limit which is about 165 m in this region (McDonald, 1968; Parent and Occhietti,
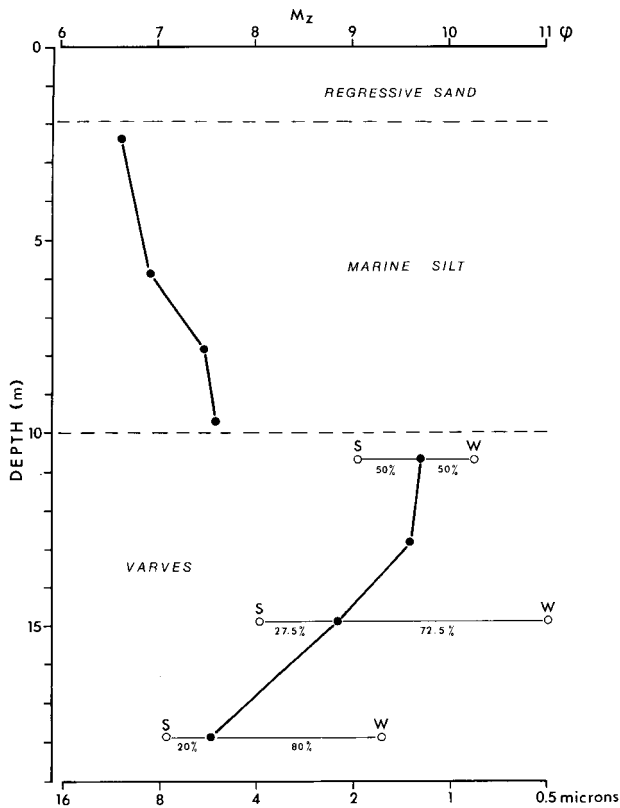
FIGURE 8. Main granulometric trends at the Rivière Landry section. The varves show a fining-upward trend while the marine silts show a coarsening-upward trend; notice the sharp break of mean grain-size (M_Z) at the contact between the two units. Solid dots in the varve series represent weighted average for total varve couplet ; percent figures represent weighting factors based on thickness data for summer (S) and winter (W) layers.

Principales tendances granulométriques à la coupe de la rivière Landry. Les varves montrent une décroissance granulométrique vers le sommet alors que les silts marins montrent une tendance inverse ; noter la discontinuité marquée de la moyenne granulométrique (M_Z) au contact des deux unités. Les cercles pleins dans la série varvaire représentent la moyenne pondérée d'un couplet ; les valeurs en pourcentage sont les facteurs de pondération dérivés des données d'épaisseur des lits d'été (S) et d'hiver (W).

1988). The section, which was exposed on the sides of a large remnant in a gravel pit (Fig. 9), showed a 7 m-thick varve unit overlying ice-contact stratified drift.

The varves displayed two distinct fining-and-thinning-upward series. The lower series consisted of 33 varves, while the upper series consisted of at least 23 more varves but was truncated at the top as a result of pit operations. As at Rivière Landry, the varves were counted on the basis of their distinctive, clay-rich winter layer. In the lower series, total varve thickness decreased exponentially from 110 cm in varve #1 to 2.5 cm in varve #33. The lower varves are typical turbidites commonly consisting of (B)CDE$_{(t,h)}$-divisions, with the summer layer of varve #1 consisting of 4 partial Bouma sequences with (A)BCD divisions. Ripple laminations show that paleocurrents were toward the southeast, up

the Rivière Saint-François valley, thus suggesting that the turbidity currents were generated by subglacial meltwaters rather than on the foreslope of deltas built by streams entering the lake.

In the upper series, total varve thickness decreases from 15 cm (varve #34) to 2 cm. Several summer layers of the upper series contained grazing traces (*Pascichnia*) similar to those observed at Rivière Landry, as well as numerous discoid calcareous concretions. Paleocurrents were also toward southeast during deposition of the upper varve series. In both varve series, winter layer thickness decreases only slightly, ranging from about 3 cm to 2 cm near the top of the section.

