Larsen (1985) saw the Nipissing and Algoma Great Lakes in a different light. He viewed these as largely climatically controlled water levels during a period of relatively steady outlet incision rather than being due to isostatic adjustment.

Modern levels of both lakes Huron and Superior were substantially achieved by about 2 ka. Further discussion of Great Lakes history is provided by Karrow (1989).

VEGETATION HISTORY

Revegetation of the study region following disappearance of glacier ice, glacial lake waters, and marine waters spans the period from 12 ka to present. McAndrews (1981) summarized fossil pollen assemblages along a north-south transect across Ontario. Another report on the vegetation history of the Hudson Bay Lowlands is based on a pollen profile from Sutton Ridge (McAndrews et al., 1982). Chapter 7 of this volume reports on the general development of the Boreal Forest in Canada (Ritchie, 1989), the Holocene changes in vegetation (Anderson et al., 1989) and patterns of vegetation colonization in adjacent Quebec (Richard, 1989).

In general, as deglaciation proceeded in Ontario, a brief period of tundra vegetation existed. This was followed by spruce forest, forests in which pine pollen dominated, and then the modern zonal vegetation. The pine forest period was more subject to latitudinal effects than was the spruce forest period. In central Ontario the spruce-pine transition occurred between 10.5 and 10.0 ka. Northward of 46°30′N the spruce minimum and white pine peak are related to the Hypsithermal period. At this time, about 5 ka, temperatures were 1-2 °C higher than present. White pine macrofossils dated at 5.0 ka indicate that this species existed nearly 100 km north of its present range during this episode. Sutton Ridge vegetation history includes a succession from sparse coastal tundra through shrub tundra to modern spruce woodland between the time of emergence and 6.5 ka, as the Tyrrell Sea maritime effects decreased (McAndrews et al., 1982). Plant macrofossils indicate the presence of *Najas flexilis* between 6.5 and 3.0 ka, suggesting that the Hypsithermal range for this species was 300 km north of the known modern range. Post-Hypsithermal cooling contracted the ranges of species such as white pine to their present range about 2.5-3.0 ka.

Throughout much of the region the postglacial growth of muskeg was an important bioclimatic and economic development. Large scale muskeg expansion in northern Ontario did not occur until the climatic decline following the Hypsithermal, when more favourable colder, moister conditions prevailed (Terasmae, 1977). The oldest date on basal peat in the Manitoba part of the Hudson Bay Lowlands is 6490 ± 70 BP (GSC-1738) at Recluse Lake and the oldest date on spruce trunks is 5960 ± 100 BP (BGS-980).

QUATERNARY GEOLOGY OF THE SOUTHEASTERN CANADIAN SHIELD

J-S. Vincent

INTRODUCTION

The southeastern Canadian Shield lies west of the Labrador Sea; north of the Gulf of St. Lawrence, Saint-Narcisse Moraine, and Ottawa River; east of the Harricana Interlobate Moraine and Hudson Bay; and south of Hudson Strait (Fig. 3.1). The Quaternary deposits of this region bear witness, with few exceptions, to events that occurred during and since the Wisconsin Glaciation. The southeastern Canadian Shield area is particularly interesting since it is presumably there that ice from the Labrador Sector of the Laurentide Ice Sheet first accumulated. After coalescing with other ice masses, it subsequently extended to the Canadian Interior Plains and the northern United States. It is also in the centre of the southeastern Canadian Shield that one of the last remnants of continental ice finally disappeared about 6.5 ka ago.

Detailed or reconnaissance surficial geology maps have been completed for only about 10% of the area (Fig. 3.38). The principal references for the mapped areas and for regional or topical reports are listed in Table 3.5. Several comprehensive studies deal specifically with particular topics such as glacial limits and weathering zones in Labrador (Ives, 1978; Clark, 1984a), glacier movements of different ages (Veillette, 1986a), the Quebec North Shore morainic systems (Dubois and Dionne, 1985), the Sakami and Harricana moraines and the Cochrane surges (Hardy, 1977, 1982b), glacial lakes Barlow and Ojibway (Vincent and Hardy, 1977, 1979; Veillette, 1988), and postglacial seas (Hillaire-Marcel, 1979). Finally, Prest (1970), Prest et al. (1968), Hillaire-Marcel and Occhietti (1980), Mayewski et al. (1981), Peltier and Andrews (1983), and Parent et al. (1985) provide syntheses on the Quaternary of the area.

Vincent, J-S.
1989: Quaternary geology of the southeastern Canadian Shield; in Chapter 3 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1).

Relief and drainage

The present physiographic aspect of the southeastern Canadian Shield results from a long series of events. In Precambrian time at least one erosion surface was developed. This surface was later submerged by Paleozoic seas in which thick sequences of sediments were deposited. During the Mesozoic and Cenozoic, the sedimentary cover was stripped by running water, and during the Quaternary, glaciers modified the old erosion surface.

Bostock (1970b) described the area as a vast undulating lake-studded plateau. The elevation of the plateau gradually rises towards the central interior from the peripheral lowlands of the St. Lawrence River basin and Hudson Bay, Hudson Strait, and Labrador Sea coasts (Fig. 3.2). Local relief is generally low, rarely exceeding 100 m, with isolated hills standing above the monotonous plateau surface. Exceptionally high terrains, with elevations in excess of 1000 m, are restricted to summit areas of the Laurentian Highlands and of the Mealey and Torngat mountains. The presence of fault-line scarps, cut here and there by deeply incised valleys in the southern part of the southeastern sector of the shield gives a mountainous aspect to the Laurentian Highlands. The most spectacular scenery is found on the Atlantic side of the Torngat Mountains. There, the ancient erosion surface, dipping west, abruptly terminates in a series of cliffs, in places more than 1000 m high, incised by valleys and fiords.

Many features of the modern physiography of the Canadian Shield relate to the presence or absence of Quaternary deposits (Surficial Materials Map of Canada, Geological Survey of Canada, in preparation). Large areas of the Laurentian Highlands and Ungava Peninsula have a

Table 3.5. References dealing with the Late Wisconsinan deposits and history of the various areas of the Canadian Shield in Quebec-Labrador

Surficial Geology Maps[1]	Regional Reports	Topical Reports
Northern Labrador and northeastern New Quebec		
(58) Barré (1984)*[2]	Clark (1984a); Evans (1984)	Andrews (1963a), Barnett (1967), Barnett and Peterson (1964), Evans and Rogerson (1986), Ives (1957; 1958a, 1960a, b, 1976), Johnson (1969), Løken (1962a, b, 1964), Matthew (1961), McCoy (1983), Peterson (1965), Short (1981), Smith (1969), Tomlinson (1963)
Central and southern Labrador		
(15) [3]Fulton and Hodgson (1980), (8, 11-13, 16-20)[3] Fulton et al. (1979, 1980a,b,c,d, 1981a,b,c,), (9-10, 14) [3]Fulton et al. (1980e,f, 1981d), (21) [3]Fulton et al. (1981e),(22) Grant (1986)	Fulton and Hodgson (1979)	Blake (1956), Fitzhugh (1973), Grant (1969, 1971), Gray (1969), Hodgson and Fulton (1972), Klassen (1983a, 1984),Morrison (1963), Rogerson (1977)
Quebec North Shore		
(24) Dredge (1983b), (25) Dubois and Desmarais (1983), (60) Dubois et al. (1984)	Dredge (1983b), Dubois (1980), Tremblay (1975)	Boutray and Hillaire-Marcel (1977), Dionne (1977), Dredge (1976a, b), Dubois (1976, 1977, 1979), Dubois and Dionne (1985), Dubois et al. (1984)
Lac Saint-Jean		
(29) Dionne (unpublished), (30-36), LaSalle and Tremblay (1978)	LaSalle and Tremblay (1978)	Dionne (1968, 1973), Dionne and Laverdière (1969), LaSalle et al. (1977a, b), Laverdière and Mailloux (1956), G. Tremblay (1971, 1973)
Laurentian Highlands		
(52) Chagnon (1969), (49) Denis (1976), (56) Gadd (unpublished), (54) Gadd and Veillette*[4], (42) Fulton (unpublished), (46) Lamothe (1977), (51) LaSalle (1978), (55) LaSalle and Gadd*[4], (50) Occhietti (1980), (48) Pagé (1977), (44-45) S.H. Richard (in press, 1984), (43), S.H. Richard et al. (1977), (47) Tremblay (1977)	Denis (1976), Occhietti (1980), Tremblay (1977)	Dadswell (1974), Denis (1974, 1976), Hardy (1970), Lamothe (1977), LaSalle et al. (1977a, b), Laverdière and Courtemanche (1960), 1980b), Romanelli Pagé (1977), Parry (1963), Richard (1978,(1975)
Upper Ottawa River basin		
(37-41) Veillette (1986b,c, 1987a,b,c), (53) Veillette and Daigneault (1987)		Antevs (1925), Veillette (1982, 1983a,b, 1986a, 1988), Vincent (1975), Vincent and Hardy (1977, 1979).
Quebec Clay Belt and area southeast of James Bay		
(24) Baker and Storrison (1979), (61) Bisson (1987), (23) Bouchard et al. (1974), (26) Chauvin (1977), (59) Hardy (1976)*,(2) Lee et al. (1960), (28a,b), G. Tremblay (1972; 1974), (3-5) Vincent (1985a,b,c,)	Bisson (1987), Bouchard (1980, 1986), Chauvin (1977), Hardy (1976), Martineau (1984a), G. Tremblay (1974), Vincent (1977)	Allard (1974), Antevs (1925), Bouchard and Martineau (1984, 1985), Bouchard et al. (1984), DiLabio (1981), Dionne (1974), Hardy (1977, 1982a,b), Hillaire-Marcel el al. (1981), Hillaire-Marcel and Vincent (1980), Ignatius (1958), Lee (1959b, 1960, 1962, 1968), Martineau (1984b), Mawdsley (1936), Norman (1938, 1939), Prichonnet et al. (1984), Shaw (1944), Vincent and Hardy (1977, 1979), Wilson (1938)
Area east of Hudson Bay		
	Allard and Séguin (1985), Hillaire-Marcel (1976)	Allard and G. Tremblay (1983), Andrews and Falconer (1969), Archer (1968), Fairbridge and Hillaire-Marcel (1977), Hillaire-Marcel (1979, 1980), Hillaire-Marcel and Boutray (1975), Hillaire-Marcel and Fairbridge (1978), Hillaire-Marcel and Vincent (1980), Plumet (1974), Portman (1972),Walcott and Craig (1975)
Ungava Peninsula		
(1) Lauriol (1982)*	Lauriol (1982), Gray and Lauriol (1985)	Blake (1966), Bouchard and Marcotte (1986), Drummond (1965), Gangloff et al. (1976), Gray et al. (1980), Lauriol and Gray (1983), Lauriol and Gray (1987),Lauriol et al. (1979), Løken (1978), Matthews (1966, 1967), Taylor (1982)
Central New-Quebec and western Labrador		
(57) Guimont and Laverdière (1982)*, (7) Henderson (1959)*, (6) Hughes (1964)*	Henderson (1959), Hughes (1964)	Barr (1969), Derbyshire (1962b), Ives (1959, 1960a,c), Kirby (1961a,b), Richard et al. (1982)

1 The number in brackets, before the author's name, refers to the number shown in Figure 3.38.
2 The reference accompanied by an asterix indicates that the map portrays major landforms not deposits.
3 Synthesis maps at a scale of 1:500 000 are also available in Fulton (1986a, b).
4 Manuscript maps prepared as part of the Quaternary Geologic Atlas of the United States of the United States Geological Survey (scale 1:1 000 000).

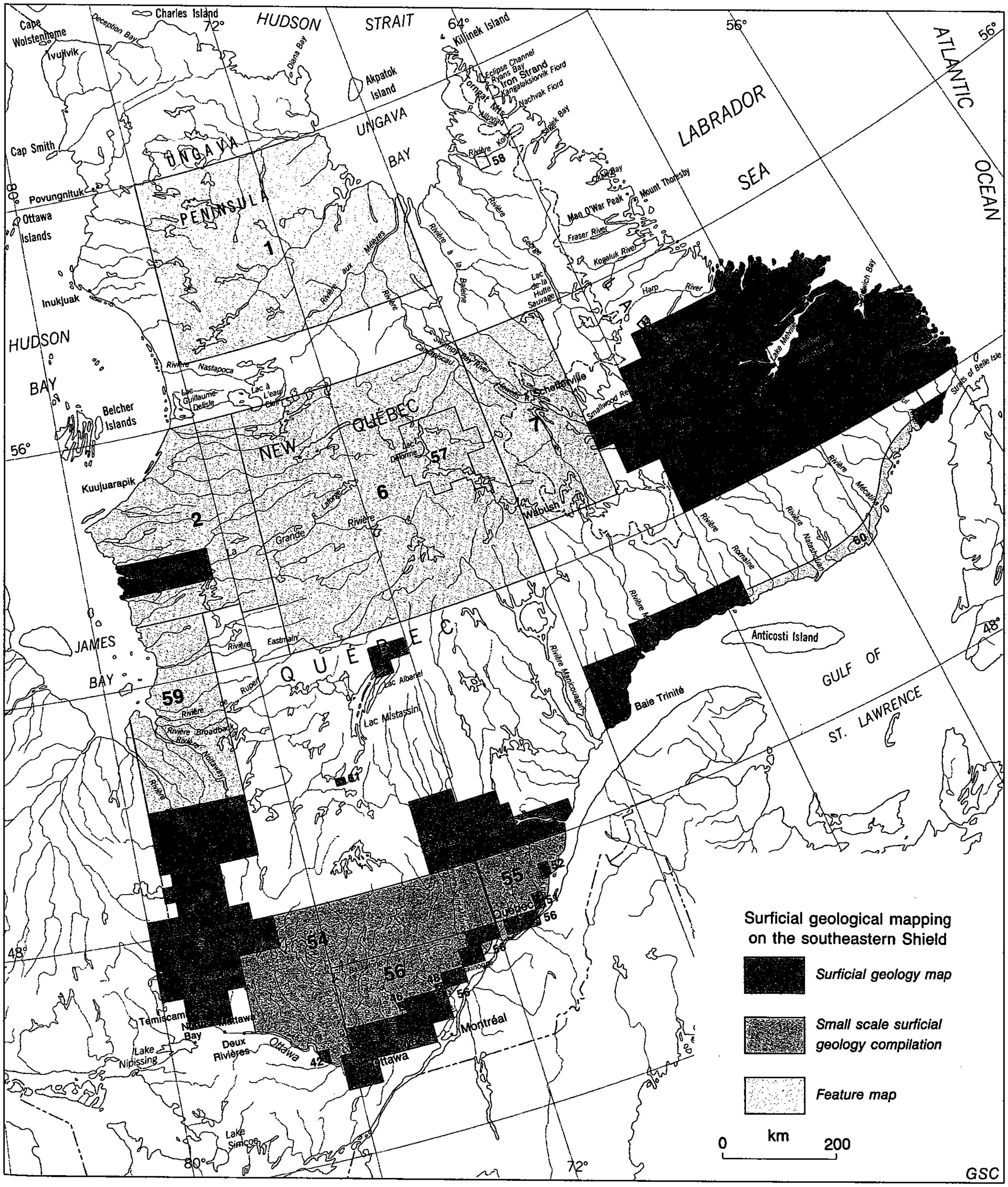


Figure 3.38. Location map for southeastern Canadian Shield and areas covered by surficial geology maps listed in Table 3.5 and by maps used as data source for Figure 3.39.

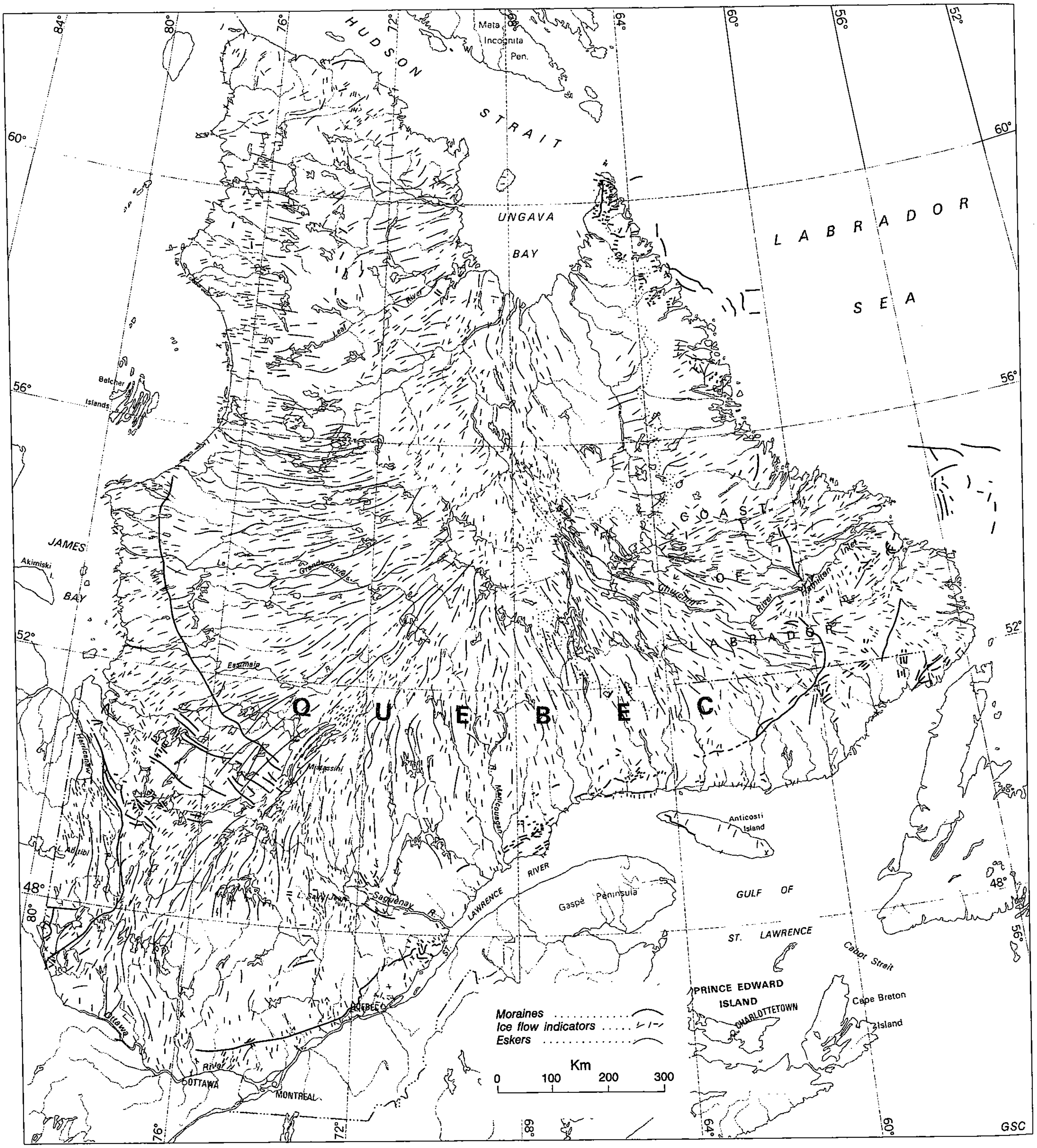


Figure 3.39. The main glacial landforms of the southeastern Canadian Shield compiled from all available sources of information.

sparse cover of Quaternary materials (Fig. 3.4), whereas the central part of the southeastern shield has a thick till sheet moulded into drumlins and ribbed moraine. Coastal areas are underlain by flat-lying marine or glacial marine sediments.