TYPE SECTION FOR PRE-CHAMPLAIN SEA VARVES

We propose the Rivière Landry section as the type section for pre-Champlain Sea varves in the St. Lawrence valley. The sedimentary and faunal records of this section (Parent, 1987; Rodrigues, 1992) are closely linked with the regional Late Wisconsinan history. The varves exposed in the lower half of the section are an important lithostratigraphic unit, as pointed out in our earlier paper (Parent and Occhietti, 1988) and also by Rodrigues (1992). We propose the name Danville Varves to designate this lithostratigraphic unit which typically contains a single microfossil species, *Candona subtriangulata*, in low abundance and which provides an unequivocal record of glaciolacustrine conditions prior to Champlain Sea incursion in the St. Lawrence valley. Correlative units have been reported from several boreholes in the Ottawa valley (Anderson *et al.*, 1985), at the Melbourne (Parent, 1987a) and Saint-Césaire (de Vernal *et al.*, 1989) sections in southeastern Québec, in the Foster Pit (Naldrett, 1986, 1988a, 1988b) and at the Twin Elm, Bearbrook, Casselman and Sparrowhawk Point sections (Rodrigues, 1992) in the Ottawa-Upper St. Lawrence Valley. In addition to its excellent sedimentological and microfaunal record, the proposed stratotype has the advantage of being readily accessible and unlikely to be ruined or lost in the foreseeable future.

CHRONOLOGY OF DEGLACIATION

The most reliable chronological marker for regional deglaciation is the beginning of the Champlain Sea episode, which was an isochronous event within the region where marine waters directly replaced glaciolacustrine waters of Lake Candona (Parent and Occhietti, 1988). Within that region (Fig. 5), available ^{14}C dates for early marine faunas may be subdivided in two groups (Table II). The first group consists of 10 dates which were obtained from faunal assemblages that are consistent with early marine conditions and that come from sites located within the region of isochronous marine incursion. The second group includes (1) early faunas from sites located north of the zone of isochronous incursion (GSC-5854, GrN-1697, TO-703) and (2) faunas recording early marine regression (UQ-290, GSC-4934, UQ-29, UQ-1429, W-2311, W-2311, TO-704). On the basis of the tight clustering of Group I dates between 12000 and 11665 years BP (mean = 11712; s.d. = 190), and allowing
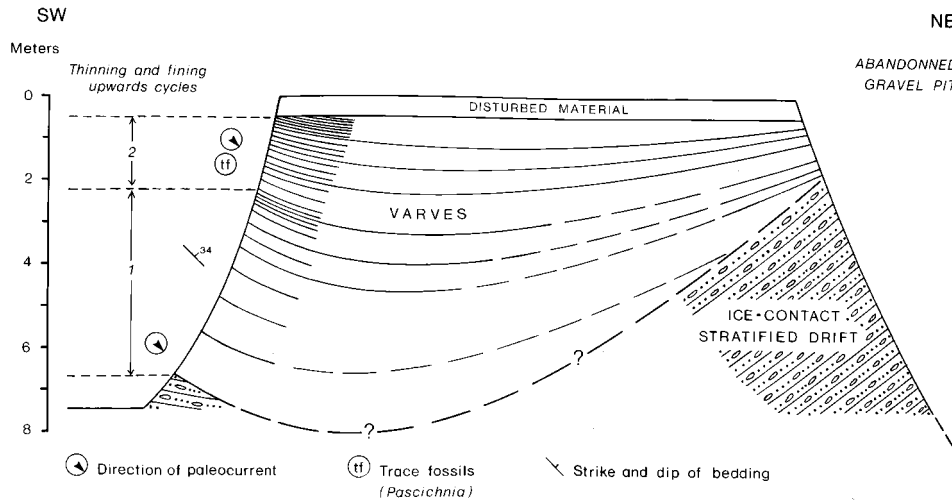
SW

Meters

Thinning and fining upwards cycles

DISTURBED MATERIAL

VARVES

34

NE

ABANDONNED GRAVEL PIT

ICE·CONTACT STRATIFIED DRIFT

?

?

?

2

1

Direction of paleocurrent

tf Trace fossils (Pascichnia)

Strike and dip of bedding

FIGURE 9. At the Melbourne section, the Danville Varves overlie ice-contact stratified drift and display two main thinning and fining upward cycles.

À la coupe de Melbourne, les varves de Danville surmontent des sédiments juxtaglaciaires et montrent deux mégaséquences positives.

for a slight delay between initial marine incursion and the development of datable faunas, the beginning of Champlain Sea is estimated at 12 000 years BP (corrected shell age).