The drainage of the southeastern shield is obviously related to the slope of the land and thus is essentially radial from the higher south-central areas. Rivers with the highest average annual discharges are the Saguenay, Ottawa, Nottaway, La Grande, Caniapiscau, and Churchill. River courses are generally dictated by the structural elements of the Precambrian Shield although locally they are controlled by glacial deposits.

NATURE AND DISTRIBUTION OF QUATERNARY DEPOSITS

Figure 3.39 shows the distribution of the principal glacial landforms of the southeastern Canadian Shield. The general distribution of Quaternary deposits is shown on the Surficial Materials Map of Canada (Geological Survey of Canada, in preparation). The definition and descriptions of glacial landforms mentioned here can be found in Prest (1968, 1983a).

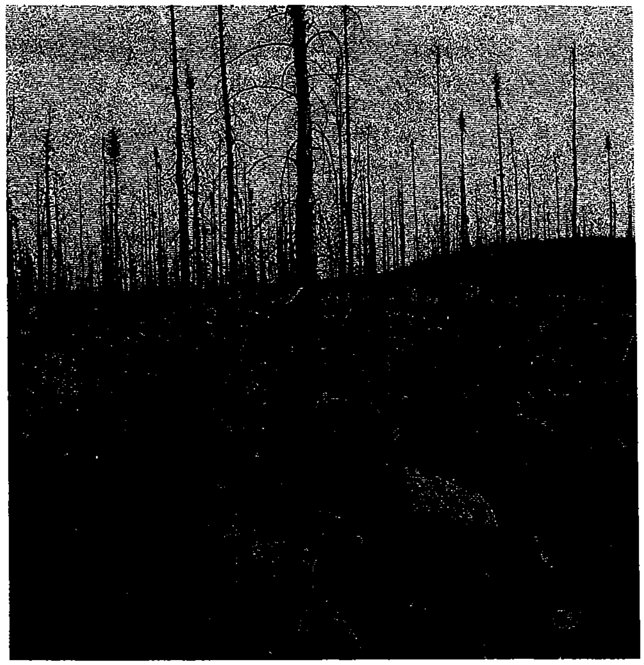


Figure 3.40. Bouldery till surface in southwestern Labrador. Courtesy of R.J. Fulton. 161608

Glacial deposits

Till and related landforms

The texture, structure, composition, and morphology of the tills of the southeastern shield are variable. Generally derived from metamorphic and igneous rocks, tills in this area are sandy, stony (Fig. 3.40) and noncalcareous. Although thicknesses in excess of 10 m are reported locally, the till is generally thin, probably less than 2 m in average. The Cochrane till southeast of James Bay is finer grained, highly calcareous, and nearly stone free; it was formed by a glacier advancing across glacial lake sediments. Finer grained till, with significant carbonate content (>5%) is locally associated with Proterozoic dolomitic sources in the Lac Mistassini area (DiLabio, 1981) and in northwestern Ungava Peninsula (Delisle et al., 1984). Figure 3.41 shows matrix textures of tills from various locations in the area. In the oxidized zone, up to 3 m thick, the tills are brown to yellowish brown whereas in the unoxidized zone they are greyish. In the Labrador Trough (Fig. 3.3) area, some tills are reddish due to incorporation of hematite. Except in the lowland southeast of James Bay and Torngat Mountains, only one till is recognized.

In central New Quebec there is a gradational series of belts of glacial depositional landform assemblages. Proceeding from Hudson Bay towards the ice divide, drumlin fields give way to ribbed moraine fields, which in turn grade into hummocky moraine. As recently portrayed by Bouchard et al. (1984) in the Lac Mistassini area (Fig. 3.42), tills in regions of ground moraine or fluted ground moraine are massive, found mainly in interfluve areas, and deposited by ice near the pressure melting temperature. Tills comprising ribbed moraine are commonly stratified, generally occur in low areas, and are thought to be formed by shearing of basal ice. Finally, tills comprising the hummocky moraine areas are massive, generally interstratified with sand and gravel, and thought to have formed at the margin of cold-based ice. Some hummocky moraines show conspicuous glacial lineations suggesting that their genesis is more related to shearing and ice-flow rather than to marginal ablation of cold-based ice. Clasts in fluted till or ribbed moraine are dominantly local whereas clasts in hummocky moraine have more distant sources. Whether these landform-sediment associations in the Lac Mistassini area are representative of all areas remains to be shown.

Extensive tracts of land are made up of ground moraine moulded into drumlins (Fig. 3.43) or plastered on the lee side of bedrock hills (crag-and-tail). In the Lac Mistassini area drumlins are 100-3000 m long, 30-600 m wide, and 10-25 m high (Bouchard, 1980). In the La Grande Rivière area drumlins are 200-2000 m long, 100-400 m wide, and 3-25 m high (Vincent, 1977). Well developed glacial lineaments are common except in the rugged terrains of the southern Laurentian Highlands, the Torngat Mountains, and in the area of final ice retreat (Fig. 3.39). Lauriol (1982) and Gray and Lauriol (1985) have shown that the ice divide area in Ungava is not characterized by hummocky moraine but instead by an unmoulded till sheet flanked distally first by thin and discontinuous till and then by fluted till. The northern portion of the unmoulded till sheet is made up of rounded soliflucted slopes that contrast with fresh morainic surfaces elsewhere. Perhaps this is a remnant till surface only slightly affected by the last ice advance.

Ribbed moraine (Fig. 3.44) and hummocky moraine (Fig. 3.45) occur mainly where Labrador Ice persisted latest (Hughes, 1964; Prest et al., 1968). The ribbed moraine has been described in some detail by Hughes (1964), Cowan (1968), and Bouchard (1980). Ribs are up to 1600 m long, 200 m wide, and 27 m high; generally the rib crests are spaced 90-300 m apart. In many areas radial belts of ribbed moraine alternate with belts of drumlins as they do in

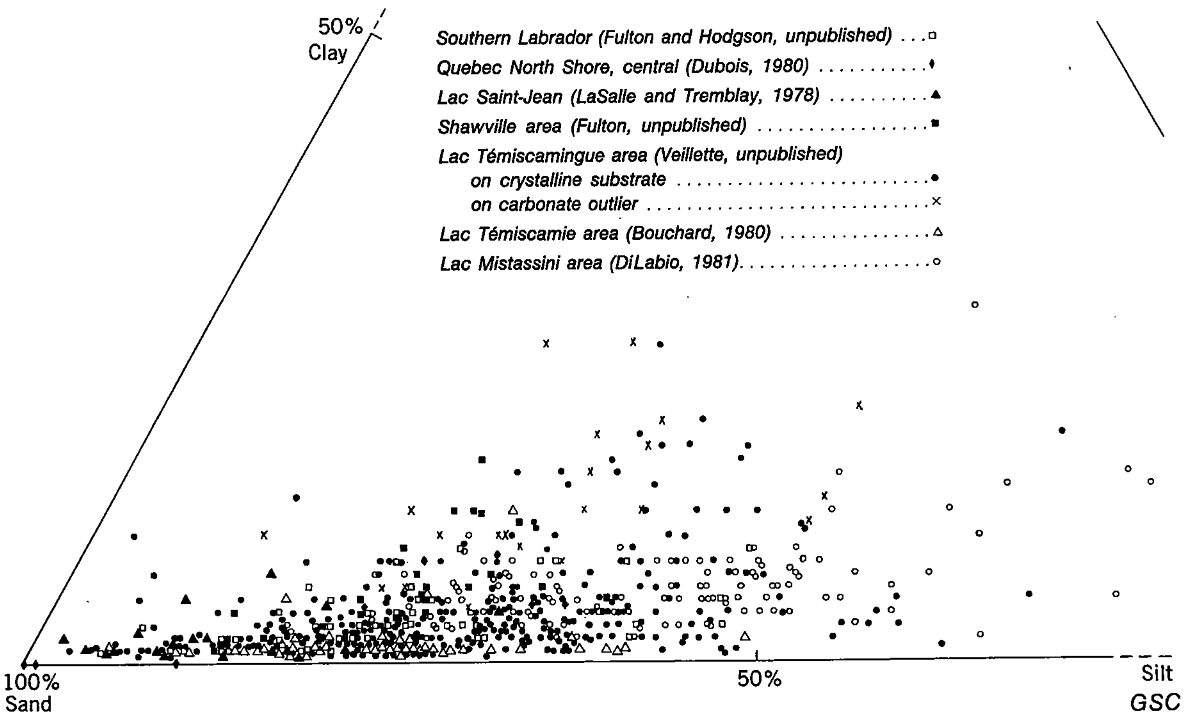


Figure 3.41. Ternary diagram showing grain size composition of till matrixes in the southeastern Canadian Shield

Keewatin and Manitoba (Shilts et al., 1987; this chapter, Dyke and Dredge, 1989). This attests to the complexity of ice flow regimes even over small areas. Large areas of hummocky moraine, which consist of irregularly shaped mounds of melt-out till, and "drift pressed features" such as moraine plateau, are common particularly in the final retreatal ground of Labrador Ice (Fig. 3.45). A belt of hummocky moraine, 100 km long, also occurs in southeastern Labrador (Fulton et al., 1981c, e). No detailed studies have been completed on these in the southeastern Canadian Shield.

De Geer moraines (Fig. 3.46) are restricted to low relief areas where the margin of the retreating ice sheet was in contact with lacustrine or marine water. They are remarkably well formed and abundant in the northeastern glacial Lake Ojibway basin and in the Tyrrell Sea basin. The moraines are composed largely of till and are largest and most consistently aligned where the ice sheet retreated in deep water. Where formed in shallow water, the moraines are smaller and their alignment is in places erratic. In La Grande Rivière area, De Geer moraines are 1-10 m high, 5-150 m wide, and 50-1500 m long. Spacing between moraine crests varies between 50 and 400 m but is most commonly between 100-200 m, presumably indicating annual retreat rates (Vincent, 1977). East of Hudson Bay, Lauriol and Gray (1987) measured a similar average spacing of 100-200 m between moraine crests.

Glaciofluvial deposits

Glaciofluvial deposits are widespread on the southeastern Canadian Shield. They occur as major end and interlobate moraines (Fig. 3.39, 3.47, 3.48), as eskers (Fig. 3.39), as kames, and as subaqueous and subaerial outwash deposits. Glaciofluvial deposits consist mostly of stratified sands and gravels.

The Saint-Narcisse (see Occhietti, 1989), Quebec North Shore, and Sakami end moraines mark major halts during recession (Fig. 3.39, 3.47). The Quebec North Shore

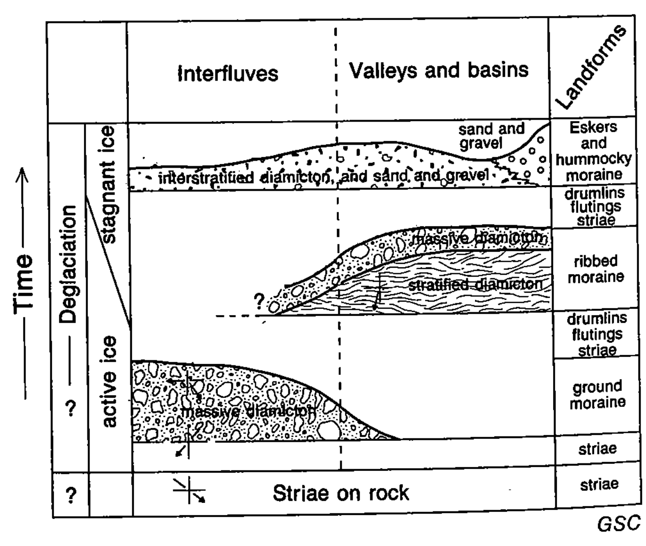


Figure 3.42. Relative stratigraphic and topographic occurrences of till lithofacies in the Lac Albanel area (after Bouchard et al., 1984).

Moraine system extends for 800 km (including gaps) between Rivière Manicouagan in Quebec and south of Lake Melville in Labrador (Dubois and Dionne, 1985). It consists of discontinuous till and ice contact ridges, up to 4 km wide and 50 m thick (Fig. 3.48), with outwash deltas and plains on its distal side. The Sakami Moraine extends over a distance of 630 km between Kuujjuarapik (Great Whale, Poste-de-la-Baleine) on the southeastern coast of Hudson Bay to Lac Mistassini (Hillaire-Marcel et al., 1981; Hardy, 1982a). Asymmetric ridges of ice contact and proglacial deposits, up to 6 km wide and 40 m thick, make up the moraine. The Harricana Interlobate Moraine (Hardy 1977, 1982b; Veillette, 1986a) marks the final zone of convergence of Labrador Ice and Hudson Ice. It can be traced for 1000 km from islands in eastern James Bay to Lake Simcoe in Ontario (Veillette, 1986a). The moraine is composed of a complex series of ridges, up to 10 km wide and 100 m high, of glaciofluvial deposits profusely pocked with elongated kettles.

Eskers are conspicuous features of the region everywhere except in the more rugged Torngat Mountains and the central and eastern parts of the Laurentian Highlands. Eskers form a radiating pattern from the central divide area. A characteristic feature of the eskers is that they commonly cross the grain of the landscape, ascending and descending hills and ridges. Where the ice retreated in contact with water bodies, eskers consist of subaqueous outwash deposits and may be buried by thick sequences of fine grained sediments.

Subaerial outwash deposits are less common because over much of the area the ice retreated in contact with either glacial lakes or the sea. Outwash terraces or delta complexes are present along some major valleys, such as those draining southward from the Laurentian Highlands. Meltwater channels are particularly abundant near the final retreat centre (Fig. 3.49); they were used by Ives (1959) to designate the sites of final ice sheet disintegration.

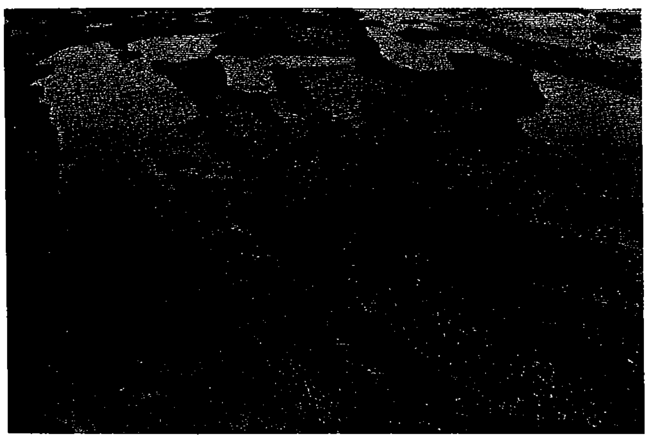


Figure 3.43. Drumlin field northeast of Lac Mistassini area. Courtesy of J-C. Dionne.

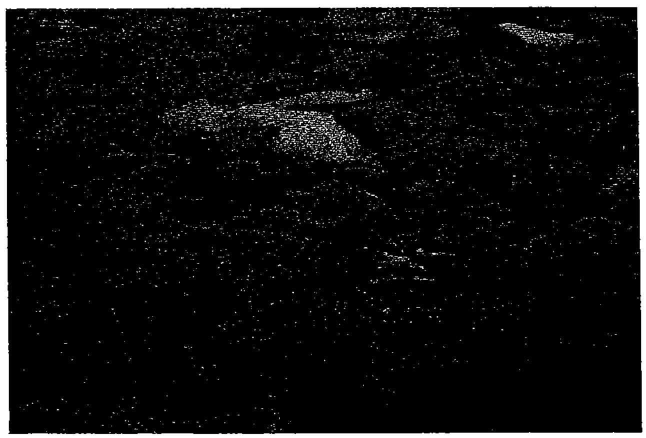


Figure 3.45. Hummocky moraine in the area northeast of Lac Mistassini. Courtesy of C. Laverdière.

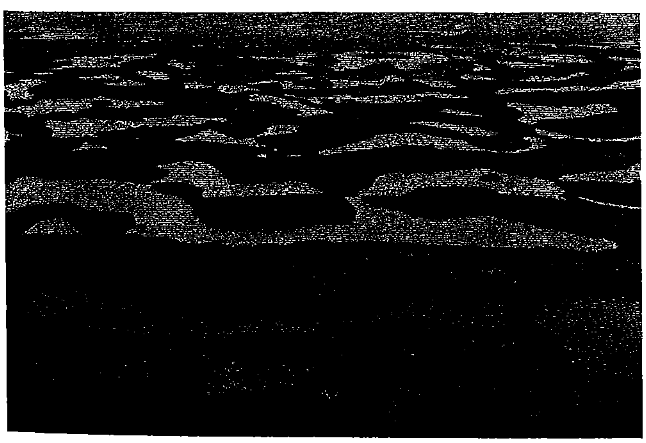


Figure 3.44. Ribbed moraine field northeast of Lac Mistassini. Courtesy of C. Laverdière.

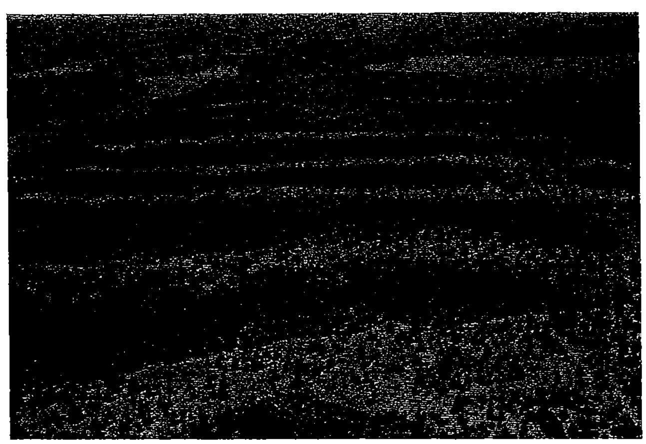


Figure 3.46. Suite of De Geer moraines in the area north of La Grande Rivière and east of Sakami Moraine. 204061-M

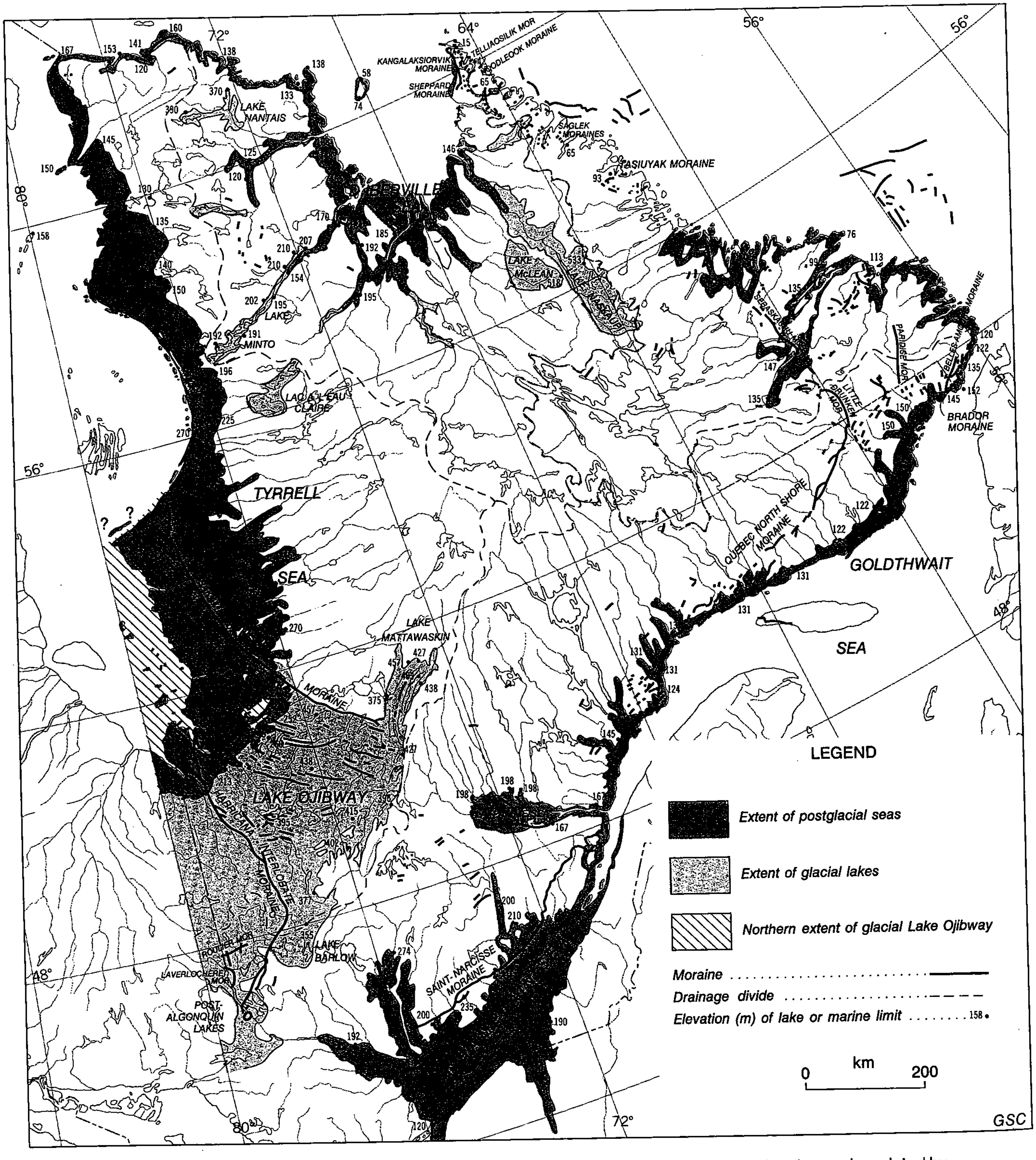


Figure 3.47. Major end and interlobate moraines of the southeastern Canadian Shield and areas inundated by glacial lakes and marine waters during the last deglaciation.