This age estimate is based on shell dates that were published by different laboratories and that do not constitue a rigorously homogeneous data set, as shown in Table II. At the time that many of these dates were reported, $\delta^{13}C$ values were not commonly measured and most laboratories reported uncorrected ages, *i.e.* they did not normalize shell ages to $\delta^{13}C = -25‰$ (+ 410 years) and did not incorporate a correction for the marine reservoir effect (−400 years). The rationale was that the two corrections cancelled each other out. On the other hand, the Geological Survey of Canada Radiocarbon Dating Laboratory, which has carried out a large number of the Champlain Sea dates, has been correcting shell ages to $\delta^{13}C \approx 0‰$ (measured or estimated). Because the uncorrected and corrected ages for marine shells fall very close to each other in most cases, this lack of standard reporting procedure has had only a minor impact on the homogeneity of previously published listings of ^{14}C dates for this part of the Champlain Sea (*e.g.* Parent and Occhietti, 1988: their Table III). More recently, in order to be consistent with the reporting procedures of most of the other laboratories, the GSC Laboratory has added a normalized age ($\delta^{13}C = -25‰$) to its published shell dates, and has undertaken to update its existing shell date database, depending on the availablility of $\delta^{13}C$ measurements for previously dated samples.

Of the dates available within the region (Table II), a single date (QC-474: 12 480 ± 240 years BP, uncorrected), obtained on *Mya* sp. shells (Prichonnet, 1982), has been rejected. Detailed investigation of this site indicates that the enclosing material is a regressive silty sand and that the dated specimens were *Mya arenaria*, a species which inhabits exclusively sandy littoral environments, a setting which is incompatible with an age of 12 480 years BP at an elevation of only 63 m. Re-dating of the site has yielded an age of 11 250 ± 100 years BP (UQ-1429), which is in much better agreement with the local sedimentologic context (Parent and

Occhietti, 1988) but is still several hundred years older than other *Mya arenaria* assemblages from the Champlain Sea basin.

Given an estimated age of 12 000 years BP for the beginning of the Champlain Sea and given the 103 year varve record at the Rivière Landry stratotype, the age of the Ulverton-Tingwick Moraine, which lies 5 km southeast (downglacier) of the Rivière Landry site, may be estimated at 12 100 years BP. Since we also know that Glacial Lake Candona drained when the ice front retreated from the vicinity of Warwick (Figs. 4, 5; see also Parent, 1987; Parent and Occhietti, 1988), the rate of ice retreat during the Lake Candona episode may be inferred. Since the glaciolacustrine episode lasted slightly more than 103 varve years at the Rivière Landry site (refer to section on the Danville Varves) and since the lake drained when the ice front had retreated about 20 km northwest of this site, the estimated rate of ice retreat is about 200 m·a^{-1}.

The deglacial record of the region indicates that the earlier morainic belts (Mont Ham, Cherry River–East-Angus, Dixville) are essentially recessional features; their age may also be inferred on the basis of a retreat rate of 200 m·a^{-1}. The age of the Dixville Moraine, which lies 80 km southeast of the Ulverton-Tingwick Moraine, may thus be estimated at about 12 500 years BP. Applying the same procedure yields an approximate age of 12 325 years BP for the Cherry River–East-Angus Moraine, and 12 200 years BP for the Mont Ham Moraine. Since the Frontier Moraine of the nearby Lac Mégantic region (Shilts, 1981) is thought to be slightly older than the Dixville Moraine (Parent and Occhietti, 1988), its age may range between 12 550 and 12 600 years BP.

Because appropriate local marine ^{14}C reservoir corrections have yet to be established for the Champlain Sea, the estimated ages for the recessional moraines of the southeastern Québec uplands are based on the beginning of the marine incursion, 12 000 years BP, which is an age corrected for a mean global surface ocean reservoir age (R) of approximately 400 years (Stuiver and Brazunias, 1993). Local deviations (ΔR), caused for instance by the incomplete

TABLE II

Radiocarbon dates on early Champlain Sea shells from the Appalachian piedmont and adjacent regions

Laboratory number	Published age * (years BP)	Elevation (m ASL)	Locality	Dated material (species)	Enclosing sediment	Collector	Reference
Group I - Earliest Champlain Sea fauna, southeastern Québec and adjacent Lake Champlain valley							
I-13342 (50% leach)	11 700 ± 170	127	Warwick, Qc	*Hiatella arctica*	Reworked esker gravel	Parent	Parent and Occhietti, 1988
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
GSC-505 (10% leach)	*11 880 ± 180	122	L'Avenir, Qc	*Macoma balthica* (90%)	Pebbly gravel and sand	McDonald	Lowdon and Blake, 1970
GSC-936 (50% leach)	*12 000 ± 230						
GSC-475 (10% leach)	*11 530 ± 160	145	Ste-Christine, Qc	*Hiatella arctica* (mostly)	Silt	McDonald	Lowdon and Blake, 1970
GSC-475-2 (50% leach)	*11 500 ± 160						
I-4489	11 740 ± 200	145	Frelighsburg, Qc	*Macoma balthica*	Silty lens in deltaic sediment	?	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
Group II - Other early Champlain Sea fauna, southeastern Québec and adjacent Lake Champlain valley							
GSC-5854	*11 500 ± 100	79	St-Sylvère, Qc	*Hiatella arctica*	Stratified silt	Occhietti	Hétu *et al.*, 1995
GrN-1697	11 490 ± 110	171	Mont Royal, Qc	Unidentified shells	Unspecified	Elson	Elson, 1962, 1969
TO-703	11 530 ± 90	15	Ste-Monique-de-Nicolet, Qc	*Portlandia arctica*	Massive mud	Rodrigues	Rodrigues, 1992
UQ-290	11 370 ± 200	149	Danville, Qc	*Macoma balthica*	Deltaic sand (offlap)	Parent	Parent and Occhietti, 1988
GSC-4934	*11 100 ± 90	137	Warwick, Qc 45°56'35" N; 72°00'09" W	*Hiatella arctica*	Sand and gravel (offlap)	Parent	Unpublished
UQ-29	11 360 ± 110	105	Adamsville, Qc	*Macoma balthica*	Offlap sand	Prichonnet	Prichonnet, 1982, 1984
QC-475	12 480 ± 240	61	St-Dominique, Qc	*Mya* sp.	Unspecified	Prichonnet	Prichonnet, 1982
UQ-1429	11 250 ± 100	63	St-Dominique, Qc	*Mya arenaria*	Sand and silt (offlap)	Occhietti	Parent and Occhietti, 1988
W-2311 (Shells) W-2309 (Wood)	11 420 ± 350 10 950 ± 300	58	Burlington, Vt	Unidentified pelecypods Unidentified wood	Silt and caly	Hunt and Wagner	Wagner, 1972 Spiker *et al.*, 1978
TO-704	10 970 ± 60	40	St-Césaire, Qc	*Balanus hameri*	Pebbly sandy mud	Rodrigues	Rodrigues, 1992

* Previously published GSC dates on shell material were corrected to $\delta^{13}C \simeq 0$‰, which assumes a marine reservoir age of 400 years. Dates from the other laboratories were published without correction for fractionation (normalization to $\delta^{13}C = -25$‰) or for marine reservoir age.

mixing between incoming marine waters and freshwaters from Lake Candona, may explain some of the discrepancies between our proposed chronology and the chronology of deglacial features in adjacent northern New England. Our marine-based age of 12 500 years BP for the Dixville Moraine is indeed almost the same as the age assigned to the Bethlehem Moraine (12 400 years BP; Thompson *et al.*, 1996) on the basis of terrestrial dates. If a rate of ice retreat of about 200 m·a^{-1} is again applied to the 80 km distance separating the two moraines, the offset between the two chronologies is on the order of 400 to 500 years, which gives us an indication of what the local ΔR value might be. This value is about the same as the age discrepancy noted by Anderson (1988) on the basis of pollen zones within Cham-

plain Sea sediments in the Ottawa region. The application of a provisional local marine reservoir correction (ΔR) of 450 years would yield ages of 11650, 11750, 11875 and 12050 years BP respectively for the Ulverton-Tingwick, Mont Ham, Cherry River–East-Angus and Dixville moraines.

The application of this local marine reservoir correction would also reduce to less than 1000 years the time gap between the inferred age of deglaciation and the onset of organic deposition in lakes and ponds of the southeastern Québec uplands, about 11000 years BP (Richard, 1977; Mott, 1977). This age was based on conventional basal dates that have been recently confirmed by new AMS dates, *e.g.* 11140 ± 120 years BP (TO-6330) at Lac Albion (P.J.H. Richard, written comm., 1998). In spite of the attractiveness of these locally corrected ages (R + ΔR) for the main regional deglacial events, it seems preferable to continue using the chronology based on the mean global surface ocean reservoir correction (R), at least until the question of local marine ^{14}C reservoir corrections within the Champlain Sea basin has been investigated more thoroughly.