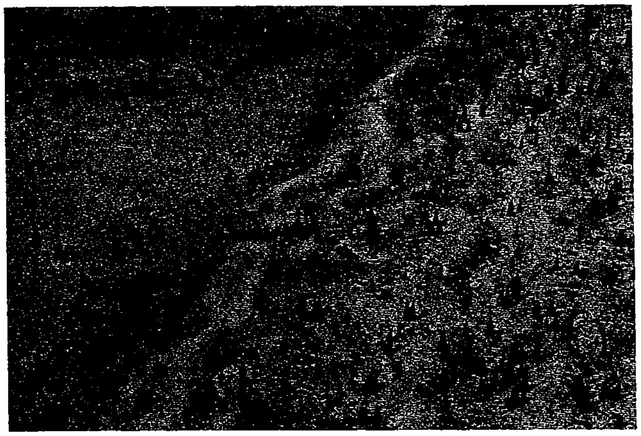


Figure 3.48. Quebec North Shore Moraine in the Rivière Natashquan area. Courtesy of J-C. Dionne.

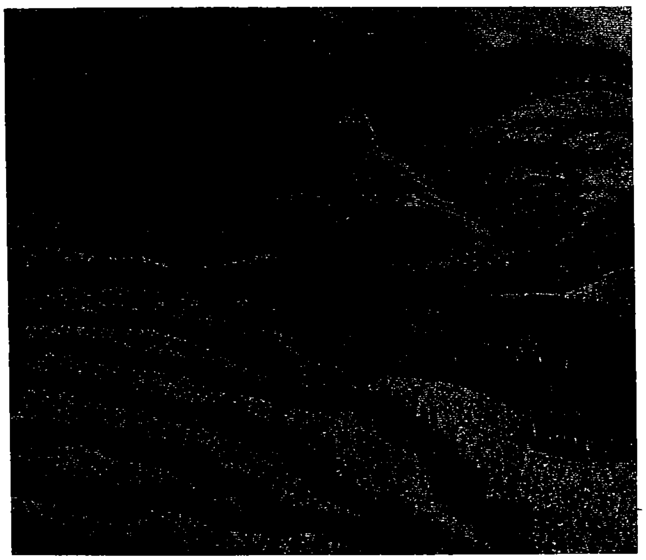


Figure 3.49. Meltwater channels 70 km north of Schefferville, near the retreat centre of Labrador Ice. Courtesy of J.D. Ives.

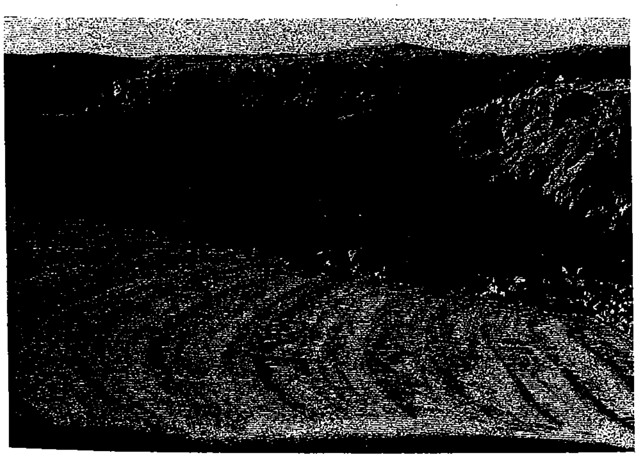


Figure 3.50. Flights of raised beaches of the Tyrrell Sea in the Lac Guillaume-Delisle area. Courtesy of C. Hillaire-Marcel.

Glacial lake and marine deposits

Extensive areas were flooded by glacial lakes and seas during deglaciation (Fig. 3.47). Glacial Lake Ojibway and the Tyrrell Sea were by far the largest water bodies.

Fossiliferous silts and clays, generally massive, were deposited offshore in the seas, whereas varved silts and clays were laid down in glacial lakes. Glaciomarine facies (rainout sediment from glacial ice) are recognized locally but in most places glacial marine rhythmites occur at the base of the marine offshore sediment suite especially in proximal locations or in estuaries. At the shoreline of basins, littoral facies were also developed resulting in beaches, spits, and other related features. As the seas or lakes regressed or were drained, nearshore facies were produced in places where waves and currents could rework glacial, particularly ice contact, deposits. Hence, thick sequences of nearshore sands, overlying the offshore facies, are found near moraines and eskers. Flights of raised beaches (Fig. 3.50) can be spectacular, particularly in the Tyrrell Sea basin and in many locations along the Labrador coast. Estuarine and deltaic sediments were deposited at mouths of rivers. Since rivers entered the sea at progressively lower elevations as isostatic uplift progressed, there is downstream overlapping of younger coarser estuarine and deltaic deposits onto older fine grained deposits. Most older and higher delta deposits, which generally overlie fine grained offshore materials, were later incised by rivers; thus they occur as terrace remnants capping fine grained sediments.

Eolian, fluvial, and organic deposits

Eolian deposits are found in small areas where the wind has reworked sands and silts of glaciofluvial, marine, or deltaic origin. Large dune fields are found on outwash terraces and deltas in the Lac Saint-Jean area, the Quebec North Shore area, and in lower Churchill River valley (David, 1977); on reworked glacial lake materials near the Harricana Interlobate Moraine (Fig. 3.51); and locally on glaciofluvial and deltaic deposits elsewhere. The largest occurrence is south of Harp River in east-central Labrador.

Present day rivers are reworking older, mainly glacial and marine sediments and depositing materials on their

Figure 3.51. Parabolic dunes outlining bog areas on the east flank of the Harricana Interlobate Moraine in the area east of Lac Témiscamingue. Courtesy of J.J. Veillette. 203506-C

floodplains and as deltas in lakes and the sea. Most drainage basins contain an abundance of lakes which trap sediments entrained by streams. As a consequence, floodplains of more than limited extent are found only on the lower reaches of some of the larger rivers.

Organic deposits have accumulated in wetlands on glacial lake and marine plains, low relief till areas, and in depressions on bedrock surfaces. The peat cover is widespread in the area adjacent to James Bay and according to the National Wetlands Working Group (1981) wetlands in which organic sediments accumulate cover more than 50% of the surface in an area extending roughly 250 km inland. This same group also shows a belt of wetlands about 500 km wide stretching from James Bay to the Labrador Coast. Because climate varies from north to south, the type of wetland in which organic deposits accumulates also varies from north to south. At the southern fringe of the Canadian Shield, wetlands are dominantly domed raised bogs and ladder fens. Farther north, and throughout much of the central part of the region, organic deposits are being laid down in domed, flat, basin and string bogs and ribbed fens. In the northern part of the area, peat plateaus and patterned fens give way to wetlands characterized by low and high centred polygons containing ice wedges and lenses and underlain by permafrost. Palsas are common in coastal and northern bogs and thermokarst depressions are locally common in ice-rich organic deposits.

QUATERNARY HISTORY

Sediments on the southeastern Canadian Shield mostly record late Quaternary deglacial events. Older deposits are exposed along rivers to the southeast of James Bay and are found in the Schefferville area. Older tills have also been reported at the surface in the Torngat Mountains and coastal summits to the south of these, in the Mealy Mountains, possibly in southeastern Labrador, and in northern Abitibi. Age control on deglacial events is based mainly on radiocarbon dates on marine shells and basal lake sediments (Table 3.6). Major moraines, glacial lakes, postglacial seas and morphostratigraphic (weathering) zones have been named, but few lithostratigraphic units have been carefully defined and formally named.

In this section the limited information on pre-Late Wisconsinan events is discussed first. Then, deglacial events and chronology are discussed for a series of subregions and correlations are suggested (Fig. 3.52).

Pre-Late Wisconsinan events

Southeast James Bay area

Sediments recording pre-Late Wisconsinan events are known from eight locations (Hardy, 1982b; Bouchard et al, 1986; Fig. 3.31; Table 3.1). Thin laminae of peat underlying till, interbedded with compact clayey silt rhythmites, in a 10 m sequence, have been dated at >42 000 BP (Y-1165, Stuiver et al, 1963; Prest, 1970) along Rivière Harricana. In four sections along the banks of lower Rivière Nottaway, compact shell-bearing silts, wood-bearing clayey silts, and organic-bearing rhythmites and silty sands underlie the surface till. Near the mouth of Rivière de Rupert, compact organic-bearing silts and sandy silts underlying till have been observed in a core. All these sediments record marine or lacustrine facies similar to those of the interglacial Missinaibi

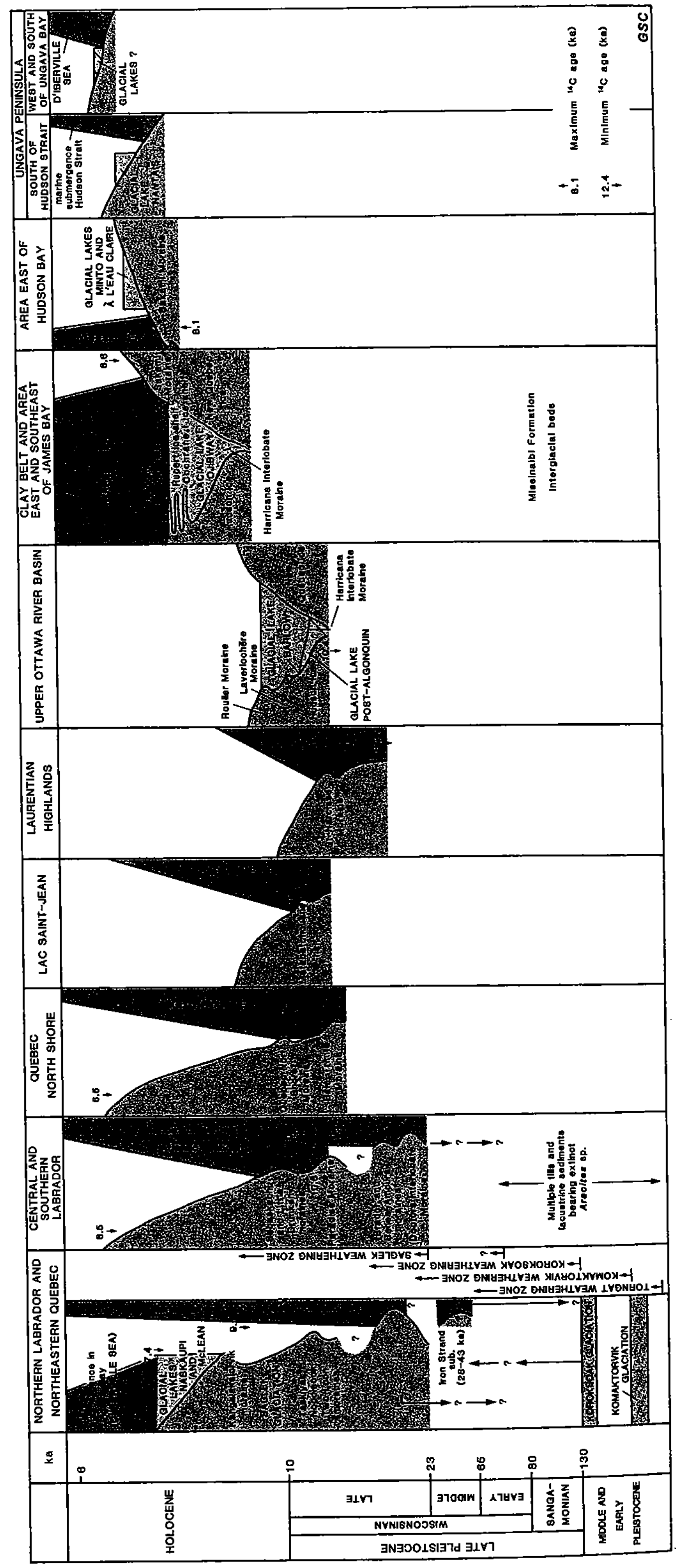


Figure 3.52. Correlation of Wisconsinan events of the southeastern Canadian Shield area. Each column is constructed from references given in the text.

Formation in the Hudson Bay Lowlands of Ontario (Skinner, 1973; this chapter, Dredge and Cowan, 1989). Compact waterlain sediments, similar to those described above, occur beneath glacial deposits in a borrow pit situated 115 km southeast of the mouth of Rivière Nottaway (50°39′N, 77°34.5′W; Hillaire-Marcel and Vincent, unpublished). The location and elevation (275 m) of this site suggest that one or the other of the water bodies recorded by deposits of the Missinaibi Formation may have extended far inland in Quebec. Finally, at the Selbaie Mine in northern Abitibi, Bouchard et al. (1986) have noted the presence of sand, gravel, and till under definitely Late Wisconsinan tills.

Schefferville-Wabush area

Recent investigations by Klassen and Thompson (1987) and Klassen et al. (1988), have indicated that several lithologically distinct till units are present in the Wabush area. In one section glaciofluvial and organic-bearing lacustrine sediments were identified. The significance of these finds is still not known but the deposits may well record pre-Sangamonian events since fruits of the extinct *Aracites* have been identified in lake sediments buried by three distinctively different tills.

During a drilling operation 25 km northwest of Schefferville a twig was obtained from a peat layer which was overlain by 18 m of glaciofluvial materials and till. This small piece of wood gave a radiocarbon age of 24 250 ± 600 BP (Granberg and Krishnan, 1984); the date, if valid, may record a nonglacial pre-Late Wisconsinan event. This is a surprising discovery because it may suggest that an area of the southeastern shield near the heart of the Laurentide Ice Sheet may have been ice free during the Middle Wisconsinan.

Torngat Mountains and adjacent coastal areas

Surface materials on summits of the Torngat Mountains, and on coastal summits as far south as Fraser River, apparently record glacial and nonglacial events of pre-Late Wisconsinan and pre-Wisconsinan age. A long lasting controversy exists between researchers who feel extensive nunatak areas existed during Late Wisconsinan and those who feel the areas were ice covered. The former link weathering zones to different glacial events; the latter believe that weathering zones result from progressive deglaciation of summit areas during the Late Wisconsinan, represent zones developed under different subglacial thermal regimes, or record subaerially weathered terrains that were preserved under cold-based ice. The argument between the two groups of researchers persists because of equivocal field evidence, poor chronological control of events, and a tendency to fit the facts to theory (Ives, 1974; Clark, 1984a). Comprehensive reviews of this equivocal evidence can be found in Ives (1974, 1978), Mayewski et al. (1981, p. 106-108), Brookes (1982), Clark (1984a), Evans (1984), and Evans and Rogerson (1986).

Ives (1978), mainly using his own work (Ives, 1957, 1958a,b, 1963, 1974, 1975, 1976; Ives et al., 1975) and that of Tomlinson (1958 and 1963), Wheeler (1958), Løken (1962a, b), Andrews (1963a), and Johnson (1969), proposed that three regional ice sheet glacial events are recorded in the Torngat Mountains, on the basis of clearly defined weathering zones. The weathering zone limits (or glacial boundaries, Figure 3.53) slope seaward and ice was progressively thinner and less extensive during successive glaciations.

During the oldest **Komaktorvik Glaciation,** all of northern Labrador, except for higher summits in the central and northern Torngat, was ice covered and glaciers extended far out onto the continental shelf. The nunataks of this glaciation were assigned to the Torngat Weathering Zone which is characterized by mature felsenmeer, tors, and deep weathering pits, and by the assumed absence of glacial erratics. Areas affected by the Komaktorvik Glaciation, and not later covered by ice, comprise the Komaktorvik Weathering Zone. This zone is similar in character to the Torngat Weathering Zone but contains erratics.

During the **Koroksoak Glaciation** ice again covered most of Labrador and extended out onto the shelf, but nunataks were larger than during the Komaktorvik Glaciation. Areas covered by the ice during Koroksoak Glaciation, and not later covered by ice, comprise the Koroksoak Weathering Zone. In this zone there is abundant evidence of glaciation and only limited (incipient) felsenmeer development. Weathering characteristics are intermediate between those of the Komaktorvik Weathering Zone above and those of the areas covered by ice during Saglek Glaciation. The boundary between the Koroksoak and Komaktorvik zones is a trimline which is obscured in places by periglacial slope processes.

Abundant and fresh looking glacial landforms characterize the area covered by the **Saglek Glaciation.** Presumably during this glaciation ice reached the outer coast adjacent to the Torngat Mountains, only at mouths of fiords. The Saglek glacial limit, is marked by fresh lateral and end moraines (Ives, 1976; Clark, 1984a). Felsenmeer and advanced weathering features are generally absent in the area covered by Saglek ice. Locally, due to lithology or preservation beneath the Saglek ice, small areas of extensive weathering have been reported (Gangloff, 1983).

The absence of detailed studies comparing the weathering characteristics of each zone and the lack of chronological control makes it difficult to assign ages to the different glaciations. Andrews (1974) argued on the basis of comparisons with the Baffin Island weathering zones, the development of clay minerals, and increase in ferric oxides, that the Koroksoak and older terrains on Baffin Island must be at least mid-Quaternary in age. Even though many workers, including J.D. Ives (University of Colorado, personal communication, 1985), believe the Saglek Moraines were built during the Late Wisconsinan maximum, the absolute age of the Saglek Glaciation is still not known (Andrews, 1977; Mayewski et al., 1981), although a Late Wisconsinan age is likely. Clark (1984a) corrrelated his Late Wisconsinan Two Loon drift with the deposits of Saglek Glaciation; Clark and Josenhans (1985), after studying the Saglek Moraines in their type area, argued that they are Late Wisconsinan.