THE ST. LAWRENCE ICE STREAM

The fact that glacial striae generated by the Appalachian ice-flow reversal extend into a region of well-documented northwestward ice retreat, as shown in Figure 2, has remained a dilemna. To resolve it, some authors (Gauthier, 1975; Shilts, 1981) have favored a readvance to the position of the Highland Front Morainic System, a series of ice-contact deposits originally delineated by Gadd *et al.* (1972). This explanation had the advantage of reconciling several key elements of the deglacial record, as known then. However this explanation has met with a major objection since it was first formulated: there was no stratigraphic evidence to suggest that the ice front readvanced southeastward across an early marine embayment. And subsequent investigations across key areas of the northwest Appalachian margin, from the Granby region to the Québec City region (Prichonnet, 1984; Prichonnet *et al.*, 1982; Parent, 1987; LaSalle and Chapdelaine, 1990; Shilts, 1997), have contributed to confirm that objection. Moreover, these investigations have led to the collapse of the concept of the Highland Front Morainic System, which was in fact an amalgamation of diachronous ice-contact sediments bodies, a question that was discussed at length by Parent and Occhietti (1988).

However recent investigations on the north shore of the St. Lawrence River (Lanoie, 1995; Dionne and Occhietti, 1996; Cloutier *et al.*, 1997) have revealed that regional southward flow was reoriented toward east and northeast along the foothills of the Laurentian Highlands (Fig. 10). These observations of crosscutting glacial striae were recorded in a region extending for about 100 km both upstream and downstream from Québec City and at elevations up to about 300 m. It is reasonable to assume that these late re-orientations of ice flow are coeval with those observed south of the St. Lawrence River. The late-glacial ice-flow patterns shown in Figure 10 are based on the known distribution and trend of these sets of glacial striae in the

region extending from the Laurentian Highlands to the White Mountains (Chauvin *et al.*, 1985; Lortie and Martineau, 1987; Parent, 1987; Lowell, 1985; Rappol, 1993; Lanoie, 1995). These patterns clearly define a major late-glacial ice stream, which we will simply call the St. Lawrence Ice Stream (SLIS). In this context, the direct cause of the Appalachian ice-flow reversal is the development of an ice stream in the St. Lawrence valley, which may have been associated with rising sea level in the Gulf of St. Lawrence prior to about 13000 years BP.

Known ice-flow patterns suggest that the St. Lawrence Ice Stream had a width of about 25 km with its central axis offset toward the south shore of the St. Lawrence estuary in the region between Québec City and Rivière-du-Loup (Fig. 10). Beyond that point, the ice stream was probably centered in the estuary. Recognition of the SLIS provides an explanation for other regional geomorphic features, for instance the occurrence of eastward-trending crag-and-tail ridges in the Portneuf region (Parent *et al.*, 1998), on the north side of the St. Lawrence River about 50 km upstream from Québec City. Another example is the narrow belt of large streamlined strike ridges extending along the south shore of the St. Lawrence between Québec City and Rivière-du-Loup and which were probably swept clean of their cover of glacial sediments by the SLIS, well before the late-glacial marine incursion. The absence of subglacial streamlined features in till plains of the Appalachian piedmont east of the Rivière Saint-François valley (Figs. 2, 10) may have been caused by ice-flow reorientation in the upper "catchment" area of the SLIS.

However the most significant aspect of the SLIS is that it indicates the Appalachian ice flow reversal event occurred several hundred years prior to marine incursion in the vicinity of Québec City, thus allowing the reversal event to be reconciled with regional deglaciation patterns and chronology.