Interpretations based on weathering zones are still being challenged. Gangloff (1983) saw no significant differences between the matrix texture, clay content, mineralogy, or surface morphology of quartz sand grains in Saglek Glaciation till and in felsenmeer of the Koroksoak Weathering Zone. Gangloff also documented the presence of tors in valley bottoms which were not destroyed by Saglek Glaciation ice, and of weathering pits and incipient tafoni which developed in Holocene time. Systematic mapping of

Table 3.6. Pertinent radiocarbon ages of the Canadian Shield area of Quebec-Labrador

Laboratory Dating no.	Age (years BP)	Locality	Reference	Material	Significance
Beta-9516	6 600 ± 100	Rivière Laforge, Que.	P.J.H. Richard (personal communication, 1985)	gyttja	"Oldest" minimum age for deglaciation of Rivière Laforge area (LG 4 Reservoir).
Beta-11121	9 800 ± 220	Rivière Deception, Que.	Gray and Lauriol (1985)	marine shells	With I-488, oldest shell date for deglaciation of the south shore of Hudson Strait.
DIC-517	42 730 +6680/-9970	Iron Strand, Labrador	Ives (1977) Short (1981)	marine shells	Date implies that Iron Strand site records a Middle Wisconsinan Interstade.
Gif-424	10 250 ± 350	Metabetchouan, Que.	LaSalle and Rondot (1967)	marine shells	"Oldest" minimum age for deglaciation of Lac Saint-Jean area.
Gif-3770	10 230 ± 180	Rivière-à-la-Chaloupe, Que.	Dubois (1980)	marine shells	"Oldest" minimum age for deglaciation of area west of Rivière Romaine, Quebec North Shore.
GSC-672	7 970 ± 250	Sugluk Inlet, Que.	Matthews (1967) Lowdon and Blake (1968)	marine shells	"Oldest" reliable minimum age for deglaciation of the south shore of Hudson Strait.
GSC-706	7 430 ± 180	Ottawa Islands, N.W.T.	Andrews and Falconer (1969) Lowdon and Blake (1968)	marine shells	"Oldest" minimum age for deglaciation of the Ottawa Islands and best approximation for the age of marine limit.
GSC-1337	9 140 ± 200	Rivière Moisie, Que.	Dredge (1983b) Lowdon et al. (1971)	marine shells	"Oldest" minimum age for deglaciation and age of the Quebec North Shore Moraine in the Rivière Moisie area.
GSC-1533	12 400 ± 160	Charlesbourg, Que.	LaSalle et al. (1977a) Lowdon and Blake (1973)	marine shells	"Oldest" minimum age for deglaciation of St. Lawrence estuary east of Québec City and begining of Champlain Sea Episode.
GSC-1592	6 460 ± 200	Michikamau Lake	Fulton and Hodgon (1979) Lowdon and Blake (1973)	peat	"Oldest" minimum age for deglaciation of the Upper Churchill River area.
GSC-1646	12 200 ± 160	Cantley, Que.	Romanelli (1975) Lowdon and Blake (1973)	marine shells	"Oldest" minimum age for deglaciation of lower Rivière Gatineau valley and incursion of Champlain Sea there.
GSC-1772	11 900 ± 160	Martindale, Que.	Romanelli (1975) Lowdon and Blake (1973)	marine shells	"Oldest" minimum age for deglaciation of middle Rivière Gatineau valley and incursion of Champlain Sea there.
GSC-2101	10 300 ± 100	Shawinigan, Que.	Occhietti (1980)	marine shells	"Oldest" minimum age for deglaciation in area between Montrealand Québec City.
GSC-2825	10 900 ± 140	Pinware, Labrador	Grant (1986) Lowdon and Blake (1979)	marine shells	"Oldest" minimum age for deglaciation of area north of the Straits of Belle Isle and for Goldthwait Sea on the north shore of the St. Lawrence.
GSC-2946	9 120 ± 480	Hudson Strait	Fillon and Harmes (1982) Lowdon and Blake (1980)	marine shells	"Oldest" minimum age for deglaciation of eastern Hudson Strait.
GSC-2970	7 600 ± 100	Northwest River, Labrador	Lowdon and Blake (1980)	marine shells	"Oldest" minimum age for deglaciation of the western end of Lake Melville.
GSC-3022	10 400 ± 140	Lake Hope Simpson, Labrador	Blake (1982); Engstrom and Hansen (1985)	gyttja	"Oldest" likely reliable minimum age for deglaciation of Alexis River area.
GSC-3067	9 640 ± 170	Moraine Lake, Labrador	Blake (1982) Engstrom and Hansen (1985)	gyttja	"Oldest" minimum age for deglaciation of upper St. Paul River area.
GSC-3094	6 320 ± 180	Lac Delorme, Que.	Richard et al. (1982) Blake (1982)	gyttja	"Oldest" minimum age for deglaciation of the ice devide area in the Lac Caniapiscau region of central New Quebec.
GSC-3241	6 500 ± 100	Border Beacon, Labrador	Blake (1982)	gyttja	"Oldest" minimum age for deglaciation of area northeast of Smallwood Reservoir.
GSC-3460	10 400 ± 200	Montreal River, Ont.	Veillette (1988)	gyttja	"Oldest" minimum age for deglaciation of southern portion of Lac Témiscamingue area and minimum age for McConnell Lake Moraine (part of Harricana Interlobate Moraine) in that area.
GSC-3467	10 100 ± 180	Lac Kipawa, Que.	Veillette (1988)	gyttja	With GSC-3460 "oldest" minimum age for deglaciation of area just south of Lac Témiscamingue.
GSC-3615	6 510 ± 110	Lac Gras, Que.	King (1985)	gyttja	"Oldest" minimum age for deglaciation of upper Rivière Moisie area.

Table 3.6. (cont.)

Laboratory Dating no.	Age (years BP)	Locality	Reference	Material	Significance
GSC-3644	6 200 ± 100	Lac Starkel, Que.	King (1985)	gyttja	"Oldest" minimum age for deglaciation of the wide-mouthed U-shaped retreat center northeast of Lac Opiscotéo.
GSC-3947	7 130 ± 100	Deception Bay, Que.	B. Lauriol (personal communication, 1986)	marine shells	Shells dated are from immediate vicinity and about at same altitude as other collection of Matthews (I-488) which gave an age of 10 450 ± 250 BP.
GX-5522	11 160 ± 520	Moraine Lake, Labrador	Short (1981)	lake mud (low organic content)	Lakes situated on distal site of Saglek moraines. With GX-6362 date provides support for ice free areas in northern Labrador in the Late Wisconsinan?
GX-6345	10 275 ± 225	Makkovik Harbour, Labrador	Barrie and Piper (1982)	foraminifera in marine muds	"Oldest" minimum age for deglaciation of area north of Lake Melville.
GX-6362	18 210 ± 1900	Square Lake, Labrador	Short (1981) Clark et al. (1986)	lake mud (low organic content)	Lake situated on distal side of Saglek moraines. With GX-5522 date provides support for ice free areas in northern Labrador in the Late Wisconsinan?
GX-8240	34 200 +2100/-1600	Iron Strand, Labrador	Clark (1984a)	marine shells	Date implies that Iron Strand site records a Middle Wisconsinan Interstade.
GX-8241	28 200 +1200/-1000	Iron Strand, Labrador	Clark (1984a)	marine shells	Date implies that Iron Strand site records a Middle Wisconsinan Interstade.
GX-9293	9 110 ± 470	Shoal Cove, Labrador	Clark (1984a)	marine shells	With L-642 and TO-305 "oldest" minimum age for deglaciation of the Labrador coast east of the Torngat Mountains.
I-488	10 450 ± 250	Deception Bay, Que.	Matthews (1967)	marine shells	Oldest shell date for deglaciation of the south shore of Hudson Strait, but is likely erroneous (see GSC-3947).
I-5922	10 400 ± 150	Sacre-Coeur-de Saguenay, Que.	Dionne (1977)	marine shells	Oldest age for deglaciation of the mouth of Rivière Saguenay area. Oldest age determination for Laflamme Sea.
I-8363	8 230 ± 135	Kuujjuarapik, Que.	Hillaire-Marcel (1976)	concretion	Maximum age for construction of Sakami Moraine and incursion of Tyrrell Sea east of Hudson Bay.
I-9632	6 990 ± 150	Payne Bay, Que.	Gray et al. (1980)	marine shells	"Oldest" minimum age for deglaciation of south-western Ungava Bay and incursion of D'Iberville Sea.
L-642	9 000 ± 200	Eclipse Channel, Labrador	Løken (1962b)	marine shells	With GX-9293 and TO-305 "oldest" minimum age for deglaciation of the Labrador coast east of the Torngat Mountains.
QU-122	7 880 ± 160	Rivière La Grande, Que.	Hardy (1976)	marine shells	Maximum age for drainage of Lake Ojibway and oldest available minimum age for construction of Sakami Moraine and incursion of Tyrell Sea.
QU-574	9 970 ± 130	Rivière des Anglais, Que.	Dubois (1980)	marine shells	"Oldest" minimum age for deglaciation of the Rivière Manicouagan area.
SI-1737	10 240 ± 1240	Saint-John Island, Labrador	Jordan (1975)	gyttja	"Oldest" minimum age for deglaciation of eastern Lake Melville area.
SI-1959	6 815 ± 125	Pyramid Hills Lake, Que.	Short (1981)	lake mud	"Oldest minimum age for deglaciation of area southeast of Ungava Bay.
SI-3139	10 550 ± 290	Eagle River, Labrador	Lamb (1980)	gyttja	"Oldest" minimum age for deglaciation of area south of the Mealy Mountains and for construction of the Paradise Moraine.
SI-4131	34 360 ± 850	Iron Strand, Labrador	Ives (1977) Short (1981)	marine shells	Date implies that Iron Strand site records a Middle Wisconsinan Interstade.
TO-200	7 970 ± 90	Central Lake Melville, Labrador	Vilks et al. (1987)	marine shells	"Oldest" reliable minimum age for the deglaciation of central Lake Melville.
TO-305	9 830 ± 70	Nachvak Fiord, Labrador	R.J. Rogerson (personal communication, 1986)	marine shells	"Oldest" minimum age for deglaciation of the Labrador coast east of the Torngat Mountains.
UQ-547	6 700 ± 100	Rivière Nastapoca, Que.	Allard and Seguin (1985)	marine shells	"Oldest" minimum age for deglaciation of the Nastapoka Sound area.
Y-1165	>42 000	Rivière Harricana, Que.	Stuiver et al. (1963)	peat	Minimum age for Missinaibi Formation deposits in the Quebec James Bay Lowlands.

glacial deposits and weathering zones, augmented by weathering and stratigraphic studies, must be completed before the Quaternary geology of the Torngat Mountains can be more clearly interpreted.

Recent studies by Clark (1984a) in the Iron Strand area of northern Labrador, and by Evans (1984), Evans and Rogerson (1986), and by Bell et al. (1987) in the Nachvak Fiord area, provide further insight into the Quaternary history of the Torngat Mountains. On the basis of field mapping and ice sheet profile reconstructions, Clark (1984a) and Clark and Josenhans (1986) recorded two glacial advances. The older, which they correlate with the Koroksoak Glaciation of Ives (1978), deposited the Iron Strand drift, left nunataks in the Torngats, and extended onto Saglek Bank. The younger advance, which they correlate with the Saglek Glaciation of Ives (1978), deposited Two Loon drift, left large ice-free areas, and extended according to them out to the shelf edge through the deep areas on the shelf (Clark and Josenhans, 1986). Clark (1984b) estimated that the ice surface on the drainage divide was 818 m a.s.l. west of Kangalaksiorvik Fiord and 606 m a.s.l. west of Eclipse Channel.

At Iron Strand, in an area known to have been covered by the ice that deposited Iron Strand drift but not by the younger ice, massive marine clayey sands are overlain by a silty sand till or glaciomarine diamicton containing abraded marine shells, and by fossiliferous nearshore marine sands. South of this site, a soil thicker than that developed during the Holocene is present above similar beds. Shells from the diamicton dated 34 200 +2100/−1600 BP (GX-8240); likely in situ shells from the overlying marine sands dated 28 200 +1200/−1000 BP (GX-8241); and shells from colluvium derived from the same bluffs are 42 730 +6680/−9970 BP (DIC-517) and 34 360 ± 850 BP (SI- 4131; see Ives, 1977; Short, 1981). The shell dates indicate that the diamicton records a pre-Late Wisconsinan advance of continental ice which reached the sea. Based on radiocarbon dating and amino acid analyses, Clark (1984a) tentatively corelated the Iron Strand drift with the Loks Land Member of the Clyde Foreland Formation of Baffin Island and suggested that it represents an event which occurred between 100 ka and 34 ka. Andrews and Miller (1984) suggested that the Iron Strand fossiliferous deposits represent a Middle Wisconsinan nonglacial interval because the amino acid racemization of shells was significantly less than those which typify the Kogalu aminozone, possibly of Sangamonian age (this chapter, Andrews, 1989). If the radio-carbon dates are considered to lie beyond the limit of the method, as is commonly done with shell dates in this age range, the Iron Strand drift could be older than Middle Wisconsinan. Despite the problem of absolute age, the record in the Iron Strand area and the minimum ages support the Quaternary framework established on the basis of weathering zones and confirm that parts of the north coast of Labrador remained ice-free throughout the Late Wisconsinan (Ives, 1977).

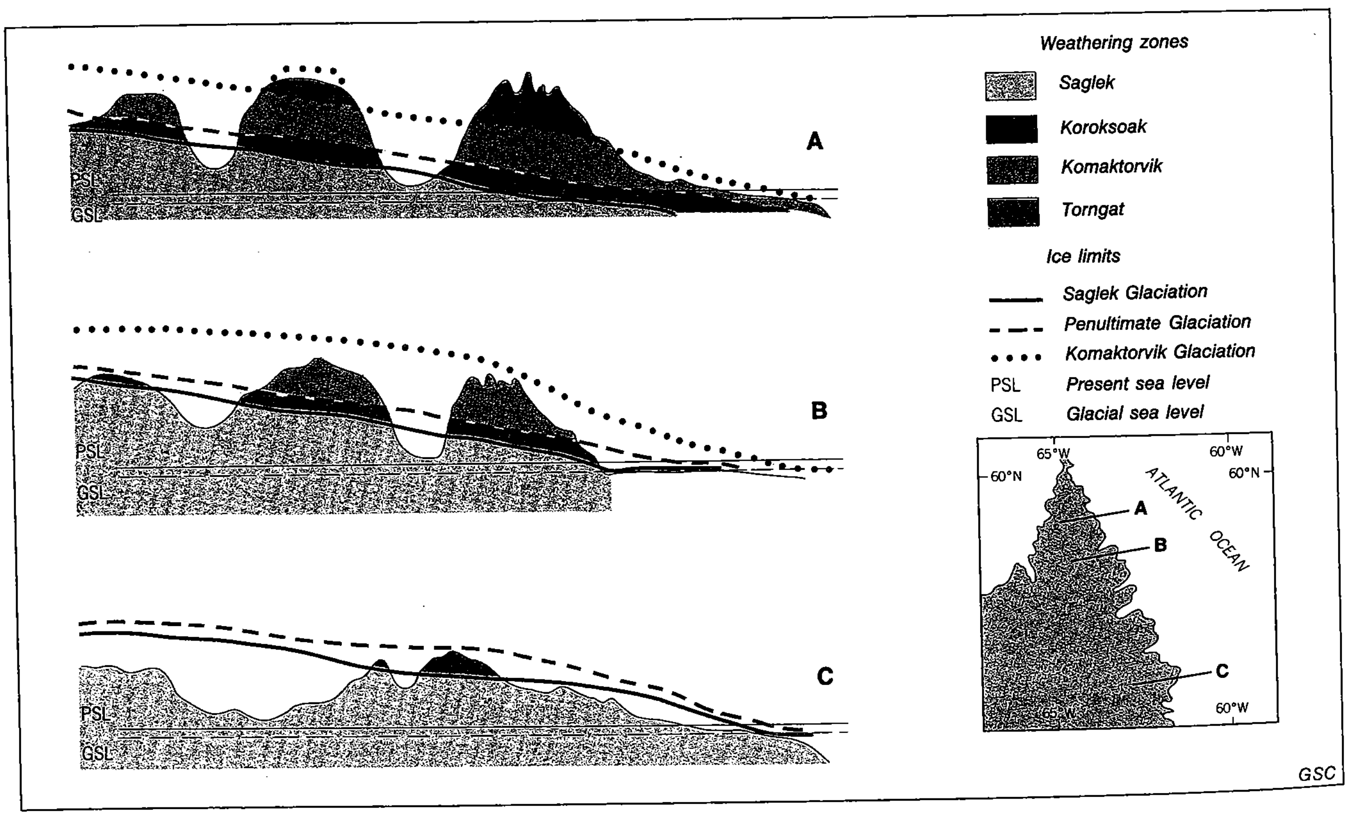


Figure 3.53. Cross-sections, drawn perpendicular to the general trend of the Laurentide Ice Sheet margin, showing ice profiles and weathering zones of the northern and central Labrador highlands. Inset shows locations of A, B, and C (after Ives, 1978).

Based on detailed mapping, Evans (1984), Evans and Rogerson (1986), Rogerson and Bell (1986), and Bell et al. (1987) recognized two advances of Laurentide ice in the Nachvak Fiord area. Their Ivitak phase is the oldest and most extensive; nunataks were present but the ice extended onto the shelf. As in the Iron Strand area, shells from a glacial marine diamicton assigned to the Ivikak phase have provided four finite radiocarbon ages in the Middle Wisconsinan time range (Bell et al, 1987). The subsequent Nachvak phase was less extensive: the ice only extended to the mouths of fiords and is marked west of the head of Nachvak Fiord by the Tinutyaruik Moraine. On the basis of soil development studies and the older radiocarbon dates, the workers believed that deposits laid down during Ivitak phase glaciation are >40 ka old and are equivalent in age to the Iron Strand drift of Clark (1984a). They further believed that their younger Nachvak phase is equivalent in age to the Saglek Glaciation. Even though correlations are tentative, the studies also confirm that parts of northern Labrador remained ice free at least during the Late Wisconsinan and that the ice may not have extended onto the Labrador Shelf at that time.

Additional support for ice-free areas during the Late Wisconsinan may be provided by a radiocarbon age of 18 210 ± 1900 BP (GX-6362) on organic sediments at the base of a core from a lake on the distal side of and dammed by the Saglek Moraines (Short, 1981; Clark et al., 1986). However, since the organic content in the dated core was very low (<0.1% total organic matter) the accuracy of the age determination can be questioned (see discussions in Short, 1981). Whatever the case the date probably provides a maximum age for the Saglek Moraines (Clark et al., 1986). Another age determination of 11 160 ± 520 BP (GX-5522) dates the recession of the ice front from the Saglek Moraines but may also be too old.