Figures 11 and 12 show regional schematic cross-sections (A-A' and B-B') at two key time intervals: 12500 years BP and 12100 years BP, respectively about 500 years and 100 years prior to Champlain Sea incursion. At 12500 years BP, section A-A' shows the ice margin shortly after it had retreated from marine waters in east-central Maine, while the northeastward flowing St. Lawrence Ice Stream was fully active some 200 km upglacier from the southern ice margin. The SLIS had drawn down the surface of the ice sheet surface by an unknown amount, presumably several hundred meters, and an ice divide had formed in northern Maine (Lowell, 1985). Although the base of the ice stream was only about 25 km wide, its surface width may have been twice as large. Judging from the width of the terrain in which strong northeastward ice-flow reorientation has been observed, the shear zones bordering the ice stream were probably several kilometers wide. At 12100 years BP, a residual ice cap in the Appalachian uplands had become detached from the ice sheet and, as suggested by several radiocarbon dates from the region between Québec City and Rivière-du-Loup (see Table III *in* Parent and Occhietti, 1988; also TO-948: 12450 ± 160 years BP *in* Rappol, 1993), an arm of the Goldthwait Sea had invaded the estuary. The Saint-Jean-Port-Joli
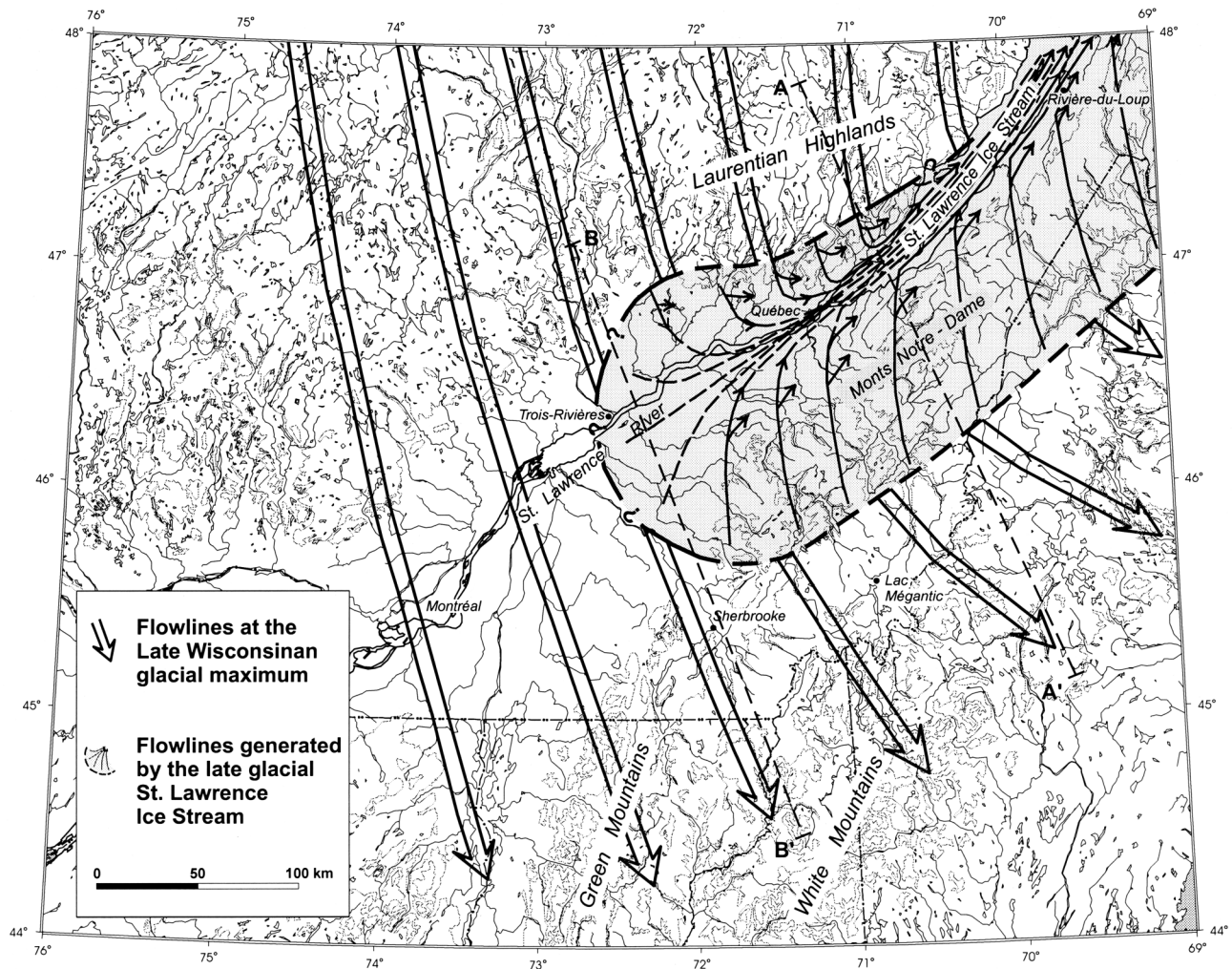
FIGURE 10. Reorientation of Late Wisconsinan flowlines generated by the St. Lawrence Ice Stream. Lines A-A' and B-B' show the location of schematic cross-sections of Figures 11 and 12.

Ré-orientation des écoulements glaciaires du Wisconsinien supérieur engendrée par le Courant glaciaire du Saint-Laurent. Les lignes A-A' et B-B' montrent la localisation des coupes schématiques des figures 11 et 12.