Mealy Mountains and southeastern Labrador

Various researchers (Gray, 1969; Rogerson, 1977; Fulton and Hodgson, 1979) have speculated that summits of the Mealy Mountains were not overtopped by Late Wisconsinan ice, and Fulton and Hodgson (1979) have suggested that the Paradise Moraine (Fig. 3.47) marks the Late Wisconsinan glacial limit. Differences are readily apparent in the character of the terrains separated by moraine systems. Radiocarbon ages, to 21 ka obtained from sediments in a lake core near Alexis River by H.E. Wright, Jr. (University of Minnesota, personal communication, 1979) and in a core taken offshore from Lake Melville (Vilks and Mudie, 1978) appear to lend support to the idea of a Late Wisconsinan ice-free area in southeastern Labrador. The reliability of the dates can be questioned, however, since the offshore dates are based on total organic content (Fillon et al., 1981) and contamination from older carbon could account for the old gyttja dates on land.

Late Wisconsinan buildup and limit

The entire southeastern Canadian Shield, except for some nunataks in northern Labrador, possibly summits of the Mealy Mountains, and possibly the region lying beyond the Paradise Moraine, was covered by Late Wisconsinan ice of the Labrador Sector of the Laurentide Ice Sheet. In this section the sparse evidence on glacial buildup and the problems of defining the glacial limit on the Atlantic seaboard are discussed.

It is generally agreed that initial development of the Labrador Sector of the Laurentide Ice Sheet in the Late Wisconsinan (or much earlier?) was on the interior uplands and that glacial flow was radial to limits on the Labrador Shelf and in the Gulf of St. Lawrence, and was radial to zones of confluence with adjacent ice masses in the west and in the north. In two areas of the southeastern Canadian Shield, ice flow indicators record movements different from those thought to have been produced during the last flow phase. Whether these flows are related to Late Wisconsinan dispersal centres or whether they are older is not known.

In southern Labrador, Klassen (1983a, 1984) and Klassen and Bolduc (1984), postulated a dispersal centre situated between Churchill River and the Gulf of St. Lawrence. This is based on striae indicating northward and north-northeastward flow in the area north of Lake Melville and in the upper Churchill River region. The striae predate the Late Wisconsinan regional northeastward, eastward, and southeastward flow . The age of this dispersal centre is not known but it could be as young as "early" Late Wisconsinan. Occhietti (1982) argued for an Early Wisconsinan accumulation centre on the central Laurentian Highlands to explain the Quaternary record in the St. Lawrence Lowlands, and Quinlan and Beaumont (1982) have suggested independently, on the basis of the sea level record in Atlantic Canada, that a major ice dome must have existed in the same area during the "early" Late Wisconsinan. The observations of Klassen and Bolduc may be field evidence for Occhietti's and Quinlan and Beaumont's suggestions. Farther north in the Schefferville area, Klassen and Thompson (1987) recently identified five distinctive phases of ice flow requiring different or major shifts of dispersal centres. Because, according to Klassen and Thompson (1987, p. 65) "the ice flow features recorded on outcrop surfaces do not show differential weathering, the phases ..... appear to have been produced during a period of continuous ice cover and may represent changes during one or more stadials of Wisconsin Glaciation".

In Abitibi, Lac Témiscamingue, and adjoining regions of western Quebec, striae and glacially transported materials indicate that ice flowing southwest from New Quebec extended west of the Harricana Interlobate Moraine (the feature that marks the final contact between Labrador Ice and Hudson Ice; Chauvin, 1977, p.16; Kish et al., 1979, p. 8, Veillette, 1982, 1986a). Time of this event is not known. There is also evidence that at one time ice flowing southeast out of James Bay extended at least 20 km east of the Harricana Moraine in the area between Rivière Harricana and Rivière Nottaway (L. Hardy, Poly-Géo. Inc., Longueil, Quebec, personal communication, 1985). In the area east of James Bay, the existence of eastward moving ice proposed by Lee (1959b) was later rejected by Dionne (1974) who attributed the presence of erratics from a western provenance, noted by Lee, to sea ice rafting. On Ungava Peninsula, Bouchard and Marcotte (1986) saw no evidence for ice flowing eastward from Hudson Bay. Much farther inland, in the Lac Mistassini area, Martineau (1984b), Prichonnet et al. (1984), and Bouchard and Martineau (1985) have documented southeastward ice flow which preceded the regional southwestern ice flow from New Quebec recorded on the eastern and western side of the Harricana Moraine. At this time it is impossible to say which flow directions relate to the Late Wisconsinan maximum and which ones might be older. Whatever the case, at some time before final deglaciation, ice from central New Quebec carried materials west of the Harricana Interlobate Moraine, whereas ice from a dispersal centre west of the east coast of James Bay flowed southeastwardly as far as Lac Mistassini. The identification along the middle St. Lawrence estuary of stromatolitic dolomite from the Lac Mistassini and Lac Albanel area (Dionne, 1986) indicates that the southeasterly flows could have been even more extensive.

The Late Wisconsinan extent of Laurentide ice on the Atlantic seaboard of Labrador is not definitely known and is the subject of much controversy (Fig. 3.54). Even though early workers such as Bell (1884), Daly (1902), and Coleman (1921) suggested that parts of northeastern and southeastern Labrador had not been glaciated, the tendency until recently, as expressed by Flint (1957) and Prest (1969), has been to portray the Late Wisconsinan ice sheet as extending well offshore (Fig. 3.54). Based on the weathering zone concept, areas of eastern Labrador have more recently been shown as having escaped Late Wisconsinan glaciation (Ives, 1960a, 1978; Hughes et al., 1981; Fig. 3.54). Based in part on the location of the Paradise Moraine, which Fulton and Hodgson (1979) speculated might mark the limit of Late Wisconsinan ice in southeastern Labrador, and on data from offshore investigations (Vilks and Mudie, 1978), Rogerson (1981, 1982), and Prest (1984) have shown relatively large ice-free areas on land (Fig. 3.54). Finally, using the mapped extent of a distinctive offshore till sheet assigned to the "early" Late Wisconsinan on the basis of total organic radiocarbon ages (deemed anomalously old by Fillon et al., 1981), Josenhans et al. (1986) and Clark and Josenhans (1986) have reasserted that Late Wisconsinan ice extended well offshore onto the continental shelf (Fig. 3.54). They proposed that outlet glaciers in the Torngat Mountains in fact advanced onto the shelf, coalesced and deposited a till sheet.

None of the models portrayed in Figure 3.54 can be validated at this time, but the best interpretation at the moment is that areas of northern Labrador, summit areas of the Mealy Mountains, and possibly areas distal to the Paradise Moraine were not covered by Late Wisconsinan ice, and that the ice that covered the rest of the landmass, flowed onto the shelf. Until (1) glacial limits and weathering zones are mapped everywhere in detail, (2) in depth studies are undertaken, as on Baffin Island, to document the relative differences between all weathering zones, and (3) stratigraphic studies accompanied by unequivocal radiometric dates are undertaken, the location of the Late Wisconsinan limit will remain in question.

Deglaciation history

Ice initially retreated in water (Fig. 3.47) in all peripheral areas of the southeastern Canadian Shield. This circumstance greatly influenced the style of deglaciation. A series of paleogeographic maps (Fig. 3.55) portrays the ice cover and location of major moraines, glacial lakes, and postglacial seas for the period 12 ka to 7 ka. Figure 3.52 provides a tentative correlation of the events in each of the regions discussed and is referred to in the following discussions.

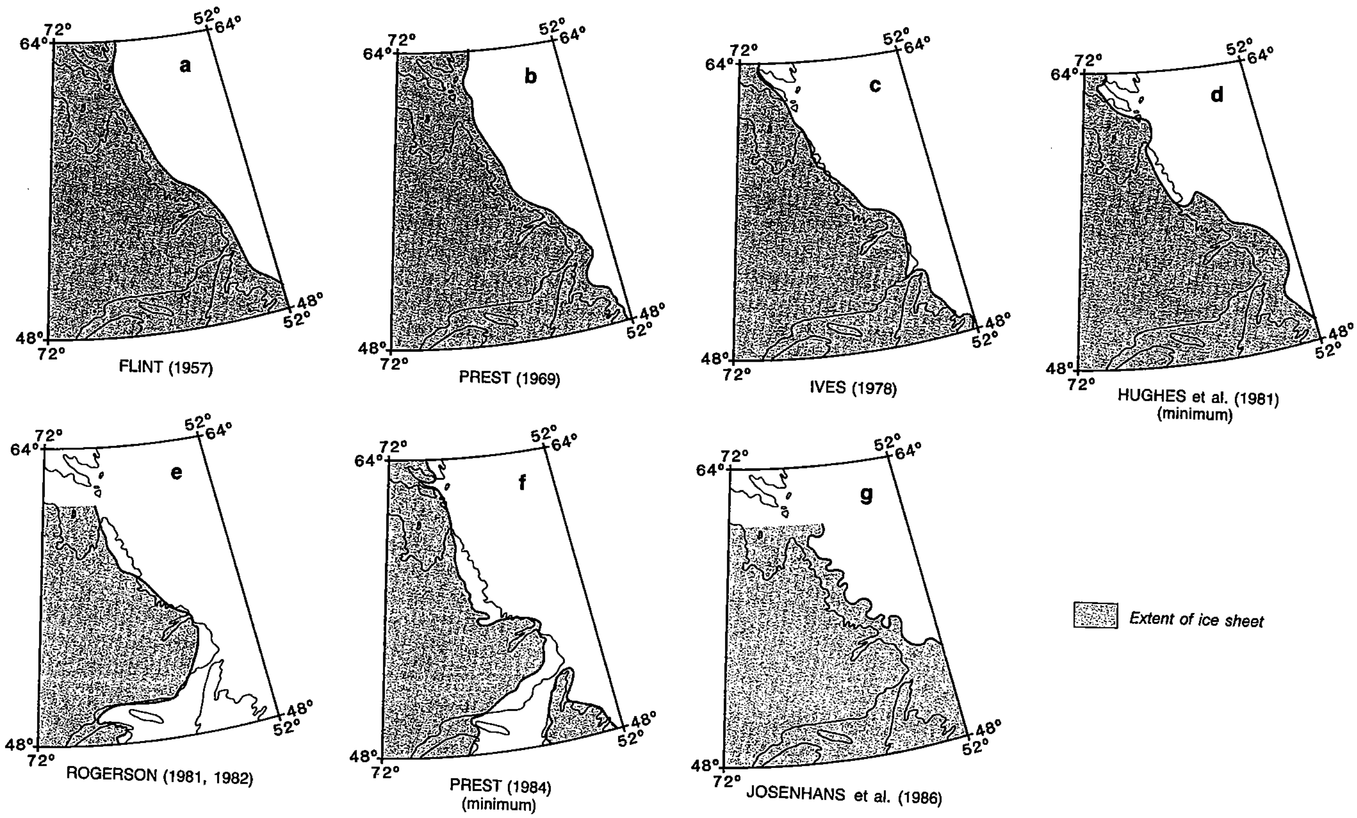


Figure 3.54. Speculative Late Wisconsinan limits of the Laurentide Ice Sheet on the Atlantic seaboard of Labrador as proposed by various authors.

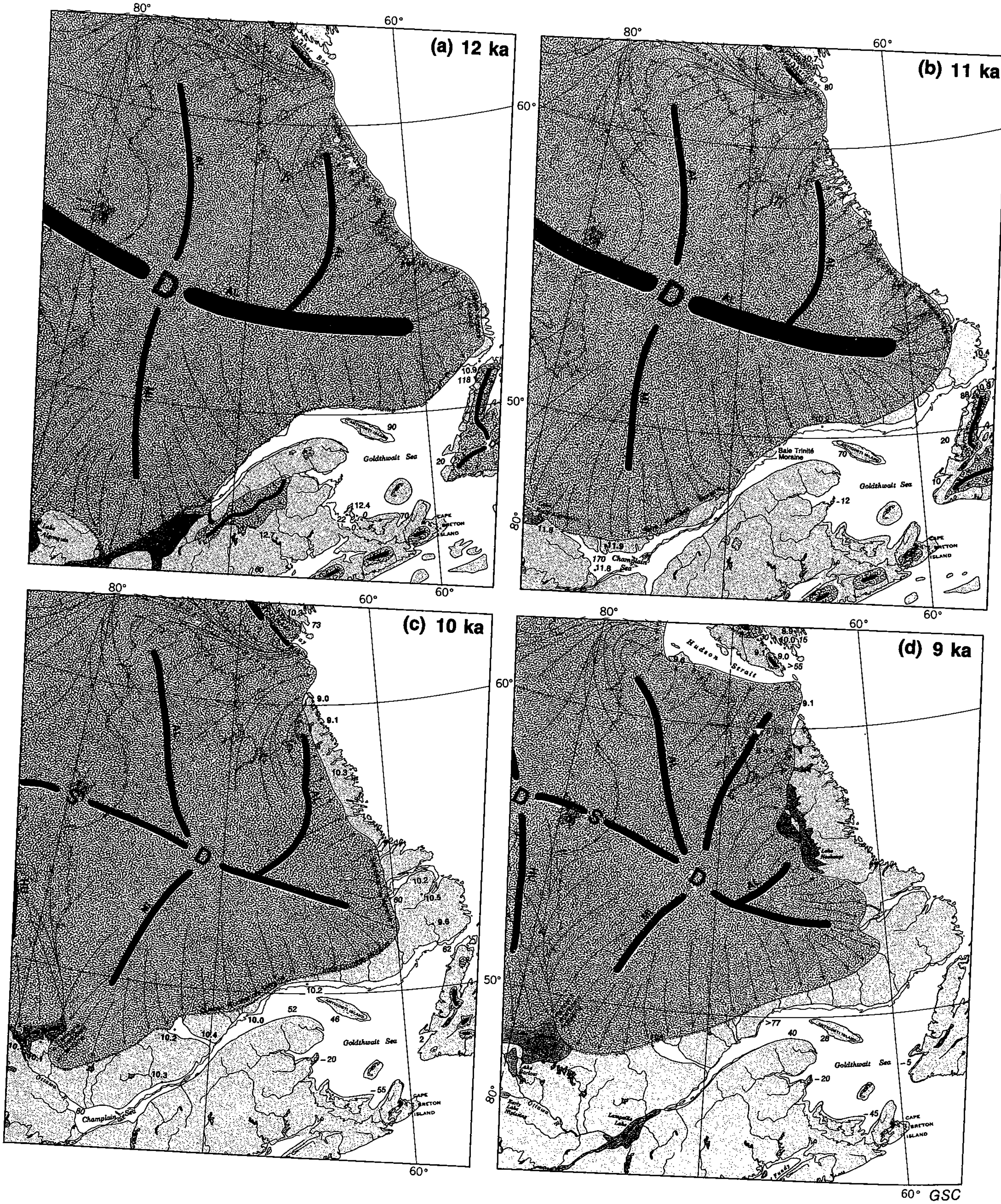
Northern Labrador and northeastern New Quebec

During the Late Wisconsinan, ice of the Labrador Sector advanced east northeasterly from central New Quebec and easterly from Ungava Bay (Ives, 1957, 1958a; Løken, 1962a) towards the Labrador Shelf over northeastern New Quebec and northern Labrador (Saglek Glaciation in the Torngat Mountains; Andrews, 1963a). Higher coastal summit areas from Fraser River north and much of the more northerly Torngat Mountains and adjacent coastal forelands were not overtopped by ice. Clark (1984a; Fig. 3.56) recently illustrated that ice from west of the watershed crossed the Torngat Mountains in the form of outlet glaciers, extended to fiord mouths, and did not override large nunatak areas comprising the Koroksoak and higher weathering zones. The probable upper limit of extent of Late Wisconsinan ice is documented in several local studies but, except in Andrews (1963a), Clark (1984a), Evans (1984) and Evans and Rogerson (1986), no attempt has been made at mapping precisely the limit of ice cover. Generally the upper trimline of Saglek Glaciation not only slopes from the divide to the ocean but also slopes down towards the north. In the area between Fraser River and Okak Bay, ice reached a maximum elevation of about 700 m (Andrews, 1963a). Mount Thoresby and Man O'War Peak are the most southerly nunataks (Andrews, 1963a; Johnson, 1969). The limit in the Saglek Fiord area is 615 m (Smith, 1969); in the Ryans Bay region, Clark (1984a) stated that the ice passing through the Torngat Mountains did not reach elevations of more than 800 m on the drainage divide. In summit areas lying above the Saglek Glaciation level, cirque glaciers or small local ice caps existed which were independent of the Laurentide Ice Sheet (Fig. 3.56; Ives, 1960a; Løken, 1962a; Evans, 1984; Clark, 1984a; Evans and Rogerson, 1986; Bell et al., 1987). The question of the actual extent of ice on the shelf is the subject of much controversy. Initial studies by Clark (1984a) and the studies of Evans (1984), Evans and Rogerson (1986), and Rogerson and Bell (1986) indicated that the Late Wisconsinan ice only extended to the mouths of the fiords. In opposition other studies, as mentioned above (Josenhans et al., 1986; Clark and Josenhans, 1986), stated that the same ice extended well offshore to near the shelf edge.

Deglacial events in rugged coastal areas are complex and have been discussed for various locations particularly by Ives (1958a, 1960a), Løken (1962b, 1964), Tomlinson (1963), Andrews (1963a), Johnson (1969), Clark (1984a), Evans (1984), and Evans and Rogerson (1986) . At the limit reached by the Late Wisconsinan ice, extensive systems of lateral moraines and kame complexes were built (Saglek Moraines of Ives, 1976; Fig. 3.52). Several end and lateral moraines mark positions of halts or local readvances of the ice front during retreat from the Saglek Moraines towards an ice mass located west of the Atlantic/Ungava Bay watershed (Fig. 3.47). Notable examples are the Tasiuyak Moraines in the Fraser River-Okak Bay area (Andrews, 1963a), and the well correlated Noodleook, Two Loon, and Kangalaksiorvik (= Sheppard) moraines of Løken (1962b, 1964) on Torngat Peninsula. Andrews (1977) suggested that the Kangalaksiorvik Moraines, which extend from Killinek Island to Ryans Bay may be correlative with moraines of Cockburn age elsewhere in Arctic Canada (Andrews and Ives, 1978). J.D. Ives (University of Colorado, personal communication, 1985) has observed ice flow indicators between the southwestern Torngat Mountains and lower Rivière George, and along the east coast of Ungava Bay which clearly indicate a late reversal of ice flow into Ungava Bay. He interpreted this as resulting from the development of a calving bay in northern Ungava Bay which extended rapidly southward until the entire coastline was ice free.

Because coastal areas were glacial isostatically depressed, waters of the Atlantic Ocean submerged both ice-free and newly deglaciated low-lying areas. Little is known about sea level history in the area, but generally marine limit progressively decreases in elevation towards the north. Marine waters may have reached elevations of about 93 m south of Okak Bay (Andrews, 1963a), 65 m in the Saglek Bay and Ryans Bay areas (Løken, 1962b; Smith, 1969), 73 m at the head of Nachvak Fiord (Bell et al., 1987), 42 m north of Ryans Bay (Løken, 1964), and 16 m on Killinek Island (Løken, 1964). Shells collected by R.J. Rogerson (Memorial University personal communication, 1986) from the head of Nachvak Fiord gave an age of 9820 ± 70 BP (TO-305), the oldest Holocene age so far obtained on the north coast. Other shells collected by Løken (1962b) north of Ryans Bay gave an age of 9000 ± 200 BP (L-642) and by Clark (1984a) from south of Iron Strand gave an age of 9820 ± 470 (GX-9293). The presence in northernmost Labrador of tilted shorelines truncated by a 15 m high horizontal shoreline is considered by Løken (1962b) as recording an early Holocene transgression and also suggests that thick continental ice did not overlie the northern tip of Labrador.