Moraine (Chauvin *et al*., 1985) was presumably being deposited at about that time along the northern margin of the residual ice cap. Although the ice sheet had thinned substantially over the Laurentian highlands, the ice front still remained well within the Goldthwait Sea.

Section B-B' is located about 120 km to the southwest (Fig. 10). The reconstruction for 12 500 years BP shows the ice margin standing in the vicinity of the Dixville Moraine (Fig. 12), while the SLIS had generated reorientation of flowlines throughout a vast region upglacier from the southern ice margin. In this reconstruction, the ice sheet surface in the Laurentian highlands is placed at about the same height as in section A-A'. The Appalachian ice flow reversal event may have been still taking place at about 12 500 years BP. Subsequent thermo-latitudinal ice retreat caused rapid dissipation of the ice divide; thus, the ice flow reversal event had likely already ceased by the time the ice margin retreated to the position of the Cherry River–East-Angus Moraine. By 12 100 years BP, as mentioned earlier, the ice margin had retreated

to the position of the Ulverton-Tingwick Moraine and the level of ice-dammed lakes in the Appalachian piedmont had fallen to that of Lake Candona. This reconstruction (Fig. 12) has the advantage of accounting for the ice-flow reversal event as well as all key aspects of the regional deglaciation patterns, including the persistence of southeastward meltwater flow in subglacial tunnels that deposited glaciofluvial sediment bodies such as the Warwick-Asbestos esker (Fig. 2). Moreover this reconstruction indicates the ice-flow reversal event was short-lived, a conclusion which is in agreement with the minute size of the northward striae throughout much of the region (Lamarche, 1971, 1974; Lortie, 1976).

CONCLUSION

Deglaciation of the Appalachian uplands north of the White Mountains took place mainly by backwasting of an active ice-front; however the meltwater-dominated record of ice-marginal landform-sediment assemblages, as well as the low ice-sur-
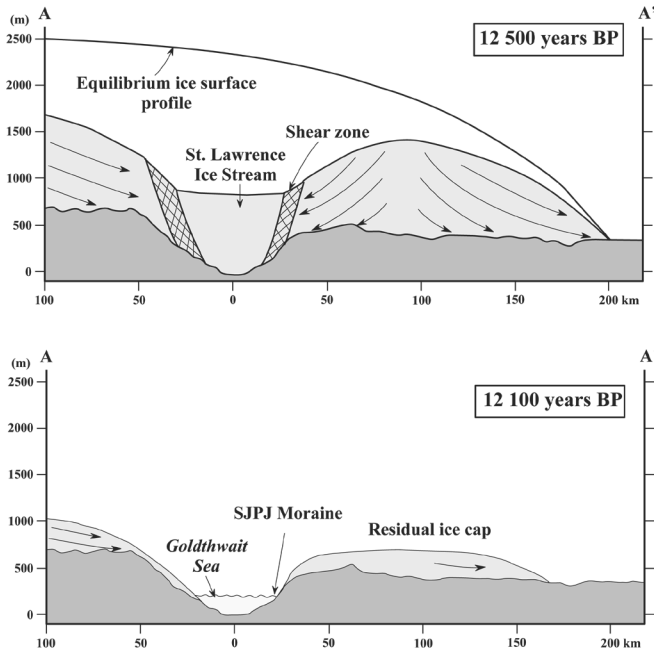
FIGURE 11. Schematic reconstruction of deglacial events at 12 500 and 12 100 years BP across the St. Lawrence estuary and adjacent highlands.

Reconstitution schématique des événements tardiglaciaires vers 12 500 et 12 100 ans BP à la hauteur de l'estuaire du Saint-Laurent et des hautes terres adjacentes.

face profiles revealed by these assemblages, suggest that the marginal zone of the Laurentide Ice Sheet was undergoing extensive ablation and was not very active. The lack of well-defined end-moraines, such as the St-Narcisse, further supports this interpretation. The pattern of thermo-latitudinal ice retreat in the region was extensively modified by ice-flow reorientation generated by the St. Lawrence Ice Stream, a major ice stream which formed in the St. Lawrence River estuary prior to 13 000 years BP and lasted until at least 12 400 years BP. Two main types of deglacial terrains may thus be recognized in the region: a) the southwestern part of the region was characterized by northward ice-marginal retreat and coeval development of large ice-dammed lakes; b) the northeastern part of the region was characterized by the downwasting of large, detached or partly detached ice masses, such as the Bois-Francs Residual Ice Cap and other residual ice masses east of the Chaudière valley.