During deglaciation of the Torngat Mountains, numerous small and possibly ephemeral glacial lakes were ponded on the east side of the drainage divide in tributary valleys blocked by ice tongues. As the ice margin receded downslope on the west side of the mountains, larger and longer lasting glacial lakes were created. Lakes in tributary valleys of Rivière Alluviaq drained into a fiord south of Iron Strand (Ives, 1957), whereas other lakes, in the Rivière Koroc basin drained towards Saglek Bay (Ives, 1958a; Fig. 3.47). The most important lakes were the Naskaupi and McLean glacial lakes (Ives, 1960a,b; Matthew, 1961; Barnett and Peterson, 1964; Barnett, 1964, 1967; Peterson, 1965; Fig. 3.47), which extended over large areas in the upper Rivière George and Rivière à la Baleine drainage basins. These lakes were dammed between the watershed crest on the east, the southwestward retreating main body of Labrador Ice on the west, and the inferred late ice ridge or dome over Ungava Bay. The configuration of the ice required to hold the lakes is discussed by Prest (1970, 1984). Glacial Lake Naskaupi cut a series of well defined strandlines, some of which are incised in bedrock. Probable outlets, which permitted drainage of water to the Labrador Coast, are the headwaters of Fraser, Kogaluk, Harp, Kanairiktok, and Naskaupi rivers. Glacial Lake McLean, in upper Rivière à la Baleine basin, was separate from Lake Naskaupi but drained into it by a channel just west of Lac de la Hutte Sauvage. Both lakes finally drained into the D'Iberville Sea (postglacial Ungava Bay) when ice had retreated sufficiently, perhaps through the development of a calving bay, to allow penetration of marine waters.


(a) 12 ka
(b) 11 ka
(c) 10 ka
(d) 9 ka
80°
60°
50°
Goldthwait Sea
CAPE BRETON ISLAND
Baie Trinité Moraine
Champlain Sea
Hudson Strait
GSC

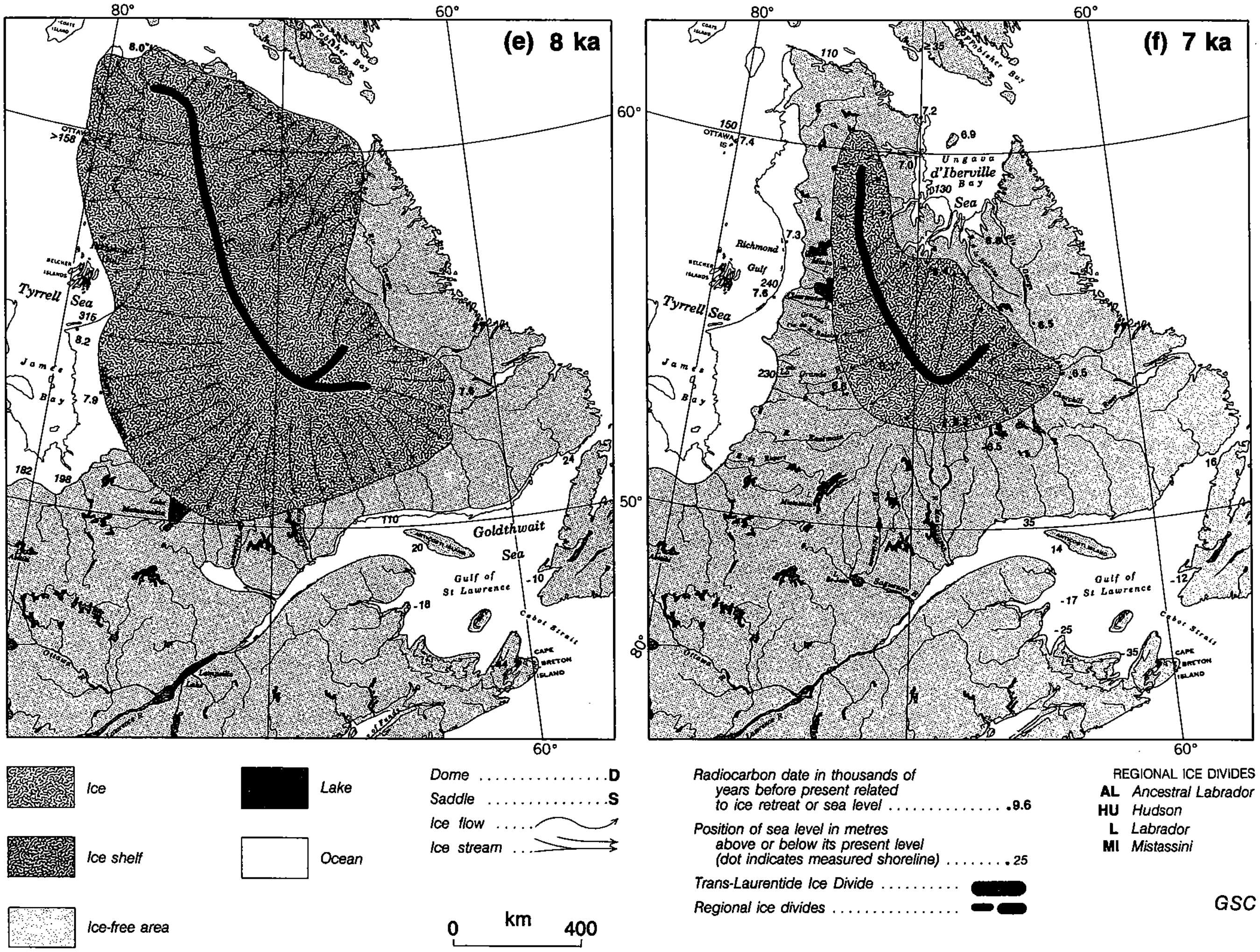


Figure 3.55. Paleogeographic maps of the southeastern Canadian Shield area showing ice cover, location of major end moraines, glacial lakes, and postglacial seas for (a) 12 ka, (b) 11 ka, (c) 10 ka, (d) 9 ka, (e) 8 ka, and (f) 7 ka.

Figures 3.55a to 3.55f present a speculative reconstruction for the deglaciation of northern Labrador and northeastern Quebec. Ice is shown as remaining for a long time east of Ungava Bay in such a way that the damming of glacial lakes Naskaupi and McLean and other lakes to the north can be accounted for. The time when the ice receded from the Saglek Moraines is unknown. Numerous radiocarbon age determinations on lake cores or bogs are reported by Short (1981). Many of the dates appear suspect because of anomalously old ages, the very low organic carbon content of many dated samples, age reversals in the cores, and the possibility of incorporation of old organic carbon. It seems preferable to rely solely on the few shell dates that give a minimum age for deglaciation of the coastal areas of about 9 ka. Inland, in the lower Rivière George basin, an age of 6815 ± 125 BP (SI-1959; Short, 1981) may provide a minimum age for deglaciation and drainage of glacial Lake Naskaupi. Abandoned moraines and lichen-kill areas adjacent to present cirque glaciers, and glacier-free cirques provide a record of neoglacial expansion of glaciers in the Torngat Mountains (McCoy, 1983; Clark, 1984a; Evans, 1984, Evans and Rogerson, 1986).

Central and southern Labrador

Ice from the Labrador Sector advanced towards the coast of central and southern Labrador. As previously mentioned, summit areas in the Mealy Mountains and a large area on the distal side of the Paradise Moraine may not have been overridden by Laurentide Ice during the Late Wisconsinan. Based on lateral moraines sloping to the east at the limit of what appears to be different weathering zones, Late Wisconsinan ice may have reached elevations of only 710-555 m in the Mealy Mountains south of Lake Melville

(Gray, 1969), 500-300 m in the Mealy Mountains north of Sandwich Bay, and 275-122 m near the coast east southeast of Sandwich Bay (Rogerson, 1977).

Ice flow during deglaciation was generally northeastward in central Labrador, eastward in the Lake Melville area, and southeastward in southeastern Labrador. Flow was controlled to a great extent by the Mealy Mountains which parted the flow, and by Lake Melville where a calving bay may have existed (Fulton and Hodgson, 1979). Major end moraines were built during the retreat phase. If Laurentide Ice covered all of southeastern Labrador and joined with the Newfoundland Ice Cap, then the Bradore and Belles-Amours moraines (D.R. Grant, unpublished; Fig. 3.47) are likely Late Wisconsinan retreatal features. If, on the other hand, southeastern Labrador remained ice free, these moraines must be older and possibly date from a Middle or Early Wisconsinan stade, with the Paradise Moraine farther west representing the Late Wisconsinan stadial maximum. As shown in Figure 3.47 and 3.55a-f, many moraines were built during retreat of ice from the area. The longest of these is the Sebaskachu-Little Drunken Moraine System (Blake, 1956; Fulton and Hodgson, 1979), which apparently is part of the Quebec-North Shore System of Dubois and Dionne (1985). Later retreat towards central New Quebec was marked largely by construction of numerous eskers and fluted landforms (Fig. 3.39). Generally free drainage to the Atlantic prevented major glacial lake development. Rhythmites, observed in the Naskaupi River basin (Blake, 1956) and elsewhere (surficial geology maps of Fulton, 1986a, b and Fulton et al., 1979-1981), may be marine rhythmites developed in estuaries with low salinity.

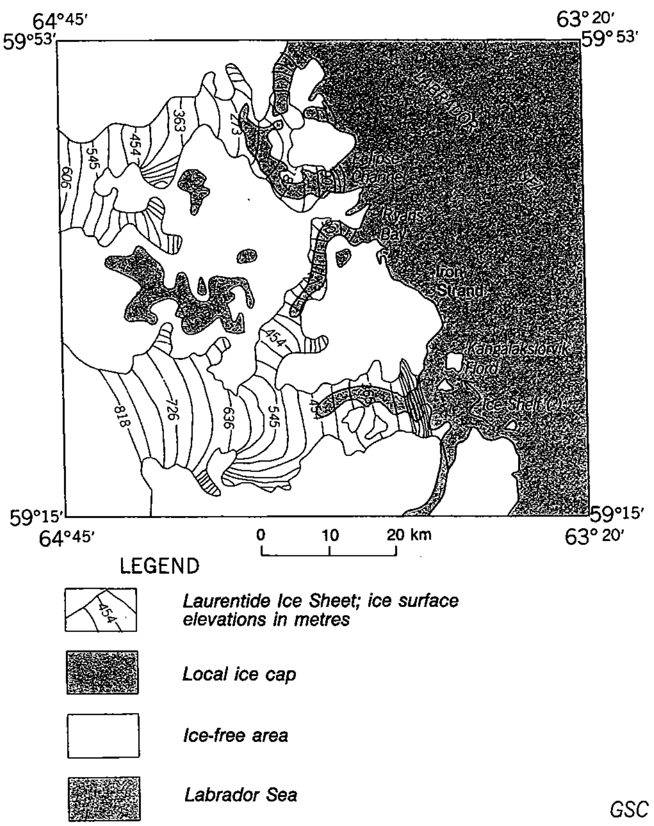


Figure 3.56. Ice sheet reconstruction in northern Labrador during the Late Wisconsinan (after Clark, 1984a).

Atlantic Ocean waters submerged glacial isostatically depressed coastal areas of central and southern Labrador during deglaciation. Marine limit has been traced by R.J. Fulton (Geological Survey of Canada, personal communication, 1984) and is shown on Figure 3.47. Along the coast, north of Straits of Belle Isle it may have reached 150 m (D.R. Grant, Geological Survey of Canada, personal communication, 1985), between Sandwich Bay and Lake Melville 113 m (Rogerson, 1977), and northeast of Lake Melville 80-85 m (Hodgson and Fulton, 1972). In the Lake Melville area marine limit increases in elevation inland from about 75 m on the outer coast to 150 m west of the lake (Fitzhugh, 1973).

Based on few radiocarbon dates, and on the location of major moraines, the paleogeographic maps (Figs. 3.55a-f) provide a simple reconstruction of deglaciation. On the basis of radiocarbon ages of shells, the area north of the Straits of Belle Isle was ice free by at least 10 900 ± 140 BP (GSC-2825), the coastal area east of Kanairiktok River by 10 275 ± 225 BP (GX-6345), central Lake Melville by 7970 ± 90 BP (TO-200), and the west end of Lake Melville by 7600 ± 100 BP (GSC-2970)(Table 3.6). On the basis of radiocarbon-dated lake cores, the area south of Alexis River was ice free by at least 10 400 ± 140 BP (GSC-3022). The age determinations of 9640 ± 170 BP (GSC-3067), 10 550 ± 290 BP (SI-3139), and 10 240 ± 1240 BP (SI-1737) on lake sediments provide minimum ages for the deglaciation of the upper St. Paul River area, the southern Mealy Mountains, and the eastern end of Lake Melville, respectively, and provide a minimum age for the construction of the Paradise Moraine. Farther inland a date of 6460 ± 200 BP (GSC-1592) on peat provides a minimum age for the deglaciation of upper Churchill River area and a 6500 ± 100 BP (GSC-3241) date on lake sediments, a minimum age for the deglaciation of the upper Harp River basin of central Labrador.

Quebec North Shore

The Quaternary history of the coastal fringe north of the Gulf of St. Lawrence between the mouth of Saguenay River and the Quebec/Labrador border is relatively well known but little data are available for the area farther inland.

During the Late Wisconsinan, south flowing ice of the Labrador Sector covered the Quebec North Shore and extended to its limit in the Gulf of St. Lawrence (see Grant, 1989), without apparently overrunning the eastern tip of Anticosti Island (Gratton et al, 1984). Ice retreated generally northwestward during deglaciation (Fig. 3.55a-f). The eastern extremity of the North Shore and the headland in the Baie-Trinité region were likely the first areas to become ice free. Moraines were built in the Baie-Trinité area between 13.5 and 9 ka (Dredge, 1976b, 1983b). The Baie-Trinité Moraines are correlated with the about 11 ka old Paradise and St-Narcisse moraines (Fig 3.52, 3.55b).

The marine submergence (Fig. 3.47) in the Gulf of St. Lawrence east of Québec City (Gadd, 1964, p. 1253) was named Goldthwait Sea by Elson (1969) and details on its development in Quebec can be found particularly in Dionne (1977), Hillaire-Marcel (1979), and Dubois (1980). Maps showing the extent of the sea can be found in Dubois et al. (1984). Marine limit varies from 150 m in the Quebec/

Labrador border area (Boutray and Hillaire-Marcel, 1977) to 130 m and 122 m in the Rivière Mécatina and Rivière Natashquan areas, respectively (Dubois et al., 1984). Farther west in the Rivière Romaine to Rivière Moisie area it lies between 128-131 m (Dubois, 1977, 1980), in the Rivière Moisie to Baie Trinité between 100-130 m (Dredge, 1976a, 1983b), in the Rivière Manicouagan area between 138-145 m (J-M. Dubois, Université de Sherbrooke, personal communication, 1985), and finally near the mouth of Rivière Saguenay it is at about 167 m (J-M. Dubois, personal communication, 1985).

Radiocarbon dates on shells provide the best available minimum ages for deglaciation of the Quebec North Shore. In addition to the 10 900 $\pm$ 140 BP (GSC- 2825) date in Labrador near the Quebec border, the oldest reliable age determinations are from the area west of Rivière Romaine (10 230 $\pm$ 180 BP, Gif-3770), the Rivière Moisie area (9140 $\pm$ 200 BP, GSC-1337), and the Rivière Manicouagan area (9970 $\pm$ 130 BP, QU-574). Three radiocarbon dates 10.0 to 10.6 ka obtained by Tremblay in the Rivière Moisie area are erroneous since redated samples gave ages 1000 years younger (Dubois, 1980).

As ice retreated farther north onto the Canadian Shield, the Quebec North Shore Moraine (Dubois and Dionne, 1985), more than 800 km long, was built in areas generally lying between 200 and 400 m elevation (Fig. 3.47, 3.48). It extends from Rivière Manicouagan to south of Lake Melville in Labrador (Dubois, 1979, 1980; Fulton and Hodgson, 1979). The feature includes segments in the Rivière Manicouagan area (Sauvé and LaSalle, 1968), west of Rivière Moisie (Lac Daigle Moraine of Dredge, 1976b, 1983b), in the Rivière Moisie to Rivière Romaine area (the Manitou-Matamec Moraine of Dubois, 1976, 1977, 1979, 1980), and in the Rivière Natashquan to south of Lake Melville area (the Little Drunken Moraine of Fulton and Hodgson, 1979, and the Aguanus-Kenamiou Moraine of Dionne and Dubois, 1980). The moraine is considered by Dubois and Dionne (1985) to represent a halt of the ice sheet during a cooler climatic phase. Another possible explanation is that this is a re-equilibration moraine (Andrews, 1973; Hillaire-Marcel et al., 1981) which was built as the ice established a new equilibrium profile after retreating onto land from a major topographic depression (the Gulf of St. Lawrence) where it had retreated by calving. The age of the moraine is uncertain. The Goldthwait Sea was in contact with the moraine only locally in the Rivière Moisie and Rivière Manicouagan areas. On the basis of the paleogeographic setting of Goldthwait Sea radiocarbon-dated shells, Dubois (1980) and Dubois and Dionne (1985) suggested an age of 9.5 to 9.7 ka.

The ice margin retreated northward and northwestward from the Quebec North Shore Moraine towards central New Quebec leaving eskers and fluted landforms (Fig. 3.39). Rhythmites in river valleys were either deposited in lakes dammed by ice or ice contact deposits (Dubois, 1980) or more likely in low salinity estuaries of the Goldthwait Sea extending up the valleys. According to Dubois (1980), the Goldthwait Sea in the middle North Shore area attained its inland limit (128-131 m) at about 9.5 ka and had regressed to 106 m by 9.1 ka, to 75-76 m by 7.7 ka, to 45-46 m by 7.2 ka, and to 15 m by 5.2 ka. Inland the oldest minimum age for deglaciation, 6510 $\pm$ 110 BP (GSC-3615), was obtained from a lake core in the upper Rivière Moisie area (King, 1985).

Lac Saint-Jean area

Generally, south flowing ice of the Labrador Sector covered all the region and extended south of the Gulf of St. Lawrence. Tremblay (1971), Dionne (1973), and LaSalle and Tremblay (1978) have shown that the ice moved in a southerly direction except along Rivière Saguenay, where ice flow was generally southeasterly along the axis of the valley (Fig. 3.39). As the ice retreated from the uplands south of Lac Saint-Jean, small glacial lakes were dammed and De Geer moraines were built (LaSalle and Tremblay, 1978). A late ice lobe occupied the Lac Saint-Jean depression and ice contact materials were deposited at its receding margin (LaSalle et al., 1977a; LaSalle and Tremblay, 1978; Fig. 3.55c). Some of these deposits, such as those on the south side of Lac Saint-Jean, have been assigned to the Metabetchouane "halt" by LaSalle et al. (1977a) and should be considered as end moraines.

As the ice receded northwesterly up Saguenay Valley and into the Lac Saint-Jean basin, marine waters from the Gulf of St. Lawrence extended over lower lying, newly deglaciated land. This arm of the Goldthwait Sea was referred to as the Laflamme Sea by Laverdière and Mailloux (1956). Marine limit lies at 167 m (J-M. Dubois, Université de Sherbrooke, personal communication, 1985) at the mouth of Rivière Saguenay and at about 167 m and 198 m south and north of Lac Saint-Jean, respectively (LaSalle and Tremblay, 1978). The "oldest" minimum ages for marine invasion and deglaciation are 10 400 $\pm$ 150 BP (I-5922) at the mouth of Rivière Saguenay and 10 250 $\pm$ 350 BP (Gif-424) south of Lac Saint-Jean. Following deglaciation of the Lac Saint-Jean basin, the ice retreated northward leaving behind numerous eskers and fluted landforms (Fig. 3.39, 3.55d,e,f).