While the recognition of deglacial patterns has proven difficult in the northeastern region, mainly because of the paucity of ice-marginal features, the pattern of northward ice-marginal retreat is well documented in the southwestern region, where a series of four recessional morainic belts were formed. From oldest to youngest, these are the Dixville Moraine, the Cherry River–East-Angus Moraine, the Mont Ham Moraine and the Ulverton-Tingwick Moraine. The level of the first major glacial lake to form in the region, Glacial Lake Memphremagog, had fallen to its Sherbrooke Phase shorelines (240-275 m) when the ice-margin stood at the position of the Cherry River–East-Angus Moraine, about
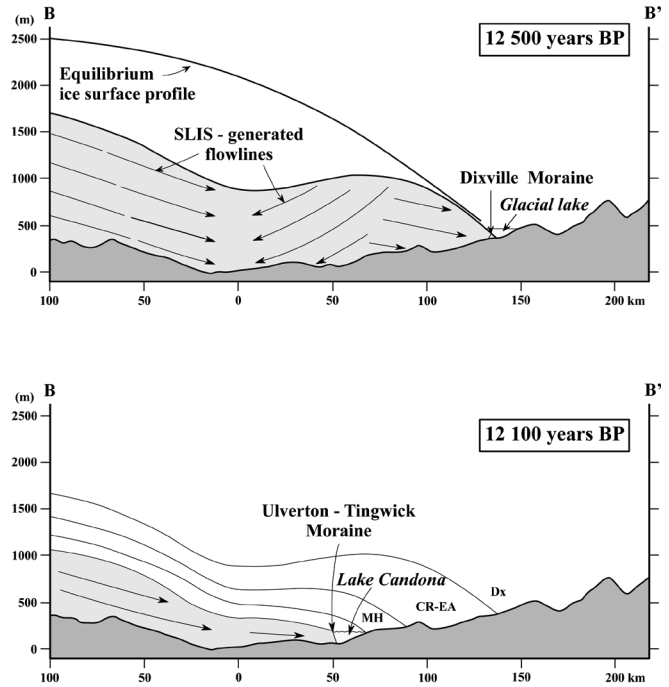
FIGURE 12. Schematic reconstruction of deglacial events at 12 500 and 12 100 years BP across the Central St. Lawrence Lowlands and adjacent Appalachian uplands of southern Québec.

Reconstitution schématique des événements tardiglaciaires vers 12 500 et 12 100 ans BP à la hauteur du centre des Basses terres du Saint-Laurent et des hautes terres appalachiennes du Québec méridional.

12 325 years BP. When the ice margin retreated from the Mont Ham Moraine, about 12 200 years BP, a new spillway was uncovered and water levels in the Saint-François valley fell by about 35 m to the level of a transitional phase. Further ice-marginal retreat to the position of the Ulverton-Tingwick, about 12 100 years BP, caused water levels to drop to the level of Glacial Lake Candona (190-230 m), a large glacial lake that had formed by coalescence of three water bodies: Lake Iroquois, Glacial Lake Vermont and Glacial Lake Memphremagog. Within about 100 years following deposition of the Ulverton-Tingwick Moraine, ice retreat along the Appalachian piedmont resulted in the final drainage of Lake Candona, thus allowing Champlain Sea waters to invade much of these glaciolacustrine terrains by about 12 000 years BP (corrected ^{14}C years, $\delta^{13}C \simeq 0‰$). Water levels had then fallen to 165 m, which is the local marine limit (Parent and Occhietti, 1988).

The sedimentary and faunal record of the transition from Lake Candona to Champlain Sea incursion is exposed at the Rivière Landry section, which is the proposed type section for the Danville Varves. This stratotype provides a unique, 103 year-long varve series which records the migration of a characteristic ostracod species, *Candona subtriangulata*, into the eastern part of the glaciolacustrine basin. The Rivière Landry varve record serves as a basis for estimating regional rates of ice retreat (about 200 m·a^{-1}). Given the lack of datable material in glaciolacustrine sediments depos-

ited during deglaciation, this rate of retreat also provides pre-liminary ages for recessional morainic belts in the region. These estimated ages are in broad agreement with the deglacial chronology proposed for the adjacent regions of northern New Hampshire (Thompson *et al.,* 1996; Gerath *et al.*, 1985) and for the Lake Champlain valley (Connally and Sirkin, 1973).

ACKNOWLEDGEMENTS

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