Western Laurentian Highlands

The western Laurentian Highlands are situated north of St. Lawrence River and Ottawa River between Rivière Saguenay and Lac Témiscamingue. Only events of the area situated north of the St. Narcisse Moraine (Fig. 3.47) are discussed here. The events that preceded and accompanied construction of the moraine are discussed by Parent and Occhietti (1988) and Occhietti (1989).

Ice flow during advance and retreat was controlled by topography with ice funnelled into the Rivière Saint-Maurice basin, and into the low general area centred on basins of Petite Nation, du Lièvre, Gatineau, and Coulonge rivers. The characteristic deglacial deposits are short segments of end moraines, isolated ice contact deposits, and outwash trains (Parry, 1963; Hardy, 1970; Denis, 1974; Lamothe, 1977; Pagé, 1977; Tremblay, 1977; Occhietti, 1980). In this moderate relief area higher summits first became ice free; at places this led to formation of small isolated glacial lakes perched on the flanks of higher hills, dammed by ice still occupying depressions. Such lakes are documented by Laverdière and Courtemanche (1960), Parry (1963), Lamothe (1977), and Pagé (1977). This style of deglaciation also led to development of typical kame and kettle topography in the Lac Maskinongé area where a remnant ice mass disintegrated (Denis, 1974). Fluted till terrains and extensive esker complexes are not as common as elsewhere on the southeastern Canadian Shield.

As ice retreated north of the St. Narcisse Moraine, the

Champlain Sea, which already flooded the glacial isostatically depressed St. Lawrence River basin, invaded the lower parts of the main valleys between about Rivière Sainte-Anne and Rivière Ouareau, in Rivière Petite Nation, and in the du Lièvre and Gatineau river basins (Fig. 3.47). The marine limit on the north shore of the Champlain Sea lies at about 192 m in the western end of the basin (R.J. Fulton, Geological Suvey of Canada, personal communication, 1984), 200 m in Rivière Petite Nation valley (Richard, 1980b), 235 m north of Montréal (Lamothe, 1977; Pagé, 1977), and 200 m in the Saint-Maurice valley (S. Occhietti, Université du Québec à Montréal, personal communication, 1984). Waterlain deposits of problematic (marine, estuarine, or freshwater) origin are found in the upper parts of Saint-Maurice valley (Occhietti, 1980), as well as in Gatineau and du Lièvre valleys. Wilson (1924) initially interpreted waterlain sediments in the latter valleys as Champlain Sea sediments. Gadd (1972), mainly because of the presence of rhythmites, proposed that the valleys were occupied by a glacial lake. Dadswell (1974) used the distribution of modern day crustaceans to define a former water body connected to the Champlain Sea which extended to north of Lac Baskatong. There is no sill or morainic dam that might have produced a lake of this extent but northward tilting of 0.4 m/km would be sufficient to extend the Champlain Sea from its limit of 210 m near Ottawa to the 274 m upper limit of the water body north of Lac Baskatong (Fig. 3.47). It is therefore likely that the rhythmites were in fact deposited in a marine estuary rather than a glacial lake.

On the basis of radiocarbon ages on Champlain Sea and Goldthwait Sea shells the southern periphery of the Laurentian Highlands was ice free by at least 12 400 ± 160 BP (GSC-1533) in the Québec City area, and possibly as early as 12 200 ± 160 BP (GSC-1646) or 11 900 ± 160 BP (GSC-1772) in lower Rivière Gatineau valley (Romanelli, 1975; Parent and Occhietti, 1988; Occhietti, 1989.) As discussed by Occhietti (1989), the timing of the initiation of Champlain Sea is controversial but it probably occurred about 12 ka. Figure 3.55a shows the paleogeography just before the ice dam near Québec City was breached. Several Goldthwait Sea shell samples have given radiocarbon ages that vary between 11 ka and 12 ka in the area east of Québec City. Likewise, several Champlain Sea shell samples have been dated in that age range in the area between Montréal and the western end of the basin. The oldest shell date between Montréal and Québec City however is 10 300 ± 100 BP (GSC-2101). This unusual distribution of dates has led Hillaire-Marcel (1981) to question the validity of the old dates near Ottawa and has, in part, prompted Gadd (1980) to propose a calving bay mechanism for deglaciation of the western basin of the Champlain Sea. The deglacial chronology of the adjacent Lac Saint-Jean and upper Ottawa River areas indicates that the Laurentian Highlands were completely ice free by 10 ka.

Upper Ottawa River basin

In the Late Wisconsinan, ice from the Labrador Sector flowed southward to southwestward across the upper basin of Ottawa River (Veillette, 1983a, 1986a). Ice flow during deglaciation was complex because it involved the uncoupling of two major ice masses along the Harricana Interlobate Moraine (Fig. 3.47). South of the Hudson Bay-St. Lawrence divide, the moraine has been mapped and identified mostly on the basis of ice flow indicators by Veillette (1983b, 1986a), who argued that the McConnell Lake Moraine (Boissonneau, 1968) southwest of Lac Témiscamingue, and the Boulter "esker" and other extensive glaciofluvial deposits (Chapman, 1975) from south of North Bay to the vicinity of Lake Simcoe constitute an extension of the Harricana Moraine. Vincent and Hardy (1979) inferred a similar extension of the interlobate zone to North Bay based on regional ice flow features.

Ice retreated initially in a north-northeasterly direction but in the northwestern part of the area retreat was to the northwest. The region northeast of Témiscaming and northwest towards Lac Témiscamingue became ice free first (Veillette, 1983b, 1986a, 1988). Later a lobe of ice occupied the northern Lac Témiscamingue trough. The Laverlochère Moraine (Veillette, 1983a,b, 1986a, 1988) was built on the borders of this lobe (Fig. 3.47). West of the Harricana Interlobate Moraine and north of Lac Témiscamingue, but south of the Hudson Bay drainage divide, the Roulier Moraine (Fig. 3.47; Vincent and Hardy, 1977, 1979) was formed as a result of a halt or as a re-equilibration moraine formed when ice halted on the edge of higher ground near the height of land after retreating in a glacial lake basin (Hillaire-Marcel et al., 1981).

Glacial lakes abutted the ice margin in low-lying western and northern areas (Vincent and Hardy, 1977, 1979; Veillette, 1983b, 1988). The earliest glacial lake phase is thought to be related to the northeast extension of the Post-Algonquin glacial lake from the Great Lakes basin (Harrison, 1972). Possibly during the Sheguiandah and certainly during the Korah Phase, the lake was still dammed by ice blocking drainage down Ottawa River in the Mattawa area, and extended northeast of Témiscaming along a re-entrant in the ice front along the Harricana Interlobate Moraine and on the west side of Ottawa River to Lac Témiscamingue (Vincent and Hardy, 1977, 1979; Veillette, 1988; Fig. 3.57a). When ice withdrew from Ottawa River valley, east of Mattawa, water levels dropped, and glacial Lake Barlow (Wilson, 1918) occupied Lac Témiscamingue basin. Glacial Lake Barlow was initially controlled first by a morainic dam at Deux-Rivières (the Aylen Phase) and later, as differential uplift occurred, by a rock and morainic sill at Témiscaming (the Témiscaming Phase, Fig. 3.57b; Vincent and Hardy, 1979). Higher elevations of maximum water planes west of the Harricana Interlobate Moraine indicate that the area west of the moraine became ice free first. Maximum glacial Lake Barlow levels in the area east of Lac Témiscamingue were at about 300 m and rise to the northeast to about 380 m in the vicinity of the present Hudson Bay watershed where the Harricana Moraine crosses it (Veillette, 1988). The northeastern tilt of glacial Lake Barlow water planes suggests thicker ice in New Quebec rather than in Hudson Bay (Hillaire-Marcel et al., 1980).

Numerous minimum ages for deglaciation are available from dates on lake cores. Age determinations of 10 400 ± 200 BP (GSC-3460) and 10 100 ± 180 BP (GSC-3467) in the area east of Témiscaming provide the best estimate for the time of deglaciation in the Harricana Interlobate Moraine re-entrant (Veillette, 1988).

Quebec Clay Belt/James Bay area

During deglaciation, southwest flowing Labrador Ice and southeast flowing Hudson Ice separated along the Harricana Interlobate Moraine. The moraine, named by Hardy (1976; Fig. 3.47) and originally recognized by Low (1888) as an end

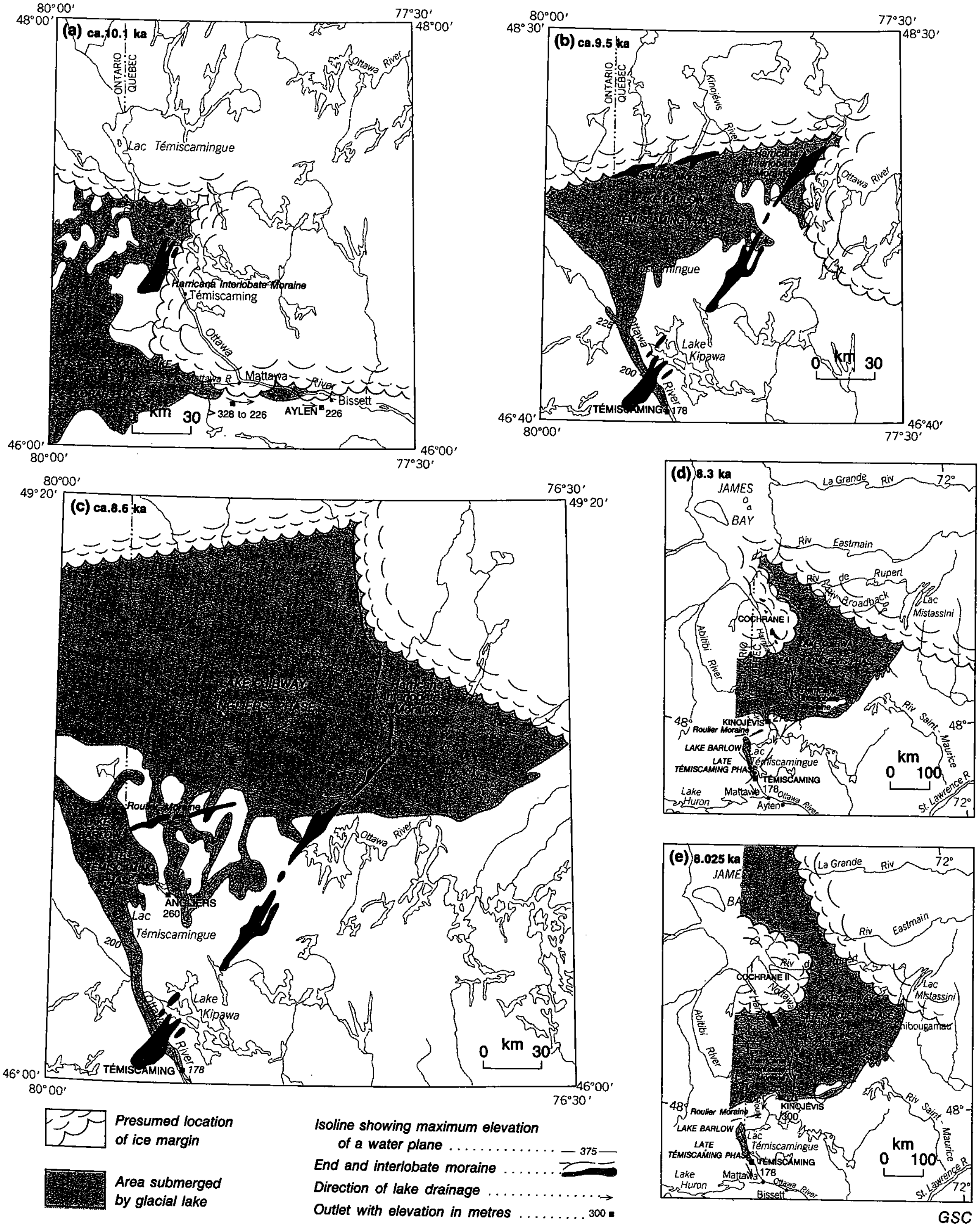


Figure 3.57. Paleogeographic maps showing successive ice frontal positions and extent of water bodies during different phases of the Post-Algonquin, Barlow, and Ojibway glacial lakes in the western part of the southeastern Canadian Shield (after Vincent and Hardy, 1979).

moraine and by Wilson (1938) as "a moraine between two ice-sheets" was studied by L.P. Tremblay (1950), Allard (1974), G. Tremblay (1974), Hardy (1976, 1977, 1982b), and Veillette (1982, 1983a,b, 1986a, 1988). The latter two authors carefully documented the two converging ice flows.

East of the Harricana Interlobate Moraine ice flow during deglaciation was towards the south-southwest near the continental divide and it swung progressively to the west until, in the area north of La Grande Rivière, it was distinctly westerly (Fig. 3.39; Hardy, 1976). In the vicinity of the interlobate moraine, the ice flow was everywhere deflected to the south, and in many areas ice flowed parallel to the moraine axis. In the La Grande Rivière area, superposed drumlins and striae (Lee et al., 1960) show that Labrador Ice was free to move in a more westerly direction following separation of the ice masses (Vincent, 1977). West of the Harricana Interlobate Moraine, in Abitibi and Lac Témiscamingue region, Veillette (1986a) has carefully documented late glacial southeasterly flow.

As Labrador Ice and Hudson Ice retreated, glacial Lake Ojibway (Coleman, 1909) was dammed between the ice front and the continental divide to the south (Fig. 3.47). The extent, as well as the history, of this lake in Quebec is discussed in detail by Vincent and Hardy (1977, 1979). The maximum lake limit rises northward from about 380 m at the divide to more than 450 m in the interfluve area between Rivière de Rupert and Rivière Broadback. The maximum depth of Lake Ojibway was more than 500 m on the east coast of James Bay. Lake waters flooded the lowlands east of James Bay as far as the Sakami Moraine (Hardy, 1976) and as far north as just east of Kuujjuarapik (Hillaire-Marcel, 1976; Fig. 3.57e). Glacial Lake Obijway became separated from glacial Lake Barlow with the emergence of the Angliers sill east of the head of Lac Témiscamingue (Vincent and Hardy, 1977, 1979; Angliers Phase, Fig. 3.57c). This sill was then the lowest point on the Hudson Bay-St. Lawrence drainage divide, which was displaced south by isostatic tilting of the crust. With differential uplift, the sill of the outlet migrated north along Rivière Kinojévis (Early Kinojévis Phase, Fig. 3.57d) until it reached the most northerly lowest point on the present divide between the upper Kinojévis and upper Harricana drainage basins (Late Kinojévis Phase, Fig. 3.57e). Large channels on the divide and wide and deeply incised Rivière Kinojévis valley were cut by overflowing glacial lake waters.

Extensive fields of De Geer moraines were built in glacial Lake Ojibway (Mawdsley, 1936; Norman, 1938; Shaw, 1944; Ignatius, 1958). Norman (1938) calculated that the rate of ice front retreat was 173-239 m/a.

A varve chronology has been established for lakes Barlow and Ojibway. Antevs (1925) counted 2027 varves beginning with varve 1 in a river cut southwest of Lac Témiscamingue. Hughes (1965) confirmed the upper part of Antevs varve diagrams, in the Clay Belt of Ontario, and added 58 varves to the sequence. Hardy (1976) correlated some 625 varves in the James Bay Lowlands of Quebec with Hughes' sequence and added 25 varves which register the most northerly and northeasterly extent of Lake Ojibway. Hardy's varve sequence extends to the time of final drainage of the lake into Tyrrell Sea. This varve chronology indicates that a total of 2110 years elapsed from the time varve 1 was laid down southwest of Lac Témiscamingue to the time Lake Ojibway drained. Using this, Hardy (1976) calculated that the rate of ice retreat increased from about 320 m/a southeast of James Bay to possibly 900 m/a in the La Grande Rivière area in the deep lake basin just before it drained.

As in Ontario (Prest, 1970; this chapter, Dredge and Cowan, 1989), late glacial ice surges occurred into glacial Lake Objiway. Hardy (1976) has documented three "Cochrane" surges in the lowlands southeast of James Bay. His reconstruction is based on: (1) mapping the extent of the clayey and carbonate-rich Cochrane till; (2) measuring ice flow features related to each movement; (3) logging sections in which Cochrane tills are interstratified with Lake Ojibway sediments, and (4) studying the effect of the Cochrane surges on the sedimentology of the varves. His work enabled him to reconstruct ice profiles by showing precisely where the ice was grounded or floating in the lake basin. The extent of ice during Cochrane I surge is shown in Figure 3.57d and during Cochrane II in Figure 3.57e. A third surge, intermediate between the Cochrane I and II was called the Rupert surge. These originated to the northwest and, hence, came from Hudson Ice. The Cochrane I of Hardy (1976) in Quebec is probably somewhat younger than the Cochrane I of Prest (1970) in Ontario. During Cochrane I and II surges, Lake Ojibway varves became coarser and thicker (effect of the ice advance) and richer in carbonate (effect of overriding Paleozoic beds south and southeast of James Bay) (Hardy, 1976). The maximum Cochrane I and II surges occurred 300 years (varve year 1810 and correlative with Hughes' (1965) Frederick House varve sequence) and 75 years (varve year 2035 and correlative with Hughes' (1965) Connaught varve sequence) before glacial Lake Ojibway drained. Changes in sedimentology of the varves can be attributed entirely to the effect of the surges. Hence there is no need to invoke fluctuating lake levels, as suggested by Hughes (1965) or different lake phases such as the Antevs or Opemiska lakes suggested by Prest (1970).

When Labrador Ice had retreated to the approximate position of the Sakami Moraine, waters from Hudson Strait penetrated Hudson and James bays and flooded the isostatically depressed lowlands. Opening to the sea allowed drainage of glacial Lake Ojibway and led to formation of the Sakami Moraine (Hardy, 1976). The moraine is a major feature extending over a distance of 630 km from the vicinity of Kuujjuarapik on the southeastern coast of Hudson Bay to southern Lac Mistassini (Fig. 3.47). Hardy (1976) has shown that the Sakami Moraine marks the position of Labrador Ice at the time glacial Lake Ojibway drained and the postglacial sea, called the Tyrrell Sea by Lee (1960), submerged coastal areas. Hillaire-Marcel et al. (1981) considered the Sakami Moraine as a re-equilibration moraine, that is, a moraine that results from the stabilization of the ice front when the glacier, which was previously calving and in part floating in a deep water basin, grounded. Ice did not resume its retreat until a new equilibrium profile had been established. Locally the ice may have readvanced. Drainage of glacial Lake Ojibway led to a slight readvance in the Lac Mistassini area (Bouchard, 1980), previously recognized as the Waconichi ice advance by DiLabio (1981). Drainage of glacial Lake Ojibway and submergence by the Tyrrell Sea was likely a catastrophic event; it is recorded in sections where post-Cochrane II Lake Ojibway varves are overlain by a thin diamicton with carbonate debris, marking the drainage horizon, and by fossiliferous marine clays (Hardy, 1976). The sequence is similar to that described by Skinner (1973) in the Hudson Bay Lowlands of Ontario. He interpreted the diamicton as resulting from a sudden widespread density flow. Higher hills in the Lake Ojibway basin also record the seemingly sudden drainage; on these hills both upper and lower wave washing limits exist (Norman, 1939; Hardy, 1976). Below the lower limit of washing, no strandlines are present because the lake level fell too rapidly for strandlines to form. The lower washing limit represents a long synchronous shoreline that is now tilted, and it offers an excellent opportunity for measuring the amount of delevelling accomplished since it formed.

The Tyrrell Sea followed the retreating ice front after construction of the Sakami Moraine. Marine limit is at about 198 m in the southern part of its basin (Hardy, 1976) and rises northward to the Kuujjuarapik area where it reaches elevations of 315 m (Hillaire-Marcel, 1976). In La Grande Rivière area marine limit decreases eastward from close to 270 m in the area just east of the Sakami Moraine to 246 m farther up river (Vincent, 1977). Extensive swarms of De Geer moraines, many of which overlie drumlins and eskers, were built east of the Sakami Moraine (Fig. 3.46). Rates of ice retreat averaged 217 m/a (Vincent, 1977) . In the Lac Mistassini area, a relatively shallow glacial lake, named glacial Lake Mattawaskin by Bouchard (1986) followed the retreating ice front (Fig. 3.55e). Discharge of lake waters was first via Rivière Broadback, and then via Rivière de Rupert and Rivière Eastmain as lower outlets became free of ice (Vincent and Hardy, 1977, 1979; Bouchard, 1980, 1986). Rates of ice recession in this lake basin were estimated at 220 to 260 m/a (Bouchard, 1980).

The late Quaternary chronological framework for the Quebec Clay Belt and the area southeast and east of James Bay rests on the age assigned the Sakami Moraine. Shells collected in clay lying on the proximal flank of the Sakami Moraine date 7880 ± 160 BP (QU-122). This date led Hardy (1976) to assign an age of 7.9 ka, or slightly older, to the moraine. Hillaire-Marcel (1976) assigned a 8.1-8.0 ka maximum age to the moraine on the basis of a 8230 ± 135 BP (I-8363) age (corrected to 8.1-8.0 ka to account for isotopic fractionation) on concretions in glacial Lake Ojibway clays closely associated with formation of the moraine near Kuujjuarapik. The approximate age of ca. 8.0 ka for the Sakami Moraine (therefore also for the drainage of glacial Lake Ojibway and incursion of the Tyrrell Sea) agrees well with the age proposed for the drainage of glacial lakes Agassiz and Ojibway and for the incursion of Tyrrell Sea on the west side of Hudson Bay (see this chapter, Dyke and Dredge, 1989; Dredge and Cowan, 1989). Assuming an 8.0 ka age for the Sakami Moraine, and using the varve chronology, the area west of Lac Témiscamingue was deglaciated some 10 ka ago and the height of land at the Quebec-Ontario border, 9.2 ka ago; the Cochrane I reached its maximum 8.3 ka ago and Cochrane II 8025BP. East of the Sakami Moraine, based on the De Geer moraine chronology (Vincent, 1977), the Tyrrell Sea reached its eastern limit in the La Grande Rivière area about 7.5 ka ago. Farther inland the oldest radiocarbon age of 6600 ± 100 BP (B-9516) was obtained by P.J.H. Richard (Université de Montréal, personal communication, 1984) on basal gyttja from a lake in the Rivière Laforge area.

Area east of Hudson Bay

Ice from central New Quebec and central Ungava Peninsula flowed generally westward and southwestward into Hudson Bay (Hillaire-Marcel, 1979; Gray and Lauriol, 1985). This ice flow direction is also recorded on the Belcher Islands (Jackson, 1960). On the Ottawa Islands the direction of last ice movement was west-southwest, but Andrews and Falconer (1969) have documented an earlier northeasterly flow, associated by Dyke et al. (1983) with Hudson Ice.

When Labrador Ice finally separated from Hudson Ice in Hudson Bay, only the area west of the Sakami Moraine in the Kuujjuarapik area, and perhaps the extreme northwestern part of Ungava Peninsula, was ice free (Fig. 3.55e). On the basis of the progressively shifting ice flow directions recorded by striae on the Ottawa Islands , the drainage corridor is inferred to have been located west of the islands (Andrews and Falconer, 1969).

As the ice margin retreated to the east, the Tyrrell Sea covered the newly deglaciated areas and swarms of De Geer moraines were built at the ice front in a belt extending northward from Inukjuak to south of Ivujivik. Marine limit decreases northward from about 315 m on the southeastern coast of Hudson Bay to possibly as low as 105 m east of Povungnituk. From there it rises northerly to about 170 m near Hudson Strait (Gray and Lauriol, 1985). As in the area east of James Bay, marine limit also declines inland; it falls from 248 m to 196 m along Rivière Nastapoca (Allard and Seguin, 1985), and from 158 m on the Ottawa Islands (Andrews and Falconer, 1969) to 105 m east of Povungnituk (Gray and Lauriol, 1985) (Fig. 3.47). Rates of uplift, as measured in the Lac Guillaume-Delisle (Richmond Gulf) area, were 9.6-10 m/century at the time of deglaciation (Allard and Seguin, 1985). According to Hillaire-Marcel (1976), by 7 ka the rate of uplift was 6.5 m/century and it decreased linearly to the present rates estimated to be of the order of 1.1 m/century. Fairbridge and Hillaire-Marcel (1977) and Hillaire-Marcel and Fairbridge (1978) recognized a 45 year cycle in beach building in the Lac Guillaume-Delisle area that they related to the "double Hale" solar cycle; thus they deduced a cyclic record of storminess.

East of marine limit, ice continued its retreat towards central Ungava Peninsula in the north or towards central New Quebec farther south. Beyond the Tyrrell Sea limit, a shallow glacial lake was formed in the Lac à l'Eau Claire area between an uplifted sill west of the lake and the ice front (Allard and Seguin, 1985; Fig. 3.47, 3.55f). At the head of Rivière aux Mélèzes, glacial Lake Minto formed between the Hudson Bay-Ungava Bay drainage divide and the receding ice front (Lauriol, 1982; Lauriol and Gray, 1983; Fig. 3.47, 3.55f).

The time of deglaciation for the area west of the Sakami Moraine, in the Kuujjuarapik area, is estimated at 8.1 ka by Hillaire-Marcel (1976). In the Rivière Nastapoca area, the oldest dated shells, 45 m below marine limit, were 6700 ± 100 years old (UQ-547). Shells 17 m below marine limit on the Ottawa Islands (Andrews and Falconer, 1969) were dated at 7430 ± 180 BP (GSC-706), indicating rather late deglaciation of western Ungava. Marine shells from the Cape Smith area originally provided radiocarbon ages of ca. 8 ka but were later redated at ca. 6.8 BP (Lauriol and Gray, 1987).

Ungava Peninsula

Both northerly flow into southern Ungava Bay, from central New Quebec, and flow towards Hudson Bay, western Hudson Strait, and western Ungava Bay, from a central north-south ice divide on Ungava Peninsula are recorded

(Gray and Lauriol, 1985; Bouchard and Marcotte, 1986). Ice flowing westward and northward from this ice flow centre, called the Payne center by Bouchard and Marcotte (1986), apparently coalesced with ice moving northeastward in Hudson Bay and eastward in Hudson Strait (flow in offshore area from Andrews and Falconer, 1969; Shilts, 1980; Laymon, 1984; Gray and Lauriol, 1985).

The northwestern and northern extremities of the area were deglaciated first (Fig. 3.55e). Marine waters submerged the Ungava coast along Hudson Strait where marine submergence generally decreases from west to east (167 m at Cape Wolstenholme, Matthews, 1967; 138 m near Diana Bay, Gray et al., 1980), and from north to south (170 m on Charles Island and 120 m at the head of Deception Bay, Gray and Lauriol, 1985). Rates of uplift at about 8 ka were estimated at 7.9 m/century (Matthews, 1967). The age of deglaciation of the south shore of Hudson Strait has been the subject of some controversy. Most researchers would agree that some areas were ice free at least 7970 ± 250 BP (GSC-672), but some, on the basis of three "older" radiocarbon age determinations, postulate a much earlier deglaciation. A 10 450 ± 250 BP (I-488; Matthews, 1966, 1967) age may in fact be erroneous since shells collected by B. Lauriol (University of Ottawa, personal communication, 1985) from the immediate vicinity of the Deception Bay site originally sampled by Matthews were dated at 7130 ± 100 BP (GSC-3947). Notwithstanding this, two other age determinations, the oldest of which is 9800 ± 220 BP (Beta-11121), have been obtained from the Deception Bay area by dating in situ *Portlandia arctica* and *Nuculana minuta* shells collected in glacial marine sediments overlying till (Gray and Lauriol, 1985; Lauriol and Gray, 1987). This conflicts with evidence on Meta Incognita Peninsula of Baffin Island which requires that ice extend to the mouth of Hudson Strait until 8.6 ka or later (this chapter, Andrews, 1989). Miller et al. (1988) have suggested that a late readvance of Labrador Ice across Hudson Strait, between 9 ka and 8.2 ka, could account for the late presence of ice on southern Baffin Island. Upon further retreat of the ice front towards the interior of Ungava Peninsula, glacial lakes were dammed between the ice front and higher ground on the Hudson Bay-Hudson Strait and Hudson Bay-Ungava Bay drainage divides (Fig. 3.47; Prest et al., 1968; Prest, 1970). Of these lakes, the best documented is glacial Lake Nantais (Lauriol and Gray, 1987; Fig. 3.47).

Ice filling Ungava Bay retreated both westward towards central Ungava Peninsula and south towards central New Quebec perhaps through the development of a calving bay. Eastern Hudson Strait was deglaciated by 9.1 ka on the basis of a radiocarbon date on shells collected from a seabed core (9120 ± 480 BP, GSC-2946). The D'Iberville Sea (Laverdière and Bernard, 1969) followed the retreating ice front. Blake (1976), Gangloff et al. (1976), Lauriol et al. (1979), Gray et al. (1980), Lauriol (1982), and Gray and Lauriol (1985) provide detailed accounts of sea level history in this sector. On the west coast of Ungava Bay marine limit rises from 138 m near Diana Bay to 195 m in upper Rivière aux Mélèzes drainage basin (Gray and Lauriol, 1985). On Akpatok Island, 75 km offshore, marine limit is much lower (58-74 m; Løken, 1978). Several radiocarbon dates on shells indicate that the west coast of Ungava Bay was ice free by 7 ka (e.g., 6990 ± 150 BP, I-9632; Fig. 3.55e). Standing water bodies extending far inland in the lower parts of the valleys of Rivière aux Mélèzes (Gray and Lauriol, 1985) and Rivière Caniapiscau (Drummond, 1965), thought to be glacial lakes, could well have been brackish estuaries of the D'Iberville Sea.

Central New Quebec and western Labrador

Since Low (1896) first recognized central Labrador-Ungava Peninsula as one of the final centres of ice disintegration, much controversy has surrounded the identification of the exact locations of the last glacial masses. As portrayed in Wilson et al. (1958), Ives (1960a), Prest et al. (1968), and Prest (1969), ice flow indicators provide evidence that Labrador Ice flowed radially from the horseshoe-shaped ice divide extending from northwest of Lac Delorme in the west, to northern Smallwood Reservoir in the east. Whether the ice divide had been stable for a considerable part of the Late Wisconsinan, or whether its location fluctuated considerably, is not definitely established. For example, Hughes (1964) and Richard et al. (1982) suggested that the ice divide must at one time have been situated well to the northeast of its final location because rocks from the Labrador Trough were transported to areas west of the ice divide. Equally, whether or not the final location of the divide coincides with the location of the last ice remnants is not clearly established. Recent fieldwork by Klassen and Thompson (1987), who have recognized up to five distinct phases of flow in the area, should help clarify these points.

Field investigations by Perrault (1955), Grayson (1956), Henderson (1959), Ives (1959, 1960a, c, 1968, 1979), Kirby (1961a, b), Derbyshire (1962a), in the Schefferville area of the Labrador Trough, brought conclusive evidence for the presence of small ice remnants in the low-lying basins of Howells River (Kivivic ice divide; just west of Schefferville), and Swampy Bay River valleys (north-northwest of Schefferville). Other authors, basing their arguments on glacial ice flow indicators and landforms (Low, 1896; Hughes, 1964; Laverdière, 1967; Richard et al., 1982) and on the intersection of projected strandline tilt directions of glacial lakes as an indicator of the location of the maximum ice thickness (Ives, 1960b; Harrison, 1963; Barnett, 1964; Barnett and Peterson, 1964), proposed that the final ice masses disintegrated in the general area of the "horseshoe-shaped" ice divide. Much controversy has ensued between the different proponents (Ives, 1968; Barnett and Peterson, 1968; Bryson et al., 1969; Laverdière, 1969a,b; Laverdière and Guimont, 1982), but it is likely that there were numerous retreat centres both along the final location of the divide and in adjacent low-lying basins where discrete ice masses finally melted.

Dates from numerous lake cores have been used to date the final disappearance of ice (Grayson, 1956; Morisson, 1970; McAndrews and Samson, 1977; Short, 1981; Stravers, 1981; Richard et al., 1982; King, 1985). On or near the final location of the ice divide, basal dates of 6320 ± 180 BP(GSC-3094; Richard et al., 1982) in the Lac Delorme area, and 6200 ± 100 BP(GSC-3644; King, 1985) in the Lac Stakel area provide the best minimal estimates. In the Schefferville area of the Labrador Trough numerous dates have been obtained (Short, 1981; Stravers, 1981), including some as old as 16 ka; however, these dates are suspect because of the possible presence of redeposited older organic matter in samples containing less than 1% organic carbon (Stravers, 1981). Apart perhaps from small remnant ice masses in depressions, it is probably safe to assume that Labrador Ice had completely melted by 6.5 ka.

Postglacial history

Postglacial emergence

Much data on sea level history have been presented in the sections dealing with the deglaciation of each major region. Minimum sea level curves have been constructed for six areas (Fig. 3.58) and are discussed here. Although a considerable number of shell samples has been dated, few can be tied to specific sea level positions. Also, in most cases, only part of an emergence curve can be plotted since no material has been found to date both the lower and the upper parts of curves. The curves are nevertheless useful since they provide a value for the minimal sea level which could have existed at any given time in different areas.

The family of curves in Figure 3.58 illustrates the relative isostatic response of different areas to the ice load. As would be expected, the middle Quebec North Shore and Deception Bay curves, farthest away from the centres of loading, exhibit the least glacial isostatic recovery, whereas those which are closest (La Grande Rivière and Lac Guillaume-Delisle) exhibit most. The intersection of the La Grande Rivière and Lac Guillaume-Delisle curves at about 6 ka may indicate the switch from the initial dominant influence of the Hudson Ice load to a later dominant influence of the Labrador Ice load as proposed by Hillaire-Marcel (1980).

The shape of the La Grande Rivière curve and its extrapolation to marine limit from two dates on sediments associated with the construction of the Sakami Moraine interestingly support the age estimate, of about 8 ka, for the drainage of glacial Lake Ojibway and incursion of Tyrrell Sea east of James Bay.

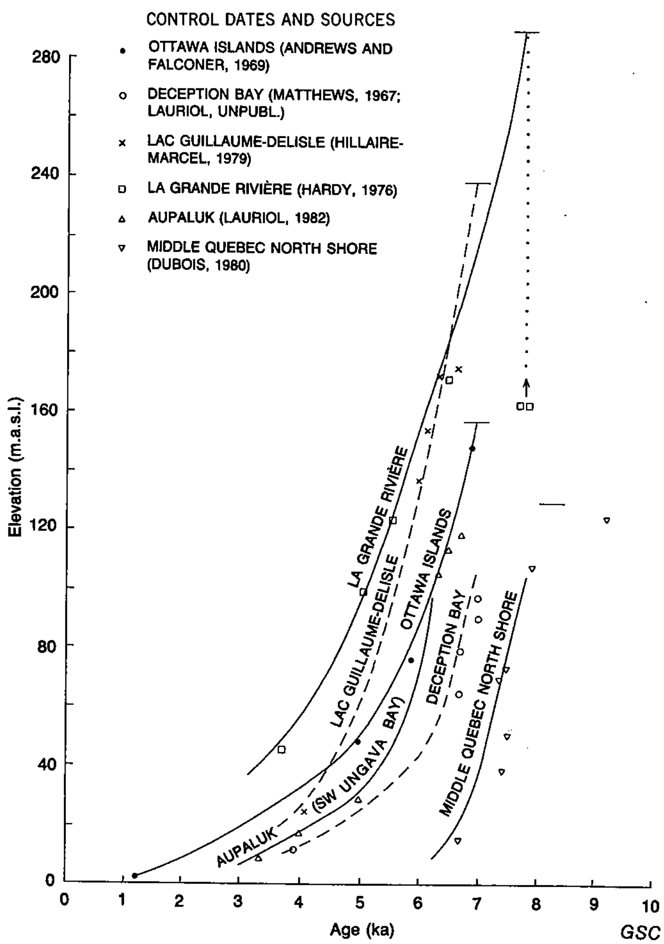


Figure 3.58. Minimum emergence curves for six areas of the southeastern Canadian Shield.

Late Wisconsinan and Holocene vegetational history

P.J.H. Richard

Thanks to data from pollen analyses, the vegetation landscapes are known to have been diversified at the margin of the retreating Laurentide Ice Sheet, but in the central part of the southeastern Canadian Shield, since about 5 ka, as well as in peripheral areas, since about 8 ka, changes have been generally minor. Detailed accounts of vegetational history and lists of the numerous available studies are given in Richard (1977a, 1981, 1985, 1989).

At about 11 ka in the south (Richard, 1977b; Savoie and Richard, 1979), and at about 7.5 ka in Ungava Peninsula (Richard, 1981), nearly unvegetated and then tundra environments existed. In Labrador, shrub-tundra occupied the newly deglaciated terrain until about 8 ka in the south, and 5 ka in the north-central part of the area (Short, 1978; Lamb, 1980, 1984). Elsewhere, in the central portion of the southeastern Canadian Shield, trees colonized the land soon after ice retreat, the drainage of glacial lakes (Richard, 1980a), or emergence from postglacial seas (Richard, 1979). *Larix, Populus, Picea,* and *Betula,* accompanied by *Alnus,* were the first colonizers. In southern Labrador and on the Quebec North Shore, *Picea glauca* and *Abies balsamea* were the first trees present (Lamb, 1980, 1984). In the Laurentian Highlands tree colonization of the higher areas was slow; *Populus* played a dominant role there for more than a millennium (Richard, 1977b).

During the early Holocene, a complex history occurred on the southern margin of the Canadian Shield. Large fluctuations in the abundance of *Pinus divaricata, Pinus strobus,* and *Tsuga canadensis* on one hand, and the northward migration of *Acer saccharum, Fagus grandiflora,* and other thermophilous species on the other hand, characterize this period. The presence of spruce, then fir, led the way to the present day mixed forests. *Pinus strobus* was 50 km farther north than today about 5 ka ago in the middle Holocene (Terasmae and Anderson, 1970). In New Quebec, *Pinus divaricata* migrated into the spruce dominated area east of James Bay about 3 ka (Richard, 1979) barely reaching Rivière Caniapiscau to the east (Richard et al., 1982). In the north, however, the position of treeline has fluctuated little since deglaciation (Gagnon and Payette, 1981; Richard, 1981). In the last 1000 years the boreal forest has progressively opened up, possibly in response to climatic cooling in the late Holocene (Short, 1978).

Richard, P.J.H.
1989: Late Wisconsinan and Holocene vegetational history; in Chapter 3 of Quaternary Geology of Canada and Greenland, R.J. Fulton (ed.); Geological Survey of Canada, Geology of Canada, no. 1 (also Geological Society of America, The Geology of North America, v. K-1)